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Key Points:

- Andean (~323–306 Ma) and syn-collisional (~304–288 Ma) granitoid magmatism in the Gissar arc was accompanied by middle–upper-amphibolite-facies metamorphism
- Post-collisional delamination of the Gissar arclogite root induced asthenospheric upwelling, lower-granulite-facies metamorphism, and alkaline magmatism
- The M–HT/LP Garm massif and M–HP/LT Fan-Karategin belt represent divergent P–T–t paths of the peri-Gondwanan Karakum terrane, which accreted to Asia by ~300 Ma

Supporting Information:

- Supporting Information S1
- Tables S1–S16 and Data Sets S1–S8

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Birth, life, and demise of the Andean–syn-collisional Gissar arc: Late Paleozoic tectono-magmatic-metamorphic evolution of the southwestern Tian Shan, Tajikistan

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Abstract The amalgamation of the Central Asian Orogenic Belt in the southwestern Tian Shan in Tajikistan is represented by tectono-magmatic-metamorphic processes that accompanied late Paleozoic ocean closure and collision between the Karakum-Tarim and Kazakh-Kyrgyz terranes. Integrated U–Pb geochronology, thermobarometry, pseudosection modeling, and Hf geochemistry constrain the timing and petro-tectonic nature of these processes. The Gissar batholith and the Garm massif represent an eastward, along-strike increase in paleodepth from upper-batholith (~21–7 km) to arc-root (~36–19 km) levels of the Andean–syn-collisional Gissar arc, which developed from ~323–288 Ma in two stages: (i) Andean, I-type granitoid magmatism from ~323–306 Ma due to northward subduction of the Gissar back-arc ocean basin under the Gissar microcontinent, which was immediately followed by (ii) syn-collisional, I–S-type granitoid magmatism in the Gissar batholith and the Garm massif from ~304–288 Ma due to northward subduction/underthrusting of Karakum marginal-continental crust under the Gissar microcontinent. A rapid isotopic pull-up from ~288–286 Ma signals the onset of juvenile, alkaline-syenitic, post-collisional magmatism by ~280 Ma, which was driven by delamination of the Gissar arclogite root and consequent convective asthenospheric upwelling. Whereas M–HT/LP prograde metamorphism in the Garm massif (650–750°C/6–7 kbar) from ~310–288 Ma was associated with subduction-magma inundation and crustal thickening, HT/LP heating and decompression to peak-metamorphic temperatures (~800–820°C/6–4 kbar) at ~288 ± 6 Ma was driven by the transmission of a post-collisional, mantle-derived heat wave through the Garm-massif crust.

1. Introduction and Geologic Setting

1.1. Introduction

The Tian Shan (Tien Shan) orogenic belt (Figure 1), which stretches ~2500 km E–W from easternmost Xinjiang in China to central Uzbekistan, is the longest and southernmost component of the Central Asian Orogenic Belt (CAOB; also “Altaids,” after Şengör *et al.* [1993]). The orogen is divided, to a first order and somewhat schematically, into the North, Middle, and South Tian Shan by E–W-trending sutures that reflect a complex, mostly Paleozoic assembly (Figure 1) and provide insight into suturing and terrane-accretion processes. Diverse mineral assemblages and textures in related igneous and metamorphic belts attest to divergent but related dynamics of subduction, accretion, and collision throughout the orogen. The NW–SE-striking, ~900-km-long, dextral Talas-Fergana fault (TFF) divides the Tian Shan into western and eastern parts and has a protracted history and variable slip rate, whereby significant dextral transcurrent motion during the late Paleozoic-early Mesozoic [Şengör *et al.*, 1993; Şengör and Natal’in, 1996; Sobel, 1999; Allen *et al.*, 2001; Alexeiev *et al.*, 2009a; Konopelko *et al.*, 2013; Rolland *et al.*, 2013; Konopelko *et al.*, 2017] transitioned into the modern, low-slip rate (<2 mm/a [Mohadjer *et al.*, 2010]), dextral strike-slip setting. The primary Paleozoic structural framework has been strongly overprinted by far-field strain related to the successive, Mesozoic-Cenozoic accretion of “Tethyside” Asia [after Şengör and Natal’in, 1996], which renders assessment of the Paleozoic framework a challenging task.

The goals of this study are to: (i) document the late Paleozoic magmatic and metamorphic evolution of the western South Tian Shan in Tajikistan (henceforth the “southwestern Tian Shan” for simplicity); (ii) compare our findings to the synchronous evolution of the Tian Shan along-strike, to the east and across the Talas-Fergana

Figure 1. Paleozoic magmatic rocks of the Tian Shan, modified from *Li et al.* [2006]. Regionally significant magmatic bodies colored to emphasize Paleozoic spatiotemporal magmatic patterns. Sutures are indicated with dashed black lines, ages adopted from *Burtman* [2010]. Plutonic rock bodies are indicated with bright colors, whereas volcanic rock bodies are indicated with corresponding faded colors. Sedimentary rock units (variably metamorphosed “strata”) indicated in gray. Approximate locations of (U)HP/LT and M–HP/LT belts indicated with light-blue stars, (M)–HT/LP complexes indicated with red stars. Carboniferous–Permian plutonic and volcanic rocks are widespread throughout the Tian Shan, indicating widespread magmatism throughout the Tian Shan during the late Paleozoic. Early Paleozoic (Cambrian–Silurian) magmatism is primarily concentrated in the north Tian Shan (also “Issyk-Kul terrane” and “Yili block”), STS, south Tian Shan; MTS, middle Tian Shan; NTS, north Tian Shan; NETS, northeast Tian Shan; o.m., ophiolitic mélange.

fault, where the orogen has been more thoroughly studied; and (iii) develop a preliminary tectonic framework for the Paleozoic geological evolution of the Tian Shan. Our study specifically focuses on the Gissar batholith, the Garm massif, and the Zeravshan subduction-accretion complex (Figure 2). Our approach integrates insight from investigation of 14 granitoids, six amphibolite-granulite facies schists gneisses (sedimentary and igneous protoliths), nine greenschist–blueschist-facies schists (sedimentary and igneous protoliths), two Cretaceous sandstones, and four modern river sand samples (summarized in Table 1). Our analytical data set comprises U–Pb igneous (15 samples) and detrital (12 samples) zircon geochronology, U–Pb monazite geochronology (one sample), igneous (five samples), and detrital (three samples) zircon Hf geochemistry, petrology, and thermobarometry. We integrate our data with that from previous investigations to document the late Paleozoic birth, life, and demise of the Andean–syn-collisional Gissar arc in the context of the southwestern Tian Shan. We then assess our results more broadly within the framework of the Tian Shan along strike and across the Talas–Fergana fault.

We use the International Chronostratigraphic Chart [Cohen *et al.*, 2013; updated] to integrate absolute ages into a Carboniferous and Permian stratigraphic framework. We refer to the “lower/early” and “upper/late” Carboniferous in lieu of the time-equivalent Mississippian and Pennsylvanian. In our usage, the “middle Carboniferous” primarily encompasses the Serupkhovian and Bashkirian stages, which span the transition between the lower/early and upper/late Carboniferous. We refer to component stages within the Carboniferous and Permian to discuss geologic time in greater detail.

1.1.1. The Central Asian Orogenic Belt

The CAOB is a broad accretionary-orogenic system [e.g., Şengör *et al.*, 1993; Cawood, 2009] and comprises a collage of oceanic-island and continental-margin arcs, ophiolitic mélanges, and microcontinents that amalgamated to Asia during the Neoproterozoic–Permian. Şengör *et al.* [1993] proposed that this amalgamation occurred along a single subduction zone (the postulated, 7000 km long Kipchak arc) that underwent pervasive oroclinal bending and strike-slip duplication. Most recent studies, however, favor the view that the CAOB comprises a wider array of island-arc terranes with more diverse origins, including arcs built on older crust [e.g., Windley *et al.*, 2007; Wilhem *et al.*, 2012]. Irrespective of terrane origins, petro-tectonic logic renders the abundance of island-arc complexes in the CAOB an unambiguous representation of significant generation of new, juvenile continental crust [e.g., Schulmann and Paterson, 2011]. Many regional studies of the CAOB (many of which cite juvenile Sm–Nd and Lu–Hf isotopic signatures as evidence) recognize significant crustal growth via the generation of juvenile crust as the orogen’s first-order feature, but quantitative estimates of the fraction of juvenile crust that comprises the CAOB are debated. An upper estimate for the fraction of juvenile continental crust in the CAOB is >50% [Şengör *et al.*, 1993; Hong *et al.*, 2004], which Kröner *et al.* [2014, 2017] argue is an overestimate.

Assembly of the CAOB culminated during the late Paleozoic with several collisions between cratons, microcontinents, and island-arc complexes throughout Central Asia [e.g., Allen *et al.*, 2001]. This final phase of assembly in the Tian Shan involved collision between the (perhaps linked) Karakum and Tarim cratons with the southern margin of paleo-Kazakhstan. Xiao *et al.* [2013], for instance, advocate this first-order regional tectonic evolutionary scheme by defining the Tian Shan overall as comprising an early–middle Paleozoic network of subduction-accretion systems that transitioned to an arc-continent collision zone (via an Andean stage of development) in the late Carboniferous–Permian and finally to a Mediterranean-type orogen in the late Permian–Triassic.

1.1.2. E–W Subdivision of the Tian Shan: The Talas–Fergana Fault

The Talas–Fergana dextral strike-slip fault (Figure 1) is a major structure that cuts the late Paleozoic–early Mesozoic, east–west tectonic framework of the Tian Shan. Within the context of the CAOB, the TFF is part of a system of NW-striking, dextral and ENE-striking, sinistral faults that may have accommodated N–S shortening and lateral escape during terrane accretion [Rolland *et al.*, 2013]. Although minor sinistral movement likely occurred during the Ordovician, Permian dextral motion accounts for the majority of the ~180–200 km maximum dextral offset along the Talas–Fergana fault [Ognev, 1939; Burtman, 1975; Şengör *et al.*, 1993; Şengör and Natal’in, 1996; Sobel, 1999; Allen *et al.*, 2001; Alexeiev *et al.*, 2009a; Konopelko *et al.*, 2013; Rolland *et al.*, 2013; Konopelko *et al.*, 2017]. Therefore, the primary Permian motion of the Talas–Fergana fault overlaps in space and time with the final assembly of the Tian Shan and accommodated major regional kinematic reorganization associated with the transition from accretionary to collisional tectonics in the CAOB. Notwithstanding that the Talas–Fergana fault is active with <2 mm/a dextral slip [Mohadjer *et al.*, 2010], net post-Triassic offset along the fault is trivial in comparison to the primary, earlier offset.

Figure 2. Geologic map of the southwestern Tian Shan, modified from Soviet 1:200,000 geologic maps [Yushin *et al.*, 1964] and with tectonic delineations adopted from Brookfield [2000]. Sample locations for this study indicated with colored circles that correlate with lithology. Weighted-mean zircon $^{206}\text{Pb}/^{238}\text{U}$ “crystallization” ages for igneous and metamorphic rocks from this study and related studies provided for igneous rocks and metamorphic/migmatitic rocks. Catchments for modern-river-sand samples (this study) outlined in light blue. All ages from our study are $^{206}\text{Pb}/^{238}\text{U}$ weighted-mean ages. *Kaibner et al.*’s [2016b] reported concordia ages used because their conservative 90–110% concordance-filtering scheme renders concordia ages comparable to weighted-mean ages. We use *Konopelko et al.*’s [2016] reported $^{206}\text{Pb}/^{238}\text{U}$ weighted-mean ages and calculated such ages from *Konopelko et al.*’s [2015] analyses that they incorporated into concordia ages. All ages without symbols reflect crystallization ages. Zircon age interpretations: ‘, age of metamorphism; ‘’, age of migmatization; *, age of dike intrusion. Monazite age interpretation: m’, age of prograde metamorphism; M’, age of peak metamorphism. Non-underlined numerical values in italics indicate pressure estimates; underlined numerical values in italics indicate temperature estimates. Mean $\pm 2\sigma$ zircon ϵ_{Hf} values [Kaibner *et al.*, 2016b; this study] are indicated for relevant samples with “Hf” prefix; mean whole-rock ϵ_{Nd} values [Kaibner *et al.*, 2016b; Konopelko *et al.*, 2016] are indicated for relevant samples with “Nd” prefix.

Table 1. Summary of All Samples and Analyses Performed in This Study^a

Sample	North Longitude	East Latitude	Name	U-Pb ^b	zHf ^c	EPMA ^d
<i>Gissar/Garm Granitoids</i>						
12TS06	38.76310	68.81825	Gissar granite	iZrn 301 ^e	-	-
12TS08	38.79043	68.82535	Gissar Bt granite	iZrn 300 ^e	-	-
12TS09	38.91470	68.81835	Gissar Amp + Bt Qz diorite	iZrn 309 ^e	-	6.5/709 (Al-in-Hbl/Amp-Pl)
12TS12	38.94645	68.79977	Gissar Amp + Bt granite	iZrn 313 ^e	iZrn +0.6 ^e	-
12TS23A	39.10133	70.10305	Garm Bt granodiorite	iZrn 292 ^e	-	-
12TS30	39.05867	70.10673	Garm Grt + Bt granodiorite	iZrn 294 ^e	iZrn -7.3 ^e	830 (GARb)
12TS58	39.00558	70.31005	Garm Bt Qz monzonite	iZrn 299 ^e	-	-
12TS91	39.29060	70.69147	Garm Grt Qz syenite	iZrn 297 ^e	-	-
12TS96	39.15250	70.70563	Garm Bt + Cal + Ne + Ccn syenite	iZrn 280 ^e	iZrn +3.3 ^e	-
12TS102	39.16390	70.64622	Garm Bt granite	iZrn 295 ^e	-	-
12TS111	39.15575	70.98645	Garm Amp + Bt + Ep Qz monzonite	iZrn 292 ^e	iZrn -2.7 ^e	5.9/733 (Al-in-Hbl/Amp-Pl)
12TS114	39.19403	71.10930	Garm Amp + Bt granite	iZrn 295 ^e	-	7.9/686,714 (Al-in-Hbl/Amp-Pl,Zr-in-Ttn)
12TS121	39.22633	71.33342	Garm Bt + Ms granite	iZrn 291 ^e	-	-
14TS02	38.75891	69.28602	Gissar Bt granite	iZrn 298 ^e	-	-
<i>Garm Gneisses</i>						
12TS34	39.16600	70.33405	Garm amphibolite facies paragneiss	-	-	665 (GARb)
12TS47	39.06325	70.26753	Garm amphibolite facies metabasite	-	-	716 (Amp-Pl); X _{CPx}
12TS59	39.06478	70.38363	Garm amphibolite facies metabasite	-	-	731 (Amp-Pl); X _{CPx}
12TS62	39.06510	70.38640	Garm granulite facies paragneiss	dZrn ^e ; Mnz 296/287 ^f	-	6.8/750, 829/4 (GASP/GARb); 650-750/6-7, 800-820/4-6 (PS); X _{Spl} , Ilm
12TS73	39.14975	70.58440	Garm amphibolite facies paragneiss	dZrn ^e	-	678 (GARb)
12TS101	39.14038	70.69742	Garm amphibolite facies paragneiss	dZrn ^{e,g} ; mZrn 291 ^g	-	-
<i>Fan-Karategin Greenschists/Blueschists</i>						
12TS13	38.97868	68.76338	Fan-Karategin greenschist (metasedimentary)	dZrn ^e	-	-
12TS22	39.10465	70.10037	Fan-Karategin greenschist (metasedimentary)	dZrn ^e	-	-
12TS38	39.16423	70.19075	Fan-Karategin greenschist (metasedimentary)	dZrn ^e ; dTtn ^f	-	-
12TS83	39.22430	70.85343	Fan-Karategin greenschist (?metasedimentary)	-	-	-
12TS88	39.29850	70.68265	Fan-Karategin greenschist (?metasedimentary)	-	-	-
12TS89	39.29850	70.68265	Fan-Karategin greenschist (?metasedimentary)	-	-	-
12TS94	39.31452	70.65002	Fan-Karategin greenschist (meta-andesite)	iZrn 450 ± 5 ^e	iZrn -3.5 ^e	X _{Amp}
12TS112	39.17175	71.10832	Fan-Karategin Bt schist (?metasedimentary)	-	-	-
13TS17	39.13960	68.85925	Fan-Karategin blueschist (metabasite)	-	-	X _{Amp} , Bt, Ep
<i>Cretaceous Sandstones</i>						
12TS55	39.00822	70.05650	Turonian (94-90 Ma) sandstone	dZrn ^e	dZrn 0.0 ^e	-
12TS57	39.01948	70.22380	Barremian-Aptian (131-113 Ma) sandstone	dZrn ^e	dZrn +4.5/+0.5 ^e	-
<i>Modern River Sands</i>						
12TS32RS	39.05608	70.11232	Sangikar river sands	dZrn ^e	-	-
12TS54RS	39.14775	70.18965	Sorboq river sands	dZrn ^e	-	-
12TS107RS	39.17367	70.82340	Yasman river sands	dZrn ^e	dZrn -4.2 ^e	-
13TS29RS	39.18763	68.57669	Yagnob river sands	dZrn ^h	-	-

^aMineral abbreviations follow *Whitney and Evans* [2010]. i, igneous zircon; d, detrital; m, metamorphic zircon.

^b(U-Th)/Pb geochronology. Numerical values indicate (re)crystalliation ages (Ma).

^czircon Lu/Hf geochemistry. Numerical values indicate mean $\varepsilon_{\text{Hf}}(t)$.

^dEPMA, electron probe microanalyzer data. Numerical values indicate pressure (kbar) and/or temperature (°C) estimates. Al-in-Hbl, Al-in-Hbl barometry; Amp-Pl, Amp-Pl thermometry; Zr-in-Ttn, Zr-in-Ttn thermometry; GARb, Grt-Bt thermometry; GASP, Grt-Als-Pl-Qz barometry; PS, pseudosection; X_{mineral}, additional mineral composition data acquired.

^eAnalyzed using Nu Plasma LA-MC-ICPMS (Faraday mode).

^fAnalyzed using laser-ablation split stream ICPMS.

^gAnalyzed using Nu Plasma LA-MC-ICPMS (ion-counter mode).

^hAnalyzed using Element2 LA-SC-ICPMS (Faraday mode).

1.1.3. N-S Subdivisions of the Tian Shan: Sutures and Terranes

Several suture zones transect the Tian Shan E-W (Figure 1) and provide a vestige of component ocean basins of the vast Paleo-Asian ocean, the closure of which defines the orogen's first-order architecture (Figure 1) [e.g., *Burtman*, 2010]. Whereas the Terskey and Yili sutures represent ocean closure during the middle Paleozoic, the Turkestan, Junggar, and Gissar sutures represent late Paleozoic closure of individual ocean

basins that comprise the broader closure of the Paleo-Asian ocean. The Turkestan (South Tian Shan) ocean basin arose in the Neoproterozoic [Burtman, 2010] and/or the Cambrian [Dolgoplova et al., 2016], spanned ~1800 km during the Devonian [Klischevich and Khranov, 1994], and was subducted northward under the Kazakh-Kyrgyz continent in two phases during the Early Silurian–Middle Devonian and during the Carboniferous [Seltmann et al., 2011; Dolgoplova et al., 2016; Konopelko et al., 2016]. The Turkestan (South Tian Shan) suture records the late Paleozoic, diachronous closure of the Turkestan ocean basin via northward subduction [Xiao et al., 2013, and references therein], and the accompanying collision between the Tarim-Karakum (Tarim-Alai) terrane(s) in the south and the Kazakh-Kyrgyz continent (Middle Tian Shan; mainland Asia) in the north. The Gissar suture, which is comparatively localized (along-strike) and situated at the southwestern extremity of the Tian Shan, represents the short-lived Gissar back-arc ocean basin, which opened and closed during the Carboniferous [Portnyagin, 1974; Brookfield, 2000; Burtman, 2010; Konopelko et al., 2016].

Most studies [Gao et al., 2009; Biske and Seltmann, 2010; Charvet et al., 2011; Seltmann et al., 2011; Biske et al., 2013] have adopted a North(east)–Middle–South naming scheme for the terranes separated by the Tian Shan sutures. Others [Burtman, 2010; Xiao et al., 2013], however, refer to terrane and arc names individually to emphasize the orogen's tectonic diversity. As with any large orogen that spans many countries, terrane correlation is difficult and compounded by geologic complexity and the diversity of names and terrane definitions that naturally results from a diverse set of research groups.

1.1.4. The South Tian Shan

The South Tian Shan (STS) is a 2500-km-long accretion-collision complex that developed along the southern margin of the Kazakh-Kyrgyz continent during the late Paleozoic via subduction and closure of the Turkestan ocean. The STS is bounded to the north by the Turkestan suture and comprises, from east to west, the Kumishi (Kumish-tala), Kokshaal, Alai, Gissar, and Kyzylkum segments (Figure 1). The STS extends farther E–W than the other Tian Shan terranes before pinching out along-strike, which is perhaps best reflected by the strike length of the Turkestan suture: its surface exposure stretches from as far east as north-central Xinjiang (China, ~88°E), through Kyrgyzstan and Tajikistan to central Uzbekistan (63–64°E), perhaps as far west as the Uzbek Sultan-Uvais range (~60.5°E). Geophysical evidence indicates that the Turkestan suture extends west of its surface exposure, under Cenozoic sedimentary rocks, as far as the Urals [Burtman, 1975; Brookfield, 2000; Puchkov, 2009, 2013; Xiao et al., 2013].

It is unclear whether the Karakum and Tarim terranes, which are separated by the Talas-Fergana fault in the north and overridden by the Miocene Pamir salient in the south, were part of a single cratonic terrane or two separate terranes that docked to mainland Asia during the late Paleozoic. Tectonic reconstructions in which the Karakum terrane (variably grouped with or referred to as the “Alai terrane”) comprises a long, narrow [Xiao et al., 2010, 2013, 2015], consistently wide [Goldfarb et al., 2014; Mühlberg et al., 2016], or paleogeographically undefined [Burtman, 2010; Seltmann et al., 2011] westward prolongation of the Tarim terrane have emphasized Tarim-Karakum connectivity prior to collision. Alternatively, Bakirov and Maksumova [2001] and Alexeiev et al. [2009a] consider the Tarim and Karakum/Alai terranes not to have been connected prior to collision. In particular, Louri et al. [2016] advocated a scenario in which the terranes were originally separate before being united by the TFF, which evolved from an oceanic transform fault into an intracontinental strike-slip fault during the late Paleozoic collision.

The Kokshaal segment of the South Tian Shan is notably characterized by the southward emplacement of high-ultrahigh-pressure / low-temperature ((U)HP/LT) rocks over an accretionary complex that developed along the southern margin of the Yili-Middle Tian Shan terrane (mainland Asia) during the late Paleozoic [Xiao et al., 2013]. Two blueschist–eclogite localities separated by ~500 km along strike of the Turkestan suture (Figure 1) in Kyrgyzstan (Atbashi locality [Hegner et al., 2010; Li et al., 2011]) and China (Kebuer-te-Habutengsu locality [Wei et al., 2009; Zhang et al., 2009; Su et al., 2010; Tian and Wei, 2012; Du et al., 2014]) record nearly synchronous (320–305 Ma) (U)HP metamorphism of oceanic crustal and pelitic–felsic rocks. Overlapping, precise (321–313 Ma, garnet + whole-rock Sm–Nd) [Mühlberg et al., 2016] and younger, less precise (316–286 Ma, allanite U–Pb) [Louri et al., 2016] ages for HP/LT subduction metamorphism in the eclogite-facies Kassan metamorphic complex in the Kyrgyz Chatkal range in the Middle Tian Shan suggest a similar geodynamic evolution west of the Talas-Fergana fault that was synchronous and/or younger than that in the Kokshaal STS. Separation of the Kassan metamorphic complex ~50–100 km west of the contorted Turkestan suture (rather than along the linear trace of the Turkestan suture, as for the Kokshaal HT/LP exposures) may reflect allochthonous horizontal transport due to late Paleozoic shortening. Alternatively, an

~450 Ma origin for the HP/LT Kassan metamorphic complex due to closure of the Karaterek basin via southward subduction [Alexeev *et al.*, 2016] is also compatible with its geographic separation from the Turkestan suture. The composite Kokshaal (U)HP belt and the potentially analogous Kassan metamorphic complex represent the most deeply exhumed exposures along the Turkestan suture and its bounding terranes. Rapid post-peak-pressure exhumation of (U)HP/LT rocks in all localities [Li *et al.*, 2011; Mühlberg *et al.*, 2016] is consistent with significant kinematic reorganization in response to continental collision.

