A Preliminary Framework for Magmatism in Modern Continental Back-Arc Basins and Its Application to the Triassic-Jurassic Tectonic Evolution of the Caucasus

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Abstract Extension within a continental back-arc basin initiates within continental rather than oceanic lithosphere, and the geochemical characteristics of magmatic rocks within continental back-arcs are poorly understood relative to their intraoceanic counterparts. Here, we compile published geochemical data from five exemplar modern continental back-arc basins—the Okinawa Trough, Bransfield Strait, Tyrrenhenian Sea, Patagonia plateau, and Aegean Sea/Western Anatolia—to establish a geochemical framework for continental back-arc magmatism. This analysis shows that continental back-arcs yield geochemical signatures more similar to arc magmatism than intraoceanic back-arcs do. We apply this framework to published data for Triassic-Jurassic magmatic rocks from the Caucasus arc system, which includes a relict continental back-arc, the Caucasus Basin, that opened during the Jurassic and for which the causal mechanism of formation remains debated. Our analysis of $^{39}$Ar/$^{40}$Ar and U-Pb ages indicates Permian-Triassic arc magmatism from ~260 to 220 Ma due to subduction beneath the Greater Caucasus and Scythian Platform. Late Triassic (~220–210 Ma) collision of the Iranian block with Laurasia likely induced trench retreat in the Caucasus region and led to migration of the Caucasus arc and opening of the Caucasus Basin. This activity was followed by Jurassic arc magmatism in the Lesser Caucasus from ~180 to 140 Ma and back-arc spreading in the Caucasus Basin from ~180 to 160 Ma. Trace element and Sr-Nd isotopic data for magmatic rocks indicate that Caucasus Basin magmatism is comparable to modern continental back-arcs and that the source to the Lesser Caucasus arc became more enriched at ~160 Ma, likely from the cessation of back-arc spreading.

Plain Language Summary Ocean basins are consumed by the process of subduction, in which the crust and upper mantle of one tectonic plate are dragged beneath another plate. Although subduction happens when two plates are converging, about a third of subduction zones are characterized by areas where the overriding plate is stretched and thinned, sometimes forming a back-arc basin: a small basin behind the volcanic arc associated with the subduction zone. The Caucasus Mountains within the Arabia-Eurasia continental collision formed over the past 35 million years by closure of a poorly understood ancient back-arc basin called the Caucasus Basin. This study compiles published data on the age and chemistry of igneous rocks within this basin, its associated volcanic arc, and comparable modern back-arc basins to understand how extension began, which informs how the basin closed to form the Caucasus Mountains. These data are best explained by a model, previously proposed for modern back-arc basins, in which the collision of Iran about 220–210 million years ago with Eurasia caused extension in the Caucasus region that resulted in magmatism about 180–160 million years ago. The main volcanic arc associated with the subduction zone migrated southward as a result.

1. Introduction

Subduction zones exert first-order controls on the Earth system by facilitating transfer of material between Earth’s surface and mantle and by bringing continents together via ocean basin closure (e.g., Collins, 2003; C.-T.A. Lee et al., 2015; R. J. Stern, 2002). Despite accommodating convergence between tectonic plates, subduction zones are often characterized by extension in the overriding plate, forming back-arc basins behind the main volcanic arc (e.g., Karig, 1971; Uyeda & Kanamori, 1979). These small, often short-lived ocean basins along convergent margins complicate the tectonic history of ocean basin closure (e.g., Cowgill et al., 2016; Dewey, 1977; McKenzie et al., 2019), provide a means of initiating continental breakup (e.g., Dal Zilio et al., 2018; Stampfli et al., 2013; Yang et al., 2021), and can create new lithosphere (e.g., Pearce &
Stern, 2006; Saunders & Tarney, 1984; Woodhead et al., 1993). As a result, investigating the mechanisms of back-arc basin formation and how such basins evolve over time is critical for understanding how subduction systems govern the long-term assembly and distribution of Earth’s lithosphere.

It is generally accepted that overriding-plate extension in a subduction zone requires either slab rollback in the downgoing plate with accompanying trench retreat or migration of the overriding plate away from the trench (e.g., Sdrolias & Müller, 2006; Uyeda & Kanamori, 1979). Parameters that may determine whether trench retreat or overriding-plate migration will occur include the age and dip angle of the subducting lithosphere, relative plate velocities, mantle flow dynamics, and subduction polarity (e.g., Carlson & Melia, 1984; Doglioni et al., 2007; Faccenna et al., 2001; Nakakuki & Mura, 2013; Sdrolias & Müller, 2006). Observations of modern arc systems have revealed a close correlation between the presence of back-arc basins, arc curvature, and recent along-strike entry into the subduction zone of a buoyant indenter, such as a continent or seamount (McCabe, 1984; Vogt, 1973; Wallace et al., 2009). As a result, a primary mechanism for initiating back-arc basin formation is thought to be collision of the arc with a buoyant indenter, which induces back-arc spreading along strike of the collision by fore-arc rotation and trench retreat (Magni, 2019; Magni et al., 2014; Wallace et al., 2005, 2009). Understanding how back-arc basins form in modern systems provides a critical tool for unraveling the history of past tectonics in areas that have undergone complex, protracted deformation.

Because most well-studied modern back-arc basins initiated within oceanic lithosphere as intraoceanic back-arc basins (e.g., Lau Basin, East Scotia Sea, Mariana Trough; Saunders & Tarney, 1984), current understanding of the geochemical characteristics of back-arc magmatism is largely derived from intraoceanic back-arc basins. However, involvement of continental lithosphere has the potential to substantially impact the geochemical signatures of continental back-arc basin magmatism. To explore such potential impacts, we compile published trace-element and Sr-Nd isotopic data from five exemplar continental back-arc basins. Three are well-developed modern marine back-arc basins that initiated in continental lithosphere: the Bransfield Strait, the Okinawa Trough, and the Tyrrenhian Sea (Figures 1a–1d). The fourth is a subaerial continental back-arc basin in the Patagonia plateau in the incipient stages of development (Figure 1e), and the fifth is a partially subaerial and partially marine basin in the Aegean Sea and Western Anatolia (Figure 1f). From these published data, we develop a preliminary geochemical framework linking the geochemistry of continental back-arc magmatism to tectonic processes in this setting, finding that continental back-arc basins yield geochemical signatures more similar to modern arc magmas than intraoceanic back-arc basins do. We then apply this framework to an ancient continental back-arc for which the origin and evolution are disputed: the Caucasus Basin.

Between the Black and Caspian Seas, the present-day configuration of the Caucasus region is dominated by the closure of the Mesozoic-Cenozoic marine Caucasus Basin during the Late Cenozoic Arabia-Eurasia continental collision (Figure 1g; e.g., Adamia et al., 1977; Cowgill et al., 2016; Gamkrelidze, 1986; Vincent et al., 2016; Zonenshain & Le Pichon, 1986). Situated between the Paleozoic basement of the Greater Caucasus to the north and the Mesozoic Lesser Caucasus volcanic arc to the south, the Caucasus Basin is widely understood to have evolved in the Jurassic as a back-arc basin during northward subduction beneath the Lesser Caucasus arc (e.g., Adamia, Alania, et al., 2011; Saintot et al., 2006; Zonenshain & Le Pichon, 1986). In most paleogeographic models, the Caucasus Basin experienced its main phase of extension in the Middle Jurassic (e.g., McCann et al., 2010; Vincent et al., 2016; Zonenshain & Le Pichon, 1986). Such extension is assumed to have initiated within continental lithosphere, given evidence for subaerial deposition in the earliest stages of basin development and the exposure of continental basement on either side of the basin (e.g., Ershov et al., 2003; McCann et al., 2010; Vincent et al., 2016), making the Caucasus Basin an example of a continental back-arc basin. However, the causal mechanism by which the basin opened, as well as its evolution and spatial extent, remain poorly understood, leading to divergent views of the composition of the basement within the basin prior to Late Cenozoic closure (Cowgill et al., 2016, 2018; Vincent et al., 2016, 2018). Importantly, the spatial, temporal, and geochemical patterns of Triassic-Jurassic arc-related magmatism in the Caucasus have not been fully integrated to develop a complete and testable geodynamic model for the opening of a continental back-arc basin that later played a critical role in the Cenozoic evolution of the Arabia-Eurasia collision.
Figure 1. (a) Map showing relative locations of the Okinawa Trough, Bransfield Strait, Tyrrhenian Sea, Patagonia plateau, Aegean Sea/Western Anatolia, and Caucasus. Maps of the modern continental back-arc basins of (b) the Okinawa Trough, (c) Bransfield Strait, (d) Tyrrhenian Sea, (e) Patagonia plateau, and (f) Aegean Sea/Western Anatolia indicate locations of compiled geochemical analyses with SiO$_2$ <52 wt. % (circles) and SiO$_2$ 52–57 wt. % (squares), colored according to location within the basin. (f) Overview of tectonic elements involved in the Caucasus arc system and locations of compiled geochronologic (stars) and trace-element geochemical analyses (squares, circles). Symbol color indicates method for geochronology and tectonic domain for trace-element geochemistry. U—Ustica, E—Enarete, Pr—Prometeo, Pa—Palinuro.
Here we propose a geodynamic model explaining the Mesozoic tectonic evolution of the Caucasus arc-back-arc system by interpreting published geochemical and geochronologic data from the Caucasus region in the context of the preliminary geochemical framework we develop from modern continental back-arc basins. Data from the Caucasus are best explained by a model in which Late Triassic collision of the Iranian block to the east initiated trench retreat, southward migration of the magmatic arc front in the Caucasus, and opening of the Caucasus Basin by Middle Jurassic time. Our compilation of U-Pb and ⁴⁰Ar/³⁹Ar analyses indicates major pulses of arc-related magmatism north of the Caucasus Basin in the Permian-Triassic (~260–220 Ma) and the Middle Jurassic (~180–160 Ma), as well as south of the basin in the Middle-Late Jurassic (~180–140 Ma). Trace element and Sr-Nd isotopic data in the Caucasus Basin dominantly exhibit geochemical signatures similar to modern arc magmatism, with smaller subsets of samples indicating more depleted or enriched mantle sources. This is comparable with signatures found in modern continental back-arcs, and this combination of geochemical signatures is most consistent with a marine back-arc basin that has experienced at least some seafloor spreading. Temporal shifts in the geochemistry of the Lesser Caucasus arc from more depleted incompatible element signatures in the Middle Jurassic to more enriched signatures in the Late Jurassic suggest that back-arc spreading in the Caucasus Basin depleted the mantle wedge beneath the arc in the Middle Jurassic. Collision-induced trench retreat and back-arc basin formation readily explain these geochronologic and geochemical data and serve as a useful model for understanding the evolution of relict subduction systems and basin closure throughout the western Tethyan realm.

2. Geochemistry of Modern Continental Back-Arc Basins

We begin by summarizing the tectonic setting of five modern continental back-arc basins that serve as exemplars. Next we present the methods and results of our data compilation. We conclude with a discussion of the similarities and differences in geochemical characteristics of modern continental versus intraoceanic back-arc magmatism.

