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Reply

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Key Points:

- The Greater Caucasus Basin encompassed a broader region in space and time than envisioned by *Vincent et al.* [2018], who define the basin narrowly and disregard Cenozoic shortening.
- Terminal basin closure occurred when the Greater and Lesser Caucasus collided.
- While the data cited by *Vincent et al.* [2018] are consistent with the onset of basin closure by 35 Ma, in no way do they indicate terminal basin closure at this time.

Index Terms
8102 Continental contraction orogenic belts and inversion tectonics
8104 Continental margins: convergent
8157 Plate motions: past
8169 Sedimentary basin processes
9335 Europe

Key Words
Caucasus
Arabia-Eurasia collision
Relict basin closure
Plate motion deceleration
Hard and soft collision

We appreciate the interest in our recent work [*Cowgill et al.*, 2016, C16 herein] and the opportunity to explain our differences with a paper published in the same issue [*Vincent et al.*, 2016, V16 herein] and the subsequent comment [*Vincent et al.*, 2018, V18 herein]. The points raised by V18 do not significantly affect the results of C16.

Using the Caucasus as an example, C16 illustrates that closure of a relict basin in the early phase of continental collision provides a mechanism for absorbing plate convergence with minimal plate deceleration or upper-crustal shortening. Importantly, the shortening systems accommodating such basin closure can produce cryptic suture zones that are easily overlooked or misinterpreted. We further show that total crustal shortening preserved in a collisional orogen can be significantly lower than actual post-collisional plate convergence in the case of relict-basin closure, and that deficits of upper crustal shortening relative to post-collisional convergence should be expected (Fig. 9 in C16).

Determining the pre-collisional width of the Greater Caucasus basin is essential for understanding the evolution of the Caucasus region specifically, and relict basin closure in general. This parameter defines the total magnitude of underthrusting between the Greater and Lesser Caucasus against which upper crustal shortening can be compared, thereby constraining plausible geodynamic models for this region. Determining the pre-collisional width of this now-shortened and partially subducted basin is challenging. C16 proposed using the spatial distribution of detrital zircon provenance as one possible tool to evaluate if the basin was wide enough for sediments deposited within it to be solely sourced from one flank or other.

Our specific premise as presented in C16 is that between the Black and Caspian Seas, the Greater Caucasus Basin

- extended from the Variscan basement of Greater Caucasus in the north to the Mesozoic volcanic arc basement of the Lesser Caucasus in the south,
- started opening in the Jurassic as a back-arc basin along the Eurasian continental margin and then continued opening in a back arc setting throughout the late Mesozoic and early Cenozoic [e.g., *Zonenshain and Le Pichon*, 1986],
- was at least 350-400 km wide at its maximum extent in the Paleocene/Eocene, when together with the Black Sea and South Caspian Basin it formed a large back-arc basin system within Paratethys along the Eurasian margin [e.g., *Popov et al.*, 2004],
- was characterized by distinct detrital-zircon provenance domains at the time of maximum extent, with a northern domain dominated by Variscan-aged grains and a southern domain characterized by both Mesozoic-Cenozoic grains and a marked lack of the older grains seen in the northern domain,

- started closing in the Eocene via imbrication of the sedimentary cover of the basin and underthrusting/subduction of the underlying basement,
- continued closing through the Oligocene and Miocene until the arc basement of the Lesser Caucasus collided with the Variscan basement of the Greater Caucasus at ~5 Ma, accelerating rock and surface uplift in the Greater Caucasus,
- was subjected to sedimentary infilling throughout its history, both during the opening phase and later during closure, when successor flexural foreland basins were superimposed on the relict extensional back-arc basins,
- had complex internal structure characterized by multiple sub-basins, the geometry of which almost certainly changed over time as they were filled with sediment, partitioned during progressive shortening of the basin, and subjected to large base level changes in the increasingly restricted and compartmentalized Paratethys [e.g., *Forte and Cowgill*, 2013].

In their comment, V18 make numerous points regarding multiple aspects of the above premise. Many of these comments focus on important, but particularly nuanced points, so before responding to each of the many detailed points in V18, it is useful to note five main factors underpinning their disagreement with our work.