Late Paleozoic, HT/LP granitoid-migmatite-gneiss complexes have been documented in the Kokshaal, Gissar, and Kyzylkum segments of the STS (Figure 1). Li and Zhang [2004] advocated a counterclockwise pressure-temperature-time evolution for a Crd + Grt + Sil paragneiss from the Muzhaerte locality (footwall of the overthrust Kebuerte-Habutengsu HP/LT belt) in the Chinese Kokshaal STS. Whereas these investigators interpreted a 299 ± 5 Ma age (zircon U-Pb) to represent granulite-facies peak metamorphism, Guo and Zhang [2009] interpreted a younger, 280 ± 8 Ma age (EPMA-based monazite U-Th-Pb) to represent this metamorphic episode. Despite a 5–40 Myr lag time, Tian and Wei [2012] advocate the hypothesis that the Muzhaerte HT/LP complex and the adjacent (U)HP/LT Kebuerte-Habutengsu locality represent paired metamorphic belts. Amphibolite facies, Barrovian-type metamorphism, and migmatization in the Garm massif in the Tajik Gissar STS (our study area; discussed below) was previously bracketed between 299–290 Ma [Konopelko *et al.*, 2015] and 305–270 Ma [Käbner *et al.*, 2016b]. Comparatively old, Visean–Serupkhovian HT/LP metamorphism (~343–322 Ma, zircon U-Pb SHRIMP) in the Lolabulak migmatite-granitoid-gneiss complex ~100 km to the west of our study area [Mirkamalov *et al.*, 2012], which Konopelko *et al.* [2016] note overlaps with that in the Baysunta migmatite-granitoid-gneiss complex [Abdulhameed *et al.*, 2014] ~150 km west-southwest of our study area, is of unclear tectonic origin and relationship to the development of the lithologically similar Garm massif.

1.1.5. Timing of Collision Between the Tarim-Karakum and Kazakh-Kyrgyz Continents

Four lines of evidence have been employed to bracket the timing of the collision between the Tarim and Kazakh-Kyrgyz (Middle Tian Shan) continents, which Han *et al.* [2011] note the diachronous, westward younging, “scissors-like” occurrence of during the late Carboniferous (~320–300 Ma) in the Kokshaal-Kumishi South Tian Shan. Ages of ophiolitic-mélange rocks provide maximum-age constraints for the collision, whereas post-collisional, A-type, “stitching” intrusions (discussed below) provide minimum-age constraints. Ages for (U)HP, eclogite-bearing complexes in the Kokshaal STS (along the Turkestan suture) are generally interpreted to reflect early stages of collision. Sedimentologic collisional-timing constraints for the formation of the Turkestan suture include thick upper Carboniferous–lower Permian (Bashkirian–Asselian) foredeep turbidites and molasses [Biske, 1995, 1996; Biske and Seltmann, 2010], which span the collisional timeframe. It is uncertain and contentious whether or not the westward younging trend of the closure of the Turkestan ocean can be meaningfully extrapolated farther west into the Alai-Gissar-Kyzylkum South Tian Shan.

1.1.6. Subduction-Related and Post-Collisional Late Paleozoic Magmatism

Spatially widespread, late Paleozoic, and compositionally diverse magmatism across most of the Tian Shan (Figure 1) provides a record of the significant geodynamic changes that accompanied the final amalgamation of the CAO. Late Carboniferous, calc-alkaline, subduction-related, granitoid intrusive suites and their volcanic counterparts are predominantly distributed west of the Talas-Fergana fault. The Chatkal-Kurama arc in the Middle Tian Shan was generated by northward subduction of the Turkestan ocean basin [Seltmann *et al.*, 2011; Konopelko *et al.*, 2016], whereas the Gissar arc in the Gissar segment of the South Tian Shan was generated by northward subduction of the Gissar ocean basin [Brookfield, 2000; Käbner *et al.*, 2016b; this study].

Early Permian post-collisional intrusions are distributed across the entire Tian Shan and across suture zones and are widely attributed to collision between the Karakum-Tarim and Kazakh-Kyrgyz (Middle Tian Shan) continents [Konopelko *et al.*, 2007; Biske and Seltmann, 2010; Charvet *et al.*, 2011; Glorie *et al.*, 2010, 2011; Han *et al.*, 2011; Ren *et al.*, 2011; Biske *et al.*, 2013; De Grave *et al.*, 2013; Mao *et al.*, 2014]. Post-collisional magmatism in the Kyzylkum-Gissar-Alai STS is generally represented by larger igneous bodies of more diverse composition than in the Kokshaal STS [Seltmann *et al.*, 2011]. The Kyzylkum-Gissar-Alai STS rocks are dominantly A-type, alkali-calcic granitoids, with compositional diversity illustrated by silica-undersaturated and/or mafic alkaline magmatic rocks, as well as high-K (I-type) and peraluminous (S-type) granitoids [Seltmann *et al.*, 2011]. Post-collisional magmatism related to the Junggar-Yili collision began earlier in the northeastern Tian Shan, as evidenced by late Carboniferous plutons that cut the slightly older Junggar suture zone [Han *et al.*, 2010]. The post-collisional intrusions commonly bear juvenile geochemical signatures, which many investigators

relate to the role of lithosphere-scale, strike-slip shear zones and/or lithospheric removal in focusing and controlling mantle-derived heat and magma [Konopelko *et al.*, 2007; Han *et al.*, 2010; Konopelko *et al.*, 2011; Seltmann *et al.*, 2011; Mao *et al.*, 2014].

1.2. The Tajik Southwestern Tian Shan

The southwestern Tian Shan is an $\sim 600 \times 100$ km region of high topography that transitions into the Fergana and Tajik basins to the north and south, and into the Karakum-Kyzylkum basin to the west (Figure 1). The orogen is bounded by the TFF in the east and by sutures in the north and south. It comprises three primary lithotectonic units (Figure 2), following Brookfield [2000], from south-north: (i) the Gissar arc unit; (ii) the Zeravshan subduction-accretion complex; and (iii) the Turkestan-Alai flysch complex. The Turkestan suture to the north is cryptic but regionally significant; the Gissar suture to the south is well exposed but probably localized to the southwestern Tian Shan. The Baysunta [Brookfield, 2000; Abdulhameed *et al.*, 2014] and Lolabulak [Mirkamalov *et al.*, 2012; Dolgoplova *et al.*, 2016] HT/LP migmatite-granitoid-gneiss complexes (Figure 1) are situated ~ 150 km west-southwest and ~ 100 km west of our study area in Uzbekistan and comprise apparently similar lithologies to the Garm massif but are of presently ambiguous tectonic origin. The Tajik basin south of the Gissar suture exposes Mesozoic–Cenozoic strata within a fold-and-thrust belt related to the late Cenozoic development of the Pamir salient. Tectonic slivers of Mesozoic–Cenozoic strata are also exposed in the southwestern Tian Shan. The Fergana basin to the north exposes mostly late Cenozoic sedimentary rocks.

1.2.1. The Gissar-arc Unit

We divide the Gissar-arc unit into three genetically related subunits: (i) the Gissar batholith, (ii) the Garm massif, and (iii) the Gissar suture zone.

The Gissar batholith comprises a late Carboniferous–early Permian calc-alkaline suite of intermediate–felsic granitoids (Figure 2) and represents a magmatic arc that was generated by northward subduction of the Gissar back-arc ocean basin [Portnyagin *et al.*, 1976; Brookfield, 2000; Käbner *et al.*, 2016b; Konopelko *et al.*, 2016]. The Gissar-batholith granitoids, which intruded Ordovician–lower Permian strata of the Zeravshan complex, extend ~ 60 km west of our study area into Uzbekistan and comprise an $\sim 250 \times 20$ – 40 km, ~ 6000 km², E–W-oriented exposure overall (Figure 1). Granitoids of similar age extend ~ 400 – 700 km farther west into Uzbekistan [e.g., Dolgoplova *et al.*, 2016], although tectonic origins and relationship to the Gissar batholith are not clear. In addition to prevalent intrusive contacts, Soviet 1:200,000 geologic maps [Yushin *et al.*, 1964] indicate discontinuous, E–W-striking faults along the northern boundary of the Gissar batholith, which are likely part of a wider system of $\sim$ E–W-striking, late Cenozoic thrust and dextral strike-slip faults that modified the Paleozoic framework of the southwestern Tian Shan [Käbner *et al.*, 2016a]. Apparent northward tilting and back rotation of a block of northern Gissar-batholith crust along a Late Cenozoic, north-dipping, south-vergent thrust fault is evidenced by a trend of increasing Al-in-hornblende granitoid-emplacement pressures (deeper crustal levels) toward the thrust fault [Käbner *et al.*, 2016b] and highlights the prevalence of late Cenozoic thrust faulting within the Gissar batholith itself.

We interpret a suite of intermediate–felsic volcanic rocks, which is assigned an early Permian age in Soviet 1:200,000 geologic maps [Yushin *et al.*, 1964] and discontinuously exposed across $\sim 50 \times 10$ km (E–W) northwest of Dushanbe, to represent the volcanic component of the Gissar arc and group it with the Gissar batholith for simplicity (Figure 2). Sedimentologic evidence indicates a Middle Carboniferous age (Bashkirian–lower Moscovian) for Gissar-arc intermediate–felsic volcanic rocks [Baratov, 1976], which may represent an incomplete or erroneous age assignment in the Soviet maps. We are not aware of modern isotopic ages for the Gissar arc volcanic rocks.

The Garm massif (light pink in Figure 2) is an E–W-trending $\sim 100 \times 20$ km, ~ 1000 km² exposure consisting of: (i) discontinuous outcrops of amphibolite–granulite-facies paragneisses and orthogneisses with felsic–mafic bulk compositions; (ii) variably deformed granitoids that commonly contain igneous garnet; (iii) migmatites; (iv) feldspathoid syenites; (v) aluminous schists; and (vi) minor marble and quartzite [Konopelko *et al.*, 2015]. The northern boundary of the massif is a complex, north-dipping, south-vergent thrust zone that juxtaposes the greenschist–blueschist-facies Fan-Karategin belt, which is a component of the Zeravshan complex (discussed below), structurally over the Garm massif [Konopelko *et al.*, 2015]. This contractional system must have been active during the late Cenozoic because it includes tectonic slices of Jurassic–Neogene strata, but

an initial late Paleozoic component of movement is also possible. To the south, the Garm massif is unconformably overlain by Cretaceous sandstones of the Tajik basin.

The late Carboniferous Gissar suture zone bounds the Gissar batholith to the south (Figure 2) and represents a vestige of the short-lived Gissar back-arc ocean basin, which opened along the northern passive margin of the Karakum terrane during the early Carboniferous and closed during the Middle [Baratov, 1976] or Late [Burtman, 1975; Brookfield, 2000; Burtman, 2010] Carboniferous. Early Carboniferous rifting of the Gissar microcontinent [cf. Konopelko *et al.*, 2016] off of the northern Karakum terrane and subsequent opening of the Gissar back-arc ocean basin may be related to southward subduction of the Turkestan ocean basin [e.g., Burtman, 2010] or northward subduction of the Paleo-Tethys ocean basin [Biske, 1996; Biske and Seltnann, 2010; Dolgoplova *et al.*, 2016], the southern active margin of which corresponds to contemporaneous arcs in the northern Pamir and Kunlun. The Gissar suture zone is defined by an ophiolitic mélange consisting of lower(?) Carboniferous basalts and serpentinized ultramafic fragments and overlying carbonate and clastic sedimentary rocks. Transitional ocean ridge-supra-subduction geochemical signatures for middle Carboniferous gabbro-plagiogranite (gabbro-tonalite) intrusions from the Gissar suture zone [Konopelko *et al.*, 2016] reflect primarily mantle-derived origins collectively compatible with a back-arc, hyperextensional forearc, and/or supra-subduction setting(s). We interpret the Gissar suture zone to encompass exposures of lower-middle Carboniferous mafic volcanic and associated sedimentary rocks delineated on Soviet 1:200,000 geologic maps (C₁₋₂) [Yushin *et al.*, 1964].

A suite of early-middle Permian, alkaline granitoids comprises lobate, slightly ~E-W-elongated intrusions in the eastern part of our study area (orange in Figure 2). One small pluton (~3 × 4 km surface exposure) occurs within the Garm massif, and generally larger plutons are exposed north of the Garm massif within the Zeravshan subduction-accretion complex and the Turkestan-Alai flysch complex. Intermediate-felsic, Gissar-type and -age intrusions are found among the alkaline intrusions and are somewhat larger overall but comprise the same scattered map pattern as the alkaline intrusions. Both types of isolated intrusions continue ~200 km east of our study area into Kyrgyzstan and represent a significant contrast to the connected, linear, E-W-oriented map pattern of the Gissar batholith. The alkaline intrusions likely represent “post-collisional” magmatism documented in the Kyrgyz Tian Shan [Konopelko *et al.*, 2007; Seltnann *et al.*, 2011; DeGrave *et al.*, 2012], which immediately post-dates Gissar magmatism.

1.2.2. The Zeravshan Subduction-Accretion Complex and Fan-Karategin Belt

The Zeravshan subduction-accretion complex (Figure 2) comprises a package of early-middle Paleozoic, active- and passive-margin (volcano-)sedimentary sequences of the Karakum continent that were incorporated into an accretionary wedge along the northern margin of the continent. Accretion apparently occurred during the Visean (Early Carboniferous; 347–331 Ma), as indicated by an angular unconformity above folded Silurian–Devonian strata near the Jijikrut mine in the center of the complex (Figure 2) [Baratov, 1976; Biske, 1996]. Younger Paleozoic shortening structures in the Zeravshan complex, which imbricate strata and disrupt primary stratigraphic relationships, reflect a shift from late Carboniferous accretion to collision. The Zeravshan complex is bounded to the south by a late Paleozoic intrusive contact with the Gissar unit, which was overprinted by late Cenozoic thrust and/or dextral strike-slip faults [Käbner *et al.*, 2016a]. The northern boundary of the Zeravshan complex is cryptic in Tajikistan but clearly exposed ~200 km along strike to the west of our study area in the Zirabulak mountains in Uzbekistan, where the Carboniferous (?) Zirabulak ophiolitic mélange separates the Zeravshan subduction-accretion complex from the Turkestan-Alai flysch complex (discussed below) to the north [Kukhtikov, 1969]. We follow Brookfield’s [2000] eastward correlation of the Zirabulak ophiolitic mélange into Tajikistan, where it may coincide with the Zeravshan fault.

The Fan-Karategin HP/LT belt (Figures 1 and 2; known locally as the Yagnob, Norvat, and Gorif formations) comprises the metamorphosed, greenschist–blueschist facies component of the Zeravshan subduction-accretion complex and is bounded by the Zeravshan fault to the north and by the Gissar-Karategin fault to the south [Volkova and Budanov, 1999]. Blueschist-facies assemblages in the Fan-Karategin belt occur as sheets or small blocks/lenses that are interlayered with a dominantly greenschist-facies matrix. Peak metamorphic pressure-temperature estimates are 8 kbar/450°C (greenschist-epidote-blueschist facies), and the dominant greenschist retrogressive overprint was estimated at 3–5 kbar/300–400°C [Minaev, 1982; Minaev and Fedorov, 1984]. The protoliths for the high-pressure rocks are primarily oceanic basalts, with comparatively rare andesite, quartzite, pelite, and limestone protoliths that represent the overlying succession.

Although isotopic geochronologic constraints for the timing of HP/LT metamorphism in the Fan-Karategin belt are not available, stratigraphic constraints for metamorphism of the Yagnob formation indicate a lower–middle Visean age (early Carboniferous, ~347–337 Ma) [Shvanov, 1983] that is consistent with the purported Visean timing of Zeravshan-complex accretion [~347–331 Ma; Baratov, 1976; Biske, 1996]. We favor this early–middle Visean age assignment over Volkova and Budanov's [1999] speculative Silurian age assignment for Fan-Karategin HP/LT metamorphism.

1.2.3. The Turkestan-Alai Flysch Complex

The Turkestan-Alai flysch complex (Figure 2) comprises a Paleozoic stratigraphy that is coeval with that in the Zeravshan complex [Brookfield, 2000] and distinguished from it by an abundance of flysch and a scarcity of metamorphic rocks. This primarily active-margin complex comprises Silurian and upper Carboniferous flysch (including turbidite sequences), as well as Devonian–Middle Carboniferous limestones, all of which were deposited in the Turkestan ocean basin and tectonically consolidated during its closure, but not incorporated into the Zeravshan subduction-accretion complex. The Turkestan suture bounds the Turkestan-Alai flysch complex to the north and the Zirabulak ophiolitic mélange bounds it to the south.

2. Petrography and Mineral Chemistry

We investigated petrographic thin sections of igneous and metamorphic rock samples (summarized in Table 1) to identify mineral assemblages and textures. We use Whitney and Evans' [2010] mineral abbreviations to describe mineral assemblages. We identified and characterized minerals for a representative subset of samples using energy- and wavelength-dispersive X-ray spectroscopy detectors (EDS and WDS, respectively) on a Cameca SX100 electron probe microanalyzer (EPMA) at the Michael J. Drake Electron Microprobe Laboratory at the University of Arizona. Mineral assemblages and compositions for the igneous rocks are summarized in Tables S1 and S2 those for the metamorphic rocks are summarized in Tables S3 and S4. Oxide abundances for the full mineral chemistry data set are provided in Data Set S1 in the supporting information. Calculated mineral formulas are provided in Tables S5–S10. Mineral chemistry analytical and methodological procedures are summarized in Text S1 [Pouchou and Pichoir, 1991; Leake et al., 1997; Holdaway, 2000].

2.1. Results: Igneous Petrography and Mineral Chemistry

Granitoids from the Gissar unit (henceforth the “granitoid suite”) are generally felsic (15–40% Qz) and have low abundances of mafic minerals (0–15% Bt + Amp; Table S1). Those with moderate mafic-phase abundances (15–30% Bt + Amp) have an intermediate-felsic composition generally with more plagioclase (33–55%) and less quartz (10–20%). Modal abundances of quartz, plagioclase, K-feldspar, and feldspathoids for all the granitoids were normalized in a QAPF scheme (Figure 4) which exhibits the compositional diversity of the granitoid suite. Biotite is present, and the most abundant minor mineral (up to 25%), in all samples. Calcic amphibole is less common, occurring in only four of the granitoids. Nondeformed garnet occurs as a minor mineral in two samples (12TS30 and 12TS91) and signifies a peraluminous bulk composition, particularly for sample 12TS30 (Figure 3a), in which garnet is coarse grained (2–3 mm). The compositional diversity of the granitoid suite is characteristic of an evolved magmatic arc [e.g., Ducea et al., 2015]. Perthite and myrmekite are common in all samples. Primary opaque accessory phases include ilmenite, hematite, and pyrite. Primary, euhedral, and anhedral titanite occurs in three granitoids as a primary magmatic phase.

The Gissar granitoids exhibit varying degrees of alteration, which is primarily indicated by feldspar sericitization. Secondary chlorite appears both as neoblastic grains and as a partial replacement of biotite and amphibole. Muscovite occurs exclusively as a secondary phase with the exception of 12TS121, where it is primary. Secondary epidote is rare except in 12TS111, which contains 3% epidote and is the most altered granitoid in this study.

All but two of the granitoids are non-deformed and lack preferred orientation of any mineral. The exceptions are Garm granitoid samples 12TS23A and 12TS102, which are weakly deformed, as exhibited by weakly foliated biotite and dynamically recrystallized quartz. 12TS23A is from the northern periphery of the Garm massif, and 12TS102 is from the interior (Figure 2). Both samples exhibit low strain and are nonmetamorphosed.

Figure 3. Photomicrograph images of igneous and metamorphic rocks in this study. (a) Sample 12TS30, Garm Grt + Bt granodiorite (plane-polarized light). All phases are randomly oriented. Igneous Grt indicates a peraluminous bulk composition. Anhedral Grt grain boundary indicates resorption. (b) Sample 12TS96, Garm Bt + Cal Ne + Ccn syenite (cross-polarized light). Mineral assemblage indicates crystallization from a silica-undersaturated melt with CaCO_3 . (c) Sample 12TS47, Garm amphibolite-facies metabasite (cross-polarized light), highlighting granoblastic texture and mafic mineral assemblage (Pl + Amp + CPx). (d) Sample 12TS59, Garm amphibolite-facies mafic-intermediate gneiss (plane-polarized light). Matrix primarily comprises Act + Phl groundmass. (e) Sample 12TS62, Garm granulite-facies paragneiss (plane-polarized light). Grt is strongly resorbed and replaced by pinites, with Spl + Ilm likely tracing the former, peak-temperature, pre-resorption grain boundary. Relict Crd occurs within pinites aggregates. (f) Sample 13TS17, Zeravshan blueschist strongly overprinted by greenschist-facies mineral assemblages (plane-polarized light). Relict Mrbk-Wnc indicates blueschist-facies pressures. (g) Sample 12TS94, Fan-Karategin greenschist-facies meta-andesite (cross-polarized light). (h) Sample 12TS38, Fan-Karategin greenschist with detrital Ttn (plane-polarized light).

Calcic amphibole occurs in four of the granitoids (Table S1), three of which (12TS09, 12TS111, and 12TS114) were analyzed with EPMA (Tables S2 and S5 and Figure S1a). For Gissar diorite sample 12TS09, calcic amphibole in contact with plagioclase was analyzed both for amphibole rims and for amphibole cores that enclose inclusions of plagioclase. Amphibole in both contexts is Fhb-Fts, although average Al_2O_3 is $\sim 2.3\%$ greater for rims than for inclusions. For altered Garm monzonite sample 12TS111, fresh amphibole is Ts-Mhb-Prg and altered amphibole is uniformly Mhb. Fresh amphibole is also distinguished from altered amphibole in this sample by its higher SiO_2 , MgO, CaO, and lower TiO_2 , Al_2O_3 , FeO (total), NaO, and K_2O . Amphibole in this sample has a higher $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ overall than amphibole in the other granitoids. For Garm granite sample 12TS114, amphibole is Ts-Mhb-Prg and contains the most Al_2O_3 ($\sim 13\%$) of amphibole in these three samples. Amphibole rim analyses for 12TS09 (and for 12TS114 to a lesser extent) exhibit a pronounced negative correlation between $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ and Si atoms per formula unit (Figure S1a).

Plagioclase occurs in varying abundances in all of the granitoids (9–55%; Table S1) and was analyzed with EPMA in the same samples (12TS09, 12TS111, and 12TS114; Tables S2 and S6). For Gissar diorite sample 12TS09, plagioclase inclusions within amphibole ($\text{Ab}_{59}\text{An}_{40}\text{Or}_1$) are more mafic than matrix plagioclase ($\text{Ab}_{70}\text{An}_{30}\text{Or}_1$), which is consistent with protracted plagioclase crystallization from a fractionating melt. For Garm monzonite sample 12TS111, plagioclase composition ($\text{Ab}_{52}\text{An}_{47}\text{Or}_1$) for plagioclase in contact with fresh amphibole varies more than in the other samples but does not correlate with whether the grains are in contact with fresh or retrogressed amphibole, as indicated by oxide weight percentages that are within error of each other in both contexts. For Garm granite sample 12TS114, plagioclase composition ($\text{Ab}_{60}\text{An}_{39}\text{Or}_1$) is similarly felsic to that for matrix plagioclase in 12TS09.