2.1. Tectonic Settings

The ~150-km-wide Okinawa Trough is located between the Ryukyu volcanic arc to the southeast and the East China Sea to the northwest and extends ~900 km along strike between the islands of Taiwan to the southwest and Kyushu to the northeast (Figure 1b; C.-S. Lee et al., 1980). The trough is interpreted to have opened as a back-arc basin during northwest-directed subduction of the Philippine Sea plate beneath continental Eurasia, with an initial phase of extension in the Late Miocene and a second at ~2 Ma (e.g., Kimura, 1985; Sibuet et al., 1987). A number of studies have interpreted the Okinawa Trough as recording an along-strike shift in the manner of back-arc extension, with more northerly portions representing juvenile stages of continental rifting and more southerly portions representing more advanced rifting and possibly the initiation of seafloor spreading (e.g., Liu et al., 2016; Shinjo et al., 1999; Yan & Shi, 2014).

The ~100-km wide Bransfield Strait separates the South Shetland Islands and subducting Phoenix plate to the north from the Antarctic continent and plate to the south for ~500 km along strike (Figure 1c; Weaver et al., 1979). South-dipping subduction of the Phoenix plate beneath the Antarctic plate produced the South Shetland arc, with subduction ceasing at ~4 Ma and extension within the Bransfield Strait beginning at about the same time (Barker, 1982; Lawver et al., 1995). The exact mechanism of extension and seafloor spreading within the Bransfield Strait remains enigmatic, especially given the largely coeval shut-off of the South Shetland arc during the onset of extension. As a result, the Bransfield Strait is sometimes interpreted as an anomalous back-arc basin influenced by strike-slip motion between the Scotia and Antarctic plates (González-Casado et al., 2000) and/or cessation of spreading at the ridge between the Phoenix and Antarctic plates (González-Casado et al., 2000; Lodolo & Pérez, 2015).

Opening of the Tyrrenian Sea initiated within continental crust behind the Calabrian and Aeolian arcs during northwest-directed subduction of the African plate beneath the Eurasian plate, forming a basin that is ~450 km wide across strike but only extends for ~250 km along-strike (Figure 1d; e.g., Barberi et al., 1973; Faccenna et al., 1996; Rehault et al., 1987). Seafloor spreading was primarily confined to the northwestern Vavilov Basin from ~4 to ~2.5 Ma and has migrated eastward to the Marsili Seamount since ~2 Ma (e.g., Kastens & Mascle, 1990; Kastens et al., 1988). Additional volcanic centers on the periphery of the basin
associated with back-arc activity include the island of Ustica and the Enarete, Prometeo, and Palinuro seamounts (e.g., Trua et al., 2004, 2007).

South of ~39°S, Pliocene-Quaternary lavas in the Patagonia plateau are interpreted to have erupted in an extensional back-arc east of the Andean continental arc, which formed by eastward subduction of the Nazca and Antarctic plates beneath the South American plate (Figure 1e; e.g., Goring & Kay, 2001; C. R. Stern et al., 1990). The subaerial Patagonia back-arc is an incipient continental back-arc basin and thus provides a useful contrast to the more well-developed marine basins described above. Trace-element and isotopic geochemical signatures indicative of an enriched mantle source have typically been interpreted to result from asthenospheric upwelling in a slab window resulting from subduction of the Chile Ridge (e.g., D’Orazio et al., 2000; Goring & Kay, 2001).

Eocene-Quaternary volcanic rocks are broadly distributed in the Aegean Sea and Western Anatolia north of the active South Aegean volcanic arc, which is the result north-directed subduction of the African plate beneath the Eurasian plate (Figure 1f; e.g., Innocenti et al., 1981; Pe-Piper & Piper, 2005; Piper & Perissoratis, 2003). Extension in the Aegean-West Anatolia back-arc is long-lived but has not led to the formation of oceanic lithosphere (e.g., Agostini et al., 2010; Jolivet & Faccenna, 2000; Pe-Piper & Piper, 1989), making it a continental back-arc basin in an intermediate stage between the incipient, subaerial Patagonia plateau and the more well-developed exemplar basins noted above. Proposed causal mechanisms for extension and magmatism in the Aegean-West Anatolia system include rollback of the subducting slab (e.g., Brun et al., 2016; Le Pichon & Angelier, 1979; Philippon et al., 2014), westward extrusion of Anatolia along the North Anatolian fault following Eocene collision of Arabia (e.g., Armijó et al., 1999; Flerit et al., 2004), post-orogenic collapse in Anatolia (e.g., Gautier et al., 1999; Jolivet, 2001), and differential advancement of the Eurasian overriding plate (e.g., Agostini et al., 2010; Doglioni et al., 2002).

2.2. Methods

This study compiles and reports previously published major element, trace element, and Sr-Nd isotopic data from modern volcanic whole-rock and glass samples from the five exemplar continental back-arc basins. For general major-element classification, we use the total alkali-silica (TAS) diagram of Le Bas et al. (1986) for rock type and describe rocks as either subalkaline or alkaline according to Irvine & Baragar (1971). As an organizing framework for understanding the trace-element composition of continental back-arc magmatism, we use a Th-3Tb-2Ta ternary diagram (Cabanas & Thieblemont, 1988), which has recently been shown to be a useful tool for discriminating between basalts from a variety of tectonic settings (Xia & Li, 2019).

To minimize the effects of fractional crystallization and allow assessment using the Th-3Tb-2Ta diagram, we restrict our analysis to samples with SiO2 < 57 wt. %, for which analyses of Th, Tb, and Ta are reported. Rather than restricting our analysis to samples of only mafic composition (SiO2 < 52 wt. %), we also include some samples of intermediate composition (SiO2 52–57 wt. %), because the early stages of differentiation likely have not systematically changed the trace-element ratios for highly incompatible elements in these samples relative to their mafic counterparts, and such samples provide an important component of relatively sparse data sets. Th, Tb, and Ta are likely to behave similarly during early differentiation, rendering the position of data on a Th-3Tb-2Ta diagram relatively insensitive to the distinction between the mafic and intermediate samples.

The Th-3Tb-2Ta diagram serves as a useful shorthand for common trace element signatures in magmatic rocks, given that Th functions as a highly incompatible large-ion lithophile element, Tb functions as a slightly less incompatible heavy rare earth element (HREE), and Ta serves as a highly incompatible, strongly fluid-immobile high field strength element (e.g., McDonough & Sun, 1995; Saunders et al., 1980; Sun & McDonough, 1989). In general, rocks with depletion in highly incompatible elements will have high Tb relative to the more incompatible Th and Ta, whereas rocks influenced by subduction zone metasomatic processes will have low Ta relative to Th, and rocks enriched in highly incompatible elements will have high Ta relative to Th and Tb.

To illustrate the utility of this diagram, Figure 2 shows as fields and averages the Th-Tb-Ta compositions of depleted and enriched mid-ocean ridge basalts (D-MORB and E-MORB) and intraoceanic back-arc basin basalts from Gale et al. (2013), ocean island basalts (OIB) from Willbold & Stracke (2010), average arc
basalts from Kelemen et al. (2007), and basalts sourced from subcontinental mantle lithosphere (SCLM) in East-Central China from Wang et al. (2011) and Zhang et al. (2009). Details on the construction of these fields and averages are provided in the Supporting Information S1. In general, D-MORB plots near the 3Tb apex and arc basalts plot near the Th-3Tb join (Figure 2). OIB and SCLM-derived basalts have significant overlap in a broad field closer to the 2Ta vertex than D-MORB or arc basalt, and E-MORB defines an array from D-MORB compositions to the OIB/SCLM-derived compositions. Intraoceanic back-arc basin basalts overlap to some degree with all the other compositions, in keeping with the general understanding that rocks within these basins may have MORB-like, arc-like, or enriched-mantle geochemical characteristics.
(e.g., Pearce & Stern, 2006; Saunders & Tarney, 1984; Xia & Li, 2019). However, a kernel density estimate (KDE) indicates a dominance of the D-MORB-like signature indicative of seafloor spreading in these basins (Figure 2).

Average compositions for each of these fields plotted on primitive mantle normalized multi-element diagrams indicate the general minor/trace-element patterns corresponding to the position on the Th-3Tb-2Ta diagram (Figure 2). We include a rare earth element (REE) diagram and an additional diagram of incompatible minor/trace elements after Pearce (2014) that eschews fluid-mobile elements susceptible to secondary alteration (e.g., Cs, K, Rb, Sr), with both diagrams normalized to primitive mantle (Sun & McDonough, 1989). D-MORB exhibits depletion in both light rare earth elements (LREE) and other highly incompatible elements. Arc basalts are characterized by enrichment in LREE and other highly incompatible elements, with the addition of a negative anomaly in the fluid-immobile high field strength elements Nb and Ta due to the influence of slab fluids, as well as high Th associated with subducted sediment (e.g., Pearce & Cann, 1973; Pearce & Norry, 1979). E-MORB, OIB, and SCLM-derived basalt all exhibit pronounced enrichment in LREE and other highly incompatible elements without the negative Nb-Ta anomaly. KDEs of intraoceanic back-arc basin basalts yield a dominant signature with D-MORB and arc rock characteristics, including LREE depletion with a negative Nb-Ta anomaly. For modern continental back-arc basins, we use the Th-3Tb-2Ta diagram and the primitive mantle normalized multi-element diagrams to assess the influence of depleted-mantle (e.g., D-MORB), subduction-related (e.g., arc basalt), and enriched-mantle (e.g., E-MORB, OIB, and SCLM-derived basalts) sources and compare continental back-arcs with their intraoceanic counterparts.

As an additional means of comparison and for assessing crustal contamination, we also include Sr-Nd isotopic analyses from these continental back-arcs. We generally interpret lower initial values in the \( \frac{\text{Sr}}{86} / \frac{\text{Nd}}{144} \) isotopic ratio and higher initial values in the \( \frac{\text{Sr}}{87} / \frac{\text{Sr}}{86} \) isotopic ratio to reflect melt sources that are more depleted, whereas higher initial \( \frac{\text{Sr}}{87} / \frac{\text{Sr}}{86} \) and lower \( \frac{\text{Nd}}{144} / \frac{\text{Nd}}{143} \) reflect more enriched melt sources (e.g., DePaolo, 1979; DePaolo & Wasserburg, 1979). However, we note that Sr-Nd analyses do a poor job of distinguishing between D-MORB, E-MORB, OIB, SCLM-derived basalt, and intraoceanic back-arc basins (Figure 2). We do not include Pb or Hf isotopic analyses because such data are lacking from the Caucasus Basin. Given the young age of the modern back-arc samples, the magnitude of change between initial and present day \( \frac{\text{Sr}}{87} / \frac{\text{Sr}}{86} \) and \( \frac{\text{Nd}}{144} / \frac{\text{Nd}}{143} \) is smaller than analytical uncertainty. Thus, we take present-day \( \frac{\text{Sr}}{87} / \frac{\text{Sr}}{86} \) and \( \frac{\text{Nd}}{144} / \frac{\text{Nd}}{143} \) as sufficiently representative of initial ratios at the time of formation. We calculate eNd for all modern samples using the chondritic uniform reservoir (CHUR) \( \frac{\text{Nd}}{144} / \frac{\text{Nd}}{143} \) ratio of 0.512630 (Bouvier et al., 2008).