The first main source of disagreement regards the definition of the Greater Caucasus Basin: as shown on Fig. 1 of V18 and described in both V16 and V18, they define the Greater Caucasus Basin narrowly as solely Jurassic-Eocene deep-water deposits. This choice places several provenance samples outside of their definition of the Greater Caucasus Basin. In contrast, we use a broader definition, and consider the basin as the evolving depocenter that existed spatially between the Scythian platform in the north and the Lesser Caucasus to the south and temporally from the Jurassic until collision of the Greater and Lesser Caucasus. We therefore accept that the bathymetry and water depth of the Greater Caucasus Basin evolved over time, and was likely shallow for most of the Cenozoic due to sedimentary infilling and structural thickening of the basin fill as it was shortened. Thus, we have no problem placing shallow-water deposits inside the evolving Greater Caucasus Basin, as opposed to along its southern margin, as the model of V16 and V18 assumes. V18 fixate on distinguishing between an older rift basin (how they define the Greater Caucasus Basin) and younger successor flexural basins. We see little utility in such a genetic definition of the basin, which fails to account for possible flexure of the now-underthrust or subducted floor of the relict basin during early stages of basin closure. In our use, the Greater Caucasus Basin is the Caucasus sector of

Paratethys, regardless of subsidence mechanism.

The second main disagreement relates to the role of Cenozoic shortening in obscuring the original basin geometry and paleogeographic location of samples and stratigraphic sections. As noted in C16, the provenance samples and stratigraphic sections shown on Fig. 1 of V18 and Figs. 2 and S1 of C16 are exposed in thrust sheets within the Greater Caucasus orogen that have been detached from the underlying basement during Cenozoic shortening. Both the magnitude and timing of slip on these faults remain to be established. Thus, the original positions of these samples at the time of their deposition relative to both the basin margins, and one another within the basin, remain to be determined. Importantly, paleogeographic interpretations of the geology in this region by V18 largely ignore crustal shortening within the Caucasus region as a whole, and the Greater Caucasus in particular.

A third main disagreement centers on the meaning of terminal basin closure: V18 implicitly employ a stratigraphic definition of terminal basin closure that is only tangentially related to the tectonic model proposed in C16. In a tectonic sense, terminal basin closure occurs when a block of more buoyant crust in the lower plate is underthrust or subducted beneath an orogen and collides with the basement of a buoyant upper plate. Structurally, this would manifest as a transition from an orogen dominated by subduction with little to no accretion and upper plate-shortening to an orogen with significant accretion and upper-plate shortening, sensu Fig. 9 of C16. Dating such closure requires determining the timing of the major structures juxtaposing the basement of the upper and lower plates, such as dating thrusts that transferred basement units from the colliding lower-plate block into the overriding orogen. In the case of the Caucasus, C16 clearly refers to final basin closure as juxtaposition of the basement of the Lesser Caucasus in the lower plate to the south with the Variscan basement of the upper plate along the Scythian platform margin to the north.

In contrast, V18 appear to define terminal closure as either the onset of subaerial erosion, the cessation of deep marine deposition, or the transition to flexural foreland basin deposition along the margins of the basin. All three of these stratigraphic signals are critical for understanding the tectonic and paleogeographic evolution of a basin. However, they do not equate with terminal closure in the geodynamic/tectonic sense used in C16.

For example, onset of subaerial erosion within a convergence zone in no way requires that a basin has closed, it just means that a growing subaqueous thrust belt breached sea level, as illustrated by Nias Island [*Karig et al.*, 1978; *Moore and Karig*, 1980] and other sub-aerial portions of the outer-arc ridge of the Sunda forearc, which are adjacent to the Indian Ocean basin. Likewise, cessation of deep marine deposition in a closing basin does not directly correspond to geodynamic closure because it depends on the irregularity of the colliding margins [e.g., *Dewey and Burke*, 1974] and the topographic evolution of the intervening orogen/foreland basin system [*DeCelles*, 2011], as illustrated by the Sunda-Banda arc-continent collision [*Harris et al.*, 2009]. Similarly, conversion of a relict marine basin to a flexural foreland basin fails to date terminal collision because initiation of basin closure is expected to produce a flexural successor basin that is superimposed on those portions of the relict basin that are adjacent to the growing orogen. In an upper-plate-fixed reference frame, this foreland basin will migrate across the lower plate as that plate is subducted/underthrust [*DeCelles and DeCelles*, 2001]. Thus, once subduction/underthrusting of the Greater Caucasus Basin was underway, those parts of the relict basin to the south of the developing thrust belt would have converted to a foreland-basin setting, even before the basin had closed in a geodynamic sense.