Garm granodiorite sample 12TS30 is unique among the granitoid suite in that it contains 2% igneous garnet (Figure 3a and Table S1). The garnet is resorbed and exhibits increasing Mn and $\text{Fe}^{2+}/(\text{Fe}^{2+} + \text{Mg})$ toward rims (Figure S2), which is a typical symptom of retrograde garnet dissolution [e.g., Kohn and Spear, 2000]. On average, garnet cores are $\text{Alm}_{69}\text{Prp}_{24}\text{Sp}_{5}\text{Gr}_{5}$ and rims are $\text{Alm}_{75}\text{Prp}_{17}\text{Sp}_{5}\text{Gr}_{4}$ (Tables S2 and S7). Sample 12TS30 contains the greatest abundance of biotite in the granitoid suite (25%, Table S1). The Fe# for matrix biotite (Fe_{49} mean value) is within 1σ of biotite in contact with garnet (Fe_{48} mean value) (Tables S2 and S8).

Figure 4. QAPF diagram for Gissar/Garm igneous rocks, color coded by age to explore the correlation (or lack thereof) between increasingly alkaline composition and decreasing age (purple to orange). The correlation is weak; nevertheless, the oldest sample is a typical granite and the youngest sample is a feldspathoid syenite. Granitoids from the Gissar batholith have a dashed black outline; those from the Garm massif have a solid black outline. Modal abundances of Qz, Pl and Kfs were estimated by point-counting thin sections of each sample and normalized after Streckeisen [1979]. “ $e_{\text{Hf}} \times x \pm x$ ” indicates mean $e_{\text{Hf}}(t) \pm 2\sigma$ standard deviation for samples 12TS12, 12TS30, 12TS96, and 12TS111.

Samples 12TS96 and 12TS91 are from kilometer-scale alkaline intrusions (Figure 2) that intrude the Garm massif and the Zeravshan complex, respectively. Sample 12TS96 (Figure 3b) is a feldspathoid syenite and is unique within the granitoid suite with regard to its primary feldspathoid (nepheline + cancrinite) and calcite mineralogy (Table S1). The copresence of the feldspathoids and calcite indicates crystallization from a silica-undersaturated melt containing dissolved CaCO_3 . Cancrinite commonly occurs as rims surrounding mantles of altered K-feldspar. Grt + Qtz syenite sample 12TS91 resembles sample 12TS96 in its high degree of alkalinity and provides a silica-saturated analog to it. Garnet in 12TS91 is strongly resorbed.

2.2. Results: Metamorphic Petrography and Mineral Chemistry

Mineral assemblages for six gneisses from the Garm massif are summarized in Table S3 and mineral chemistry is summarized in Table S4. Detailed mineral chemistry results are provided in Tables S5–S10. Relevant photomicrograph images are provided in Figure 3 and further petrographic details are provided in the appendix.

2.2.1. Garm Gneisses: This Study

Two Garm gneisses (12TS47 and 12TS59) have an Amp + CPx + Pl mineral assemblage, which indicates amphibolite-facies metamorphism of a mafic-intermediate protolith (Tables S3 and S4). Four mineralogical observations of Garm gneiss sample 12TS47 (Figure 3c) indicate a mafic bulk composition: (i) Mg-rich amphibole (Act–Mhb, Table S5), (ii) Ca-rich plagioclase (Ab_{39-48} , Table S6), (iii) abundant clinopyroxene ($\text{Hd}_{88}\text{Di}_{12}$, Table S10), and (iv) absence of quartz. Stability of actinolite, which is a major mineral, indicates low metamorphic pressures for 12TS47. Granoblastic texture and compositional segregation of plagioclase and Amp + CPx indicate that any significant deformation must have occurred before peak temperature conditions. For Garm gneiss sample 12TS59, three observations indicate an intermediate bulk composition: (i) the copresence of amphibole (Mhb–Ed; Table S5) and Na-rich plagioclase ($\text{Ab}_{80}\text{An}_{18}\text{Or}_2$, Table S6), (ii) minor clinopyroxene ($\text{Hd}_{74}\text{Di}_{26}$, Table S10), and (iii) absence of quartz. Amphibole and ilmenite occur as fine-grained, fresh porphyroblasts, whereas plagioclase occurs as coarser-grained, sericitized porphyroclasts (Figure 3d).

Garm paragneiss sample 12TS62 (Figure 3e; Appendix A) has a Grt + Bt + Sil + Pl + Qz + Spl + Ilm ± Crd (relict) ± Ms mineral assemblage (Table S3) with 24% pinitite (a retrogressive relict of cordierite and perhaps K-feldspar; Text S2) that, taken together with the absence of Rt, indicate granulite-facies peak metamorphic temperature and qualitatively low metamorphic pressures. This sample is strongly foliated and weakly annealed, which we interpret to indicate continued deformation during retrogression. Foliation is defined by: (i) layering of biotite and fibrolitic sillimanite; (ii) compositional layering of pinitite and stretched, oriented plagioclase and garnet; and (iii) aggregates of oriented, elongated spinel and ilmenite that are associated with sillimanite and pinitite (Figure 3e). Biotite is rarely oblique/perpendicular to the foliation, whereas muscovite (<1% modal abundance) is commonly oblique/perpendicular to the foliation. We interpret these textures to indicate a two-stage production of partial melt by pervasive muscovite-dehydration melting and significant migration of this melt out of the rock, followed by comparatively insignificant biotite-dehydration melting. Plagioclase and biotite occur both in garnet pressure shadows and in the matrix.

Garnet in sample 12TS62 is significantly resorbed and replaced by pinitized cordierite (Text S2) [e.g., *Ogiermann and Kalt*, 2000]. Linear spinel + ilmenite aggregates that are parallel to and spaced ~300 μm from garnet grain boundaries (Figure 3e) likely trace preresorption boundaries of formerly larger garnet grains. Garnet resorption is also indicated in EPMA data by increasing Mn and Fe²⁺/(Fe²⁺+Mg) toward rims [e.g., *Kohn and Spear*, 2000] (Figure S2 and Data Set S1). Contrast-sensitive backscattered electron (BSE) images for a 3 mm garnet indicate that the retrogressive re-equilibration front encroaches ~200 μm into the grain past its current grain boundary. Average garnet compositions are Alm₆₇Prp₁₈Sps₁₂Grs₃ for cores, Alm₆₉Prp₁₇Sps₁₁Grs₃ for mantles, and Alm₇₂Prp₁₄Sps₁₁Grs₃ for rims (Table S4). Spinel and ilmenite exhibit minor solid solution between their Fe²⁺ and respective Mg²⁺/Mn²⁺ end-members but are similarly Fe²⁺-rich; average formulas are Hc₈₄Spl₁₆ and Ilm₈₈Pph₁₂ (Table S10).

The textural evidence for garnet resorption and replacement by late (pinitized) cordierite qualitatively indicates a P-T-t path involving late decompression that may have accompanied continued heating. Two important end-member reactions in this regard, after *Ganguly et al.* [2000], involve the breakdown of garnet (almandine) to spinel (hercynite) and cordierite (Fe cordierite) and suggest a likely paragenetic sequence for sample 12TS62.

Reaction (R1) may have accompanied concomitant temperature increase and pressure decrease, whereas reaction (R2) is more attributable to decompression.

We estimated the following bulk composition for 12TS62 by integrating modal abundances (estimated by point counting 1500 grains), representative mineral chemistry (from EPMA analyses), and mineral densities: 56.1% SiO₂, 2.1% TiO₂, 19.7% Al₂O₃, 10.7% FeO + Fe₂O₃, 0.2% MnO, 3.8% MgO, 0.8% CaO, 1.5% Na₂O, 4.0% K₂O, and 1.3% H₂O. We partitioned the modal abundance of pinitite in the sample (24%) into 18% cordierite and 6% K-feldspar. Overall, our estimated bulk composition most closely resembles a greywacke protolith and/or restite, which is consistent with the mineral assemblage, textures, and implied granulite-facies peak metamorphic temperatures.

Three Garm gneisses (12TS34, 12TS73, and 12TS101) have a Bt + Grt + Pl + Qz mineralogy (Table S3) consistent with amphibolite-facies metamorphism of quartzofeldspathic-pelitic protoliths. The foliation is continuous and anastomosing in all samples and is primarily defined by biotite. Garnet in all samples is fine grained (<500 μm diameter) and strongly resorbed. For sample 12TS34, garnet cores exhibit higher Alm, Sps, and lower Prp than garnet rims (in contact with biotite): Garnet cores are Alm₇₀Prp₁₅Sps₁₂Grs₃, and garnet rims are Alm₆₇Prp₁₁Sps₁₉Grs₃ (Tables S4 and S7). The Fe# for biotite in contact with garnet (Fe# = 0.49) is very similar to that in the matrix (Fe# = 0.50) (Tables S4 and S8). For sample 12TS73, garnet cores exhibit higher Prp and lower Alm and Sps than garnet rims (in contact with biotite): garnet cores are Alm₇₃Prp₁₇Sps₇Grs₃ and garnet rims are Alm₇₇Prp₁₀Sps₁₀Grs₃ (Tables S4 and S7). The average biotite Fe# for 12TS73 is slightly higher for matrix biotite (Fe# = 0.50) than for biotite in contact with garnet rims (Fe# = 0.48) (Tables S4 and S8).

2.2.2. Garm Gneisses: Comparison to Konopelko et al. [2015]

Konopelko et al.'s [2015] St + Bt schist samples "11" and "12" from the eastern Garm massif are distinguished from our Garm paragneisses by the presence of staurolite and absence (sample 11) or trace abundance (sample 12) of garnet. These samples are similar to ours in their abundance of biotite and absence (sample 11) or trace abundance (sample 12) of K-feldspar. Konopelko et al. [2015] also investigated migmatized paragneisses and several migmatite varieties, which we observed generally in the field but did not investigate in detail. These samples comprise mainly quartz and plagioclase with varying portions of K-feldspar, biotite, muscovite, and chlorite and attest to partial melting of the Garm massif.

EPMA analyses for Konopelko et al.'s [2015] St + Bt schist sample 12 provide the opportunity to compare mineral chemistry with our study. The Fe# for garnet in sample 12 is 0.83–0.85, shows no correlation between core and rim analyses (which would indicate growth or retrograde zoning), and is similar in magnitude to the garnet Fe# in our samples (Table S7). The Fe# for biotite in sample 12 is lower for core analyses (0.49) than for rim analyses (0.50–0.52), which is similar in magnitude but the reverse of the pattern that we observe in our samples 12TS30 and 12TS73 for matrix biotite and biotite rims in contact with garnet (Table S8). Plagioclase in sample 12 exhibits zoning, with $\text{Ab}_{84}\text{An}_{16}$ for cores and $\text{Ab}_{77}\text{An}_{14}\text{Or}_9$ for rims. We observe the same pattern for plagioclase in garnet pressure shadows in sample 12TS62 but at lower Ab content ($\text{Ab}_{71}\text{An}_{26}\text{Or}_3$ for plagioclase cores and $\text{Ab}_{69}\text{An}_{28}\text{Or}_2$ for plagioclase rims in contact with garnet; Table S6).

2.2.3. Fan-Karategin Greenschists and Blueschists

We investigated eight greenschists and one blueschist (13TS17) from the Fan-Karategin greenschist-blueschist belt (the metamorphosed component of the Zeravshan subduction-accretion complex). The samples are generally dominated by a greenschist facies mineral assemblage: Chl + Wm + Ep + Qz (Table S3). The exception is 13TS17, which is discussed below.

Blueschist sample 13TS17 (Figure 3f) contains ~200 μm relict blue amphibole porphyroblasts, the cores of which comprise magnesioriebeckite (Figures 3f and S1b and Tables S4 and S5) and the rims of which comprise winchite–barroisite. Zoning, observed in BSE, correlates with this change in composition for individual grains. Three observations indicate epidote-blueschist metamorphic conditions (8–13 kbar, 400–500°C) for 13TS17: (i) magnesioriebeckite amphibole cores; (ii) the absence of porphyroblastic jadeite; and (iii) abundant, porphyroblastic epidote (epidote-blueschist facies, 400–500°C; Table S11). Winchite–barroisite amphibole rims and very fine grained actinolite–magnesiohornblende intergrown with phlogopite comprise the majority of the matrix in 13TS17 and signify a strong, greenschist-facies, retrogressive overprint. Coarse-grained (1–2 mm) clinopyroxene ($\text{Hd}_{98}\text{Di}_2$) porphyroclasts and mafic plagioclase (An_{99}) indicate a basaltic protolith. A late CaCO_3 phase crosscuts the primary and retrogressive fabric.

Andalusite in sample 12TS22 (Table S3) indicates low-pressure metamorphism of a pelitic protolith. This sample was collected near the contact between the Garm massif and the Zeravshan complex and may have undergone contact metamorphism.

We interpret the presence of metamorphic Act and Czo and absence of CPx in sample greenschist sample 12TS94 (Figure 3g and Table S4), taken together with igneous-zircon U-Pb data (discussed below in section 4.1.3), to indicate an andesitic protolith. Samples 12TS13, 12TS22, and 12TS38 (Figure 3h) have characteristic greenschist facies mineral assemblages and sedimentary protoliths (demonstrated by detrital-zircon and -titanite data, discussed in section 4.3). We tentatively interpret similar mineral assemblages for samples 12TS83, 12TS88, and 12TS89 to indicate sedimentary protoliths for these samples. Samples 12TS13 and 12TS112 are phyllonites; the latter is a low-grade, Bt + Qz schist with minor tourmaline.

3. Pressure-Temperature Estimates

We implemented thermobarometry to estimate granitoid-crystallization and metamorphic pressures and temperatures for the Gissar batholith and the Garm massif (section 3.1). Additionally, we implemented pseudosection modeling for Garm massif granulite-facies paragneiss sample 12TS62 to further constrain metamorphic pressures and temperatures and cross check thermobarometric results (section 3.2). Finally, we semi-quantitatively estimate metamorphic conditions based on mineral chemistry for Fan-Karategin blueschist sample 13TS17 (section 3.3).

Figure 5. Al-in-Hbl pressure [Anderson and Smith, 1995] and Amp-PI (Ed-Rct exchange) temperature [Holland and Blundy, 1994] estimates of granitoid crystallization in the Gissar batholith and the Garm massif. (a) Plot of pressure-temperature estimates for all non-altered analyses for all three samples in this study. Rim-pair, inclusion-pair, and core-pair estimates are respectively plotted as filled circles, hollow circles and filled triangles. Whole-sample error bars represent 2σ standard deviation of the non-weighted mean of individual pressure and temperature estimates. (b) Probability density plots (PDPs) of pressure and temperature estimates for all rim-pair estimates for all three samples in this study. Dashed PDP for 12TS09 represents core-pair analyses. (c) Normalized probability-density plots of Al-in-Hbl granitoid-crystallization-pressure estimates (non-weighted mean ± 2σ standard deviation) for the Gissar batholith (gray: this study) [Käbner *et al.*, 2016b] and the Garm massif (black: this study). Shaded, high-pressure component of Gissar-batholith PDP indicates contribution to net PDP from our sample 12TS09, which is older and has a notably higher mean-pressure value and a greater 2σ uncertainty than Käbner *et al.*'s [2016b] five Gissar-batholith granitoids. Corresponding depth axis calculated from pressure assuming 2.7 g/cm³.

3.1. Thermobarometry

We used the “amphibole-plagioclase” thermometer [Holland and Blundy, 1994] and the “Al-in-hornblende” barometer [Anderson and Smith, 1995] for three amphibole + plagioclase-bearing Gissar/Garm granitoids to estimate granitoid-crystallization temperatures and pressures (Figure 5 and Table S12 and Data Set S2). We also applied the amphibole-plagioclase thermometer to two Garm amphibolite-facies metabasites to estimate metamorphic temperatures (Figure 6 and Table S12 and Data Set S2). We used the “Zr-in-titanite” thermometer [Hayden *et al.*, 2008] to estimate granitoid-crystallization temperatures for one Garm granitoid (Table S12). Methodological procedures for the amphibole-plagioclase, Al-in-hornblende, and Zr-in-titanite thermobarometers are discussed in Text S3 [Hammarstrom and Zen, 1986; Holland and Blundy, 1994; Anderson and Smith, 1995; Hayden *et al.*, 2008].

We used the “garnet-biotite” thermometer (GARB) and garnet-aluminosilicate-plagioclase-quartz barometer (GASP) [Ganguly *et al.*, 1996] for Grt + Bt- and Grt + Sil + Pl + Qz-bearing rocks to estimate peak-metamorphic temperatures and pressures (Figure 6 and Table S12 and Data Set S3). GARB/GASP methodological procedures are discussed in Text S4 [Ghent, 1976; Koziol and Newton, 1988; Elkins and Grove, 1990; Patiño Douce *et al.*, 2009; Ganguly *et al.*, 1996; Berman, 1998; Holdaway, 2000; Kohn and Spear, 2000].

3.1.1. Results (Gissar/Garm Granitoids): Amp-PI Thermometry, Al-in-Hbl Barometry, and Zr-in-Titanite Thermometry

In total, we used 43 Amp-PI analysis pairs (35 rim pairs, 6 inclusion pairs, and 2 core pairs) to estimate crystallization temperatures and pressures for granitoid samples 12TS09, 12TS111, and 12TS114 (Figure 5 and Table S12 and Data Set S2). Fresh rim pairs for all samples produce clusters in pressure-temperature space (Figure 5),

Figure 6. Garnet-biotite (GARb) temperature estimates and garnet-aluminosilicate-plagioclase (GASP) pressure estimates for Garm paragneisses (a, c, and d) and Garm peraluminous granitoid (b). Amphibole-plagioclase (Amp-PI) temperature estimates, based on Ed-Ri exchange, for a Garm paragneiss (e) and a Garm metabasite (f). For GARb and GASP (Figures 6a–6d), solid-black lines indicate “scheme (i)” retrogressive-analysis pairs (minimum-temperature and minimum-pressure estimates) and colored dashed lines represent “scheme (ii)” non-retrogressive-analysis pairs. For Amp-PI temperature estimates (Figures 6e and 6f), red line indicates mean of individual analyses (black lines). For GARb and GASP, 2σ uncertainties for temperature and pressure were computed from the $\pm 2\sigma$ end-member values of the representative mean compositions for the preferred pairing scheme. For Amp-PI, 2σ uncertainties represent the 2σ standard deviation of temperature estimates for all analyses.

for which we report the mean and 2σ standard deviation of temperatures and pressures to represent granitic melt-crystallization conditions (Table S12). We also estimated pressure-temperature conditions for amphibole and plagioclase in other contexts including plagioclase inclusions in amphibole (“inclusion pairs”) and paired amphibole and plagioclase cores (“core pairs”) but do not emphasize the results.

Rim pairs for Gissar Amp + Bt Qz-diorite sample 12TS09 (the freshest of the three granitoids) define a population with a narrow range of temperatures and wide range of pressures (Figures 5a and 5b): $709 \pm 15^\circ\text{C}/6.5 \pm 2.4$ kbar (Table S12 and Data Set S2). The near-uniform temperature and ~ 4 -kbar range in pressures are consistent with isothermal decompression of the granitoid melt. Overall, the estimated pressure (6.5 ± 2.4 kbar) and temperature ($709 \pm 15^\circ\text{C}$) ranges for Gissar granitoid 12TS09 are intermediate between and overlap with those for Garm granitoids 12TS111 (5.9 ± 0.8 kbar; $733 \pm 34^\circ\text{C}$) and 12TS114 (7.9 ± 1.8 kbar; $686 \pm 41^\circ\text{C}$) with a comparatively restricted temperature range and wider pressure range. Assuming a 2.7 g/cm^3 average overlying crustal density, the 6.5 ± 2.4 kbar crystallization pressure estimate for Gissar granitoid sample 12TS09 corresponds to a 24 ± 9 km intrusion depth.

Sample 12TS09 also contains small ($\sim 100 \mu\text{m}$), An-rich plagioclase inclusions (Table S6) within larger amphibole grains, for which we selected 6 inclusion pairs of analyses near grain boundaries between the rims of included plagioclase and the adjacent, consistently low-Al domains (Table S5) of the hosting amphibole grains. Temperature estimates are higher, and pressure estimates are lower for the inclusion pairs with respect to the rim pairs (Figures 5a and 5b). Assuming a simple crystallization scheme where amphibole

interiors and plagioclase inclusions within them represent nuclei in an early stage of a progressively crystallizing granitic melt and amphibole and plagioclase rims represent a late stage, the higher temperature estimates for the inclusion pairs are consistent with a cooling magma but the corresponding lower pressures are inconsistent with a coevally decompressing magma. An alternative explanation, maintaining that the inclusion pairs predate the rim pairs, is a two-stage crystallization history of low-pressure cooling followed by 4–5 kbar of isothermal compression, which is consistent with rapid tectonic addition of ~15–19 km of overburden during crustal thickening related to the continental collision that occurred near this time. Another alternative explanation is that the low inclusion pair pressure estimates are an erroneous artifact of an evolving, fractionation-dependent melt composition for which *Anderson and Smith's* [1995] barometer does not account.

Rim pairs for Garm Amp + Bt + Ep Qz monzonite sample 12TS111 define a tightly constrained population with mean $\pm 2\sigma$ temperature/pressure estimates of $733 \pm 34^\circ\text{C}/5.9 \pm 0.8$ kbar (Figures 5a and 5b and Table S12 and Data Set S2). Paired analyses from an amphibole core and a plagioclase core (a “core pair”) produced high-temperature (747°C) and anomalously high-pressure (7.3 kbar) estimates with respect to the rim pair population. The high-temperature estimate overlaps with the upper temperature range of the rim pair population and the anomalously high-pressure estimate overlaps in 2σ standard deviation with the uppermost pressure range for the rim pair population. This core pair estimate is therefore not statistically distinguishable from the rim pair population but rather represents an early end-member, high-temperature, high-pressure, component of a continuous stage of melt crystallization. Assuming a 2.7 g/cm^3 average overlying crustal density, the 5.9 ± 0.8 kbar crystallization pressure estimate for Garm granitoid sample 12TS111 corresponds to a 22 ± 3 km intrusion depth.

Rim pairs for sample 12TS114 define a population with mean $\pm 2\sigma$ temperature/pressure estimates of $686 \pm 41^\circ\text{C}/7.9 \pm 1.8$ kbar (Figures 5a and 5b and Table S12 and Data Set S2). A core-pair pressure-temperature estimate (Text S3) for this sample produced anomalously high-temperature (780°C) and low-pressure (6.2 kbar) estimates with respect to the rim-pair population. The low-pressure estimate overlaps with the lower pressure range of the rim pair population, but the anomalously high temperature does not overlap with the temperature range of the rim-pair population, even in calibration uncertainty and 2σ standard deviation. The higher-temperature, lower-pressure pattern for this core-pair estimate with respect to the rim-pair population resembles that found for the inclusion pairs in 12TS09. Because the contexts are different for these two samples (inclusions versus cores), however, we do not emphasize comparison. Assuming a 2.7 g/cm^3 average overlying crustal density, the 7.9 ± 1.8 kbar crystallization pressure estimate for Garm granitoid sample 12TS114 corresponds to a 29 ± 7 km intrusion depth.

Six Zr-in-titanite temperatures (at 7.9 kbar) for Garm Amp + Bt granite sample 12TS114 average $714 \pm 131^\circ\text{C}$ (Table S12), which overlaps with amphibole-plagioclase temperatures but spans a much wider temperature range.

3.1.2. Comparison to Previous Studies

Compared to our 6.5 ± 2.4 kbar Al-in-Hbl crystallization-pressure estimate for Gissar batholith granitoid sample 12TS09, *Käbner et al.'s* [2016b] Al-in-Hbl crystallization-pressure estimates for five Gissar batholith granitoids (four of which are from the same transect north of Dushanbe as our sample 12TS09) span a lower, ~1.9–5.6 kbar 2σ range (Figures 2 and 5c). Our ~5.1–9.7 kbar Al-in-Hbl crystallization pressure 2σ range for Garm massif granitoid samples 12TS111 and 12TS114 spans higher pressures than that of *Käbner et al.* [2016b] for the Gissar batholith, and assuming a 2.7 g/cm^3 average overlying crustal density indicates ~19–36 km intrusion depths for the Garm massif versus primarily ~7–21 km intrusion depths for the Gissar batholith. We consider our sample 12TS09 to represent a thermobarometric outlier with respect to the other Gissar batholith and Garm massif granitoids due to its older crystallization age (discussed below) and its higher, more imprecise pressure estimate.