### 2.3. Geochemistry

The 41 samples from the Okinawa Trough are all subalkaline and consist of basalt and basaltic andesite (Figure 3; Guo, Zhai, Yu, Wang, et al., 2018; Guo, Zhai, Yu, Zeng, et al., 2018; Li et al., 2018; Shinjo & Kato, 2000; Shinjo et al., 1999; Shu et al., 2019; Zeng et al., 2010). These samples plot dominantly near the Th-3Tb join on the Th-3Tb-2Ta diagram, exhibit slight to moderate enrichment in LREE and other highly incompatible elements, and show negative Nb-Ta anomalies, generally rendering them geochemically indistinguishable from arc basalts. Samples from the southern Okinawa Trough tend to have more pronounced negative Nb-Ta anomalies and higher Th compared to rocks from the central Okinawa Trough, causing these samples to plot closer to the Th vertex and the Th-3Tb join. The single basaltic andesite in the data compilation from the northern Okinawa Trough also exhibits LREE enrichment and a negative Nb-Ta anomaly (Shu et al., 2019). Two basalt samples from the East China Sea shelf adjacent to the northern Okinawa Trough exhibit strong enrichment in LREE and other incompatible elements with no negative Nb-Ta anomalies (Figures 1b, 3f, and 3g; Zeng et al., 2010).

The 58 samples from Bransfield Strait are dominantly subalkaline basalt, basaltic andesite, and basaltic trachyandesite, with a smaller number of alkaline basalt and tephrite samples restricted to the eastern portion of the basin (Figure 4; Fretzdorff et al., 2004; Keller et al., 2002; Martí et al., 2013). A subset of samples are depleted in LREE and other highly incompatible elements and lack a negative Nb-Ta anomaly, causing them to plot near the 3Tb apex on the Th-3Tb-2Ta diagram within the region occupied by D-MORB. The
The majority of samples have slight to moderate LREE enrichment relative to HREE and pronounced negative Nb-Ta anomalies, resulting in placement near the Th-3Tb join, similar to arc basalt. The alkaline samples from the eastern Bransfield Strait have strong enrichment in LREE and other highly incompatible elements and plot closer to the 2Ta vertex in the center of the Th-3Tb-2Ta diagram, overlapping with E-MORB, OIB, and SCLM-derived basalt.

The 44 **Tyrrenian Sea** samples comprise basalt, picrobasalt, trachybasalt, basaltic andesite, and basaltic-trachyandesite, with both alkaline and subalkaline groups well-represented (Figure 5a; Barberi et al., 1978; Gasperini et al., 2002; Iezzi et al., 2014; Tamburrino et al., 2015; Trua et al., 2003, 2007, 2010). These samples dominantly plot near the Th-3Tb join on the Th-3Tb-2Ta diagram, exhibit enrichment in LREE and other incompatible elements, and have negative Nb-Ta anomalies, similar to arc basalt (Figures 5c–5f). Samples from the area of seafloor spreading in the western Vavilov Basin are generally more
mafic, plot closer to the 3Tb apex, and have less pronounced LREE enrichment and Nb-Ta anomalies when compared to rocks from the Marsili Seamount and other volcanic centers in the eastern part of the basin.

In comparison to the above three sites, the 155 samples from the Patagonia plateau are generally more alkaline, with significant numbers of tephrite, trachybasalt, basaltic trachyandesite, and alkaline basalt samples, in addition to subalkaline basalt and basaltic andesite (Figure 5b; Bruni et al., 2008; Choo et al., 2012; D'Orazio et al., 2000, 2004; Gorrning & Kay, 2001; Gorrning et al., 1997; Kay et al., 2007, 2013; Massaferro et al., 2006, 2014; Ramos & Kay, 1992). These samples plot near the center of the Th-3Tb-2Ta diagram and closer to the 2Ta vertex than D-MORB or arc basalt, resulting in significant overlap with E-MORB, OIB, and SCLM-basalt (Figures 5g–5i). They dominantly exhibit strong enrichment in LREE and other highly incompatible elements and typically lack strong negative Nb-Ta anomalies.

The 128 samples from the Aegean Sea and Western Anatolia are also dominated by alkaline compositions and include basalt, basaltic andesite, tephrite/basanite, trachybasalt, basaltic trachyandesite, phono-tephrite, and trachyandesite (Figure 6a; Agostini et al., 2005, 2007; Akasu et al., 2008; Aldanmaz et al., 2000, 2006, 2015; Altunkaynak & Genç, 2008; Ersoy et al., 2010, 2012; Innocenti et al., 2005; Karacik...
These samples define two general populations on the Th-3Tb-2Ta diagram (Figure 6b). One population plots near the Th vertex and Th-3Tb join and exhibits LREE enrichment and negative Nb-Ta anomalies similar to arc basalt (Figures 6d and 6e). A second population plots near the center of the diagram closer to the 2Ta vertex and exhibit LREE enrichment without negative Nb-Ta anomalies, similar to E-MORB, OIB, and SCLM-basalt (Figures 6f and 6g).

Samples from the Okinawa Trough have εNd values between about +0.5 and +5.5, with, with 87Sr/86Sr ratios in the range of 0.7034–0.7048 (Figure 3c). εNd for the Bransfield Strait samples ranges from about +4 to +9 and 87Sr/86Sr ranges from 0.7025 to 0.7043 (Figure 4c). Limited data available for Tyrrhenian Sea samples with reported Th, Tb, and Ta, as well as SiO₂ < 57 wt. %, yield εNd from about +5 to +7.5 and 87Sr/86Sr ratios from 0.7030 to 0.7035 (Figure 5c). More abundant data from the Patagonia plateau yield εNd values from about −1.5 to +6.5 and 87Sr/86Sr ratios from 0.7031 to 0.7047 (Figure 5c), and Aegean-West Anatolia samples contain εNd values of about −6.5 to +6.8 and 87Sr/86Sr ratios from 0.7030 to 0.7085 (Figure 6c). In all modern continental back-arc basins, the Sr and Nd isotopic systems exhibit a clear negative correlation and overlap with values for D-MORB, E-MORB, OIB, SCLM-derived basalt, arc basalt, and intraoceanic back-arc basin...
basalt, with the exception of a population of Aegean-West Anatolia samples with high $^{87}\text{Sr}/^{86}\text{Sr}$ (>0.706) and low εNd (<−2).

2.4. Discussion

Samples from the five exemplar basins indicate a complex range of geochemical signals present in volcanic rocks from modern back-arc basins initiated in continental lithosphere. In general, the three well-developed marine continental back-arc basins (Okinawa Trough, Bransfield Strait, Tyrrhenian Sea) are dominated by geochemical signatures comparable to those observed in arc basalt, and this signature is also prominent in the partially marine Aegean-West Anatolia system. Samples with this signature feature slight to moderate enrichment in LREE and other highly incompatible elements, as well as distinct negative Nb-Ta anomalies (Figures 7e and 7f). As a result, on the Th-3Tb-2Ta diagram, they tend to plot near the Th-3Tb join (Figure 7d). Samples plotting closer to the Th vertex tend to have either greater overall incompatible element enrichment or higher Th accompanying more pronounced negative Nb-Ta anomalies, likely indicating greater influence of the subducted slab. In each of the well-developed marine basins, position along the Th-3Tb join changes in part as a function of position in the basin, with samples in the southern Okinawa Trough,
eastern Bransfield Strait, and eastern Tyrrhenian Sea (Marsili Seamount) typically exhibiting greater Th than samples from other parts of their respective basins (Figures 1, 3, and 5). This may indicate a greater influence of fluids and sediment from the subducted slab on the mantle source for melting in these subbasins.

A small subset of samples from these three basins show evidence for a depleted mantle source and therefore share some geochemical affinity with D-MORB. Only in the Bransfield Strait are some of these samples clearly distinguishable from arc basalt, with depletion in LREE and other highly incompatible elements and no negative Nb-Ta anomalies (Figures 4d and 4e). As a result, these samples plot near the 3Tb apex of the Th-3Tb-2Ta diagram, as does D-MORB and typical intraoceanic back-arc basin basalt (Figures 4b and 7a). This provides clear geochemical evidence for nascent seafloor spreading in the Bransfield Strait, which is supported by magnetic and gravity data (e.g., Catalán et al., 2013). However, despite a general consensus that seafloor spreading produced the basaltic crust in both the Vavilov Basin and Marsili Seamount in the
Tyrrenian Sea (e.g., Guillaume et al., 2010; Kastens et al., 1988; Savelli, 2015), no unequivocal MORB-type geochemical signal is present there, though samples from the older Vavilov Basin exhibit more evidence for influence of a depleted mantle source with more modest incompatible element enrichment and negative Nb-Ta anomalies (Figures 5e and 5f). In the Okinawa Trough, the presence of seafloor spreading remains contested (e.g., Arai et al., 2017; Liu et al., 2016; Yan & Shi, 2014), and clear evidence of a depleted mantle source without the influence of the subducted slab is also lacking in our compiled data, with most samples exhibiting some incompatible element enrichment and negative Nb-Ta anomalies (Figures 3d and 3e).

Two samples from the East China Shelf adjacent to the northern Okinawa Trough, a subset of samples from the eastern Bransfield Strait, nearly all samples from the Patagonia plateau, and many samples from the Aegean Sea/Western Anatolia show evidence for an enriched mantle source largely unaffected by a subducting slab. Such samples feature pronounced enrichment in LREE and other highly incompatible elements and lack negative Nb-Ta anomalies (Figures 3f, 3g, 4h, 4i, 5h, 5i, 6f, and 6g), leading them to plot near the center of the Th-3Tb-2Ta diagram (Figures 3b, 4b, 5g, and 6b). Such samples are geochemically similar to overlapping E-MORB, OIB, and SCLM-derived basalt, with regional interpretations for their mantle source depending on other lines of evidence. In the East China Shelf, such patterns have been attributed to melting of metasomatized SCLM during the earliest stages of back-arc extension on the continental side of the Okinawa Trough (e.g., Yan & Shi, 2014; Zeng et al., 2010). By contrast, in the eastern Bransfield Strait, Patagonia plateau, and Aegean-West Anatolia system, these patterns are attributed to melting of upwelling asthenosphere through slab windows or tears, which would lead to the absence of the subduction zone signature (e.g., Agostini et al., 2010; D’Orazio et al., 2000; Fretzdorff et al., 2004).

The influence of an enriched mantle source is also apparent in the Tyrrenian Sea, with several samples exhibiting only modest negative Nb-Ta anomalies and stronger incompatible element enrichment (Figures 5e and 5f). Authors have variably attributed this signature in the Tyrrenian Sea and in volcanic rocks in Italy to a deep-seated mantle plume (e.g., Bell et al., 2004), upwelling through a slab window (e.g., Gasperini et al., 2002), and/or metasomatism by the subducting slab (e.g., Peccerillo, 1999). The exact petrogenesis of rocks with enriched mantle signatures in back-arc settings and in other tectonic settings (e.g., oceanic islands, continental rifts, mid-ocean ridges) is complex and continues to be debated. For the purposes of this study, we remain agnostic on the exact source of such signatures but note that they are relatively common in modern continental back-arc settings, especially from areas in the earliest stages of rifting (e.g., northern Okinawa Trough, Patagonia, Aegean-West Anatolia), and should likewise be expected in relict continental back-arc settings.