Based on the aforementioned stratigraphic definitions, V18 argue that basin closure occurred at 35 Ma, and thus could not have driven the 5 Ma reorganization of the Arabia-Eurasia collision as argued in C16. However, as explained below, the data V16 cite as evidence of early basin closure simply indicate that shortening was underway by this time, and no data are provided to indicate that collision between the Lesser and Greater Caucasus is this old. Importantly, V16 and V18 fail to explain both progressive shortening from 35 Ma to present and robust thermochronologic evidence for a significant acceleration of rock and surface uplift in the Greater Caucasus at 5 Ma, which is attributed to collision between the Lesser and Greater Caucasus by C16.

The fourth main point regards the nature of the crust originally flooring the Greater Caucasus Basin: V18 argue that back arc rifting in the Greater Caucasus Basin did not generate oceanic lithosphere, as indicated by the lack of a Jurassic or younger oceanic suture zone with ophiolite or ophiolitic melange, subduction-related arc magmatism, or an accretionary prism. This argument presumes that the geologic expression of back-arc basin closure should be the same as that of large ocean basins. Rather than refuting subduction, we suggest that the geology of the Caucasus region serves to help inform our understanding of the geologic expression of such basin closure.

As noted by *Moores* [1981], the existence of a suture can be inferred from a variety of features in addition to ophiolites, including stratigraphic differences across deformed zones, structural discontinuities, and shear zones. C16 propose that the Greater Caucasus host such a suture zone, and interpret the thrust sheets in the Greater Caucasus and the foreland basin to the south as a type of accretionary prism, broadly defined, that formed from imbrication of the sedimentary cover of the Greater Caucasus Basin as it closed. It is unclear if closure of a small (350-400 km wide) back-arc basin should lead to the features that characterize sutures resulting from the closure of large ocean basins, particularly in the case of recent suturing.

We expect the geological expression of suturing to be a strong function of exposure depth. The Greater Caucasus have not likely been exhumed enough to expose the discrete suture with ophiolitic melange expected by V17. As an illustration, one could consider Taiwan, where the Luzon arc has collided with a relatively narrow basin along the Eurasian margin but shows little evidence of an ophiolite or ophiolitic melange at the surface. We agree that the paucity (but not absence) of subduction-related volcanism is surprising, although certainly not unique [cf. magmatic volumes in the Alps; *von Blanckenburg and Davies*, 1995]. In general, volumes of arc magmatism and crustal production vary in both space and time [*Dimalanta et al.*, 2002; *Jicha and Jagoutz*, 2015; *Paterson and Ducea*, 2015]. We suspect that the paucity of subduction-related volcanism in the Caucasus results from either subduction/underthrusting of a relatively narrow basin, or shortening of a system of sub-basins, such that the crust flooring the basin(s) may not have be subducted/underthrust deeply enough to trigger significant melting. It is also possible that late Cenozoic volcanism is present in the Greater Caucasus but has been erroneously assumed to be Mesozoic in age, based on lithostratigraphic correlation. Alternatively, subduction-related magmatism may have been low volume if the subduction angle was very shallow (e.g., Laramide), perhaps due to relatively buoyant basin lithosphere.