3.1.3. Results (Garm Metabasites): Amphibole-Plagioclase Thermometry

We estimated amphibole-plagioclase temperatures [*Holland and Blundy*, 1994] at 7 kbar for two Garm Amp + Pl $\pm$ CPx metabasites (Figure 6 and Table S12). Five temperature estimates for sample 12TS47 define a population with mean $\pm 2\sigma$ temperature estimates of $716 \pm 46^\circ\text{C}$. Eight temperature estimates for sample 12TS59 average $731 \pm 42^\circ\text{C}$. These upper-amphibolite-facies metamorphic-temperature estimates are similar to our estimated granitoid-emplacement temperatures and are ~100°C higher than *Konopelko et al.'s* [2015] pseudosection-based temperature estimates for their metasedimentary schist sample “12.”

3.1.4. Results (Garm Paragneisses and Granitoids): GARB Thermometry and GASP Barometry

Three Garm paragneisses (12TS34, 12TS62, and 12TS73) and one Garm granitoid (12TS30) have a garnet + biotite assemblage suitable for GARB thermometry (Figures 6a–6c), and Garm gneiss sample 12TS62 also has the additional Grt + Pl + Sil + Qz assemblage suitable for GASP barometry (Figure 6d). For GARB, scheme (i) (black lines) produces a range of relatively low-temperature estimates and scheme (ii) (colored lines) produces a range of relatively high-temperature estimates (see Text S4 for discussion of schemes (i) and (ii)). We interpret temperature estimates for scheme (i) to represent retrogressive re-equilibration, and the highest-temperature estimate in each case provides a minimum-temperature constraint. The Grt_{core}Fe_{avg}-Bt_{matrix}Fe_{avg} scheme (ii) estimates that we report as GARB peak-metamorphic temperature estimates are indicated by red dashed lines in Figures 6a–6d and are intermediate in value between the paragenetically irrelevant scheme (ii) overestimates and underestimates. The results are analogous for GASP: scheme (i) represents retrogressive re-equilibration (the highest of which provides a minimum peak-pressure estimate) and the Grt_{core}Fe_{avg}-Pl_{cores}An_{avg} scheme (ii) estimates represent our preferred GASP peak-metamorphic-pressure estimates.

Our Garm Grt + Bt paragneiss samples 12TS34 and 12TS73 exhibit overlapping, middle–upper-amphibolite-facies GARB peak-metamorphic temperature estimates (at 7 kbar): $665 \pm 37^\circ\text{C}$ and $678 \pm 45^\circ\text{C}$ (Figure 6 and Table S12). Assuming a 2.7 g/cm^3 average overlying crustal density, the mean temperatures correspond to a 26°C/km average paleo-geotherm. The peak-metamorphic temperature estimates span 30°C for 12TS34 and 60°C for 12TS73. The maximum rim-pair temperatures from these samples constrain the minimum peak-metamorphic temperatures to 648°C and 584°C . Because the relatively fine garnet grain size in these samples (200–500 μm) may have permitted retrogressive diffusion to affect the cores, we cannot rule out the possibility that our GARB temperature estimates for these samples represent underestimates. Our GARB temperature estimates for these samples overlap with the upper range of *Konopelko et al.*'s [2015] pseudosection-based temperature estimates for their Garm massif St + Bt $\pm$ Grt schist sample “12.”

The GASP peak-metamorphic-pressure estimate (at 750°C) for sample 12TS62 is 6.8 ± 0.1 kbar, which is similar to the tightly constrained range of other non-retrogressive, scheme (ii) GASP pressure estimates for this sample (Figure 6; Table S12). Assuming a 2.7 g/cm^3 average overlying crustal density, the mean pressure corresponds to a 30°C/km average paleo-geotherm. A maximum retrogressive-pair, scheme (i) GASP pressure of 6.3 kbar represents a minimum peak-metamorphic-pressure estimate for this sample.

Our preferred GARB peak-metamorphic-temperature estimate (at 4 kbar) for Garm Grt + Bt + Sil + Crd (Pin) + Spl paragneiss sample 12TS62 is $829 \pm 23^\circ\text{C}$, which spans lower-granulite facies (Figure 6 and Table S12). Assuming a 2.7 g/cm^3 average overlying crustal density, the mean temperature corresponds to a 56°C/km paleo-geotherm. The maximum rim-pair temperature estimate for this sample constrains the minimum peak-metamorphic temperature to 714°C . The GARB peak-temperature estimates for this sample span $\sim 110^\circ\text{C}$, which we note is the widest range in this study and we attribute to strong zoning from retrograde re-equilibration and resorption. Fe content in biotite exerts the primary control on GARB temperature estimates: Bt_{matrix}Fe_{max} and Bt_{matrix}Fe_{min} respectively produce the highest and lowest temperatures, whereas Bt_{matrix}Fe_{avg} produces intermediate temperatures. The corollary of this observation is that garnet core compositions are relatively homogeneous, which is facilitated by the coarse grain size (1–3 mm).

The ~ 3 kbar-greater peak metamorphic-pressure estimate (at an $\sim 80^\circ\text{C}$ lower temperature) for Garm paragneiss sample 12TS62, with respect to the peak-metamorphic temperature estimate, is consistent with textural and phase equilibria constraints (sections 2.2.1 and 3.2.1) that indicate late decompression during continued heating to peak-metamorphic temperatures in a clockwise P-T-t path. Our justification for estimating the GARB peak-metamorphic temperature for sample 12TS62 at 4 kbar rather than 7 kbar (as for all other metamorphic rock samples in this study) derives from the same evidence.

Because Garm Grt + Bt granodiorite sample 12TS30 is an igneous rock, our temperature-estimating scheme represents a granitoid-crystallization temperature of $830 \pm 33^\circ\text{C}$ (Figure 6 and Table S12) rather than a metamorphic temperature. The scheme (ii) GARB estimates for this sample are the most tightly clustered in this study, spanning only $\sim 20^\circ\text{C}$. The maximum rim-pair temperature estimate for this rock constrains the minimum crystallization temperature to 714°C .

Figure 7. Pseudosection model for Garm granulite-facies paragneiss sample 12TS62 in the $\text{Na}_2\text{O}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{K}_2\text{O}-\text{CaO}-\text{MnO}-\text{FeO}-\text{TiO}_2-\text{H}_2\text{O}$ thermodynamic system, generated using *Perple_X* version 6.7.2 [Connolly and Petrini, 2002]. The following bulk composition was estimated by integrating modal abundances, representative mineral chemistry, and mineral densities: 56.1% SiO_2 , 2.1% TiO_2 , 19.7% Al_2O_3 , 10.7% $\text{FeO} + \text{Fe}_2\text{O}_3$, 0.2% MnO , 3.8% MgO , 0.8% CaO , 1.5% Na_2O , 4.0% K_2O , and 1.3% H_2O . Capitalized phases indicate the implementation of a solution-phase model, whereas non-capitalized phases indicate the implementation of a pure, end-member phase from the *Holland and Powell* [2011] database. Mineral abbreviations follow *Whitney and Evans* [2010], with the addition of “Wm” for white mica. Green contours indicate isopleths of almandine component in garnet. Red contours indicate (dis)appearance of petrogenetically relevant phases. Yellow dashed lines indicate GARB pressure and GASP temperature estimates. Red path indicates interpreted, clockwise P-T-t path consistent with mineral assemblages, textures, and thermobarometry. Monazite $^{206}\text{Pb}/^{238}\text{U}$ weighted-mean, prograde- and peak-metamorphic ages in white boxes indicate our interpreted geochronologic interpretation.

3.2. Pseudosection Modeling

We used *Perple_X* [Connolly and Petrini, 2002] and the internally consistent thermodynamic database of *Holland and Powell* [2011] to model a pseudosection for Garm granulite-facies paragneiss sample 12TS62 in the system $\text{Na}_2\text{O}-\text{MgO}-\text{Al}_2\text{O}_3-\text{SiO}_2-\text{K}_2\text{O}-\text{CaO}-\text{MnO}-\text{FeO}-\text{TiO}_2-\text{H}_2\text{O}$. Pseudosection-modeling methodological procedures are summarized in Text S5 [Connolly and Petrini, 2002; White *et al.*, 2002; Holland and Powell, 2011; White *et al.*, 2014; Guevara and Caddick, 2016].

3.2.1. Results: Pseudosection Modeling for Garm Granulite-facies Paragneiss Sample 12TS62

Our pseudosection model (Figure 7) indicates a biotite maximum-temperature stability limit of 800–825°C (Table S12). Almandine isopleths indicate a similar, slightly lower temperature of ~775–800°C for Alm_{67} , which corresponds to the average garnet core composition ($\text{Alm}_{67}\text{Prp}_{18}\text{Grs}_{12}\text{Sp}_{53}$; Table S4). The ~925°C low-temperature limit of spinel in our pseudosection is significantly higher than all reasonable temperature estimates for this sample. We note, however, that spinel stability is known to expand to lower

temperatures with the incorporation of Fe^{3+} toward magnetite compositions [Waters, 1991; Kelsey, 2008], as well as with the incorporation of Zn or Cr [Hensen and Harley, 1990; Dasgupta et al., 1995; Guiraud et al., 1996].

Three other phase-stability observations provide maximum-metamorphic-pressure constraints and indicate a clockwise P-T-t path for sample 12TS62 (Figure 7). The absence of rutile in this sample due to its restricted stability at pressures greater than 6.0–7.5 kbar (depending on temperature) provides an upper-pressure limit of 7.5 kbar (at 750°C). Sillimanite and cordierite correspond to comparatively high- and low-pressure P-T-t trajectories, both of which are below rutile stability. Because petrographic observations indicate that sillimanite formed relatively early, we attribute early sillimanite growth to prograde metamorphism, which involved simultaneous heating and compression (sillimanite-stable, high-pressure trajectory). We attribute late cordierite growth, which accompanied garnet resorption, to decompression to ~6–4 kbar that was concomitant with continued heating to peak-metamorphic temperatures (cordierite-stable, decompressive trajectory). Whereas the net-transfer reaction(s) responsible for *primary* cordierite growth were more likely driven by continued *heating* during decompression to peak-metamorphic temperature, *secondary* pinitization of cordierite occurred during retrograde *cooling* and/or subsequent alteration. As a corollary to the clockwise P-T-t evolution that we advocate, we rule out the possibility of a counterclockwise P-T-t evolution typical of many granulite terranes [e.g., Waters, 1991] for the thermal evolution of the Garm massif.

Integrating our GARB/GASP thermobarometry for sample 12TS62 (Figure 6 and Table S12) with our pseudo-section model for this sample (Figure 7) indicates that prograde metamorphism reached peak-metamorphic pressure at ~750°C/7 kbar (~650–750°C/6–7 kbar range), which was followed by decompression and heating to peak-metamorphic temperature at ~820°C/4 kbar (800–820°C/6–4 kbar range; Figure 7). The prograde path that we infer follows the trajectory of our GASP peak-metamorphic-pressure estimate and satisfies phase-stability constraints. Our inferred decompression-heating path to peak-metamorphic temperature satisfies phase-stability constraints and follows the “Bt-out” isograd, which is consistent with minor Bt melting suggested by petrographic observations, from ~800°C/6 kbar to ~820°C/4 kbar. Although the intersection between the GARB peak-metamorphic-temperature slope and the “Bt-out” isograd at ~825°C/3 kbar provides a closer agreement between our GARB thermometry and pseudosection modeling, we consider the low pressure to be geologically unreasonable due to the high, implied 74°C/km paleo-geotherm (assuming 2.7 g/cm³ average overlying crustal density), which we would expect to be associated with ultrahigh-temperature mineral assemblages [e.g., Horton et al., 2016] that we do not observe. Paleo-geotherms of 29°C/km during peak-pressure metamorphism and 37–54°C/km during subsequent peak-temperature metamorphism, which are implied by our preferred P-T-t evolution, are respectively consistent with Barrovian, M-HT/LP, upper-amphibolite-facies, and subsequent HT/LP, lower-granulite-facies thermal regimes.

Compared to Konopelko et al.’s [2015] pseudosection modeling for their Garm St + Bt ± Grt schist sample “12” (600–660°C at 7–9 kbar), our analogous peak-metamorphic-pressure estimate for our Garm paragneiss sample 12TS62 overlaps with the low component of their pressure-range estimate at metamorphic temperatures that are up to 90°C greater than their estimated temperature range. On the contrary, our HT/LP, peak-metamorphic-temperature estimate is ~140–220°C higher at 1–3 kbar-lower pressures and implies a different thermal regime. We note that Konopelko et al. [2015] acknowledged the presence of granulite-facies metamorphic rocks and migmatites in the Garm massif, which qualitatively implies the higher metamorphic temperatures that we quantitatively document.

3.3. Results: Fan-Karategin Greenschist-Blueschist Facies Metamorphism

Our EPMA-based observation of the copresence of relict magnesioriebeckite and epidote in sample 13TS17 (Figures 3f and S1 and Tables S5, S10, and S11) demonstrates that the Yagnob formation component of the Fan-Karategin belt was metamorphosed under epidote blueschist facies metamorphic conditions (8–13 kbar, 400–500°C). Whereas our Fan-Karategin metabasite sample 13TS17 preserves the highest-pressure assemblage, our metasedimentary Fan-Karategin samples, for which minor mineralogic diversity indicates moderately consistent bulk composition, exclusively exhibit greenschist-facies mineral assemblages and lack relicts of blueschist-facies minerals. This discrepancy is likely, in part, a result of a greater susceptibility of metasedimentary rocks to greenschist-facies retrogression (compared to metabasites), although we cannot rule out

the possibility that it represents fortuitously greater peak pressures for the metabasites than for the metasedimentary Fan-Karategin rocks. Our pressure-temperature estimates for metamorphic rocks in the Fan-Karategin belt, which we designate as M-HP/LT, are consistent with previous assignments of upper-greenschist–lower-blueschist-facies metamorphism [Minaev, 1982; Minaev and Fedorov, 1984; Volkova and Budanov, 1999].

4. U-Pb Geochronology

We integrate U-Pb geochronology for dissimilar sample types to provide an integrated, multiperspective documentation of late Paleozoic tectono-magmatic-metamorphic processes in the southwestern Tian Shan. Igneous-zircon geochronology for granitoids documents two episodes of magmatism in the Gissar batholith and the Garm massif (Figures 8 and 9 and Table S13). Igneous-zircon geochronology for Fan-Karategin meta-andesite greenschist sample 12TS94 provides insight into the development of the Zeravshan subduction-accretion complex. Detrital-zircon geochronology for metasedimentary rocks and modern-river-sand samples that drain the catchments from which the igneous and metamorphic rocks were collected defines a map-scale geochronologic signature for the first-order lithotectonic units in our study area (Figure 10 and Table S14). Detrital-zircon geochronology for Cretaceous sandstones provides a broad, corroborative snapshot of sedimentary provenance during the Cretaceous. Detrital-titanite geochronology for a Fan-Karategin metasedimentary greenschist sample 12TS38 corroborates the detrital-zircon data for this sample. Metamorphic-zircon and -monazite geochronology for Garm amphibolite–granulite-facies paragneisses constrains and contextualizes the timing of metamorphism in the Garm massif within frameworks of sedimentary provenance and of the development of the Gissar arc (Figure 11 and Table S13).

We conducted igneous- and detrital-zircon U-Pb geochronology at the Arizona Laserchron Center. We used a Nu Plasma laser-ablation multicollector inductively coupled plasma mass spectrometer (LA-MC-ICP-MS) to analyze separated zircon grains in polished epoxy mounts for all samples except 13TS29RS, for which an Element2 LA-SC-ICPMS was used. Ablation pits are 30 μm wide for analyses that primarily utilize Faraday detectors (“Faraday analyses”) and 15 μm wide for analyses that utilize discrete dynode ion counters (“ion counter analyses”). Further zircon U-Pb geochronology methodological procedures are discussed in Text S6 [Press et al., 1986; Gehrels, 2000; Gehrels et al., 2006, 2008; Ludwig, 2008; Dickinson and Gehrels, 2009; Gehrels and Pecha, 2014; Pullen et al., 2014].

We analyzed detrital titanite in Fan-Karategin greenschist sample 12TS38 and monazite and xenotime in Garm granulite-facies paragneiss sample 12TS62 in situ in thin sections by the U-Pb laser-ablation split-stream method at the University of California, Santa Barbara [Kylander-Clark et al., 2013]. Ablation pits are 25 and 8 μm wide for titanite and monazite/xenotime. Analytical procedures for U-Pb titanite and monazite geochronology are outlined in Text S7 [Aleinikoff et al., 2006; Frost and Fanning, 2006; Mazdab, 2009; Kylander-Clark et al., 2013; Spencer et al., 2013] and Text S8 [Tomascak et al., 1996; Horstwood et al., 2003; Aleinikoff et al., 2006; Kylander-Clark et al., 2013; Palin et al., 2013].

4.1. Results: Igneous-zircon U-Pb Geochronology

4.1.1. Igneous-zircon Ages for the Gissar Arc

Zircon weighted-mean $^{206}\text{Pb}/^{238}\text{U}$ “crystallization ages” for our 13 granitoids span ~316–288 Ma (Figures 8, 9, and S3 and Table S13 and Data Set S4; see Figure 2 for sample locations and ages) and define the timing of magmatism for the Gissar batholith and the Garm massif established in this study. Our crystallization ages for the Gissar batholith (~316–294 Ma) overlap with but are generally older than those in the Garm massif (~302–288 Ma). This pattern is manifested by: (i) ~316–307 Ma crystallization ages for Gissar samples 12TS09 and 12TS12 and an absence of similarly old ages in the Garm massif; and (ii) ~300 Ma and ~295 Ma age peaks for Gissar and Garm granitoid crystallization ages. The weighted-mean age of the nepheline-syenite sample 12TS96 is ~280 Ma, which is ~10 Myr younger than the youngest rock in our granitoid suite. No clear correlation between crystallization age and mineralogy of the granitoids is evident except for the old, silica-saturated and young, silica-undersaturated end-members (12TS12 and 12TS96, respectively; Figure 4).

Using granitoid-crystallization-age density as a broad proxy for magmatic flux, our compiled Garm and Gissar granitoid data sets define four overlapping stages in the evolution of the Gissar arc and Garm massif (Figure 9): (1) earliest magmatism in the Gissar batholith from ~316–306 Ma, followed by ~2 Myr of

Figure 8. Zircon U-Pb ages of granitoid crystallization and metamorphism versus longitude across the Tian Shan terranes for this study and previous investigations of the Tian Shan. Samples from the South, Middle, and North Tian Shan (STS, MTS, and NTS) are respectively indicated with diamonds, boxes, and triangles. Shape outlines are colored to represent age interpretation as follows: G (black), granitoid crystallization; M (yellow), metamorphism; A (orange), alkaline magmatism (non-mafic); D (gray), dike/pegmatite crystallization; AG (pink), alkaline gabbro crystallization. We report our zircon weighted-mean $^{206}\text{Pb}/^{238}\text{U}$ ages and, for consistency, calculated such ages from single-grain analyses that were used for concordia ages in other studies. The exceptions are for Mirkamalov et al. [2012], Dolgoplova et al. [2016], and Kälsner et al. [2016b], whose concordia ages we used, and for Konopelko et al. [2016], whose reported weighted-mean ages we used.

Figure 9. Age distribution of magmatism and metamorphism in the Gissar arc. (a) Cumulative probability-density plot (PDP) compilations of zircon U-Pb (re)crystallization ages for granitoids (igneous zircon, shaded purple) and gneisses/migmatites (metamorphic zircon, shaded red) in the Gissar batholith (dashed component PDP) and the Garm massif (dotted component PDP). For each study, PDPs for granitoids and gneisses/migmatites are normalized with respect to one another based on the relative number of samples for either category. PDP for monazite 296 ± 6 Ma, prograde-metamorphic and 288 ± 6 Ma, peak-metamorphic $^{206}\text{Pb}/^{238}\text{U}$ weighted-mean ages for our sample 12TS62 (shaded dark yellow) is included as a chronometer for prograde, amphibolite-facies and peak, granulite-facies metamorphism in the Gissar arc. We use Dolgoplova et al.'s [2016] and Käßner et al.'s [2016b] reported concordia ages and Konopelko et al.'s [2016] reported $^{206}\text{Pb}/^{238}\text{U}$ weighted-mean ages. We calculated weighted-mean ages for five samples from the single-analysis ages that Konopelko et al. [2015] incorporated into their reported concordia ages. "NeS," nepheline syenite; "PIG," tonalite ("plagiogranite"). (b) PDPs of zircon single-analysis $^{206}\text{Pb}/^{238}\text{U}$ ages for modern-river-sands sample 12TS107RS and all analyses that comprise crystallization ages for crystallization-age PDPs in this study and those of Käßner et al. [2016b] and Konopelko et al. [2016].

Figure 10. Detrital-zircon age spectra, (represented as probability-density plots) for four modern-river-sand samples, two Cretaceous sandstones, three Zarafshan greenschists, and three Garm paragneisses. The ~300 Ma, Gissar age peak is well represented in the river sands and Cretaceous sandstones, as well as in metamorphic rims of Garm paragneiss 12TS101. The Zarafshan and Garm rocks have similar but distinguishable spectra: Neoproterozoic ages are common in both units but the Paleoproterozoic peak (~2450 Ma) is stronger in the Garm rocks. The Garm massif river-sand sample (12TS107RS) reproduces the 300 Ma peak almost exclusively. An ~450 Ma peak in one sandstone matches the ~450 Ma age in the Zarafshan meta-andesite greenschist. Detrital-titanite lower-intercept ages for 12TS38 are shown next to detrital-zircon age spectrum for comparison. Lithotectonic-unit-based compilations of detrital-zircon age spectra from this study compared with Konopelko *et al.*'s [2015] detrital-zircon data set for the Garm massif and Käßner *et al.*'s [2016b] inherited-zircon data set. *, we define inherited grains in the Gissar/Garm granitoid suite as those with a $^{206}\text{Pb}/^{238}\text{U}$ age older than 319.4 Ma, which is the oldest single-grain age incorporated into the crystallization age for 12TS12 (the granitoid with the oldest crystallization age in the Gissar/Garm granitoid suite).

magmatic quiescence; (2) rapidly increasing magmatic flux in the Gissar batholith at ~304 Ma, peaking at ~300 Ma and tapering off by ~294 Ma; (3) incipient magmatism in the Garm massif at ~302 Ma, building up to a peak-magmatic flux at ~295 Ma and tapering off by ~288 Ma; and (4) a secondary, low-flux pulse of alkaline magmatism in the Garm massif at ~280 Ma (sample 12TS96). Disregarding the early Gissar-batholith magmatism, compared to the Gissar batholith, magmatism in the Garm massif lags: (i) ~2 Myr behind in its inception; (ii) ~5 Myr behind in its peak; and (iii) ~6 Myr behind in its culmination. We interpret the overlapping, 2–6-Myr delay of most granitoid magmatism in the Garm massif behind that in the Gissar batholith, coupled with the different geologic contexts (batholith versus igneous-metamorphic complex), to indicate that the units represent shallow versus deep components of the Gissar arc that developed slightly out of sync.

4.1.2. Comparison of Our Results to Previously Determined Igneous-zircon Ages for the Gissar arc

The oldest published, modern granitoid-crystallization age of 321 ± 3 Ma for the Gissar batholith ~50 km to the west of our study area in Uzbekistan [Dolgoplova *et al.*, 2016] extends the apparent onset of Gissar arc magmatism to ~323 Ma (Figures 8 and 9). The crystallization ages of our oldest, ~316–306 Ma granitoids (samples 12TS09 and 12TS12) closely match Konopelko *et al.*'s [2016] $^{206}\text{Pb}/^{238}\text{U}$ weighted-mean ages for their tonalite ("plagiogranite") and granite samples 10 and 3 (Figures 8 and 9). The young, primary crystallization-age distribution (~304–294 Ma) of our Gissar batholith granitoids closely overlaps with and has an ~2 Myr-older peak age than that of Käbner *et al.* [2016b] and the central age for one granitoid from the Gissar batholith in Uzbekistan [Dolgoplova *et al.*, 2016]. Konopelko *et al.*'s [2016] ~294 Ma gabbro sample 2, which represents a stock in the Gissar batholith, overlaps with our younger, primary crystallization age distribution. Both Konopelko *et al.* [2016] and Käbner *et al.* [2016b] report ~290–280 Ma granitoid-crystallization ages in the Gissar batholith that we do not observe in our data set and collected many samples from the same ~N–S transect across the batholith north of Dushanbe that we did.