The Sr-Nd isotopic analyses available for the compiled data generally indicate minimal crustal contamination affecting the minor/trace-element patterns discussed above but do not allow for clear differentiation between mantle sources. The majority of samples have εNd >0 and 87Sr/86Sr <0.7045, indicating depletion of the melt source relative to CHUR. The only exceptions come from the Aegean-West Anatolia system, Okinawa Trough and Patagonia, where back-arc extension is in earlier stages of development and more enriched mantle sources or crustal contamination might be expected (Figures 3c, 5c, and 6c). Likewise, the most depleted Sr-Nd signatures come from the Bransfield Strait, the only basin exhibiting clear MORB-like trace-element signatures indicative of a depleted mantle source (Figure 4c).

Taken together, data from modern continental back-arc basins illustrate that early-stage rifting is typically dominated by enriched mantle signatures, that well-developed marine continental back-arcs are dominated by geochemical signatures similar to modern arc magmatism, and that a depleted mantle signature similar to D-MORB is likely to be minor if present at all. This contrasts with intraoceanic back-arc basins, which can feature a similar range of depleted, subduction-influenced, and enriched mantle signatures but are dominated by a depleted mantle signature similar to D-MORB, as seen in KDEs indicating the greatest density of samples near the 3Tb apex on the Th-3Tb-2Ta diagram with incompatible element depletion and modest negative Nb-Ta anomalies (Figures 7a–7c; Gale et al., 2013). In contrast, similar KDEs constructed from the three well-developed marine continental back-arcs (Okinawa Trough, Bransfield Strait, Tyrrenhian Sea) indicate a predominance of geochemical signatures similar to arc magmatism that plot near the Th-3Tb join, exhibit incompatible element enrichment, and have clear negative Nb-Ta anomalies (Figures 7d–7f). Continental back-arc basins also show a greater variability in Sr-Nd isotopic values relative
to intraoceanic back-arc systems, which are dominantly restricted to higher εNd values and lower \(^{87}Sr/^{86}Sr\) ratios indicative of a more depleted mantle source (Figures 7g and 7h; e.g., DePaolo & Wasserburg, 1979).

Thus, although back-arc basins generally may be characterized by a complex range of melt sources including depleted, subduction-influenced, and enriched mantle, those initiated in continental lithosphere should be expected to be more dominated by the subduction-influenced signature similar to modern arc magmatism than those in intraoceanic back-arc basins (e.g., Saunders & Tarney, 1984). Thus, even when seafloor spreading does occur it takes place closer to the associated arc and is more influenced by the subducting slab than in wider intraoceanic basins where the spreading center is more distant from the slab.

3. Geochronology and Geochemistry of the Caucasus Arc-Back-Arc System

In this section, we interpret geochemical and geochronologic data from Triassic-Jurassic magmatic rocks in the Caucasus in the context of our preliminary framework for understanding the petrogenesis of magmatic rocks with depleted-mantle, subduction-influenced, and enriched-mantle signatures in modern continental back-arc basins. We start with a review of the tectonic setting of the Caucasus Basin. Next we present the methods and results of our compilation of data from the Caucasus region. We conclude by proposing a geodynamic model for the formation and evolution of the Caucasus arc-back-arc system.

3.1. Tectonic Setting

3.1.1. Greater Caucasus and Scythian Platform

The Greater Caucasus Mountains host the highest topography in Europe and are the main locus of active convergence at their longitude within the Arabia-Eurasia continental collision zone (Figure 1g; Jackson, 1992; Reilinger et al., 2006; Sokhadze et al., 2018). West of 45°E, the axis of the range is defined by a Paleozoic crystalline core, which is separated from Mesozoic-Cenozoic strata of the Caucasus Basin to the south by the north-dipping Main Caucasus Thrust (Dotduyev, 1986; Mosar et al., 2010; Saintot et al., 2006; Vasey et al., 2020). The Bechasyn Zone on the north flank of the crystalline core consists of dominantly Early Paleozoic crystalline rocks intruded by Triassic-Jurassic magmatic rocks and overlain by Early Jurassic to Quaternary strata (Adamia, Zakariadze et al., 2011; Gerasimov et al., 2015; Hess et al., 1995; Somin, 2011). The strata overlying the Bechasyn Zone form part of the Scythian Platform, a region between the Greater Caucasus and the East European Craton to the north that is characterized by extensive sedimentary cover and a lack of basement exposure (Alexandre et al., 2004; Natal’ in & Şengör, 2005; Nikishin et al., 2011). A buried Triassic magmatic arc beneath the sedimentary cover of the Scythian Platform has been inferred from positive magnetic anomalies and Triassic magmatic rocks sampled from drill cores (Alexandre et al., 2004; Natal’ in & Şengör, 2005; Okay & Nikishin, 2015; Tikhomirov et al., 2004).

On the south flank of the western Greater Caucasus, Early-Middle Jurassic volcanic and volcanoclastic rocks are widespread within dominantly siliciclastic Caucasus Basin strata, whereas Cretaceous and Cenozoic strata are primarily siliciclastic and carbonate sedimentary rocks (Adamia, Alania, et al., 2011; Gudjabidze, 2003; McCann et al., 2010; Saintot et al., 2006; Vincent et al., 2016). In the central and eastern Greater Caucasus, Jurassic igneous rocks are more sparsely distributed (Lordkipanidze et al., 1989), though there are Cretaceous volcanoclastic rocks within the Vendam zone of the eastern Greater Caucasus (Khain, 2007; Kopp & Shcherba, 1985; Safarov, 2006). Jurassic volcanic rocks exposed on the southern flank have been interpreted by some authors as arc rocks (Hässig et al., 2020; Hess et al., 1995; Mengel et al., 1987), but most studies view them as resulting from extension within the Caucasus Basin (e.g., Cowgill et al., 2016; Lordkipanidze et al., 1989; McCann et al., 2010; Saintot et al., 2006; Vincent et al., 2016). Although Jurassic magmatic rocks with geochemical signatures indicating a depleted mantle source similar to that of D-MORB have been reported (Lordkipanidze et al., 1989; McCann et al., 2010), some authors have asserted that these rocks represent extreme thinning of continental lithosphere rather than true back-arc spreading.
(McCann et al., 2010; Vincent et al., 2016, 2018), though this view is contested (Bindeman et al., 2021; Coghill et al., 2016, 2018; Mumladze et al., 2015).

### 3.1.2. Lesser Caucasus

South of the Greater Caucasus, the Lesser Caucasus Mountains consist of the Paleozoic Dzirula-Khrami-Loki crystalline massifs, the Jurassic-Cretaceous Lesser Caucasus volcanic arc, and the Paleogene Adjara-Trialet belt (Figure 1g). The Lesser Caucasus are bound to the south by the Sevan-Akera suture, which is defined by a belt of Jurassic ophiolites (e.g., Galoyan et al., 2009; Rolland et al., 2009). The suture is generally understood to have formed by Jurassic-Cretaceous north-dipping subduction of an arm of the Neotethyan Ocean that lead to Late Cretaceous or Paleocene collision between the Lesser Caucasus arc and the continental South Armenian Block, onto which the Jurassic ophiolites were obducted (e.g., Hässig, Rolland, Sosson, Galoyan, Sahakyan, et al., 2013; Rolland et al., 2009; Sosson et al., 2010).

The Dzirula-Khrami-Loki crystalline massifs consist primarily of Paleozoic crystalline rocks with Paleozoic tectonic histories comparable to similar rocks of the Greater Caucasus crystalline core (Mayringer et al., 2011; Rolland et al., 2011; Somin, 2011; Vasey et al., 2020). The Adjara-Trialet belt consists of Cretaceous-Eocene volcanic and volcanoclastic rocks deformed in a north-directed fold-thrust belt and is frequently interpreted as the continuation of the eastern Black Sea back-arc basin (e.g., Adamia et al., 1977; Banks et al., 1997; Kazmin et al., 1986).

The Lesser Caucasus volcanic arc consists primarily of Middle Jurassic to Cretaceous magmatic rocks generally considered to have resulted from north-directed subduction of Neotethyan oceanic lithosphere (Figure 1g; e.g., Adamia et al., 1977; Kazmin et al., 1986; Rolland et al., 2011). The arc is typically divided into two sub-domains separated by Late Cenozoic volcanic rocks: the larger Somkhet-Karabagh zone to the northwest and the smaller Kapan zone to the southeast (Kazmin et al., 1986; Mederer et al., 2013). The Lesser Caucasus arc is frequently interpreted as an eastern continuation of the Eastern Pontides volcanic arc (Adamia et al., 1981; van Hinsbergen et al., 2020; Yilmaz et al., 2000).

### 3.1.3. Regional Triassic-Jurassic Paleogeography

Although paleogeographic models of the Caucasus region vary in the location and number of Triassic-Jurassic ocean basins and subduction zones, these reconstructions define two broad classes (Figure 8). The first set of models places both the Greater and Lesser Caucasus on the Eurasian margin in the Triassic. Northward subduction beneath the Lesser Caucasus along the Eurasian margin initiated prior to the Middle Jurassic, with back-arc extension between the Greater and Lesser Caucasus occurring as part of the evolution of the subduction system (Figures 8a–8c; McCann et al., 2010; Okay & Nikishin, 2015; Rolland, 2017; Rolland et al., 2011, 2016; Vasey et al., 2020). These models emphasize the presence of ∼330–300 Ma plutonism and high-temperature/low-pressure metamorphism in both the Paleozoic crystalline core of the Greater Caucasus and the Dzirula-Khrami-Loki massifs, suggesting these rocks were together on the Eurasian margin by the Carboniferous.

In the second set of models, the Greater and Lesser Caucasus were separated by a large ocean basin that did not close until Jurassic time or later (Figures 8d–8f; Adamia, Alania, et al., 2011; Hess et al., 1995; Natal’In & Şengör, 2005; van Hinsbergen et al., 2020). This set of models tends to place greater importance on Devonian-Triassic marine sedimentary rocks reported from the Dizi series, which separates the Greater Caucasus crystalline core from the crystalline massifs in the Lesser Caucasus (Figure 1g; Adamia, Alania, et al., 2011) and on structural correlations between the Lesser Caucasus and the Eastern Pontides (Yilmaz et al., 2000). These models generally show northward subduction beneath the Greater Caucasus in the Triassic-Eocene volcanic and volcaniclastic rocks deformed in a north-directed fold-thrust belt and is frequently interpreted as the continuation of the eastern Black Sea back-arc basin (e.g., Adamia et al., 1977; McCann et al., 2010; Vincent et al., 2016, 2018), though this view is contested (Bindeman et al., 2021; Coghill et al., 2016, 2018; Mumladze et al., 2015).