The fifth and final main disagreement is the characterization in V18 and V16 of existing thermochronometric constraints on the exhumation history of the Greater Caucasus, including: (1) the size of the thermochronometric dataset interpreted to reveal a rapid increase in exhumation at 5 Ma in the Greater Caucasus [*Avdeev and Niemi*, 2011]; (2) potential magmatic overprinting of thermochronometric ages from the central Greater Caucasus (V18 raise this concern explicitly for *Avdeev and Niemi* [2011], but samples collected by *Vincent et al.* [2011] from the same geographic area would presumably also be affected); and (3) the interpretation of thermochronometric datasets from the western Greater Caucasus [*Král and Gurbanov*, 1996; *Vincent et al.*, 2011].

Avdeev and Niemi [2011] presented a comprehensive suite of original thermochronometric data from the crest of the central Greater Caucasus, including apatite and zircon (U-Th)/He, apatite and zircon fission-track, and $^{40}Ar/^{39}Ar$ K-feldspar analyses from three transects spanning ~150 km along strike. Inverse thermal models were presented for individual samples for which multiple thermochronometric ages were determined, and age-elevation relationships were presented for samples that were collected in regions of high relief. All models revealed cooling from temperatures of >100-150°C since ~5 Ma [*Avdeev and Niemi*, 2011]. K-feldspar and zircon fission-track ages >230 Ma are consistent with ages of Caucasus basement [*Somin*, 2011], thus rocks exposed in the central Caucasus have been cooler than ~225°C since that time. Zircon (U-Th)/He ages in two of the transects are 20-30 Ma, requiring temperatures above ~180°C at that time. Together, these observations require exhumation of the central Caucasus to have occurred primarily during the Pliocene [*Avdeev and Niemi*, 2011]. The application of multi-method thermochronometry to vertical transects is a widely accepted technique for understanding the exhumation of mountainous regions globally [e.g. *Spotila*, 2005], and we are unconvinced that a high-quality data set constraining such a history for the Caucasus [e.g. *Hess et al.*, 1993; *Avdeev and Niemi*, 2011] should be discounted in favor of data sets with limited thermal history resolution albeit greater spatial extent [*Král and Gurbanov*, 1996; *Vincent et al.*, 2011].

V18 assert that published theromochronometric data from the central Greater Caucasus [*Avdeev and Niemi*, 2011; *Vincent et al.*, 2011] were collected from a "magmatically affected" region. The inclusion of this phrase appears to be intended to question the interpretation that these data record erosional or tectonic exhumation, although the authors of V18 offer no evidence to support this. Low-temperature thermochronometric data can be affected by magmatic activity either via local effects, such as thermal conduction or hydrothermal convection associated with a proximal igneous body [e.g., *Peyton and Carrapa*, 2013; *Whipp and Ehlers*, 2007; *Ault et al.*, 2014], or regionally, by an increase in the geothermal gradient of the crust [e.g., *Ehlers*, 2005]. Observational, geochemical, and numerical studies of metamorphic processes in contact aureoles typically find that local thermal effects associated with intrusions are limited to ~1 diameter of the intrusive body [e.g. *Cook et al.*, 1987; *Cui et al.*, 2001]. Studies that exploit the local resetting of low-temperature thermochronometers by shallow intrusions typically find resetting constrained to similar length scales [*Murray et al.*, 2016].

Resetting of low-temperature thermochronometers may also occur by top-down convection [*e.g. Ault et al.*, 2014], driven by hydrothermal cells beneath ignimbrite or lava flow emplacement. Such cells are short-lived and shallowly seated [e.g., *Gazis et al.*, 1996; *Keating*, 2005], but may impart distinct patterns on thermochronometric age-elevation relationships, with ages younging toward the paleosurface on which the hydrothermal system was established [e.g., *Abbey et al.*, 2018]. No samples presented by *Avdeev and Niemi* [2011] or *Vincent et al.* [2011] were collected in close proximity to known Cenozoic intrusive bodies, and no vertical sampling transects collected in the Caucasus display evidence of top-down re-setting [*Hess et al.*, 1993; *Avdeev and Niemi*, 2011].