Compared to our ~302–288 Ma granitoid crystallization age range for the Garm massif, Konopelko *et al.*'s [2015] ~292 Ma granite sample 21 and Käbner *et al.*'s [2016b] ~289 Ma granodiorite sample GW22 represent the young component (Figures 8 and 9). All other Garm massif granitoid crystallization ages reported by Käbner *et al.* [2016b] are late Neoproterozoic (~670–550 Ma) and represent the much earlier magmatic assembly of the igneous component of the Garm-massif protocrust. An ~265 Ma Ne syenite along the northern margin of the Garm massif (GX-12) [Konopelko *et al.*, 2016] is younger than and analogous to our ~280 Ma Ne syenite sample 12TS96, which intrudes the center of the Garm massif (Figure 1). Taken together, these samples reflect long-lived, alkaline magmatism subsequent to silica-saturated granitoid magmatism.

Rather than granitoid crystallization ages, it is Konopelko *et al.*'s [2015] and Käbner *et al.*'s [2016b] ages of metamorphism and migmatization that overlap with our granitoid-crystallization ages in the Garm massif (Figure 9). Konopelko *et al.*'s [2015] primarily migmatitic-zircon ages for the Garm massif overlap with our entire Garm-massif granitoid-crystallization age range and cluster primarily with the young component. Käbner *et al.*'s [2016b] Garm-massif metamorphic- and migmatitic-zircon ages span a wider range from ~307–278 Ma. Therefore, the extant modern geochronology indicates that metamorphism in the Garm massif occurred coevally with granitoid emplacement.

4.1.3. Igneous-zircon Ages for Fan-Karategin Meta-andesite Greenschist Sample 12TS94

Our Fan-Karategin meta-andesite greenschist sample 12TS94 yielded a single age population with a weighted-mean zircon $^{206}\text{Pb}/^{238}\text{U}$ age of 450 ± 5 Ma for 71 analyses of different grains (Figure S3 and Table S13 and Data Set S4). Given the monogenetic age distribution for zircons with oscillatory zoning and consistently high Th/U (0.4–1.4) that suggest magmatic- (rather than metamorphic-) zircon crystallization for sample 12TS94, we interpret the ~450 Ma zircon-crystallization age to represent crystallization of the volcanic protolith. Compared to the lower Silurian (S_1) stratigraphic age assignment for the sampling locality in Soviet 1:200,000 geologic maps [Yushin *et al.*, 1964], the volcanic protolith age is ~8–20 Myr older. The ~450 Ma volcanic-protolith age also provides a maximum-age constraint for greenschist-facies metamorphism in the Fan-Karategin belt.

4.1.4. Comparison of Our Results to Previously Determined Igneous-Zircon Ages for the Tian Shan Outside of the Gissar Arc

In the Chatkal-Kurama range (Middle Tian Shan in Uzbekistan; Figure 1), crystallization ages for calc-alkaline granitoids and volcanic rocks with subduction-zone geochemical affinities primarily span ~320–300 Ma [Seltmann *et al.*, 2011; Dolgoplova *et al.*, 2016; Konopelko *et al.*, 2016] and are similar to the older

component of our granitoid age range for the Gissar batholith. Whereas the ~320–300 Ma granitic magmas were likely generated by northward subduction of the Turkestan ocean basin, younger (~300–280 Ma) granitic magmatism in the Chatkal-Kurama range is generally attributed to post-collisional processes. In the Kyzylkum South Tian Shan 150–400 km west of our study area in Uzbekistan, generally younger crystallization ages (~295–275 Ma) for compositionally diverse granitoids [Seltmann *et al.*, 2011; Dolgoplova *et al.*, 2016] are similar to the ~280 Ma crystallization age of our feldspathoid syenite sample 12TS96. Tonalite (“plagiogranite”) intrusions from the Uzbek Sultan-Uvais ~750 km WNW along strike of our study area have broadly similar ~311 and ~277 Ma crystallization ages and bear juvenile geochemical signatures that may reflect intra-oceanic arc magmatism in the westernmost Tian Shan that was coeval with continental arc magmatism to the east [Dolgoplova *et al.*, 2016].

In the Alai South Tian Shan ~100–200 km east of our study area (west of the Talas-Fergana fault in Kyrgyzstan, Figure 8), granitoid magmatism apparently overlapped with but mostly postdated that in the Gissar arc, primarily spanning ~290–270 Ma (Figure 8) [DeGrave *et al.*, 2012; Käbner *et al.*, 2016b]. An imprecise, ~318 Ma granitoid crystallization age indicates that magmatism in the Alai range at partly overlapped with the onset of Gissar arc magmatism [Käbner *et al.*, 2016b]. Farther to the east, across the Talas-Fergana fault and spanning the North, Middle, and South Tian Shan, ~305–290 Ma and ~285–280 Ma magmatism [Konopelko *et al.*, 2007; Alexeiev *et al.*, 2009b; Konopelko *et al.*, 2009; Glorie *et al.*, 2010; Seltmann *et al.*, 2011] occurred coevally with silica-saturated and silica-undersaturated magmatism in the Gissar arc. These magmatic rocks broadly comprise granitoids and syenites (with the exception of an ~284 Ma alkaline gabbro), many of which the authors designate as “A-type” and interpret to be post-collisional.

4.2. Results: Detrital Zircon U-Pb Geochronology

4.2.1. Garm Paragneisses

Age spectra for the three Garm amphibolite–granulite-facies paragneisses are broadly similar to one another but differ in detail (Figure 10). Collectively, our Garm paragneisses primarily define a quasi-uniformly distributed suite of overlapping peaks spanning the Neoproterozoic (1000–550 Ma) and an isolated earliest older peak centered at ~2470 Ma. K-S analysis indicates that 12TS73 and 12TS101 are statistically indistinguishable from each other and that they are statistically distinguishable from 12TS62 (Table S15). All three samples have Neoproterozoic age distributions: those for 12TS62 are primarily late Neoproterozoic and those for 12TS73 and 12TS101 are primarily early Neoproterozoic. The primary peak for 12TS62 is a broad suite of overlapping subpeaks with a quasi-uniform distribution of ages from ~700–550 Ma. The primary Neoproterozoic peaks for 12TS73 and 12TS101 are narrower and centered at ~800 Ma and ~950 Ma. The only other significant peak for the Garm paragneisses is at ~2450 Ma for 12TS73 and 12TS101 but not for 12TS62. All three samples share an absence of significant peaks for the rest of the Mesoproterozoic and Paleoproterozoic, other than a few isolated Neoproterozoic analyses. Similarities and differences between the Garm paragneisses are also reflected in maximum-depositional ages (Table S14 and also indicated in Figure 10): the maximum depositional age for 12TS62 is ~541 Ma, whereas those for 12TS73 and 12TS101 are 619 Ma and 614 Ma, respectively.

The narrow ~300 Ma peak for Garm amphibolite-facies paragneiss sample 12TS101 (Figure 10) represents ion-counter U-Pb analyses from 12 homogeneously zoned, consistently low-Th/U zircon rims that differ markedly in zonation, Th/U, and age from the zircon interiors (Figure 11). Therefore, rather than representing provenance, we interpret these observations to indicate metamorphic overgrowths on the rims of detrital zircons [e.g., Rubatto *et al.*, 2001] during a protracted metamorphic event that was coeval with Gissar arc magmatism (Figure 9). These 12 concordant analyses span ~325–285 Ma and produce a 291.2 ± 6.4 Ma weighted-mean $^{206}\text{Pb}/^{238}\text{U}$ age with a high mean square weighted deviation (MSWD) of 8.5 (Figure 11 and Table S13). The high MSWD likely reflects an aggregate of multiple populations that are unresolved by the low number of analyses.

The most tightly constrained cluster of low-Th/U (0.01–0.02) zircon analyses in this study is for ~300 Ma zircon from the Garm paragneisses (sample 12TS101; Figure 12a), which are clearly segregated from 330–270 Ma zircons in other lithologies that have moderate-high Th/U (Figure 12b). Notwithstanding this segregation, we note that the low-Th/U granitoid zircons ($0.01 < \text{Th}/\text{U} < 0.10$) for the 330–270 Ma age range, although few in number compared to high-Th/U granitoid zircons ($0.10 < \text{Th}/\text{U} < 1.00$), comprise a weak continuum between the unambiguously “metamorphic” zircons from 12TS101 and the unambiguously “igneous” zircons from the granitoids. These low-Th/U, granitoid zircons ($0.01 < \text{Th}/\text{U} < 0.1$) are disproportionately represented

Figure 12. (a) Zircon Th/U versus single-grain age for the 3600–0 Ma timeframe for all sample types from this study and those of Konopelko et al. [2015] and Käßner et al. [2016b]. (b) Zircon Th/U versus single-grain age for the 330–270 Ma time frame for all sample types from this study and those of Konopelko et al. [2015] and Käßner et al. [2016b].

by Garm massif granitoids: for the 330–270 Ma range, 15 of the 19 low-Th/U analyses (80%), versus 175 of the 286 total analyses (61%), represent granitoids from the Garm massif. The presence of low-Th/U zircons in the Garm massif is consistent with their context within an igneous-metamorphic complex and indicates that they may represent crustal heating and anatexis melting, as opposed to intrusion and crystallization of purely subduction-related granitoids.

The Garm zircons also include an older group of low-Th/U analyses from ~900–540 Ma (Figure 12a) that may represent metamorphic petrogenesis. However, because the age distribution is more scattered and because Th/U for any given age range constitutes more of a continuum between “igneous” and “metamorphic” values than a cluster, the lack of a well-defined population precludes quantitative time constraints for earlier metamorphic episodes in the Garm massif.

4.2.2. Fan-Karategin Greenschists

Age spectra for the three Fan-Karategin greenschists (12TS13, 12TS22, and 12TS38) are similar to one another and resemble those for the Garm massif, particularly regarding the distribution of Neoproterozoic peaks (Figure 10). All three Fan-Karategin samples have a dominant Neoproterozoic composite peak from ~650–550 Ma that is similar to but narrower and more prominent than that for Garm paragneiss 12TS62. The greenschists also have a broadly self-similar distribution of subordinate peaks for the Paleoproterozoic–early Neoproterozoic, with similar peaks at ~2000 Ma and ~1050–800 Ma. The age distribution of the compiled Fan-Karategin greenschists closely resembles the component samples. K-S analysis indicates that our Fan-Karategin greenschist samples are statistically indistinguishable from one another (Table S15). Our Fan-Karategin greenschists are only statistically indistinguishable from Garm paragneiss sample 12TS62, whereas our compiled Fan-Karategin greenschists are statistically distinguishable from our compiled Garm paragneisses. Maximum-depositional ages for the Fan-Karategin greenschists span 590–535 Ma (Table S14).

Two age peaks are notably absent in the Fan-Karategin greenschists (Figure 10). The first is an ~450 Ma putative peak that would match the volcanic protolith age for Fan-Karategin meta-andesite greenschist sample 12TS94. The absence of this peak is consistent with (but does not require) that the Fan-Karategin greenschist samples come from a lower, >450 Ma stratigraphic horizon in the pre-metamorphic stratigraphic succession of the Zeravshan complex. The second notably absent peak is an ~300 Ma peak that would match the timing of Gissar magmatism and Garm metamorphism, the absence of which indicates that the pre-metamorphic Zeravshan-complex volcano-sedimentary sequence was sourced from neither the Gissar batholith nor the Garm massif and was not subjected to temperatures sufficient for zircon (re)crystallization along with the Garm massif. Taken together, these absent putative peaks and the ~590–535 Ma Fan-Karategin maximum-depositional ages suggest that the pre-metamorphic Zeravshan-complex volcano-sedimentary sequence was deposited during the early–middle Paleozoic. Furthermore, the 450 Ma volcanic-protolith age for 12TS94 provides an oldest age constraint for Fan-Karategin greenschist–blueschist-facies metamorphism [Minaev, 1982; Minaev and Fedorov, 1984], which is consistent with the stratigraphically constrained, early–middle Visean (~347–337 Ma) age of metamorphism prescribed by Shvanov [1983] for the component Yagnob formation.

4.2.3. Cretaceous Sandstones

Two sandstones were collected from Cretaceous strata that unconformably overlie the Garm massif and are separated by ~15 km E–W distance (Figure 2). Soviet age assignments in 1:200,000 geologic maps [Yushin *et al.*, 1964] for strata from which these samples were collected are Turonian (94–90 Ma) for 12TS55 and Barremian–Aptian (131–113 Ma) for 12TS57. Our calculated middle Permian and late Permian–early Triassic maximum-depositional ages for these samples (Table S14) are consistent with and ~120–180 Myr older than the stratigraphic ages assignments and indicate a lack of prevalent magmatism in the source region during the Triassic–Jurassic and some of the Early Cretaceous. Both sandstones have similarly sharp, narrow, ~300 Ma peaks that overlap with Gissar/Garm magmatism (Figure 10). For 12TS55, this peak reproduces the age distribution found in the entire suite of Gissar and Garm granitoids (Figure 9). The analogous peak for 12TS57 is slightly younger, however, and overlaps more closely with the ~280 Ma crystallization age of feldspathoid syenite sample 12TS96. Based on an ~450 Ma age peak in 12TS57 that matches the crystallization age for Fan-Karategin meta-andesite greenschist sample 12TS94 and indicates a Tian-Shan provenance, we assert that the Gissar era ages represent broad Cretaceous sampling of the exposed Gissar arc rather than the age-equivalent, Karakul-Mazar terrane in the Pamir.

The Cretaceous sandstone samples differ from each other in the pre-Gissar components of their age spectra, which makes them statistically distinguishable (Table S15). Sample 12TS55 has a suite of quasi-uniformly distributed, subordinate peaks from ~1050–550 Ma (most significant at ~600 Ma) and isolated Neoproterozoic–Paleoproterozoic ages that resemble the Fan-Karategin spectra but are too few in number to constitute statistical distinguishability between the samples. Sample 12TS55 is statistically indistinguishable from the compilation of inherited-zircon ages in the granitoid suite. The dominant, ~450 Ma age peak in 12TS57, which occurs in lieu of the pre-550 Ma ages found in 12TS55, makes it statistically distinguishable from all other samples and compilations in this study.

4.2.4. Modern River Sand Samples

Four modern-river-sand samples were analyzed (Figure 2): 12TS32RS, 12TS54RS, and 13TS29RS primarily drain the metamorphosed, Fan-Karategin-belt component of the Zeravshan subduction-accretion complex, whereas 12TS107RS primarily drains the eastern Garm massif. All river-sand samples were collected as far away as possible from major human settlements (as well as mines, bridges, etc.) to minimize the likelihood of anthropogenic zircon contamination. Additionally, river sands were collected broadly above and below the water level to minimize the influence of fluvial zircon-sorting bias and represent catchments in a time-averaged sense over geomorphic timescales.

Age spectra for the three river-sand samples that primarily drain the Fan-Karategin belt have Paleoproterozoic and latest Mesoproterozoic–Neoproterozoic peaks that closely resemble those for the Fan-Karategin greenschist samples (Figure 10 and Table S15). The principal similarities in this regard are: (i) a dominant late Neoproterozoic peak centered at ~650 Ma; (ii) quasi-uniformly distributed, subordinate, early-middle Neoproterozoic peaks from ~1050 to 700 Ma; and (iii) isolated and subordinate peaks in the middle Paleoproterozoic centered at ~2000 Ma. The large number of analyses for 13TS29RS ($n = 312$) means this sample provides the most thorough characterization of its catchment for all the modern river sands in this study [e.g., Pullen *et al.*, 2014].

Of the three Fan-Karategin-belt river-sand samples, 12TS32RS and 12TS54RS have ~300 Ma peaks that are subordinate (but nontrivial) compared to the Proterozoic age distribution (Figure 10). Cretaceous sandstone sample 12TS55 is statistically indistinguishable from river-sand samples 12TS32RS and 12TS54RS (Table S15) because it has the same ~300 Ma and Proterozoic age distributions, albeit with a reversed pattern of dominance and subordination. Sample 13TS29 notably lacks an ~300 Ma peak, which can be viewed as a reflection of the absence of Gissar/Garm intrusions within the catchment of this modern-river-sand sample. Samples 12TS32RS and 12TS54RS drain the northern periphery of the Garm massif and isolated Gissar intrusions within the Fan-Karategin belt, whereas the catchment for 13TS29RS includes a fraction of exposed Gissar intrusions that is trivial compared to the exposure of Fan-Karategin-belt rocks (Figure 2). The peak for 12TS54RS is a composite peak centered at ~300 Ma and is slightly older than that for 12TS32RS, which is centered at ~287 Ma (Figure 10).

To represent Fan-Karategin-belt maximum-depositional ages by proxy, we disregarded ~300 Ma zircon age clusters in our calculated maximum-depositional ages for the three Fan-Karategin-belt river-sand samples to eliminate the influence of cross-cutting Gissar-arc intrusions (Table S14). Two samples (12TS54RS and 13TS29RS) yield ~590–550 Ma maximum-depositional ages that are similar to the in-situ counterparts. Sample 12TS32RS produces younger, ~400 Ma and ~450 Ma maximum-depositional ages which, based on the assumption that pre-Gissar Paleozoic magmatism in the southwestern Tian Shan was trivial, provide youngest-age constraints for the deposition of the pre-metamorphosed volcano-sedimentary sequence that was incorporated into the Zeravshan subduction-accretion complex and the Fan-Karategin belt.

The Yasman river-sand sample 12TS107RS primarily drains the eastern Garm massif (Figure 2) and yields an asymmetric Gissar/Garm age peak centered at 292 Ma (Figure 9), in addition to a subordinate ~2500 Ma peak (Figure 10). Given the easterly catchment for 12TS107RS, comparison of the former peak to all igneous zircons in the probability-density plot in Figure 9 reinforces our observation that older plutonism is only represented in the Gissar batholith in the west, whereas younger plutonism is represented in the east. Sample 12TS107RS also has a subordinate ~275 Ma peak that is similar to the ~280 Ma age for feldspathoid syenite sample 12TS96, which comprises part of the catchment for 12TS107RS (Figure 2). The ~2500 Ma peak for 12TS107RS is trivial but consistent with Neoproterozoic ages observed in the Garm paragneisses. Assuming that biases for detrital-zircon sorting, abundance, and radiation-damage-related preservation are trivial, the

dominance of Gissar/Garm ages for 12TS107RS suggests that the Garm massif is volumetrically dominated by Gissar/Garm-age magmatic rocks and that the Garm paragneisses are volumetrically subordinate.

4.2.5. Comparison of Our Detrital-zircon Results to Recent Local Studies

Compiled, previously published detrital-zircon ages for Garm massif paragneisses [Konopelko *et al.*, 2015] are statistically indistinguishable from our Garm paragneiss sample 12TS62, our compiled Garm-massif samples, and the majority of our Fan-Karategin-belt in-situ and modern-river-sand samples (Figure 10 and Table S15). We observe the same similarities between our Garm and Fan-Karategin samples and compiled, published inherited-zircon ages for granitoids, orthogneisses, and migmatites from the Gissar batholith and the Garm massif [Käbner *et al.*, 2016b]. The exception is for our compiled Garm-massif samples, which are statistically indistinguishable from the published detrital-zircon ages but not for the *inherited*-zircon ages (Table S15), which reflects the dissimilar sourcing mechanisms. Overall similarity between the Garm and Fan-Karategin samples indicates a common provenance for the Garm massif and Fan-Karategin belt (including the encompassing Zeravshan subduction-accretion complex). The ~590 Ma maximum-depositional age for Konopelko *et al.*'s [2015] Garm massif St + Bt schist sample 12 (Table S14) is broadly similar to our Ediacaran maximum depositional ages for the Garm massif and the Fan-Karategin belt.

4.3. Results: Detrital-titanite U-Pb Geochronology for Fan-Karategin Greenschist Sample 12TS38

Overall, the titanite analyses for Fan-Karategin metasedimentary greenschist sample 12TS38 define two chords in $^{207}\text{Pb}/^{206}\text{Pb}$ - $^{238}\text{U}/^{206}\text{Pb}$ space with upper, common-Pb ($^{207}\text{Pb}/^{206}\text{Pb}$) intercepts at 0.918 and 0.909 and lower, concordia intercepts at ~652 Ma and ~584 Ma (Figure S4 and Table S13). The *titanite* lower intercept dates are equivalent within uncertainty to the ~647 Ma and ~593 Ma detrital-zircon $^{206}\text{Pb}/^{238}\text{U}$ age peaks for this sample (Figure 10). This suggests that the titanite is detrital, with ages providing information about the metamorphic/igneous history of the source of the Fan-Karategin belt protolith, and that the greenschist-blueschist-facies temperatures did not exceed the closure temperature for the U-Pb system in titanite (>550°C [Cherniak, 1993; Reiners *et al.*, 2005; Spencer *et al.*, 2013]).

4.4. Results: Monazite U-Pb Geochronology for Garm Granulite Facies Sample 12TS62

We report two $^{206}\text{Pb}/^{238}\text{U}$ weighted-mean ages for Garm granulite facies paragneiss sample 12TS62 (Figure 11 and Table S13). We interpret the older, 295.6 ± 5.9 Ma age ($n = 9$, MSWD = 0.97) to reflect prograde (likely amphibolite-facies) metamorphism and the younger, 287.7 ± 5.8 Ma age ($n = 31$, MSWD = 1.03) to reflect peak, granulite-facies metamorphism in the Garm massif. The monazite prograde- and peak-metamorphic ages overlap with; (i) the peak and culmination, respectively, of silica-saturated, granitoid magmatism in the Garm massif (Figure 9); and (ii) the young age range of protracted, amphibolite-facies metamorphism indicated by metamorphic zircons in Garm amphibolite-facies paragneiss sample 12TS101 (Figures 9 and 11). Collectively, we interpret the monazite and zircon age distributions, as well as the presence of migmatite, to indicate increasing metamorphic grade and pervasiveness from early amphibolite-facies temperatures to later granulite-facies peak temperatures.

Heavy rare earth element (HREE) abundances (using Yb/Gd as a metric for HREE enrichment) [e.g., Stearns *et al.*, 2013] and Y for the oldest five monazite analyses (~310–295 Ma) in Garm granulite-facies paragneiss sample 12TS62 overlap with the low ranges of HREE and Y in all younger analyses, which themselves do not exhibit a correlation between HREE enrichment/depletion or Y and age (Figure S5 and Data Set S7). One low-HREE and low-Y, 291.2 ± 5.1 Ma analysis for a 15 μm monazite included in garnet (Figure 11) is a component of the peak-metamorphic age and suggests, at face value, that significant garnet growth may not have preceded peak metamorphism. The significant degree of garnet resorption and replacement by late (pinitized) cordierite in this sample renders this texture-based constraint on the timing of garnet growth questionable, however, because some monazite grains may have formerly been included in garnet. Therefore, we interpret the low HREE to indicate garnet and/or xenotime growth before and during peak metamorphism [e.g., Hermann and Rubatto, 2003].

5. Zircon Hf Geochemistry

We generated a combined igneous- and detrital-zircon Hf data set for five igneous-zircon samples and three detrital-zircon samples (Figure 14 and Table S16 and Data Set S8) using a Nu Plasma LA-MC-ICPMS at the Arizona Laserchron Center. The igneous samples comprise three granitoids (12TS12, 12TS30, and 12TS111),

one Ne syenite (12TS96), and one Fan-Karategin meta-andesite greenschist (12TS94). The detrital samples comprise two Cretaceous sandstones (12TS55 and 12TS57) and one Garm modern-river-sand sample (12TS107RS).

We report $^{176}\text{Hf}/^{177}\text{Hf}(t)$ ratios as $\varepsilon_{\text{Hf}}(t)$ and emphasize the change with time to facilitate tectonic interpretations. Hf analyses were focused on our previously dated zircon domains with the strategy of spanning the $^{206}\text{Pb}/^{238}\text{U}$ age range of concordant-zircon analyses both *between* different samples and *within* individual samples to assess evolving trends. Overall, we interpret clusters of more positive values to indicate magma derivation from a relatively juvenile source and/or slightly older, juvenile crust. Conversely, we interpret clusters of more negative values to indicate magma derivation from relatively old and/or successively re-melted crustal material, with minimal involvement of juvenile magmas or juvenile crust. Further zircon Hf geochemistry analytical and methodological procedures are outlined in Text S9 [Cecil *et al.*, 2011; Gehrels and Pecha, 2014].