In one variant, northward subduction beneath the Greater Caucasus led to Early Jurassic collision between the Lesser and Greater Caucasus, initiating northward subduction beneath the Lesser Caucasus as subduction stepped south (Hess et al., 1995). A second variant proposes a Jurassic switch from south-directed to north-directed subduction beneath the Lesser Caucasus, either with (van Hinsbergen et al., 2020) or without (Adamia, Alania, et al., 2011) prior basin closure and collision between the Lesser and Greater Caucasus (Figure 8e).
3.2. Methods

We compile and analyze published geochronologic and geochemical data to explore the geodynamic evolution of the Caucasus arc-back-arc system. Specifically, we assess the duration of major magmatic events using detrital zircon U-Pb analyses and then use U-Pb and \(^{40}\)Ar/\(^{39}\)Ar analyses from Triassic-Jurassic magmatic rocks in the Greater and Lesser Caucasus to delineate the timing and spatial distribution of magmatism in the Caucasus arc-back-arc system.
the region. Revised kernel density estimates and plateau ages reported here are replotted from individual 
U-Pb grain ages and 40Ar/39Ar step ages using IsoplotR (Vermeech, 2018).

To assess the petrogenesis and tectonic setting of Triassic-Jurassic magmatic rocks, we integrate these geochronologic data with geochemical data from the Caucasus region analyzed using the framework for modern continental back-arc basins described above (Figures 2–7). As above, we compile data only from samples with SiO2 < 57 wt. % for which analyses of Th, Tb, and Ta are reported, with the exception of samples from the Bechasyn Zone (Hess et al., 1995) and the Caucasus Basin (Lordkipanidze et al., 1989) that lack Ta but include the similarly behaving Nb and provide critical additional spatial coverage. To ensure consistency when making comparisons across publications and with modern samples, we recalculate initial 87Sr/86Sr and initial 143Nd/144Nd for all Caucasus samples using the reported age of formation from the original publication. We then calculate initial εNd normalized to an initial chondritic uniform reservoir (CHUR) at the time of formation. The initial CHUR 143Nd/144Nd value appropriate for the age of each rock sample was calculated using the present-day CHUR 143Nd/144Nd value of 0.512630 and 147Sm/144Nd value of 0.1960 (Bouvier et al., 2008). Our recalculated values are comparable to the reported initial 87Sr/86Sr, 143Nd/144Nd, and εNd values in the original publications.

3.3. Scythian Platform and Greater Caucasus

3.3.1. Geochronology

Detrital zircon analyses from the southern flank of the Greater Caucasus capture Triassic-Jurassic age populations (Figure 9). Three Jurassic sandstone samples document peaks at ~240 Ma spanning ~260–220 Ma (K3 and GC-41; Allen et al., 2006; Vasey et al., 2020) and at ~170 Ma spanning ~180–160 Ma (GC-41 and NEGC; Allen et al., 2006; Cowgill et al., 2016). These ~240 Ma and ~170 Ma peaks are present in a number of detrital zircon analyses of modern river sediments draining the central and eastern Greater Caucasus, and individual zircon ages of ~260–220 Ma and ~180–160 Ma are present even in samples where they do not form discernible peaks (Cowgill et al., 2016; Tye et al., 2020). A single sample (EGC-5) draining poorly understood Cretaceous volcanoclastic rocks of the Vandam zone yields numerous ages younger than 140 Ma and may represent arc magmatism in the Lesser Caucasus or a later episode of back-arc magmatism in the Cretaceous (Tye et al., 2020).

The few 40Ar/39Ar and U-Pb analyses reported from Mesozoic igneous rocks in the Caucasus yield only rough age constraints that are nevertheless broadly comparable to the detrital zircon peaks (Figure 10). In the Scythian Platform, whole-rock 40Ar/39Ar analyses of volcanic rocks yield complicated age spectra with populations of age steps at ~260–240 Ma, ~210–190 Ma, ~180–170 Ma, and ~140 Ma (Figure 10; Alexandre et al., 2004). 40Ar/39Ar biotite and plagioclase analyses from andesitic-rhyolitic volcanic rocks in the Bechasyn Zone generally yield reported plateau ages of ~190–180 Ma, with a single rhyolite sample characterized by a biotite plateau age of 227.3 ± 2.5 Ma, although individual step ages are not reported (Figure 10; Hess et al., 1995). A muscovite 40Ar/39Ar analysis from mylonitized Carboniferous granodiorite within a shear zone along the MCT in the Kazbegi region of the central Greater Caucasus also yields step ages dominantly between ~230 Ma and ~190 Ma (Figure 10; Vasey et al., 2020). Zircon U-Pb analysis of a single gabbro sample from the Bechasyn Zone in the northern Greater Caucasus contains a youngest population of four zircons with a reported Triassic concordia age of 223.0 ± 4.6 Ma, accompanied by a series of older zircons inferred to be inherited (Gerasimov et al., 2015).

3.3.2. Geochemistry

Sparse analyses of Early Jurassic basaltic andesite in the Bechasyn Zone lack Ta and cannot be plotted on a Th-3Tb-2Ta diagram but are enriched in LREE and other highly incompatible elements and display Nb negative anomalies (Figures 11a, 11f, and 11g; Hess et al., 1995).

In contrast, Early-Middle Jurassic rocks from the south flank of the westernmost Greater Caucasus yield the full range of geochemical signatures observed in modern continental back-arc basins (Figures 7 and 11; McCann et al., 2010). Samples include both subalkaline and alkaline rocks, including basanite, basalt, trachybasalt, basaltic trachyandesite, and one trachyandesite (Figure 11a). As in the modern continental back-arc basins, geochemical signatures similar to modern arc rocks dominate, with most samples plotting near
the Th-3Tb join on the Th-3Tb-2Ta diagram and exhibiting incompatible element enrichment and negative Nb-Ta anomalies (Figures 11b, 11f, and 11g). However, four samples plot near the 3Tb apex on the Th-3Tb-2Ta diagram and exhibit depletion of LREE and other highly incompatible elements similar to D-MORB, although two samples show negative Nb-Ta anomalies and two others lack these anomalies (Figures 11b, 11d, and 11e). Three additional Middle Jurassic samples from the central and eastern Greater Caucasus lack Ta data but also exhibit affinity with D-MORB-like samples, given their flat REE spectra and lack of a Nb negative anomaly (Figures 11d and 11e; Lordkipanidze et al., 1989). Another subset of samples plot near the center of the Th-3Tb-2Ta diagram, show strong incompatible element enrichment, and lack negative Nb-Ta anomalies (Figures 11b, 11h, and 11i).

Combined Sr and Nd isotopic data are only available for samples from the southern flank of the westernmost Greater Caucasus (McCann et al., 2010). This limited data set yields initial εNd values ranging from 0 to +8 and initial 87Sr/86Sr ratios of 0.7038–0.7055 (Figure 11c). Both εNd and initial 87Sr/86Sr values are higher than those for CHUR, and overall the Caucasus Basin samples show a positive correlation of εNd, versus 87Sr/86Sr, rather than the negative correlation typically expected in the Sr-Nd mantle array (e.g., Caro & Bourdon, 2010; Jacobsen & Wasserburg, 1980), which may indicate that Sr has been affected by alteration or metasomatism after crystallization.

### 3.4. Lesser Caucasus

#### 3.4.1. Geochronology

Detrital zircon U-Pb analyses from modern rivers draining the Lesser Caucasus yield few Triassic ages, though a few grains date from ~240 to 200 Ma (Figure 9; Cowgill et al., 2016; Tye et al., 2020). Much more prominent is a Middle Jurassic peak at ~170 Ma, with continuous ages ranging from ~180 to 140 Ma (Cowgill et al., 2016; Tye et al., 2020).

We know of no Triassic igneous crystallization ages reported from the Lesser Caucasus arc. However, an amphibole 40Ar/39Ar analysis of an amphibolite within the Paleozoic Loki crystalline massif yields a plateau age of 212.0 ± 3.6 Ma (Figure 12; Rolland et al., 2011). Similarly, primary data showing Early Jurassic magmatic ages are absent from the arc, but a biotite 40Ar/39Ar analysis of a migmatic gneiss from the Khrami crystalline massif yields a plateau age of 182.0 ± 0.52 Ma (Figure 12; Rolland et al., 2011). Zircon U-Pb ages of ~165 Ma have been reported from tonalite in the Kapan zone and plagiogranite in the Somkheto-Kara-bagh zone (Galoyan et al., 2018; Mederer et al., 2013). 40Ar/39Ar analyses of amphibole and muscovite from blocks of the North Armenian flysch also yield plateau ages of ~166 Ma (Figure 12; Rolland et al., 2011). Late Jurassic U-Pb ages of ~155 Ma have been reported from tonalite and granite of the Somkheto-Kara-bagh zone (Calder et al., 2019; Galoyan et al., 2018; Hässig et al., 2020).

#### 3.4.2. Geochemistry

Jurassic samples from the Lesser Caucasus arc are almost entirely subalkaline basalt and basaltic andesite, with a small number of alkaline basalt, trachybasalt, and basaltic trachyandesite samples (Figure 13a). Early-Middle Jurassic rocks plot dominantly near the Th-3Tb join of the Th-3Tb-2Ta diagram and exhibit nearly flat REE spectra, with some slight LREE enrichment, and show negative Nb-Ta anomalies (Figures 13b, 13d, and 11e; Calder et al., 2019; Galoyan et al., 2018; Mederer et al., 2013, 2014). By contrast, Late Jurassic-Early Cretaceous rocks also exhibit negative Nb-Ta anomalies but generally plot closer to the Th vertex near the Th-3Tb join and display a greater degree of enrichment in LREE and other highly incompatible elements (Figures 13b, 13f, and 13g; Calder et al., 2019; Galoyan et al., 2018; Hässig et al., 2020; Mederer et al., 2013). εNd values from Jurassic magmatic rocks in the Lesser Caucasus dominantly range from +5 to +8.

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**Figure 9.** Detrital zircon ages from samples with Triassic-Jurassic ages from the Greater and Lesser Caucasus indicating times of peak magmatic activity at ~260–220 Ma and ~180–160 Ma (Greater Caucasus—blue bars) and ~180–140 Ma (Lesser Caucasus—green bars), inferred from widths of common age populations in samples of both Jurassic sandstone (yellow, depositional ages indicated in italics) and modern river sediment (pink) (GC-41 from Allen et al., 2006; NEGC, Kamuk, Tovuz from Cowgill et al., 2016; K3 from Vasey et al., 2020; EGC, CGC, and LC samples from Tye et al., 2020). Curves show kernel density estimates calculated using IsoplotR (Vermeesch, 2018) clipped to 260–140 Ma. Circles indicate ages of individual analyses.
to +6.5, and $^{87}$Sr/$^{86}$Sr ratios generally range from 0.7032 to 0.7038, with two Middle Jurassic outliers with higher $^{87}$Sr/$^{86}$Sr and/or lower εNd (Figure 13c; Calder et al., 2019; Mederer et al., 2013).