Changes in the regional geothermal gradient can also reset thermochronometers, but such affects are typically of concern in older samples where raising and then lowering of the geothermal gradient may impart a cooling signal that is misinterpreted as exhumational [*Ehlers*, 2005]. Such resetting, however is limited to samples that are near their relevant closure temperatures, and thus which can be subject to diffusion or annealing with minimal temperature increases. While we cannot fully discount this processes in the Caucasus, the requirement that it acts only on samples near their closure temperatures precludes this process from meaningfully changing the exhumation histories inferred by *Avdeev and Niemi* [2011] and *Vincent et al.* [2011]. In summary, we find no evidence that thermochronometric ages in the central Greater Caucasus were affected by regional magmatism, and assert that the original interpretations of *Avdeev and Niemi* [2011] and *Vincent et al.* [2010] that these data sets record tectonic exhumation, rather than magmatic processes, are sound.

Third, we will clarify our interpretation of low-temperature thermochronometric data in the western Greater Caucasus [*Král and Gurbanov*, 1996]; *Vincent et al.,* 2010]. We focus primarily on *Vincent et al.* [2011], because their data set includes location information as well as fission-track length measurements that permit the extraction of thermal histories.

At first-order, the Greater Caucasus appear to broadly define an anticline [e.g. *Forte et al.*, 2014], with the depth of exhumation greatest in the center of the range and decreasing northward and southward toward the margins [*Vincent et al.*, 2010; *Avdeev and Niemi,* 2011]. Samples towards the margins of the range (and at originally shallower depths) may have cooled below their relevant closure temperatures prior to Cenozoic tectonic exhumation related to formation of the Greater Caucasus. This is true of samples in both studies [cf. transect B in *Avdeev and Niemi*, 2011; Fig. 5 in *Vincent et al.*, 2011]. It is not the case that recent, rapid exhumation went “undetected” by these samples [V18]; rather, these samples are incapable of recording such an event. This interpretation is consistent with the relatively robust relationship between local relief and age observed in all thermochronometric samples from the Greater Caucasus [*Forte et al.,* 2016].

However, samples from the core of the western Greater Caucasus [WG150/1, WC49/1; *Vincent et al.*, 2011] were above or near the closure temperature of the apatite fission-track system during Cenozoic time. As noted by *Vincent et al.* [2011]: “*[WG150/1] displays a…uniform Oligocene to Present cooling path, although there may have been an increase in cooling from 1°C Ma^{-1} to 3°C Ma^{-1} during Middle Miocene time (c. 15 Ma).*” We largely concur with this interpretation and note that it is consistent with the results presented by *Avdeev and Niemi* [2011], in which slow cooling at rates of ~1-4°C Ma^{-1} are observed from Oligocene time until latest Miocene time, followed by rapid cooling at rates >10°C Ma^{-1} [cf. *Hess et al.*, 1993]. The spread of acceptable thermal histories for sample WG150/1 is large, given that the thermal models are constrained by only a single apatite fission-track analysis, and we note that thermal histories in which cooling at ~1°C Ma^{-1} extends until 5 Ma, followed by cooling at a rate of ~9°C Ma^{-1} are equally consistent with the available data [Fig. 5e in *Vincent et al.*, 2011]. Likewise, *Vincent et al.* [2011] note that sample WC49/1 records *“rapid cooling to surface temperatures….sometime after Early Oligocene time.”* Again, we concur, and note that the data are consistent with much of this cooling occurring in the past 5 Ma [Fig. 5d in *Vincent et al.*, 2011]. Thus, post-5 Ma rapid cooling is consistent with, and potentially detected by, apatite fission-track data from the core of the western Greater Caucasus, but such a cooling history is not uniquely constrained given the limitations of the data set of *Vincent et al*. [2011].

With these main points in mind, we now address each of the numbered points raised in Section 3 of V18. The comment contains numerous highly detailed sub-points, a number of which are extraneous and repetitive. For clarity, we focus on those points that are essential to the disagreement.

V18 Section 3.1) Closure at 5 Ma is not supported.