5.1. Results: Igneous Zircon Hf Geochemistry

Zircons from individual igneous samples yield relatively uniform $\varepsilon_{\text{Hf}}(t)$ values irrespective of single-analysis ages (Figure 13 and Data Set S8). Therefore, we report the non-weighted mean $\varepsilon_{\text{Hf}}(t) \pm 2\sigma$ standard deviation of all analyses for each igneous sample as a representation of single-sample Hf signature (Table S16), which provides a suite of focused snapshots that establish *intra-sample* trends through time.

We interpret the igneous-zircon Hf data for the 330–270 Ma timeframe (Table S16) to broadly represent three stages that document the evolution of the Gissar arc (Figure 13), albeit biased toward the Garm massif (from which three of the four samples come). Stage 1 (~320–307 Ma, $\varepsilon_{\text{Hf}}(t) = +0.6 \pm 4.4$), which is defined by Gissar Amp + Bt granite sample 12TS12 and Gissar Amp + Bt Qz diorite sample 12TS09, represents the onset of intermediate arc magmatism in the Gissar batholith in Tajikistan. An ~307–302 Ma isotopic pull-down (average decrease of ~6 ε_{Hf} units) defines the transition to intermediate-evolved magmatism in the Garm massif during stage 2 (~302–288 Ma, $\varepsilon_{\text{Hf}}(t) = -5.5 \pm 5.4$), the diversity of which is evenly partitioned between the coeval Garm Amp + Bt + Ep Qz monzonite sample 12TS111 ($\varepsilon_{\text{Hf}}(t) = -2.7 \pm 3.1$) and Garm Grt + Bt granodiorite sample 12TS30 ($\varepsilon_{\text{Hf}}(t) = -7.3 \pm 2.9$). In particular, the evolved Hf signature for sample 12TS30 reflects the peraluminous bulk composition that its mineral assemblage suggests (Table S1), which, in addition to its abundance of inherited zircons ($n = 16$; Table S13), suggests significant derivation from old and/or evolved crust during the peak-magmatic flux in the Garm massif (Figure 9). An ~288–286 Ma isotopic pull-up (average increase of ~9 ε_{Hf} units) defines the rapid transition to stage 3 (~286–270 Ma, $\varepsilon_{\text{Hf}}(t) = +3.3 \pm 2.6$), which is defined by Garm Ne syenite sample 12TS96 and represents a post-Gissar-arc shift to a comparatively juvenile melt source. Time-regressed component analyses for sample 12TS96 also define an *intra-sample*, 0.9 ε_{Hf} -unit increase over the ~11 Myr timeframe defined by the relevant U-Pb ages, which corroborates the broader shift to juvenile magmatic values reflected by the *inter-sample* trends.

Our igneous-zircon Hf data set for the Gissar batholith and the Garm massif is corroborated by previous investigations of Gissar-batholith granitoids in Tajikistan and Uzbekistan (Figure 13). Based on intermediate zircon ε_{Hf} values for a 321 ± 3 Ma granitoid from the Uzbek Gissar batholith ~50 km west of our study area [Dolgoplova *et al.*, 2016] that closely resemble those for our oldest, ~313 Ma granitoid sample 12TS12 from the Tajik Gissar batholith, we expand the onset of stage 1 earlier by ~3 Myr to ~323 Ma for a more spatiotemporally complete representation of the onset of Gissar-arc magmatism. Käßner *et al.*'s [2016b] ~305–301 Ma Tajik Gissar-batholith samples 6814E1 and 6901C1 span the $\varepsilon_{\text{Hf}}(t)$ and crystallization-age range of our implied Gissar-arc isotopic pull-down between stages 1 and 2. Their ~297–295 Ma Gissar-batholith samples 6814F1 and 6901D1, which span an overlapping, generally more evolved $\varepsilon_{\text{Hf}}(t)$ range than our Garm massif samples 12TS30 and 12TS111 during stage 2 (~302–288 Ma), reflect coeval evolved magmatism in the Gissar batholith (albeit ~40–100 km west of the Garm massif; Figure 2) during the peak magmatic flux of the Gissar arc (Figure 9). Coeval, ~298 Ma granitoid magmatism in the Uzbek Gissar batholith bears intermediate ε_{Hf} values [Dolgoplova *et al.*, 2016] that reflect collectively greater magmatic diversity for the Gissar arc as a whole at this time.

Fan-Karategin meta-andesite greenschist sample 12TS94 (~450 Ma crystallization age, $\varepsilon_{\text{Hf}}(t) = -3.5 \pm 1.8$) represents weakly evolved volcanism during the Upper Ordovician depositional stage of the Zeravshan volcano-sedimentary stratigraphic sequence. This consistent, tightly constrained $\varepsilon_{\text{Hf}}(t)$ range indicates a

Figure 13. Zircon $\epsilon_{\text{Hf}}(t)$ versus $^{206}\text{Pb}/^{238}\text{U}$ single-analysis age or concordia age for Gissar arc granitoids and post-collisional rocks, Cretaceous sandstones, Fan-Karategin meta-andesite greenschist, and Garm modern-river-sand sample 12TS107RS. Error bars for individual analyses represent 1σ analytical uncertainty; error ellipses for igneous (solid) and detrital (dashed) samples from this study represent non-weighted 2σ standard deviation of the mean zircon $\epsilon_{\text{Hf}}(t)$ and single-grain $^{206}\text{Pb}/^{238}\text{U}$ ages for each sample. Slopes of error ellipses are equal to slopes of least-squares regression lines. Red and gray, thin, dashed ellipses represent Gissar-granitoid samples from Dolgoplova et al. [2016] and Kāßner et al. [2016b], with zircon $\epsilon_{\text{Hf}}(t)$ uncertainties represented as the non-weighted 2σ standard deviation of the mean, and crystallization-age uncertainties as reported 2σ uncertainties for zircon U-Pb concordia ages. Light-gray arrows illustrate $\epsilon_{\text{Hf}}(t)$ trends for igneous samples; "Stages 1–3" indicate time intervals during which igneous samples exhibit relatively consistent $\epsilon_{\text{Hf}}(t)$. Dark-gray arrow indicates average crustal ϵ_{Hf} evolution. Light-yellow shaded ellipses represent whole-rock $\epsilon_{\text{Nd}}(t)$ data for Gissar arc granitoids, migmatites, and gneisses and proximal post-collisional rocks from previous studies (representative, $0.2 \epsilon_{\text{Nd}}$ unit 2σ uncertainty assigned). *, sample from this study; CA, zircon U-Pb concordia age; CHUR, Chondritic uniform reservoir; DM, depleted mantle reservoir; m, SGA, mean single-grain $^{206}\text{Pb}/^{238}\text{U}$ age for zircons analyzed for Hf.

single, homogeneous volcanic melt source during the Late Ordovician deposition of this stratigraphic interval in the Zeravshan complex that contrasts with the greater magmatic heterogeneity and complexity for the late Carboniferous–early Permian granitoids.

5.2. Results: Detrital-zircon Hf Geochemistry

To assess the degree to which the inter-sample trends defined by the focused igneous samples capture regional variability in arc evolution, we also analyzed detrital zircons from Cretaceous sandstones and modern river sands, which sample more broadly and exhibit more spread than their igneous counterparts (Figure 13 and Table S16). Based on similarities in detrital-zircon age spectra with the Gissar arc and Zeravshan-complex samples (Figure 10), we interpret the Cretaceous sandstone samples to represent a broad proxy for the late Carboniferous–early Permian magmatic evolution of the Gissar arc and the predominantly Late Ordovician evolution of magmatism recorded in the Zeravshan complex (Figure 10).

Intra-sample trends for detrital samples, reflected by linear regression, resemble and substantiate the *inter-sample* trends inferred for the igneous samples (Figure 13 and Table S16). Cretaceous sandstone sample 12TS55 broadly reflects juvenile intermediate–evolved magmatism from ~325–270 Ma, with a regressed decrease of ~3 ϵ_{Hf} units over this time frame that resembles the shift from intermediate to intermediate–evolved magmatism observed for the Gissar-arc igneous samples. Cretaceous-sandstone sample 12TS57 broadly reflects intermediate–juvenile magmatism from ~300 to 265 Ma and expands the post-Gissar-arc ϵ_{Hf} pull-up observed for the igneous samples between stages 2 and 3 to very juvenile values (~+12), albeit over a coeval range of intermediate values (>+4) and a minor, weakly defined regressed increase of ~1 ϵ_{Hf} unit. Garm modern-river-sands sample 12TS107RS defines a broad, well-defined regressed increase of ~4 ϵ_{Hf} units from ~305–270 Ma, which parallels the isotopic pull-up from intermediate–evolved to juvenile magmatism observed for the Garm massif in the igneous data between stages 2 and 3.

For the early Paleozoic timeframe, Cretaceous sandstone sample 12TS57 reflects evolved to juvenile magmatism from 465–435 Ma that broadly becomes more juvenile with time, as reflected by an average increase of ~4 ϵ_{Hf} units (Figure 13 and Table S16). This detrital sample and the coeval igneous data show the greatest and lowest ϵ_{Hf} spread in this study, which indicates that the igneous data represent a small component of a diverse spectrum of ~450 Ma magmatic rocks (or younger sedimentary rocks that sampled them) that were exposed during the Early Cretaceous.

6. Discussion

6.1. Upon What Continental Crust Did the Gissar Arc Develop?

A fundamental question concerns the nature of the crust that hosted the development of the Gissar arc—in particular, whether it was part of the Karakum-Tarim block or the Kazakh-Kyrgyz continent. Our detrital-zircon data set for Garm-massif and Fan-Karategin-belt metasedimentary rocks and river sands (Figure 10) bears a peri-Gondwanan signature [e.g., Nance *et al.*, 2010], which contextualizes it within the global framework of supercontinent cycles and broadly indicates provenance from old crust that subsequently rifted from Gondwana to form a microcontinent [e.g., Müller *et al.*, 2001] that eventually accreted to Asia. As Käßner *et al.* [2016b] suggested based on inherited-zircon ages and ~670–550 Ma Garm-massif granitoid-crystallization ages (which are similar to our detrital-zircon data set; Figure 10), the crust that hosted the development of the Gissar arc was most likely part of the Karakum-Tarim terrane. This is consistent with the regional context, in which continental crust south of the Turkestan suture is part of the Karkum-Tarim terrane and that north of the Turkestan suture is part of the Kazakh-Kyrgyz continent.

The first-order similarity between our detrital-zircon age spectra for the Garm-massif and Fan-Karategin-belt metasedimentary rocks (Figure 10 and Table S15) indicates that they share a common provenance and represent a single, continuous terrane, upon which the Gissar arc developed. Notwithstanding the broad similarity, we infer statistical distinguishability between some of our Garm and Fan-Karategin samples to represent a natural variation in provenance typical of a diverse, tectonically active terrane. This inference implies that the difference between the Garm massif and the Fan-Karategin belt (and the encompassing Zeravshan subduction-accretion complex) is chiefly one of contrasting late Paleozoic petro-tectonic settings and metamorphic grades, rather than different origins.

Taken at face value, the predominantly Ediacaran maximum-depositional ages for the Garm-massif and Fan-Karategin metasedimentary rocks (latest Neoproterozoic; Figure 10 and Table S14) simply indicate that both lithotectonic units represent younger-than-Ediacaran sampled stratigraphic sequences, albeit with divergent subsequent tectono-metamorphic histories. The Fan-Karategin belt and the encompassing Zeravshan subduction-accretion complex include strata with younger, Paleozoic demonstrated depositional ages than those in the Garm massif, as indicated by the ~450 Ma volcanic-protolith age for Fan-Karategin meta-andesite greenschist sample 12TS94, in addition to Ordovician–Carboniferous biostratigraphic age assignments for Zeravshan-complex strata in Soviet 1:200,000 geologic maps [Yushin *et al.*, 1964]. The dearth of evidence for post-Ediacaran depositional ages for the Garm-massif paragneisses does not preclude similar Paleozoic depositional ages for the sedimentary component of the Garm massif, however, because: (i) incomplete sampling of rocks with younger detrital zircon maximum depositional ages cannot be ruled out; (ii) Soviet biostratigraphic age assignments do not exist for strata represented by the Garm-massif paragneisses because they were likely obliterated by pervasive magma inundation and high-grade metamorphism during the development of the Gissar arc; and (iii) cross-cutting relationships between ~670 and 550 Ma Garm granitoids [Käbner *et al.*, 2016b] and the Garm-paragneiss sedimentary protoliths, which could provide minimum-depositional ages, are unclear.

In conclusion, it is clear that the Fan-Karategin/Zeravshan and Garm (meta)sedimentary rocks share a common provenance (to a first order), represent strata that were deposited during or after the Ediacaran, and have divergent late Paleozoic tectono-metamorphic trajectories. However, it is ambiguous and difficult to clarify whether or not these units share the same early–middle Paleozoic history because the lower–middle Paleozoic Zeravshan complex stratigraphy is somewhat preserved but that of the Garm massif is not. Either these lithotectonic units represent the same Paleozoic stratigraphy or the Garm strata developed prior to the Zeravshan strata, during the Ediacaran and/or the early Paleozoic. In the latter case, the Paleozoic Zeravshan-complex strata may have been derived from the (meta)sedimentary and igneous proto-Garm-massif crust. In either case, the ~670–550 Ma Garm magmatic rocks [Käbner *et al.*, 2016b] likely provided the source for the matching age peaks for one Garm (12TS62) and all Fan-Karategin metasedimentary rocks (Figure 10).

6.2. Timing and Nature of M–HP/LT Metamorphism in the Fan-Karategin Metamorphic Belt

The most plausible tectonic setting for the Fan-Karategin belt is the deep component of the Zeravshan subduction-accretion complex, which incorporated both Paleozoic strata and accreted fragments of southward-subducting Turkestan oceanic crust. The detrital-zircon age spectra for Fan-Karategin samples (Figure 10) indicate that they represent metamorphosed Paleozoic strata with an ancestral provenance similar to that of the Garm-massif metasedimentary rocks. A mafic, oceanic-crustal protolith is represented by our epidote-blueschist sample 13TS17 and was reported by Minaev and Fedorov [1984] and Volkova and Budanov [1999], who focused on metabasites in the Fan-Karategin belt and proposed that they represent subducted seamounts and related clastic rocks. Clastic oceanic rocks may have been subjected to the same lower-blueschist-facies conditions as the mafic rocks but were likely more susceptible to greenschist-facies retrogression than the metabasites, which may account for the bias for high-pressure mineral assemblages in the metabasites. Post-metamorphic, late Paleozoic tectonic imbrication due to collision-related shortening may account for some or all of the juxtaposition of Fan-Karategin M–HP/LT rocks that were metamorphosed at different pressures.

Our upper- (~450 Ma meta-andesite protolith age for sample 12TS94) and lower- (~316 Ma Gissar arc intrusion) age constraints for greenschist–blueschist-facies metamorphism in the Fan-Karategin belt are consistent with Shvanov's [1983] stratigraphically determined, lower–middle Visean age assignment (~347–337 Ma) for metamorphism of the component Yagnob formation. Furthermore, the age and tectonic setting agree with Baratov's [1976] and Biske's [1996] Visean (347–331 Ma) stratigraphic age assignments for the broader development of the Zeravshan subduction-accretion complex, of which the Fan-Karategin belt is the metamorphosed component.

Due to different timing and tectonic contexts, we believe the Fan-Karategin M–HP/LT belt is genetically unrelated to both the composite Kokshaal (U)HP/LT belt and the Kassan HP/LT belt (Figure 1). Eclogite-facies metamorphism spanning ~320–305 Ma in the Kokshaal composite (U)HP/LT belt, and that spanning ~320–290 Ma in the Kassan HP/LT belt, postdated greenschist–blueschist-facies metamorphism in the Fan-Karategin M–HP/LT belt by ~20–50 Myr. The close timing overlap between formation of the Kokshaal and Kassan

(U)HP/LT belts and the formation of the Turkestan suture, as well as their proximity to the suture, indicate that these belts represent a symptom of the Tarim-Karakum-Kazakh-Kyrgyz continental collision. The Fan-Karategin M-HP/LT belt, conversely, is better characterized as an early Carboniferous, local accretionary complex that developed within the scope of accretionary tectonics that spanned the Central Asian Orogenic Belt prior to this collision [e.g., *Xiao et al.*, 2013, and references therein]. We speculate that the Fan-Karategin metamorphism may represent the same subduction episode that caused the Gissar back-arc basin to open (discussed below).

6.3. When, How, and Why Did the Gissar Back-arc Ocean Basin Open and Close?

Although the short lifespan of the Gissar arc and the narrow range of age assignments for Gissar back-arc ocean basin rocks clearly indicate that the Gissar back-arc ocean basin opened and closed during the Carboniferous (Figure 14), the timing of the component stages of continental rifting, oceanic spreading, and ocean closure are ambiguous in detail. We interpret a Tournaisian(?)–Visean basaltic–dioritic dike swarm that partly intrudes continental rocks in the southern Gissar range [Portnyagin, 1974; Brookfield, 2000; Burtman, 2010] to chiefly represent continental rifting of the northern Karakum terrane that preceded oceanic spreading. Previous age assignments for ostensibly oceanic-crustal rocks distributed along the southern flank of the Gissar range are more variable, ranging from Visean–Serupkhovian for pillow basalts [Portnyagin, 1974; Baratov, 1976; Brookfield, 2000] to Serupkhovian–early Moscovian for undivided basalts and andesites [Burtman, 2010] and lower–middle Carboniferous for undivided basalts [Yushin *et al.*, 1964]. Overlying, late Serupkhovian limestones and Bashkirian tuffs [Baratov, 1976] are consistent with but do not require the cessation of oceanic spreading by this time.

We tentatively interpret an early Carboniferous (Tournaisian?)–Visean age assignment for pre-Gissar-arc, calc-alkaline volcanic rocks in the Gissar range [Baratov, 1976] to represent subduction magmatism that was tectonically related to the coeval, Visean development of the Zeravshan subduction-accretion complex (including the Fan-Karategin belt) [Baratov, 1976; Shvanov, 1983; Biske, 1996]. Given the apparent synchronicity of these genetically relatable Visean tectonic events with continental rifting of the northern Karakum terrane and subsequent opening of the Gissar back-arc ocean basin, we speculate that a short-lived, early Carboniferous episode of southward, slab-rollback subduction of the Turkestan ocean basin under the northern margin of the Karakum continent caused rapid back-arc rifting and extension in the upper plate [e.g., Royden, 1993a, 1993b], which ultimately gave rise to the Gissar back-arc ocean basin and the rifted Gissar microcontinent (Figure 15). Modern geochronologic and geochemical constraints in future investigations of these early Carboniferous, pre-Gissar arc, calc-alkaline volcanic rocks in the Gissar range may clarify whether or not our speculative model for the origin of the Gissar back-arc ocean basin is accurate.

The middle Carboniferous was apparently a transitional period in this region. Southward subduction of the Turkestan ocean basin may have ceased by ~330 Ma, by which time the symptomatic magmatism and accretionary-wedge development apparently ended (Figure 14). If true, this inference would imply that ~335–315 Ma rifting in the Gissar back-arc ocean basin was initially generated by but largely occurred in the absence of a counterpart subduction zone. The temporally overlapping inception of northward subduction of the Turkestan ocean basin under the Kazakh-Kyrgyz continent during the early [Konopelko *et al.*, 2016] or middle [Burtman, 2010] Carboniferous implies a change in the partitioning of regional convergence, which we speculate may have accommodated continued rifting in the Gissar back-arc ocean basin despite the possible demise of its causative subduction zone (Figure 15). Continued northward Turkestan-ocean-basin subduction through the late Carboniferous is evidenced by ~318–300 Ma calc-alkaline, Andean magmatism in the Chatkal-Kurama arc [Seltmann *et al.*, 2011; Konopelko *et al.*, 2016].

Closure of the Gissar back-arc ocean basin occurred via northward subduction under the southern margin of the Gissar microcontinent during the middle to late Carboniferous and gave rise to Andean magmatism in the Gissar arc. Bashkirian–lower Moscovian intermediate–felsic Gissar-arc volcanic rocks [Baratov, 1976], as well as the oldest modern radiometric ages for Gissar-batholith granitoids (~323–306 Ma, Figure 9) [Dolgoplova *et al.*, 2016; this study], reflect the onset of Gissar-arc subduction. The early Bashkirian onset of Gissar-arc subduction may have overlapped with continued generation of new oceanic crust in the Gissar (formerly) back-arc ocean basin for ~5 Myr (Figure 14), which we speculate to potentially reflect extension due to: (i) the continued effect

Figure 14. Summary of local geologic events in the southwestern Tian Shan and relevant regional events throughout the Tian Shan during the Carboniferous–middle Permian. Color scheme indicates purple, arc and subsequent syn-collisional magmatism; orange, post-collisional magmatism; red, (M-)HT/LP metamorphism; blue, (U-)HT/LP metamorphism; green, M-HP/LT metamorphism; black, geodynamic events. MTS, Middle Tian Shan; STS, South Tian Shan.

Figure 15. Model for the tectono-magmatic-metamorphic evolution of the southwestern Tian Shan. FK, Fan-Karategin greenschist–blueschist belt; GB, Gissar batholith; GM, Garm massif; GS, Gissar suture; TF, Turkistan flysch; TS, Turkistan suture; Z, Zeravshan subduction-accretion complex;

of the Chatkal-Kurama arc subduction zone to the north on the partitioning of regional convergence; and/or (ii) potential fore-arc hyperextension in the Gissar fore-arc [e.g., Maffione *et al.*, 2015].

Attributing Gissar-arc subduction initiation [e.g., Stern, 2004] to geodynamic changes associated with the broadly coeval Tarim-Karakum-Kazakh-Kyrgyz collision is reasonable to a first order but unclear in detail due to an ambiguous timing relationship with the formation of the Turkistan suture west of the Talas-

Fergana fault (Figure 14). On one hand, *Mühlberg et al.*'s [2016] ~321–313 Ma age for HP/LT metamorphism and rapid subsequent exhumation of the co-longitudinal Kassan metamorphic complex (Figure 1) indicates that the Turkestan suture formed synchronously with the inception of the Gissar arc, which suggests that the Tarim-Karakum-Kazakh-Kyrgyz collision directly induced Gissar-arc subduction at a proximal location along strike. Alternatively, a geodynamically broader, far-field influence of the Tarim-Karakum-Kazakh-Kyrgyz collision on Gissar arc subduction initiation is consistent with *Han et al.*'s [2011] model, in which the collision was underway by ~315 Ma in the Chinese South Tian Shan and diachronously propagated westward along strike. Furthermore, a hypothetical, younger, ~300 Ma age for Turkestan-suture formation at the longitude of the Gissar arc is compatible with ~318–300 Ma Andean magmatism in the Chatkal-Kurama arc (a by product of the northward-subducting Turkestan ocean basin) [Konopelko et al., 2016], as well as *Loury et al.*'s [2016] imprecise, ~316–286 age for Kassan HP/LT metamorphism. Irrespective of the cause of Gissar-arc subduction initiation, stratigraphic constraints indicate that closure of the Gissar back-arc ocean basin was complete by the Kasimovian or Gzelian (~307–299 Ma) [Burtman, 1975, 2010]. With the Gissar back-arc and Turkestan ocean basins fully consumed by the latest Carboniferous, the Tarim-Karakum-Kazakh-Kyrgyz collision was underway in the southwestern Tian Shan, giving way to syn-collisional and post-collisional processes in the Gissar arc and throughout the Tian Shan.

6.4. Timing and Nature of Late Paleozoic Andean, and Syn-collisional Magmatism in the Gissar Continental Arc

The broadly similar timing, mineralogy, and major-element geochemistry [Dolgoplova et al., 2016; Käbner et al., 2016b; Konopelko et al., 2016] of voluminous granitoid magmatism in the Gissar batholith and the adjacent Garm massif attests to subduction-related origins for both units as components of the Gissar arc, the lifespan of which spanned ~35 Myr from ~323 to 288 Ma (Figures 9, and 15). Granitoid magmatism in the Gissar batholith collectively spanned ~323–292 Ma (~299 Ma peak age) and preceded and temporally overlapped with that in the Garm massif, which spanned ~302–288 Ma (~293 Ma peak age).