3.5. Discussion

3.5.1. Triassic Evolution of the Caucasus Arc System

Despite the general paucity of exposed Triassic magmatic rocks in the Caucasus region, available geochronologic data provide strong evidence for a Triassic volcanic arc in the northern Greater Caucasus and Scythian Platform. A single zircon U-Pb analysis from the Bechasyn Zone yields a crystallization age of ∼225 Ma (Gerasimov et al., 2015), and $^{40}$Ar/$^{39}$Ar analyses from magmatic rocks yield Triassic step ages in the Scythian Platform (∼260–200 Ma) and the Bechasyn Zone of the northern Greater Caucasus (∼225 Ma) (Figure 10; Alexandre et al., 2004; Hess et al., 1995). Given the complexity of the $^{40}$Ar/$^{39}$Ar results and the lack of multiple U-Pb ages, the exact duration of this magmatism is not well-defined by these data. However, detrital zircon U-Pb analyses in Early-Middle Jurassic sandstone from the central and eastern Greater Caucasus yield a prominent Triassic peak centered at ∼240 Ma, with the bulk of ages in this peak between ∼260 and 220 Ma (Figure 9; Allen et al., 2006; Vasey et al., 2020), and analyses from an additional Middle Jurassic sandstone and modern rivers in the region yield a minor component of similar ages (Figure 9; Cowgill et al., 2016; Tye et al., 2020). We interpret these data to indicate voluminous Permian-Triassic arc magmatism in the northern Greater Caucasus and Scythian platform from ∼260-220 Ma, with arc rocks well-exposed in the region during Jurassic sandstone deposition but now largely buried beneath the cover of the Scythian platform. The minor component in modern rivers is likely due to recycling of grains from Jurassic sedimentary rocks containing Triassic zircon and/or erosion of minor exposures of Triassic magmatic rocks, such as those documented in the Bechasyn Zone (Gerasimov et al., 2015).

Our interpretation of a Triassic volcanic arc in this region is consistent with previous workers, who have observed both Triassic magmatic rocks in drill holes and a subsurface positive magnetic anomaly in the Scythian Platform (e.g., Alexandre et al., 2004; Natal’ in & Şengör, 2005; Okay & Nikishin, 2015; Tikhomirov et al., 2004). Like these workers, we attribute this arc to north-directed subduction beneath the margin of Laurasia, given widespread consensus that the Greater Caucasus and Scythian Platform formed part of the Laurasian margin by the Carboniferous (Figures 14a and 15a; e.g., Okay & Topuz, 2016; Rolland, 2017; Vasey et al., 2020). Although the exact timing and mechanism of subduction initiation remain poorly understood, one plausible explanation is that northward subduction accommodated continued convergence following Carboniferous accretion of rocks in the crystalline core of the western Caucasus to Eurasia (Rolland et al., 2011, 2016; Vasey et al., 2020).

Notably, clear evidence for Triassic arc magmatism is absent from the southern flank of the Greater Caucasus and the Lesser Caucasus, apart from two metamorphic $^{40}$Ar/$^{39}$Ar ages of ∼230–200 Ma (Figures 10 and 12; Rolland et al., 2011; Vasey et al., 2020) and a handful of detrital zircons with ages of ∼240–200 Ma in rivers draining the Lesser Caucasus (Figure 9; Cowgill et al., 2016; Tye et al., 2020). Thus, we infer that the main magmatic front of the Triassic Caucasus arc was north of the Greater Caucasus crystalline core and thus also north of the subsequent location of the Caucasus Basin and the Lesser Caucasus volcanic arc in the Jurassic (Figures 14 and 15). Triassic metamorphic $^{40}$Ar/$^{39}$Ar ages in these more southern areas may reflect the thermal effects of nearby subduction and arc magmatism at this time.

3.5.2. Jurassic Evolution of the Caucasus Arc System

The Early-Middle Jurassic marks a series of shifts in the Caucasus arc system. In the period ∼190–180 Ma, few $^{40}$Ar/$^{39}$Ar step ages are reported from the Scythian Platform (Figure 10; Alexandre et al., 2004), whereas the bulk of Mesozoic magmatic rocks from the Bechasyn Zone to the south have reported $^{40}$Ar/$^{39}$Ar

Figure 10. Plots comparing peak magmatic intervals inferred from detrital zircon U-Pb analyses shown in Figure 9 (blue bars) with $^{40}$Ar/$^{39}$Ar step and plateau ages (1σ) from Triassic-Jurassic magmatic rocks on the Scythian Platform and in the Greater Caucasus (Alexandre et al., 2004; Hess et al., 1995; Vasey et al., 2020). See Figure 1g for sample locations. For samples AN94-4 and AN94-5, red line with 95% confidence interval shown in gray indicates weighted mean plateau ages calculated using step ages in green. Plot in lower left shows individual plateau ages for 12 samples reported by Hess et al. (1995), for which individual step ages were not reported.
plateau ages from this time period (Figure 10; Hess et al., 1995). These Early Jurassic magmatic rocks are primarily andesitic in composition, and the handful of samples with SiO$_2$ < 57 wt. % exhibit LREE enrichment and negative Nb anomalies (Figure 10; Hess et al., 1995). We tentatively interpret these results to indicate a shut-off of arc magmatism in the Scythian Platform and southward migration of subduction-zone magmatism into the Bechasyn Zone in the Early Jurassic (Figure 15b). However, we acknowledge that the complex nature of the $^{40}$Ar/$^{39}$Ar ages and the lack of published primary $^{40}$Ar/$^{39}$Ar data from the Bechasyn Zone make this pattern somewhat speculative, and magmatism could have been largely coeval in both regions.

Arc magmatism resulting from subduction beneath the Lesser Caucasus had initiated by $\sim$180 Ma, as suggested by U-Pb and $^{40}$Ar/$^{39}$Ar data (Figures 14c and 15c). For example, modern rivers draining the Lesser Caucasus...
Figure 12. Same as Figure 10 but for metamorphic rocks in the Lesser Caucasus with Triassic-Jurassic $^{40}$Ar/$^{39}$Ar plateau ages (Rolland et al., 2011).
Geochemistry, Geophysics, Geosystems

Lesser Caucasus

Caucasus contain zircons as old as \( \sim 180 \) Ma in the Middle Jurassic zircon age peaks centered at \( \sim 170 \) Ma (Figure 9; Cowgill et al., 2016; Tye et al., 2020), and a metamorphic \(^{40}\text{Ar}/^{39}\text{Ar}\) age of \( \sim 182 \) Ma is reported from the Khrami massif in the Lesser Caucasus (Figure 12; Rolland et al., 2011). Trace element patterns in Middle Jurassic rocks from the Lesser Caucasus are consistent with petrogenesis in an island arc setting, with relatively flat REE patterns and negative Nb-Ta anomalies, although the flat REE patterns suggest a relatively depleted mantle source (Figure 13; Galoyan et al., 2018; Mederer et al., 2013). In contrast, by Late Jurassic time the mantle source for the subduction zone magmas had become more enriched, as indicated by Late Jurassic rocks in the Kapan zone with stronger LREE and incompatible element enrichment but similar negative Nb-Ta anomalies (Figure 13; Galoyan et al., 2018; Mederer et al., 2013). Positive \( \varepsilon\text{Nd} \) values and \(^{87}\text{Sr}/^{86}\text{Sr} \) values less than 0.7045 suggest that the Nb-Ta anomalies are dominantly derived from the mantle source, rather than contamination by continental crust (Figure 13; e.g., Ernst et al., 2005). The major episode of Jurassic arc magmatism in the Lesser Caucasus occurred between \( \sim 180 \) and 140 Ma, as suggested by the numerous and continuous zircon ages spanning this age range in the \( \sim 170 \) Ma detrital zircon peaks (Figure 9; Cowgill et al., 2016; Tye et al., 2020).

Figure 13. Geochemical plots as used in Figure 3 for samples from the Jurassic Lesser Caucasus with SiO\(_{2} < 52\) wt. % (circles) and SiO\(_{2} \geq 52\) wt. % (squares), colored according to time (light green—Early-Middle Jurassic; dark green—Late Jurassic-Early Cretaceous) and compared with values from Figure 2. (a) TAS diagram. (b) Th-3Tb-2Ta ternary diagram. (c) Diagram of \( \varepsilon\text{Nd} \) versus \(^{87}\text{Sr}/^{86}\text{Sr} \). (d and e) Primitive mantle-normalized rare earth element and immobile element plots for Early-Middle Jurassic samples. (f and g) Diagrams as in (d and e) for Late Jurassic-Early Cretaceous samples.
Collectively, these data suggest an overall shift from arc magmatism focused in the Scythian Platform and northern Greater Caucasus in the Triassic to arc magmatism focused in the Lesser Caucasus by the Middle-Late Jurassic (Figures 14 and 15). There are two fundamentally different ways of interpreting this spatio-temporal pattern: coeval termination of a Greater Caucasus arc and initiation of the Lesser Caucasus arc (e.g., Adamia, Alania, et al., 2011; Hess et al., 1995) or southward migration of the magmatic front of a single arc system (Figure 8). We prefer the second interpretation, based on the evidence presented below suggesting back-arc spreading in the Caucasus Basin.

Figure 14. Schematic paleogeographic maps of our model for the Caucasus arc system compared to map-view stages of collision-induced back-arc spreading in numerical models of Magni (2019). (a) Continental arc magmatism prior to collision. (b) Entry of a buoyant indenter (continental block) into the subduction zone. (c) Trench retreat, fore-arc rotation, and back-arc spreading along-strike of the collision zone.
Deposition and subsidence in the Caucasus Basin are generally accepted to have started in the Early Jurassic and then accelerated in the Middle Jurassic (Figures 15b and 15c; Ershov et al., 2003; Nikishin et al., 2011; Saintot et al., 2006; Vincent et al., 2016). Likewise, minor magmatism is reported in the Early Jurassic, with greater volumes of magmatism in the Middle Jurassic (e.g., Adamia, Alania, et al., 2011; Lordkipanidze et al., 1989; McCann et al., 2010; Saintot et al., 2006), although primary geochronologic data providing precise, verifiable ages for these magmatic rocks are lacking.

### 3.5.3. Back-Arc Spreading in the Caucasus Basin

Deposition and subsidence in the Caucasus Basin are generally accepted to have started in the Early Jurassic and then accelerated in the Middle Jurassic (Figures 15b and 15c; Ershov et al., 2003; Nikishin et al., 2011; Saintot et al., 2006; Vincent et al., 2016). Likewise, minor magmatism is reported in the Early Jurassic, with greater volumes of magmatism in the Middle Jurassic (e.g., Adamia, Alania, et al., 2011; Lordkipanidze et al., 1989; McCann et al., 2010; Saintot et al., 2006), although primary geochronologic data providing precise, verifiable ages for these magmatic rocks are lacking.
The limited minor/trace element data available for these rocks exhibit significant variability (Figure 11; Lordkipanidze et al., 1989; McCann et al., 2010). Rocks plotting near the 3Tb apex on the Th-3Tb-2Ta diagram exhibit flat or depleted LREE relative to HREE and only in some cases have a negative Nb-Ta anomaly, rocks plotting near the Th-3Tb join exhibit enrichment of highly incompatible elements and clear Nb-Ta anomalies, and rocks plotting closer to the 2Ta vertex exhibit strong incompatible element enrichment and no Nb-Tb anomalies (Figure 11). We interpret the widespread negative Nb-Ta anomalies in the majority of Caucasus Basin samples to indicate the formation of these rocks broadly in a subduction zone setting, given the fluid immobility of these elements and the critical role that fluids play in subduction zone magmatism (e.g., Pearce & Cann, 1973; Pearce & Norry, 1979).