Broad mineralogic and lithologic differences between the Gissar batholith and the Garm massif indicate that they represent different but related components of the Gissar arc. Whereas granitoids in the Gissar batholith have a simple, intermediate-felsic granitoid mineralogy typical of I-type granites, granitoids from the Garm massif exhibit more mineralogic diversity that indicates a continuum between I- and S-type granites, with generally lower quartz abundances, greater mafic-phase abundances, and the addition of primary garnet and white mica typical of S-type granites for two samples (Figure 4 and Table 1). Käbner et al.'s [2016b] major-element, whole-rock geochemical analyses for five ~301–286 Ma Gissar arc (four Gissar batholith and one Garm massif) granodiorites and monzogranites define a continuum between I- and S-type granites [Chappel and White, 2001] and exclusively normalize to magnesian Cordilleran granites [Frost et al., 2001], which, in addition to evidence for significant fractional crystallization (as opposed to only to crustal assimilation), indicates that both I- and S-type orogenic magmatism occurred coevally in both components of the Gissar arc during its apparently highest-flux time frame (Figure 9). Further evidence for a subduction-related origin for Gissar-batholith and Garm-massif granitoids that crystallized during both the early (~323–306 Ma) and apparently high-flux (~304–288 Ma) stages of the Gissar arc is provided by the dominant calc-alkalic and subordinate alkali-calcic major-element classification [Frost et al., 2001] of the granitoids [Käbner et al., 2016b; Konopelko et al., 2016]. Taken together, these observations indicate that ~323–306 Ma, I-type magmatism in the Gissar batholith was succeeded by ~304–288 Ma, I-S-type magmatism in both the Gissar batholith and the Garm massif. Magnesian, dominantly calc-alkali granitoid magmatism attests to an arc and/or “post-collisional” setting [Frost and Frost, 2008], the latter of which we consider to be *syn-collisional* in this case due to continued subduction and convergence during the post-Andean, ~304–288 Ma lifespan of the Gissar arc. Non-weighted, linear regression of Andean-syn-collisional Gissar-arc granitoid-crystallization ages (excluding post-collisional rocks) with eastward distance (~67.3–71.3°E) illustrates an eastward-younging pattern of Gissar arc magmatism by ~4 Myr/°E, which is equivalent to ~0.05 Myr/km (Figure 8) and similar to Käbner et al.'s [2016b] previous observation 0.035 Myr/km.

Overall, the modern exposure of the Gissar arc defines a broad, eastward, along-strike increase in paleo-depth from upper-batholith levels to what we interpret to represent the root of the Gissar arc. One line of evidence is generally low, ~1.9–5.6 kbar crystallization-pressure estimates for ~303–292 Ma Gissar-batholith granitoids that imply ~7–21 km intrusion depths [Käbner et al., 2016b], which contrast with generally

high, ~5.1–9.7 kbar crystallization-pressure estimates for ~298–289 Ma Garm-massif granitoids that imply ~19–36 km intrusion depths (this study; Figure 5c and Table S12). Peak-metamorphic pressure estimates of ~7–9 kbar [Konopelko *et al.*, 2015] and ~7.5 kbar (this study; Figures 6 and 7) for Garm-massif paragneisses corroborate our igneous geobarometry. Additionally, the close association of the commonly deformed Garm-massif granitoids with amphibolite–granulite-facies metamorphic rocks and migmatites indicate a continental-arc-root tectonic setting [e.g., Ducea *et al.*, 2015]. The greater diversity of applied pressure-estimation techniques for granitoids and paragneisses from the Garm massif (Al-in-Hbl, GASP, and pseudosection modeling) compared to granitoids from the Gissar batholith (only Al-in-Hbl) reflects the greater lithologic diversity of the Garm massif. We speculate that exposures of rare, deformed mafic and ultramafic rocks reported in the Garm massif [Budanov, 1993; Brookfield, 2000] may represent the upper MASH zone of the Gissar arc [e.g., Walker *et al.*, 2015].

We interpret the zircon ε_{Hf} pull-down (Figure 13), which is exhibited by a shift from an early (~323–308 Ma), intermediate melt-source signature in the Gissar batholith to a main-stage (~302–288 Ma), intermediate-evolved melt-source signature in the Gissar batholith and the Garm massif and corroborated by whole-rock ε_{Nd} data [Konopelko *et al.*, 2015, 2016; Dolgoplova *et al.*, 2016; Käbner *et al.*, 2016b; this study], to reflect an arc-wide change in the nature of the subducting lower plate from oceanic to marginal-continental lithosphere that accompanied the coevally completed consumption of the Gissar back-arc ocean basin and the onset of formation of the Gissar suture (Figures 14 and 15) [Baratov, 1976; Brookfield, 2000; Burtman, 2010]. Considering granitoid-crystallization ages (Figure 9) and geochemical affinities [Dolgoplova *et al.*, 2016; Käbner *et al.*, 2016b; Konopelko *et al.*, 2016] in this light, the Gissar continental arc was apparently Andean in nature from ~323–306 Ma and transitioned into a syn-collisional magmatic arc from ~304–288 Ma during continued convergence, when the previously rifted, thinned Karakum continental margin was subducted/underthrust northward below the Gissar microcontinent (Figure 15). Broadly negative whole-rock ε_{Nd} values for syn-collisional Gissar-arc granitoids and gneisses, particularly those from the Garm massif, corroborate the isotopic pull-down exhibited by the zircon ε_{Hf} data [Konopelko *et al.*, 2015; Dolgoplova *et al.*, 2016; Käbner *et al.*, 2016b; Konopelko *et al.*, 2016].

The apparent ~304–288 Ma, syn-collisional magmatic flare-up in the Gissar arc suggests that the changing nature of the subducting lower plate, from oceanic to continental, may have catalyzed magmatism in the Gissar arc. Two lines of evidence in this regard suggest significant derivation of the ~304–288 Ma, syn-collisional Gissar arc magmas from subduction of the lower plate: (i) major and trace-element patterns that indicate a significant role of fractional crystallization in the evolution of syn-collisional granitic magmatism [Käbner *et al.*, 2016b]; and (ii) “arc-related-magma” geochemical and isotopic signatures for syn-collisional (post-collisional) Gissar-arc granitoids [Konopelko *et al.*, 2016]. However, the onset of syn-collisional magmatism in the Gissar arc also coincided with Barrovian-type metamorphism and migmatization in the Garm massif (Figure 9) [Konopelko *et al.*, 2015; Käbner *et al.*, 2016b; this study], which demonstrates that upper-plate-derived, old and/or evolved crustal melts were available to be assimilated into the net Gissar-arc magmatic budget. Notwithstanding the development of anatectic melts in the upper plate, we propose that voluminous, lower-plate-derived magmas may have developed due to subduction of sediments and/or Karakum-margin continental-crustal rocks to mantle depths. Volatile, felsic, and/or carbonate-rich continental-margin sediments, in particular, may have formed buoyant diapirs that ascended through the mantle wedge and relaminated to the base of the Gissar arc lower crust [e.g., Hacker *et al.*, 2011], triggering further melting and assimilation in the upper plate. Although a significant fraction of the Gissar-arc magmatic budget was likely derived from the upper plate during its early Andean phase to begin with [e.g., Ducea *et al.*, 2015], we note that the expanded net-magmatic budget that accompanied the transition to syn-collisional orogenesis in the Gissar arc could have been caused not only by increased melt production in the upper plate due to anatexis and crustal assimilation, but also by an increased melt input from the mantle wedge above the lower plate.

Our interpretation of the early Andean and late, syn-collisional nature of granitoid magmatism (and volcanism) in the Gissar arc is similar to that of Savchuk *et al.* [1991] and Brookfield [2000], who accounted for gradually changing Gissar-arc magma compositions with the evolving nature of the subducting lower plate from oceanic, through transitional-oceanic continental crust, to marginal-continental crust. We dispute the local tectonic framework they suggest, however, in which the lower plate comprised (in its late history) a continuous crustal block that linked the Baysunta and Garm (M–HT/LP massifs (Figure 1), which would imply that

the Gissar suture bounds the Garm massif to the north rather than to the south, as we interpret (Figure 2). Because we interpret the Garm massif to represent the root of the Gissar arc that developed during its late, syn-collisional magmatic phase (Figures 9, 14, and 15), we consider the ~302–288 Ma Garm-massif melts to have been derived from both the subduction of melt fertile, marginal-continental crust and assimilation of Gissar-microcontinent crust in the upper plate. Whereas we interpret the Garm massif to represent an arc-root component of the short-lived, rifted, and resutured Gissar microcontinent, this is not the case for the Baysunta massif, which is situated south of the Gissar suture and represents a crustal block that underwent an independent, albeit perhaps similar tectonic evolution that has not been temporally constrained by published, modern geochronologic ages.

6.5. Post-collisional Lithospheric Foundering and Magmatism

An indirect line of evidence for a thick southwestern Tian Shan crust is the corroborative igneous- and detrital-zircon Hf pull-up from intermediate–evolved to juvenile values subsequent to the end of Gissar-arc magmatism at ~288–286 Ma (Figure 13), which indicates a rapid shift from an evolved/crust-derived to a juvenile/mantle-derived melt source at this time. We interpret this change in melt source to indicate asthenospheric upwelling due to lithospheric foundering (Figures 14 and 15) [e.g., Zandt *et al.*, 2004], which implies the prior existence of a crust that was thick enough for its root to have densified beyond a gravitational steady state via the formation of an “arclogite” root [e.g., Lee and Anderson, 2015]. This thick crust apparently developed over ~10–15 Myr, subsequent to the onset of formation of the Turkestan and Gissar sutures and the Karakum-Tarim-Kazakh-Kyrgyz continental collision in the southwestern Tian Shan. We note that Permian mantle-plume-related magmatism (discussed below) cannot account for the change in melt source because its possible (but unlikely) arrival in the southwestern Tian Shan occurred ~30 Myr after our observed zircon Hf pull-up.

The proposed ~288–286 Ma lithospheric-removal event in the southwestern Tian Shan is directly related to the onset of post-collisional magmatism in the southwestern Tian Shan, which was part of a broader, regional, geodynamic change that accompanied the cessation of oceanic subduction in the Central Asian Orogenic Belt. The onset of post-collisional magmatism throughout the Tian Shan is the preeminent observable symptom of this event and is defined by the interterrane, spatially widespread but low-volume emplacement of compositionally diverse but commonly alkaline, “stitching” plutons subsequent to the formation of the Turkestan (South Tian Shan) suture. Although post-300 Ma magmatism in the western Tian Shan is commonly considered to represent the post-collisional setting, post-collisional magmatism may have begun earlier along strike to the east in the Kyrgyz and Chinese Tian Shan, when and where the collision commenced earliest [Han *et al.*, 2011]. Development of crustal-scale strike-slip faults including the dextral Talas-Fergana fault [Alexeev *et al.*, 2009a; Rolland *et al.*, 2013; Konopelko *et al.*, 2017], which accommodated the kinematic reorganization from contractional, shortening-dominated to strike-slip-dominated tectonics, accompanied post-collisional magmatism and may have controlled the emplacement of juvenile magma, providing effective conduits for its ascent with little crustal assimilation [e.g., Konopelko *et al.*, 2007; Han *et al.*, 2010; Konopelko *et al.*, 2011; Seltnann *et al.*, 2011; Mao *et al.*, 2014].

We consider local post-collisional magmatism in the Tajik southwestern Tian Shan to have been caused by sub-Gissar-arc lithospheric foundering at ~288–286 Ma and underway by ~280 Ma, ~15–20 Myr after the early Permian (post-299 Ma) onset of post-collisional magmatism regionally throughout the Tian Shan [Konopelko *et al.*, 2007; Seltnann *et al.*, 2011; Konopelko *et al.*, 2016]. Whereas voluminous ~304–288 Ma, I-S-type, intermediate–evolved, silica-saturated Gissar-arc granitoid magmatism (and metamorphism) occurred in a syn-collisional setting during ongoing convergence and northward subduction/underthrusting of marginal-continental crust below the Gissar microcontinent, the ~280 ± 3 Ma crystallization age for our juvenile, alkaline, Ne syenite sample 12TS96 (Figures 9 and 13), which crosscuts Gissar arc granitoids in the Garm massif (Figure 2), represents post-collisional magmatism subsequent to the local cessation of convergence that accompanied a regional transition to strike-slip-dominated tectonics [e.g., Alexeev *et al.*, 2009a; Konopelko *et al.*, 2013; Rolland *et al.*, 2013].

The local, juvenile (zircon $\varepsilon_{\text{Hf}} = +3.3$), alkaline, ~280 Ma post-collisional magmatism that we document in the Garm massif quickly following lithospheric foundering in the Gissar arc was apparently succeeded by intermediate–evolved, alkaline and granitic magmatism in the proximal southwestern Tian Shan

(Figure 13). Whole-rock $\epsilon_{\text{Nd}}(t) = -7.4$ for an ~ 267 Ma Ne syenite representing an alkaline pluton intruding the Zeravshan complex ~ 20 km north of the Garm massif (sample GX-12; Figure 2) [Konopelko *et al.*, 2016] suggests a local shift from juvenile to evolved post-collisional magmatism, albeit observed between the Lu/Hf and Sm/Nd systems (Figures 2 and 9). Weakly negative zircon $\epsilon_{\text{Hf}}(t) = -2.0$ and moderately negative whole-rock $\epsilon_{\text{Nd}}(t) = -5.3$ for a coeval, ~ 269 Ma Kfs granite from the Lolabulak complex in Uzbekistan [Dolgoplova *et al.*, 2016] provides further indication of an isotopic pull-down by ~ 270 Ma in the southwestern Tian Shan proximal to the quiescent Gissar arc. We interpret this local, late post-collisional apparent isotopic pull-down from juvenile, alkaline magmatism to intermediate–evolved, alkaline and granitic magmatism to reflect the petro-tectonic integration of the southwestern Tian Shan into the regional, post-collisional setting of the Central Asian Orogenic Belt.

At a regional scale, post-orogenic magmatism was likely induced by broad mantle upwelling from ~ 300 – 270 Ma (Figure 14) [Han *et al.*, 2016], mostly under the Tarim craton but affecting the Tian Shan more broadly (the “Tarim mantle plume”). The full extent of the Tarim plume outside of the Tarim basin is not clear, however, which limits our ability to attribute post-orogenic magmatism to plume activity beyond the proximal south Tian Shan into the middle and north(east) Tian Shan. Käßner *et al.* [2016b] for example, interpreted younger, 256–237 Ma $^{40}\text{Ar}/^{39}\text{Ar}$ biotite and hornblende ages for small-volume basalt dikes and pipes in the southwestern Tian Shan to indicate a late, westward propagation of the Tarim plume, although its arrival would have been ~ 20 Myr too late to account for the juvenile, alkaline, post-orogenic magmatism that we document intruding the Garm massif (sample 12TS96). Furthermore, the delamination-induced origin for 288–266 Ma small mafic–ultramafic intrusions and mafic dikes in the Beishan orogen (easternmost Tian Shan) [Xue *et al.*, 2016], which overlap with the Tarim plume in time but not in space, challenges the reliability of extrapolating documented plume magmatism to account for regional, synchronous, ultramafic–mafic magmatism. Therefore, we suggest that mantle-plume-induced, post-orogenic magmatism may have been common throughout the Tian Shan but did not necessarily account for all of it. Other documented post-collisional geodynamic processes with distinct magmatic symptoms include sub-arc lithospheric removal to generate ~ 283 – 277 Ma juvenile alkaline magmas in the southwestern Tian Shan (this study), as well as slab break-off-facilitated interaction between asthenosphere and a subduction-enriched mantle wedge to generate ~ 293 – 280 Ma shoshonitic magmas in the western middle Tian Shan [Konopelko *et al.*, 2016]. Post-orogenic magmatism related to these local geodynamic events evidently occurred simultaneously to regional, Tarim-plume-induced post-orogenic magmatism and demonstrates that alternative and perhaps localized post-collisional geodynamic mechanisms likely generated similar magmatic symptoms in regions of the Central Asian Orogenic Belt distant from the Tarim basin.

6.6. Timing and Nature of Amphibolite–Granulite-facies Metamorphism in the Garm Massif

The timing overlap between protracted, amphibolite–granulite-facies metamorphism (and migmatization) in the Garm massif and granitoid magmatism in the Gissar arc (Figures 9 and 14) indicates that these processes were closely related. Taken at face value, we interpret the broad synchronicity between amphibolite-facies metamorphism and arc magmatism to represent contact metamorphism and melting of the framework crustal rocks in the Gissar microcontinent due to pervasive heating that accompanied the intrusion of the voluminous Gissar-arc melts [e.g., Käßner *et al.*, 2016b]. However, because the Karakum–Tarim–Kazakh–Kyrgyz continental collision was underway in the southwestern Tian Shan by the time the Turkestan and Gissar back-arc ocean basins closed in the latest Carboniferous [Burtman, 2010], bulk strain due to a continuation or non-instantaneous slowdown of the pre-collisional, regional convergence rate would likely have been accommodated by crustal thickening in the Gissar microcontinent. Therefore, we infer that a component of M–HT/LP, Barrovian metamorphism due to tectonic burial and crustal thickening [Konopelko *et al.*, 2015] also contributed to Garm-massif metamorphism and migmatization during the middle–late stages of the Gissar arc. The individual incompleteness of these equally compelling and mutually compatible end-member models for early Permian metamorphism in the Garm massif is illustrated by the ~ 20 – $29^\circ\text{C}/\text{km}$ paleo-geotherm range implied by our and Konopelko *et al.*'s [2015] prograde–peak metamorphic–pressure estimates (Table S12). Whereas conductive heating alone can account for the lower–middle paleo-geotherm range, the upper range (28 – $29^\circ\text{C}/\text{km}$) implies a component of intracrustal advection [e.g., Brown, 2008] that we attribute to inundation by the Gissar-arc granitoid melts, for which our granitoid thermobarometry implies an ~ 23 – $34^\circ\text{C}/\text{km}$ average paleo-geotherm range.

We infer a causative link between the overlapping timing (Figure 14) of peak-temperature, granulite-facies metamorphism ($\sim 288 \pm 6$ Ma; Figure 11), and the shift from syn-collisional, Gissar-arc, granitoid magmatism to post-collisional, juvenile, alkaline magmatism in the southwestern Tian Shan (~ 288 – 286 Ma; Figure 13). Compared to the early syn-collisional timeframe, the late, 37 – $54^\circ\text{C}/\text{km}$ geotherm range implied by our peak-metamorphic temperature thermometry and pseudosection modeling for our Garm paragneiss sample 12TS62 (Table S12) is best accounted for by a dominance of intracrustal advection, rather than conductive heating, in controlling the thermal evolution of the notably hotter Garm-massif crust [e.g., Brown, 2008]. Given the geodynamic context, we attribute the stronger post-collisional intracrustal advection to the delamination of our inferred sub-Gissar arclogite root at ~ 288 – 286 Ma, which gave rise to convective asthenospheric upwelling and was likely associated with a shallow (~ 50 – 60 km deep) lithosphere-asthenosphere boundary at this time [e.g., Hermann and Rubatto, 2003] (Figure 15).

Our overall interpretation of metamorphism in the Garm massif is that it occurred by a combination of overprinting mechanisms from ~ 310 – 282 Ma that together define a clockwise P-T-t evolution. The early majority of Garm metamorphism (~ 310 – 288 Ma; Figures 9 and 11) represents long-lived, prograde, amphibolite-facies, M-HT/LP metamorphism with peak-pressure conditions of $\sim 750^\circ\text{C}/7$ kbar (650 – $750^\circ\text{C}/6$ – 7 kbar range) that we attribute to both inundation by the Andean-syn-collisional Gissar arc melts (contact metamorphism) and syn-collisional crustal thickening (Barrovian metamorphism) (Figures 6 and 7 and Table S12). Late (288 ± 6 Ma), lower-granulite-facies, HT/LP metamorphism with peak-temperature conditions of $820^\circ\text{C}/4$ kbar (800 – $820^\circ\text{C}/6$ – 4 kbar range) accompanied the transmission of a mantle-derived heat wave through the Garm-massif crust, which we attribute to post-collisional convective asthenospheric upwelling that accompanied the delamination and foundering of a gravitationally unstable Gissar arclogite root subsequent to the magmatic quiescence of the Gissar arc (Figure 15). Irrespective of timing constraints, resolved metamorphic-temperature estimates for the Garm massif from this study and Konopelko *et al.* [2015] correlate, as a whole, inversely with distance from the town of Garm (Figure 2), the name of which serendipitously means “hot” in the Tajik language.

6.7. Tectonic Continuity (or Lack Thereof) Across the Talas-Fergana Fault

Whereas the ~ 1500 km along-strike extent of the Turkestan suture along the Tian Shan indicates continuous along-strike subduction of the Turkestan ocean basin, this is apparently not the case for the Gissar back-arc ocean basin because the Gissar suture does not cross the Talas-Fergana fault (Figure 1). The late Paleozoic tectonic setting that characterized the Gissar arc—collision-induced, short-lived, northward subduction of a back-arc ocean basin to generate voluminous granitoid arc magmas—may be analogously represented by the Kule-Heiyingshan ophiolitic mélange ~ 1000 km to the east in China (Figure 1), which Alexeiev *et al.* [2015] argue represents a back-arc basin that opened along the northern Tarim terrane during the Silurian due to southward subduction of the Turkestan ocean basin and closed from ~ 320 – 300 Ma via southward subduction. With respect to the Gissar back-arc basin, this would imply that the putatively analogous Kule-Heiyingshan back-arc basin opened earlier and was subducted with an opposite polarity during an overlapping (late Carboniferous) timeframe. This analogy is inescapably contentious, however, because in addition to the significant separation along-strike from the southwestern Tian Shan, the Kule-Heiyingshan ophiolitic mélange may represent allochthonous, southward transported fragments of the Turkestan suture [Windley *et al.*, 1990; Gao *et al.*, 1998; Biske, 1998; Burtman, 2008] rather than a separate back-arc ocean basin.

Our tectonic model is consistent with delayed Turkestan suture formation west of the Talas-Fergana fault compared to east of the Talas-Fergana fault. Whereas ~ 320 – 305 Ma ages for (U)HP/LT metamorphism and rapid cooling along the Turkestan suture in the Kokshaal segment of the South Tian Shan [Wei *et al.*, 2009; Zhang *et al.*, 2009; Hegner *et al.*, 2010; Su *et al.*, 2010; Li *et al.*, 2011; Tian and Wei, 2012; Du *et al.*, 2014] attest to the onset of the Tarim-Kazakh-Kyrgyz collision in the east, the documented Andean nature of coeval, ~ 318 – 300 Ma magmatism in the Chatkal-Kurama arc [Seltmann *et al.*, 2011; Dolgoplova *et al.*, 2016; Konopelko *et al.*, 2016] attests to the persistence of the Turkestan ocean basin in the west until the latest Carboniferous. Given this constraint, because the Tarim-Karakum-Kazakh-Kyrgyz collision followed the formation of the Turkestan suture, HP/LT metamorphism and exhumation of the eclogite facies Kassan metamorphic complex in the Middle Tian Shan (Figure 1), which is adjacent to the Chatkal-Kurama arc and a proxy for the collision, must have followed Andean magmatism in the Chatkal-Kurama arc. Therefore, Loury *et al.*'s [2016] imprecise, $\sim 301 \pm 15$ Ma age assignment for eclogite-facies, HP/LT metamorphism in the Kassan

metamorphic complex (U-Pb lower intercept age for allanite inclusions in garnet) supports our model, whereas *Mühlberg et al.*'s [2016] precise, ~321–311 Ma age assignment (garnet/whole-rock Sm-Nd and mica $^{40}\text{Ar}/^{39}\text{Ar}$) conflicts. Our tectonic model is also consistent with a westward extrapolation of *Han et al.*'s [2016] model for the diachronous, westward-propagating formation of the Turkestan (South Tian Shan) suture beyond the Kokshaal South Tian Shan, across the Talas-Fergana fault, and into the western Tian Shan. We note that potentially unresolved geologic complexity in the western middle Tian Shan, including the displacement of the Chatkal-Kurama arc and the Kassan metamorphic complex ~100 km from the Turkestan suture, may account for the apparently incompatible age constraints.

The primary phase of dextral movement along the Talas-Fergana from ~300–270 Ma [*Alexeev et al.*, 2009a; *Voytenko and Khudoley*, 2012; *Konopelko et al.*, 2013; *Rolland et al.*, 2013; *Konopelko et al.*, 2017], by which time the Karakum-Tarim-Kazakh-Kyrgyz collision had commenced and/or was complete (Figure 14), indicates that the Talas-Fergana fault accommodated syn-collisional-post-collisional kinematic reorganization at a regional scale, from convergent to strike-slip tectonics. Whereas the ~323–306 Ma, Andean phase of the Gissar arc overlapped with a proposed early, ~315–300 Ma phase of low-slip, top-to-the-south, dextral-transpressional deformation along the Talas-Fergana fault [*Rolland et al.*, 2013], the ~304–288 Ma, syn-collisional phase of the Gissar arc coincided with the early component of primary dextral slip along the fault. Sub-Gissar-arc lithospheric foundering from ~288–286 Ma, as well as post-collisional magmatism from ~283–265 Ma in the Gissar range, coincided with the late component of primary dextral slip along the Talas-Fergana fault. We speculate that the Talas-Fergana fault may have provided a partial kinematic barrier between the western and eastern Tian Shan during the Tarim-Karakum-Kazakh-Kyrgyz collision, which facilitated syn-collisional convergence and magmatism in the Gissar arc under the recently doubly sutured Gissar microcontinent. In this light, we tentatively interpret the apparent ~288–286 Ma geodynamic transition from syn-collisional to post-collisional tectonics in the Gissar range to reflect an exceeded threshold of the ability of the Talas-Fergana fault to maintain partial kinematic segregation between the western and eastern Tian Shan. Finally, we propose that kinematic integration of the southwestern Tian Shan into the post-collisional framework of the Central Asian Orogenic Belt ensued for the remaining duration of dextral slip along the Talas-Fergana fault.

7. Conclusions

Our investigation of the late Paleozoic tectono-magmatic-metamorphic evolution of the southwestern Tian Shan thematically centers on the birth, life, and demise of the Andean–syn-collisional Gissar arc, revealing the following conclusions:

1. Our detrital-zircon data indicate that the M-HT/LP Garm massif and the M-HP/LT Fan-Karategin belt (which is a component of the Zeravshan subduction-accretion complex) share a similar provenance (Figure 10) and represent the peri-Gondwanan, Karakum-Tarim terrane, which underwent divergent tectono-metamorphic histories and hosted the development of the Gissar arc. Despite the observation that detrital-zircon maximum-depositional ages for metasedimentary rocks span the Ediacaran for both units, it is ambiguous whether or not they represent a single stratigraphic sequence because the Zeravshan complex comprises a lower–upper Paleozoic stratigraphy, whereas that for the Garm massif is unconstrained. Primarily Ediacaran magmatism, represented by ~670–550 Ma granitoids in the Garm massif [*Käbner et al.*, 2016b], represents a likely source for matching age peaks in the Garm massif and the Fan-Karategin belt.
2. M-HP/LT metamorphism in the Fan-Karategin belt, which is a metamorphosed component of the Zeravshan subduction-accretion complex (Figure 2), spanned upper-greenschist–lower-blueschist facies, is bracketed between ~450 and 316 Ma and may have occurred during the early Carboniferous (Viséan; ~347–331 Ma) [*Baratov*, 1976; *Shvanov*, 1983; *Biske*, 1996] due to southward subduction of the Turkestan ocean basin under the northern Karakum terrane margin. Relict sodic amphibole (magnesian riebeckite) in a retrogressed, epidote-blueschist-facies metabasite (sample 13TS17; Figure 3f) provides evidence for HP/LT metamorphism, whereas such evidence is absent in Fan-Karategin metasedimentary rocks, which were dominantly retrogressed under greenschist-facies conditions. The Fan-Karategin belt is best contextualized within the framework of accretionary tectonics that spanned the Central Asian Orogenic Belt during the Paleozoic and does not represent a lower-pressure analog to the Kassan HP/LT or composite Kokshaal (U)HP/LT belts in Kyrgyzstan and China (Figure 1) because its formation preceded that of the Turkestan suture.

3. Southward, slab-rollback subduction of the Turkestan ocean basin under the northern Karakum terrane may have caused back-arc continental rifting in the northern Karakum terrane during the early Carboniferous (Tournaisian–early Serupkhovian; ~350–330 Ma) that evolved into oceanic spreading and opening of the Gissar back-arc ocean basin during the middle Carboniferous (early Serupkhovian–Bashkirian or early Moscovian; ~330–315 Ma) and rifted off the Gissar microcontinent (Figures 14 and 15) [Brookfield, 2000; Burtman, 2010]. Closure of the Gissar back-arc ocean basin via northward subduction under the Gissar microcontinent occurred during the middle to late Carboniferous (Bashkirian–Gzhelian; ~320–304 Ma) [Baratov, 1976; Brookfield, 2000; Burtman, 2010] and gave rise to Andean magmatism in the Gissar arc from ~323–306 Ma. (Figure 9).
4. The Gissar batholith and the Garm massif represent an eastward, along-strike increase in paleodepth from upper-batholith to root levels of the Andean–syn-collisional Gissar arc (Figure 2), which developed from ~323–288 Ma in two stages [Dolgoplova et al., 2016; Käbner et al., 2016b; Konopelko et al., 2016; this study]: (1) Andean, I-type granitoid magmatism in the Gissar batholith from ~323–306 Ma due to northward subduction of the Gissar back-arc ocean basin under the Gissar microcontinent; which was immediately followed by (2) syn-collisional, I–S-type granitoid magmatism in the Gissar batholith and the Garm massif from ~304–288 Ma due to northward subduction/underthrusting of previously thinned Karakum marginal-continental crust under the Gissar microcontinent (Figures 9, 14, and 15 and Table S13). Syn-collisional magmatism in the Gissar batholith spanned ~304–292 Ma (~299 Ma peak age) and preceded and temporally overlapped with that in the Garm massif, which spanned ~302–288 Ma (~293 Ma peak age). We propose that the input of melt-fertile, continental-margin crust into the subduction zone catalyzed syn-collisional arc magmatism, which granitoid-crystallization age densities suggest was higher-flux than the preceding Andean magmatism. Al-in-Hbl geobarometry indicates that early, Andean, I-type Gissar-batholith magmas were emplaced at middle-crustal levels (~33–15 km), whereas syn-collisional, I–S-type magmas were emplaced at middle–upper-crustal levels (~21–7 km) in the Gissar batholith [Käbner et al., 2016b] and lower–middle-crustal levels (~36–19 km) in the Garm massif. Whereas zircon ε_{Hf} signatures (Figure 13) indicate moderate assimilation of old and/or evolved continental crust during the early, Andean phase the Gissar arc, significant crustal assimilation occurred during the late, syn-collisional phase of the Gissar arc.
5. A rapid zircon ε_{Hf} pull-up from ~288–286 Ma (Figure 13) defined by Ne syenite sample 12TS96 and corroborated by sedimentary samples indicates a shift from intermediate–evolved, apparently high-flux, syn-collisional Gissar-arc magmatism from 302–288 Ma to juvenile, alkaline-syenite, apparently low-flux, post-collisional magmatism in the Garm massif by ~280 Ma (Figures 2, 3b, and 9). We interpret the rapid shift from intermediate–evolved to juvenile magmatism to reflect lithospheric foundering under the Gissar arc, which was accompanied by convective asthenospheric upwelling, low-volume melt generation, and heat transmission through the crust. Juvenile, alkaline-syenitic, post-collisional magmatism in the southwestern Tian Shan was followed by evolved, alkaline-syenitic [Konopelko et al., 2016], and other geochemically diverse post-collisional magmatism that more broadly characterized the Central Asian Orogenic Belt.
6. Metamorphism and migmatization in the Garm massif, which is represented by amphibolite–granulite-facies paragneisses, metabasites, and migmatites, occurred by a combination of overprinting mechanisms from ~310–282 Ma that together define a clockwise P–T–t evolution. Integrated metamorphic-zircon and monazite U–Pb ages, thermobarometry and pseudosection modeling indicate that the early majority of Garm metamorphism (~310–288 Ma; Figures 9 and 11) represents long-lived, prograde, amphibolite-facies, M–HT/LP metamorphism to peak-metamorphic-pressure conditions of ~750°C/7 kbar (650–750°C/6–7 kbar range for this study, which overlaps with a previous estimate of 600–660°C/7–9 kbar) [Konopelko et al., 2015; Käbner et al., 2016b; this study]. We attribute early, amphibolite-facies metamorphism to both inundation by the Gissar arc melts (contact metamorphism) and crustal thickening (Barrovian metamorphism) due to the Karakum–Tarim–Kazakh–Kyrgyz collision (Figures 6 and 7 and Table S12). Integration of a 288 ± 6 Ma monazite U–Pb age with thermobarometry and pseudosection modeling for our Garm paragneiss sample 12TS62, in which late Crd + Spl replaced early Grt + Sil, indicate that heating and decompression from upper-amphibolite to lower-granulite facies accompanied the transmission of a mantle-derived heat wave through the Garm-massif crust and defined a transition to HT/LP metamorphism, with peak-metamorphic-temperature conditions of 820°C/4 kbar (800–820°C/6–4 kbar range). We attribute late, HT/LP metamorphism to post-collisional convective asthenospheric upwelling that accompanied the delamination and foundering of a gravitationally unstable Gissar arclogite root subsequent to the magmatic

quiescence of the Gissar arc (Figure 15). Comparatively poor modern constraints for the petro-tectonic evolution of the Lolabulak and Baysunta HT/LP migmatite-granitoid-gneiss complexes in Uzbekistan render a potential genetic relationship with the lithologically similar Garm massif ambiguous.

7. Our model is consistent with a westward extrapolation of *Han et al.*'s [2011] model for the diachronous, westward propagation of the formation of the Turkestan suture from ~320–300 Ma beyond the Kokshaal South Tian Shan, across the Talas-Fergana fault, and into the western Tian Shan. The Turkestan and Gissar sutures formed more or less at the same time, ushering in the onset of the Karakum-Tarim-Kazakh-Kyrgyz collision in the southwestern Tian Shan by ~300 Ma.
8. Previously proposed phases of early, low-magnitude, dextral transpressive slip (~315–300 Ma) and the early stage of a primary phase of ~200 km dextral slip along the Talas-Fergana fault (~300–270 Ma), respectively, temporally overlap with Andean and syn-collisional magmatism in the Gissar arc. We interpret this broad timing overlap to indicate that the Talas-Fergana fault may have provided a partial kinematic barrier between the western and eastern Tian Shan during the Tarim-Karakum-Kazakh-Kyrgyz collision that controlled the distribution of syn-collisional convergence and magmatism in the South Tian Shan to the west (in the Gissar arc) but not to the east. Subsequent to local, sub-Gissar arc lithospheric removal and juvenile, alkaline-syenitic, post-collisional magmatism, the southwestern Tian Shan was kinematically integrated into the regional post-collisional tectonic setting of the Central Asian Orogenic Belt for the remaining significant duration of dextral slip along the Talas-Fergana fault.

Appendix A: Thin-section-based Petrographic Descriptions of Igneous- and Metamorphic-rock Samples

A1. Gissar-batholith Granitoids

12TS06: Gissar Granite. Anhedral Qz and sericitized Fsp.

12TS08: Gissar Bt granite. Fresh and chloritized Bt. Sericitized Fsp and myrmekite present.

12TS09: Gissar Amp + Bt Qz diorite. Fresh Amp + Bt. Sericitized Fsp present. Ilm (with Mn) and Py verified with EDS. Py exhibits rust/hematite coronas. No Ttn found in thorough EDS search.

12TS12: Gissar Amp + Bt granite. Weakly retrogressed Amp + Bt. Feldspar commonly perthitic and sericitized.

14TS02: Gissar Bt granite.

A2. Garm-massif Granitoids

12TS23A: Garm Bt granodiorite. Weakly deformed. Bt is commonly replaced by Chl. Pl shows common concentric zoning. Fsp is commonly sericitized. Ttn occurs in small isolated domains. Qz shows domains of weak dynamic recrystallization, indicating low strain.

12TS30: Garm Bt + Grt granodiorite. Likely peraluminous bulk composition. Fresh Bt 700–1500 μm long. Two to three millimeter Grt is fractured and embayed. Myrmekite is present. Fsp sericitization is uncommon and weak.

12TS58: Garm Bt Qz monzonite. Bt is commonly chloritized. Fsp is commonly sericitized, forming coarse-grained, irregular, secondary Ms.

12TS91: Garm Grt Qz syenite. Fractures in anhedral-subhedral, 300–800 μm Grt are filled with secondary mica. Fsp is commonly sericitized, forming coarse-grained, irregular, secondary Ms. Bt is commonly chloritized. Perthite and myrmekite are present.

12TS96: Garm Bt + Cal Ne + Ccn + Nsn(?) syenite. Bt + Ccn + Cal are fresh and appear to be primary. Fsp has strong perthitic texture. Primary feldspathoid and carbonate indicate crystallization from a magma undersaturated in SiO_2 and containing CaCO_3 . Cal appears primary but likely crystallized late in the sequence.

12TS102: Garm Bt granite. Weakly deformed. One to three millimeter Fsp grains are weakly stretched, show undulatory extinction and are moderately sericitized. Red 400–1200 μm Bt grown between Fsp and Qz. Qz shows undulatory extinction and weak dynamic recrystallization.

12TS111: Garm Amp + Bt + Ep granite. Strongly retrogressed. Bt and Ep are strongly chloritized. Amp is retrogressed. Fsp is strongly sericitized. Ilm presence confirmed with EDS.

12TS114: Garm Amp + Bt granite. Bt is weakly chloritized. Brown-green Amp is moderately fresh. Fsp is moderately sericitized. Zoned Pl is common. Euhedral Ttn. Ilm (confirmed with EDS) occurs both as acicular growths parallel to amphibole cleavage and as ~100 μm inclusions in Amp.

12TS121: Garm Bt + Ms granite. Bt and Ms appear primary. Fsp is moderately sericitized. Pl commonly zoned.

A3. Garm-massif Gneisses

12TS34: Retrogressed Bt + Grt quartzofeldspathic paragneiss. Anastomosing foliation is defined by aggregates of 200–1200 μm Bt grains. Feldspar clasts are sericitized and 40–1200 μm wide. Qz is dynamically recrystallized. Two hundred to five hundred micrometers of Grt has embayed grain boundaries.

12TS47: Fresh, meta-mafic Amp + CPx amphibolite. Foliation defined by interlayered 1–2 mm thick Amp (Mhb-Act) and Pl (An_{52–61}) aggregates. Olive-green Amp grains are euhedral and prismatic, 200–2000 μm long, and weakly aligned. CPx (Hd_{86–90}Di_{10–14}) intergrown with Amp. Pl forms granoblastic aggregates with internally consistent grain size ranging from 100–500 μm . Overall texture indicates static (likely contact) metamorphism. Low Zr content in Ttn indicates a mafic protolith.

12TS59: Retrogressed Amp (Mhb-Ed) + CPx (Aug) + Ilm paragneiss. Brown Amp is abundant, fresh, 100–400 μm long, weakly aligned, and defines foliation. CPx is much less abundant, fresh, and 100–200 μm in diameter. Abundant aggregates of aligned 50–100 μm Ilm grains are intergrown and aligned with Amp. Pl clasts are subhedral, strongly sericitized and 500–1000 μm wide.

12TS62: Retrogressed, non-annealed, Grt + Crd + Sil + Bt + Hc paragneiss. Foliation defined primarily by 200–800 μm long, red-brown Bt and aggregates of yellow pinite. Grt is coarse-grained (1–3 mm long), elongated (length:width ratio up to 2), and strongly resorbed (replaced by pinite), with grain boundaries that are euhedral to first order but anhedral in detail. Grt resorption is also reflected in “kickups” for Mn and Fe# observed in EPMA transects (Figure S2). Contrast-sensitive BSE images for an ~ 3 mm garnet indicate that the retrogressive re-equilibration front encroaches ~ 200 μm into the grain past its current grain boundary. Average Grt compositions are Alm₆₇Prp₁₈Sps₁₂Grs₃ for cores, Alm₆₉Prp₁₇Sps₁₁Grs₃ for mantles, and Alm₇₂Prp₁₄Sps₁₁Grs₃ for rims. Bt is primarily foliation-parallel but also rarely occurs in a foliation-oblique or foliation-perpendicular context. The Mg# for Bt is, on average, 0.43 adjacent to Grt and 0.40 in the matrix. Foliation-parallel Hc aggregates occur with Sil + pinite $\pm$ Ilm and often trace the former grain boundaries of resorbed Grt. Hc exhibits minor solid solution with Spl, having an average composition of Hc_{83–84}Spl_{16–17}. Sil (fibrolite) is parallel to foliation. Ilm exhibits minor solid solution with Pph with an average composition of Ilm₈₈Pph₁₂. Average Pl compositions are Ab₆₉An₂₈Or₂ for Pl rims in contact with Grt and Ab₇₁An₂₆Or₃ for matrix Pl. Anhedral 100–1000 μm Qz and subhedral 200–500 μm Pl form aggregates that define compositional layering with Bt. Trace Ms is associated with Bt and pinite and is both foliation parallel and oblique. Relict Crd and Kfs within pinite were identified with EDS and indicated in X-ray element maps. Pinite comprises Crd retrogressed/alterated to Ms + Prl/Als + Chl + Qz in the presence of a K⁺-bearing fluid (Pg commonly present). Retrogressive pinite involves a breakdown of Crd to Prl/Als + Chl + Qz followed by sericitization from a K⁺-bearing fluid to form Ms, whereas alteration involves a direct change of mineralogy by a K⁺-bearing fluid to form Ms + Prl/Als + Chl + Qz. Relict Kfs observed within pinite with EDS. X-ray element maps indicate that pinite contains intermediate K + Al, intermediate-low Si, and low Fe + Mg.

12TS73: Weakly retrogressed Bt + Grt paraschist. Aligned, 200–800 μm brown Bt aggregates define a slaty cleavage. One to two millimeter Qz grains are elongated parallel to foliation and have undulatory extinction. Three hundred to four hundred micronmeter Grt is anhedral, fractured perpendicular to foliation, and embayed. Kfs is commonly sericitized where associated with Grt + Bt

12TS101: Weakly retrogressed Bt $\pm$ Grt paragneiss. Anastomosing foliation defined by aggregates of 300–1200 μm Bt grains with isolated domains of chloritization. Aggregates of subhedral Qz + moderately sericitized Pl define compositional layering with grain size 200–800 μm . One hundred to three hundred micronmeter Grt is fractured and embayed.

A4. Zeravshan Subduction-Accretion Complex and Fan-Karategin M-HP/LT Belt Greenschists and Blueschists

12TS13: Zeravshan/Gissar boundary-zone Bt + Qz + Ms metasedimentary phyllonite. Foliation defined by Qz + Ms + Bt. Fine-grained Bt is ~ 50 μm long. Relict 1–2 mm long porphyroclasts are replaced by randomly oriented Qz + Ms + Bt + Chl.

12TS22: Fan-Karategin/Garm boundary Bt + Ms + Chl + And metasedimentary greenschist. Foliation is defined by mica, moderately stretched Qz, and spaced 0.5–1.5 mm Qz bands. Bt and Ms are ~300–500 μm long and intergrown with each other, with common retrogressive Chl replacement. Two hundred to five hundred micrometer-sieved andalusite porphyroclasts are enveloped by Bt + Ms foliation. Porphyroblastic Ttn is 100–200 μm in diameter, Rt is ~100 μm in diameter. An elongate opaque mineral is parallel to the foliation. Ep and Amp (Act), expected for greenschist-facies conditions, are notably absent. Given the location of this sample along the boundary between the Garm massif and the Fan-Karategin belt, it may represent a telescoping of both units and a Chl-facies overprint of a M–HT/LP assemblage.

12TS38: Fan-Karategin Chl + Ms + Ep metasedimentary greenschist. Moderately stretched/aligned Qz comprises most of this rock. Chl + Ms $\pm$ Bt aggregates define disjunctive, anastomosing foliation. Common 10–50 μm Zrn grains occur as scattered inclusions in 200–500 μm Fsp. Fine-grained (~50 μm), euhedral, detrital Ttn lacks evidence of resorption and occurs in layers that indicate relict sedimentary layering. Detrital character for Ttn is based on Ttn Tera-Wasserburg lower intercept U–Pb ages that match Zrn age peaks in this sample. Euhedral, 100–200 μm Ep and Czo grains indicate greenschist-facies metamorphism.

12TS83: Fan-Karategin fine-grained Ms + Chl + Ep metasedimentary(?) greenschist. Foliation is defined by Ms + Chl, moderately stretched ~30 μm Qz, and elongated aggregates of a fine-grained brown mineral. 200–500 μm Pl porphyroclasts and 50–200 μm Ep porphyroclasts have strong pressure shadows with Qz + Chl $\pm$ Ms. Rare post-kinematic ~70 μm Rt lacks pressure shadows.

12TS88: Fan-Karategin (Gorif-type) Ms + Bt $\pm$ Chl + Act + Ep + Czo metasedimentary(?) greenschist. Abundant 300–700 μm Ms and acicular 200–1500 μm Act, define the foliation. Fifty to two hundred micrometer Ep abundant. One hundred to four hundred micrometer Ttn. Five hundred to three thousand micrometer Pl porphyroclasts have abundant inclusions. Ms may represent retrogressed Phe. This sample was likely an epidote-blueschist-facies rock that was strongly retrogressed under greenschist-facies conditions.

12TS89: Fan-Karategin (Gorif-type) Chl + Ms metasedimentary(?) greenschist. Disjunctive foliation is defined by anastomosing Chl + Ms aggregates and stretched Qz. Approximately 100- μm opaque mineral (likely Py) scattered among Chl/Ms aggregates. Ms may represent retrogressed Phe. Yellow staining common in fine-grained Chl. Ttn commonly elongated up to 600 μm . Late Cc vein crosscuts primary fabric.

12TS94: Fan-Karategin (Yagnob-type) meta-andesitic Act + Ep + Czo + Ms + Chl greenschist. Semi-continuous foliation is defined by Ms + Chl, stretched Qz/Pl, and stretched Amp. Amp is exclusively Act, elongated, and up to 1.5 mm long. Pl appears porphyroblastic. Rare 50–200 μm Ep and elongated, foliation-parallel, 100–200 μm Czo. Ttn porphyroblasts contain pressure shadows with Ms + Qz. Monogenetic Zrn U–Pb age (449.9 ± 4.6 Ma) indicates a volcanic protolith and lack of Px in mineral assemblage indicates an andesitic–basaltic–andesitic bulk composition. Yagnob-type context indicated by abundant Ordovician–Silurian fossils.

12TS112: Low-grade Zeravshan-complex Bt metasedimentary(?) phyllonite. Alternating Bt- and Qz-dominated domains define foliation. Bt is 100–400 μm long and acicular. Euhedral 100–200 μm Tur occurs in Bt domains. Subhedral Qz forms aggregates of weakly aligned grains with consistent grain size ranging from 50 to 200 μm .

13TS17: Fan-Karategin meta-basaltic, strongly retrogressed blueschist (Yagnob-type). CPx + Ep + Czo + Act + Mrbk + Wnc–Brs + Cc + Chl + Ms + Pl assemblage indicates strong greenschist-facies overprint of peak blueschist-facies metamorphic pressures. Foliation is defined by Chl + Ms + Amp + Ep + Czo groundmass and elongated CPx ($\text{Hd}_{99}\text{Di}_{01}$) porphyroclasts, which are 500–2000 μm long and associated with Cc + Chl + Pl in pressure shadows. Groundmass Amp is likely Act. Relict blue Amp is Mrbk–Wnc, which indicates blueschist-facies metamorphic pressures. Ttn is common and fine-grained (~30 μm).

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