Given that continental crustal rocks typically exhibit a negative Nb-Ta anomaly in primitive mantle-normalized diagrams (e.g., Barth et al., 2000; Rudnick & Fountain, 1995), crustal contamination could impart a geochemical signature similar to that of subduction-related magmas to rocks that were not formed in a subduction zone setting (e.g., Ernst et al., 2005; Xia & Li, 2019). However, Sr-Nd isotopic data indicates that these rocks are dominated by mantle signatures or at least crustal rocks very recently derived from the mantle. εNd values for Caucasus Basin rocks range from ~0 to +8 (Figure 11c), indicating a depleted source relative to CHUR, consistent with formation in a subduction zone setting (e.g., DePaolo & Johnson, 1979; White & Patchett, 1984; Xia & Li, 2019), rather than the enriched source and negative εNd values typically observed in older continental crust (e.g., Allègre & Othman, 1980). Although εNd values are notably more elevated than expected for a typical mantle array for these εNd values (Figure 11c; e.g., DePaolo, 1979; DePaolo & Wasserburg, 1979), the Sr isotopic system is susceptible to hydrothermal alteration from seawater or later metasomatism (e.g., Hart et al., 1974; Kawahata et al., 2001), which could have produced these elevated values as a secondary process. Elevated εNd values due to crustal contamination should also result in lower εNd values; thus, we emphasize the Nd data in assessing the influence of this process.

A subset of Caucasus Basin rocks exhibit LREE depletion relative to HREE comparable to that observed in D-MORB (Figure 11d; Lordkipanidze et al., 1989; McCann et al., 2010), which is generated by decompression melting of an already-depleted mantle source (Figure 2; e.g., Gale et al., 2013; Sun & McDonough, 1989; Sun et al., 1979). The combination of D-MORB and subduction zone geochemical characteristics is commonly interpreted as a signature of back-arc basin spreading, given that melt in such a setting is produced by a similar mechanism to that at mid-ocean ridges but can also be affected by fluids released from the subducting slab (e.g., Pearce & Stern, 2006; Saunders & Tarney, 1984; Xia & Li, 2019). However, McCann et al. (2010) interpret these Caucasus Basin rocks to have resulted from extreme thinning of the continental lithosphere that did not lead to full back-arc spreading, an interpretation that has been subsequently cited to argue that the magmatic record in the Caucasus Basin provides no evidence for back-arc spreading (Vincent et al., 2016, 2018).

We disagree with the interpretations of McCann et al. (2010) and Vincent et al. (2016, 2018) and argue that geochemical data from the Caucasus Basin, when compared with similar data from modern continental back-arc basins, are entirely consistent with back-arc basin spreading in the Middle Jurassic (Figures 14c and 15c). Importantly, Caucasus Basin samples with D-MORB affinity are comparable to a similar subset of samples in the Bransfield Strait (Figures 4 and 11) but lack analogs in other modern continental back-arc basins. Given that Bransfield Strait has experienced the initial stages of seafloor spreading (e.g., Catalán et al., 2013; Keller et al., 2002; Lodolo & Pérez, 2015; Weaver et al., 1979), Caucasus Basin samples with D-MORB signatures likely also resulted from seafloor spreading.

Despite the general consensus that seafloor spreading has also taken place in the Tyrrenian Sea (e.g., Guillaume et al., 2010; Kastens et al., 1988; Savelli, 2015) and the possibility that seafloor spreading has initiated in the southern Okinawa Trough (Liu et al., 2016; Shinjo et al., 1999; Yan & Shi, 2014), both of these basins lack samples with D-MORB-type signatures (Figures 3 and 5). Thus, modern continental back-arc basins may experience seafloor spreading yet still lack geochemical evidence for a D-MORB-like depleted mantle source, making the presence of D-MORB-type signatures in the Caucasus Basin even more indicative of some degree of seafloor spreading via upwelling of depleted asthenosphere. The Aegean-West Anatolia system and Patagonia plateau also lack evidence for D-MORB-type signatures, in keeping with the lack of seafloor spreading in these regions.
A handful of Caucasus Basin samples yield strong incompatible element enrichment and lack negative Nb-Ta anomalies, similar to samples along the continental margin of the northern Okinawa Trough, the eastern Bransfield Strait, the Patagonia plateau, and the Aegean Sea/Western Anatolia (Figures 3–6 and 11). This signature has been attributed to melting of SCLM in the Okinawa Trough (Yan & Shi, 2014; Zeng et al., 2010) and to upwelling asthenosphere through slab windows and/or tears in the other three basins (e.g., Agostini et al., 2010; D’Orazio et al., 2004; Fretzdorff et al., 2004). Insufficient data are available in the Caucasus Basin to distinguish between these possible sources, and all but one of the samples with this signature also overlap with E-MORB compositions, so seafloor spreading is an additional possibility (Gale et al., 2013). In the absence of additional geologic constraints and considering that this signature appears to be common in continental back-arc basins, we take the conservative view that these rocks indicate some type of enriched mantle source in the Caucasus Basin but remain agnostic as to the exact source.

Considered together, the compiled data suggest that Middle Jurassic rocks in the Caucasus Basin capture a range of geochemical processes observed in modern continental back-arc, including (incipient?) seafloor spreading. Although it is perhaps plausible that D-MORB-like signatures could be produced by sourcing depleted asthenosphere beneath extremely-thinned continental lithosphere (McCann et al., 2010; Vincent et al., 2016, 2018), we see no evidence to favor this interpretation given that the Bransfield Strait also has this signal and shows evidence of seafloor spreading, whereas modern continental back-arc basins without seafloor spreading (Patagonia plateau, Aegean-West Anatolia) fail to produce this signature. Additionally, although most samples from the Caucasus Basin are indistinguishable from those expected in an arc setting (Figure 11; e.g., Jakes & Gill, 1970; Kelemen et al., 2007), the examples of Bransfield Strait, the Tyrrenian Sea, and the Okinawa Trough demonstrate that such signatures can also be produced in a continental back-arc spreading setting where the influence of subduction fluids is dominant (Figures 3–5). Even a subset of rocks found in intraoceanic back-arc spreading centers exhibit characteristics similar to main-arc rocks, though such samples still generally exhibit incompatible element depletion (Figure 7; e.g., Gale et al., 2013; Xia & Li, 2019).

3.5.4. Collision-Induced Mechanism for the Caucasus Arc-Back-Arc System

Overall, geochronologic and geochemical evidence delineate Triassic-Early Jurassic arc magmatism in the Scythian Platform and northern Greater Caucasus, Middle-Late Jurassic arc magmatism in the Lesser Caucasus, and Middle Jurassic back-arc spreading in the southern Greater Caucasus (Caucasus Basin) (Figures 14 and 15). These three magmatic events require either the termination of a northern Greater Caucasus arc and subsequent initiation of the Lesser Caucasus arc (Figure 8; e.g., Adamia, Alania, et al., 2011; Hess et al., 1995; van Hinsbergen et al., 2020) or southward migration of a single magmatic arc front (Figure 8; e.g., Okay & Nikishin, 2015; Rolland et al., 2016). We contend that the geochronologic and geochemical data presented above best fit with the latter model, with southward arc migration and back-arc extension best explained by along-strike collision of the Iranian block (Figures 14 and 15). Additional geochronologic and geochemical analyses of magmatic rocks in the area could test this model.

Along-strike collision has been proposed as one of the dominant mechanisms for producing back-arc extension in the overriding plate (e.g., Magni, 2019; Magni et al., 2014; McCabe, 1984; Wallace et al., 2005, 2009). Introduction of continental lithosphere or another buoyant indenter (e.g., seamount) into part of a subduction zone locally slows or halts subduction (Figure 14b). As the indenter deforms the overriding plate, it induces fore-arc rotation and trench retreat in areas along strike where subduction of oceanic lithosphere can continue unimpeded. Trench retreat is accommodated by extension within the overriding plate, resulting in back-arc rifting and spreading (Figure 14c). Observations of modern and ancient back-arc systems are consistent with this model (McCabe, 1984; Wallace et al., 2005, 2009). Three-dimensional numerical models illustrate the feasibility of this mechanism by introducing one or two blocks of continental lithosphere into a preexisting subduction zone and producing fore-arc rotation, trench retreat, and back-arc spreading along strike (Figure 14; Magni, 2019; Magni et al., 2014; Wallace et al., 2009).

The southward shift in the locus of arc magmatism in the Caucasus region corresponds in time with a major collision along-strike to the east (Figure 14). The Iranian block, generally considered part of the larger Cimmerian continental ribbon, collided with the Eurasian margin in the Triassic, closing a portion of the Paleotethys Ocean (e.g., Muttoni et al., 2009; Şengör, 1979). Maximum depositional ages in sedimentary rocks (Horton et al., 2008) and crystallization ages in post-collisional granitoids (Zanchetta et al., 2013) of
~220–210 Ma that cross-cut and thus post-date the Paleotethys suture in Iran indicate that collision had occurred by this time. This time of ~220–210 Ma corresponds with the end of the major episode of Triassic arc magmatism in the northern Greater Caucasus and Scythian Platform at ~220 Ma, the metamorphic 40Ar/39Ar ages in the Greater and Lesser Caucasus (Rolland et al., 2011; Vasey et al., 2020), and the beginning of the relative magmatic lull in the Caucasus region (Figure 14b). We propose that Late Triassic metamorphism and arc shut-off were due to trench retreat and rollback of the subducting slab following indentation of the Eurasian margin by the Iranian collision (Figures 14b and 15b).

Numerical models provide estimates on the magnitude and duration of trench retreat following collision of a buoyant indenter. In models where oceanic subduction continues between two continental indenters, trench retreat reaches maxima of ~4–6 cm/yr about 15 Myr after collision and may persist for an additional ~15 Myr, depending on the width of the oceanic subduction zone (Magni et al., 2014). This results in a total magnitude of ~1,000 km of trench retreat over an ~30-Myr period. In models where arc and back-arc melting are also considered (Magni, 2019), back-arc spreading initiates ~10 Myr after initial collision and becomes spatially distinct from arc magmatism after ~20 Myr and ~500 km of trench retreat (Figure 14). These timing estimates are broadly consistent with the spatial and temporal evolution of the Triassic-Jurassic Caucasus arc-back-arc system. Specifically, if collision of Iran with Eurasia occurred at ~220–210 Ma, overriding-plate extension may have led to initial back-arc magmatism at ~190 Ma (~20–30 Myr after collision) and spreading that was spatially distinct from the Caucasus arc between ~180 and 160 Ma, as suggested by geochronologic and geochemical data, with hundreds of km of trench retreat moving the locus of arc magmatism from the Scythian Platform and northern Greater Caucasus to the Lesser Caucasus (Figure 15c).

Moreover, the numerical models of Magni (2019) predict an important geochemical shift in the arc itself as back-arc spreading proceeds. Convection in the mantle wedge accompanies trench retreat and back-arc extension, bringing melt sources first to the back-arc and then to the arc. As a result, for an ~15 Myr period following initiation of back-arc spreading, melt at the arc is more depleted than would normally be expected as a result of prior melting in back-arc (Figure 15c), in agreement with results from previous observational (Cooper et al., 2010; Woodhead et al., 1993) and modeling (Hall et al., 2012) studies. Eventually, the distance between the back-arc spreading center and the arc itself becomes great enough that the melt sources separate, allowing the arc melt source to return to its previous fertile state (Figure 15d; Magni, 2019).

Arc melt depletion due to prior back-arc melting provides an explanation for the distinct difference in Lesser Caucasus arc geochemistry in the Middle versus the Late Jurassic (Figure 13; Calder et al., 2019; Galoyan et al., 2018; Hässig et al., 2020; Mederer et al., 2013, 2014). Relative depletion in LREE and other highly incompatible elements in Middle Jurassic rocks compared to Late Jurassic rocks suggest a more depleted mantle source in the Middle Jurassic than in the Late Jurassic. Importantly, within the ~170 Ma detrital zircon age peak seen in samples from both the Greater and Lesser Caucasus, the youngest individual zircon ages are ~140 Ma in the Lesser Caucasus, whereas in the Greater Caucasus few grains are found that are younger than ~160 Ma, near the Middle-Late Jurassic boundary (Figure 9). Thus, the shift to a more fertile melt source in the Lesser Caucasus in the Late Jurassic may have corresponded to a lull in magmatic activity in the Caucasus Basin as the spreading center migrated away from the arc (Figure 15d).

In summary, spatial, temporal, and geochemical patterns of Triassic-Jurassic magmatism within the Caucasus region are well-explained by a model of trench retreat and short-lived sub-arc mantle depletion following along-strike collision of the Iranian block with Eurasia (Figures 14 and 15). Late Triassic collision induced trench retreat to the west, moving the magmatic front of the Caucasus arc from north of the present-day Greater Caucasus to the Lesser Caucasus. As the trench retreated, overriding-plate extension resulted in Middle Jurassic back-arc spreading within the Caucasus Basin and depleted the mantle material sourcing the Lesser Caucasus arc. By the Late Jurassic, a lull in back-arc spreading led to relative enrichment of the mantle source for ongoing magmatism within the Lesser Caucasus arc. In this view, Caucasus Basin spreading occurred over a brief period (~180–160 Ma).

We contend that this model provides a better explanation of the data than one involving multiple subduction systems. Such a model would require either that the Greater and Lesser Caucasus crystalline massifs remained separate until the Jurassic (e.g., Hess et al., 1995; van Hinsbergen et al., 2020), despite their shared
Carboniferous history (e.g., Mayringer et al., 2011; Okay & Topuz, 2016; Somin, 2011; Vasey et al., 2020), or that the Greater and Lesser Caucasus crystalline massifs rifted apart after the Carboniferous and collided again in the Jurassic. Although neither of these possibilities can be completely eliminated, only sparse Permian-Triassic sedimentary rocks have been reported between the Greater and Lesser Caucasus (e.g., Adamia, Alania, et al., 2011), and we are not aware of any evidence of a Triassic-Early Jurassic suture. As a result, we prefer a single-subduction-zone model that provides a clear mechanism of collision-induced overriding-plate extension observed in modern arc-back-arc systems (e.g., Wallace et al., 2009) to explain the Triassic-Jurassic evolution of the Caucasus arc-back-arc system.

4. Regional Implications for the Western Tethyan Realm

Within the western Tethyan realm, the Mesozoic Lesser Caucasus arc is frequently assumed to be part of a larger arc system that includes coeval magmatism in the Pontides. Such correlations are based upon two lines of evidence: similar Paleozoic magmatic/metamorphic histories in crystalline rocks of the Lesser Caucasus and Pontides (e.g., Okay & Topuz, 2016; Rolland et al., 2016) and similar Jurassic-Cretaceous stratigraphic sections (e.g., Yilmaz et al., 2000). Many aspects of the Triassic-Jurassic evolution of the Pontides remain subjects of debate beyond the scope of this study, including whether Triassic deformation is best explained by continental collision or seamount accretion (e.g., Şengör, 2013; Topuz et al., 2013) and whether Triassic-Jurassic subduction beneath the Pontides was north-directed, south-directed, or both (e.g., van Hinsbergen et al., 2020). However, a key difference between the Lesser Caucasus and Pontides is the lack of Late Triassic accretionary complexes and coeval high-pressure/low-temperature metamorphism in the Lesser Caucasus, which are well-documented in the Pontides (e.g., Okay, 2000; Okay et al., 2002, 2006; Topuz et al., 2014). We speculate that this absence could indicate either that the Lesser Caucasus and Pontides do not share a Triassic tectonic history or that a proposed seamount accreted to the Pontides in the Triassic (the Karakaya Complex) could have served as a second buoyant indenter promoting overriding-plate extension in the Caucasus region to the east (Okay, 2000; Okay & Göncüoğlu, 2004; Robertson & Ustaömer, 2012).

The Middle Jurassic marks a major period of ophiolite formation in the western Tethyan realm, with $^{40}$Ar/$^{39}$Ar ages for both crystallization and metamorphic sole formation clustering at ∼175 Ma (e.g., Maffione & van Hinsbergen, 2018; Robertson, 2002). Similar ages are reported in ophiolites of the Lesser Caucasus south of the arc in the Sevan-Akera suture zone (Figure 1f; Galoyan et al., 2009; Hässig, Rolland, Sosson, Galoyan, Müller, et al., 2013; Rolland et al., 2010). The Sevan-Akera ophiolites are typically interpreted to have formed in the back-arc to a north-dipping, intra-oceanic subduction zone within the Neotethyan Ocean to the south of the Lesser Caucasus arc, based on subduction zone geochemical signatures such as negative Nb-Ta anomalies (e.g., Rolland et al., 2010; Sosson et al., 2010). However, the record of this intra-oceanic volcanic arc to the south of these ophiolites has not been identified. We speculatively note that fore-arc extension during trench retreat within the Caucasus arc system could also serve as the mechanism for producing the Sevan-Akera ophiolites (Figure 15c; e.g., Dilek & Flower, 2003; Dilek & Furnes, 2011; Metcalf & Shervais, 2008), without needing to invoke an additional intra-oceanic arc to the south. A similar trench retreat mechanism has been proposed to explain the Xigaze forearc ophiolite in southern Tibet, where subduction had already been ongoing prior to ophiolite formation (e.g., Butler & Beaumont, 2017; Dai et al., 2013).

Debate persists in the western Tethyan realm regarding the presence of the Cimmerian continental ribbon and Paleotethyan suture west of Iran (e.g., Şengör, 2013; Topuz et al., 2013). In our model, there is no Triassic-Jurassic Cimmerian collision/Paleotethyan suture in the Caucasus and thus locally no distinction between Paleotethyan and Neotethyan subduction (Figure 8). Thus, in the Caucasus region, north-directed Tethyan subduction likely initiated in the Permian-Triassic following collision of the Caucasus basement with Laurussia and continued until closure of the Sevan-Akera suture in the Cretaceous-Eocene. The Black Sea and South Caspian basins to the west and east of the Caucasus, respectively, likely began as part of the Middle Jurassic Caucasus Basin and then experienced their primary phases of extension in the Cretaceous (e.g., Zonenshain & Le Pichon, 1986). The mechanism for Cretaceous extension in the Black Sea remains enigmatic, with proposals including trench retreat/slab rollback attributed either to toroidal flow along the western edge of the Neotethyan slab (Stephenson & Schellart, 2010) or to subduction of a small spreading center (Sosson et al., 2016). Cenozoic extension in the South Caspian occurred following closure of the
Sevan-Akera suture (Brunet et al., 2003); thus, it may be worth exploring whether collision of the South Armenian Block could have facilitated along-strike back-arc extension in the South Caspian.

5. Conclusions

Geochronologic and geochemical data from the Caucasus region are consistent with Early-Middle Jurassic (~180–160 Ma) opening of the Caucasus Basin back-arc as the result of along-strike collision of the Iranian block with Laurasia in the Late Triassic (~220–210 Ma). \(^{40}\)Ar/\(^{39}\)Ar and U-Pb geochronology in the Greater Caucasus and Scythian Platform dominantly yield ages of ~260–220 Ma, indicating a Permian-Triassic pulse of arc magmatism to the north of the Caucasus Basin. As proposed by previous workers (Alexandre et al., 2004; Natal’In & Şengör, 2005; Okay & Nikishin, 2015; Tikhomirov et al., 2004), this likely reflects north-directed subduction beneath Laurasia during this time, prior to accretion of the Iranian block.

\(^{40}\)Ar/\(^{39}\)Ar and detrital zircon U-Pb data from the Greater Caucasus and Lesser Caucasus arc record a second population of ages beginning at ~180 Ma, which we interpret to reflect opening of the Caucasus Basin and initiation of arc magmatism in the Lesser Caucasus arc, respectively. Early-Middle Jurassic magmatic rocks in the Greater Caucasus yield trace-element geochemical signatures primarily comparable to modern arc rocks with some samples yielding patterns more similar to D-MORB and basalts derived from enriched mantle sources (e.g., E-MORB, OIB, SCLM-derived basalt). We interpret these patterns to reflect development of a continental back-arc spreading center, given their similarity to geochemical patterns observed in the modern continental back-arc settings. Sr-Nd isotopic data indicate a mantle source for these rocks, and detrital zircon U-Pb data suggest the duration of magmatism was ~180–160 Ma.

Early-Middle Jurassic rocks in the Lesser Caucasus arc exhibit trace element patterns characteristic of subduction zone magmas but are relatively depleted in LREE and other highly incompatible elements. By contrast, Late Jurassic rocks also show geochemical affinities to arc basalt but exhibit stronger incompatible element enrichment, which we interpret to reflect an increasingly fertile mantle source following cessation of back-arc spreading at ~160 Ma. Sr-Nd isotopic data indicate a depleted source for these rocks, and detrital zircon U-Pb data suggest the duration of Jurassic arc magmatism was ~180–140 Ma.

Taken together, these geochronologic and geochemical data are best explained by trench retreat induced by the Late Triassic (~220–210 Ma) collision of the Iranian block with Eurasia. Trench retreat ended the ~260–220 Ma pulse of arc magmatism in the Greater Caucasus and Scythian Platform and led to extension in the overriding plate of the subduction system, opening the Caucasus Basin back-arc with associated magmatism from ~180 to 160 Ma. The magmatic front of the arc system migrated to the Lesser Caucasus arc, with a new pulse of arc magmatism lasting from ~180 to 140 Ma and transitioning from a relatively depleted mantle source caused by back-arc spreading from ~180 to 160 Ma to a more enriched mantle source from ~160 to 140 Ma. This collision-induced, trench-retreat mechanism is commonly cited for modern back-arc systems (e.g., Wallace et al., 2009), provides a clear explanation for the Triassic-Jurassic magmatic record in the Caucasus region, and may be useful for understanding the evolution of other ancient arc-back-arc systems.

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Data Availability Statement

Supplementary text, figures, and tables of compiled published data for this manuscript are provided in the Supporting Information S1 and are available in the repository Dryad (Vasey et al., 2021) at https://doi.org/10.25338/B8N03H. Python scripts used in the data analysis and figure construction are archived at Zenodo (Vasey, 2021) at https://doi.org/10.5281/zenodo.4741366.

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