We do not agree. As noted above, V18 assume a stratigraphic definition of terminal basin closure whereas we refer to collision between the Lesser Caucasus and Variscan basement along the southern edge of Scythian platform (e.g., p. 2937 in C16). Terminal basin closure around 5 Ma is supported by thermochronometric data clearly showing early Pliocene onset of rapid exhumation of the central Greater Caucasus [*Avdeev and Niemi*, 2011]. Deposits in the Chanis River section as young as mid-Miocene [*Vincent et al.*, 2014] and are folded within the mountain-front flexure along the southern flank of the Greater Caucasus [*Banks et al.*, 1997], indicating this structure has formed since mid-Miocene time. Progressive basin closure from Eocene to late Miocene time is indicated by the changing provenance patterns described in C16 and summarized below, with samples WG22/5 (Tortonian, Middle Sarmatian) and WG15/5 (Tortonian-Messinian, Meotian) attesting to a growing orogen to the north. The Tskhenis River section (Fig. 1 in V18) crosses deposits at the western end of a syncline that continues from west of Tsageri to east of Ambrolauri and that contains marine deposits as young as middle to upper Sarmatian [*Dzhanelidze and Kandelaki*, 1957; *Gudjabidze*, 2003; *Kandelaki and Kakhadze*, 1957], i.e., 10.5 to ~8.2 Ma [*Jones and Simmons*, 1996], attesting to a topographic low between the Lesser and Greater Caucasus until at least this time. In the following we refer to this syncline as the Tsageri-Ambrolauri belt. Younger strata are not preserved in this belt, so the minimum age of marine deposition in this part of the basin remains poorly determined.

V18 Section 3.2) Western segment of the basin closed at the Eocene-Oligocene transition.

In no way do the data cited by V18 document terminal basin closure by early Oligocene. Rather, they show that the basin had at least started closing by this time, as was indicated in C16 and is reflected in the thermochronometric data of *Avdeev and Niemi* [2011]. We agree that some portion of the early orogen was likely subaerial, given evidence of surface weathering in heavy mineral data and sediment recycling [*Vincent et al.*, 2013; *Vincent et al.*, 2007]. However, we note that subaqueous slope failures on an actively deforming orogenic wedge built from imbricated sheets of Mesozoic-Paleogene fill of the back-arc basin could also contribute to recycling of the fill into younger deposits to the south.

V16 and V18 estimated early Oligocene paleoelevations around 2 km, based on the preponderance of montane pollen. However, the link between the pollen and adjacent topography is tenuous. Pollen is widely disseminated, as illustrated by measureable concentrations of terrestrial palynomorphs in marine cores collected 200-450 km off the coasts of west Africa [*Hooghiemstra et al.*, 2006] and the Pacific northwest of North America [*Heusser and Balsam*, 1977]. Subaerial volcanic edifices in the Lesser Caucasus may have been another potential source of montane pollen adjacent to the basin.

Most importantly, in no way does the presence of early Oligocene subaerial sediment sources in the western Greater Caucasus require the Lesser and

Greater Caucasus to have collided by this time. As noted above, the Sunda forearc on the eastern edge of the Indian Ocean provides a clear example of the subaerial exposure of an accretionary orogen that does not equate with closure of the adjacent basin. Oligo-Miocene deposits in the southern Greater Caucasus (Tskhenis R. section and Tsageri-Ambrolauri belt to the east, see Fig. 1 of V18) and along the southern flank of the range (Chanis R. section, Fig. 1 on V18) clearly attest to an intervening basin between the orogen to the north and the Lesser Caucasus to the south. Paleogeographic maps of the region show this basin existing through middle late Miocene (late Tortonian) time [*Popov et al.*, 2004], although these maps do not appear to fully account for subsequent shortening. We infer that this basin was a successor flexural foreland basin superimposed on the relict back-arc basin.

a) Flexural modeling is incompatible with a ~350-400 km wide Eocene basin.
This is a flawed argument. First, V18 incorrectly cite Fig. 7b of *Shillington et al.* [2008] as implying that flexure of the Tuapse margin will only be observed when the load is within 70 km of an observation point. This figure is very peripheral to the study and shows a non-unique flexural model that is not explained. *Shillington et al.* [2008] do not demonstrate that the modeled profile of the Shatsky Ridge is an originally horizontal surface that has been deformed by flexural loading, nor do they account for possible block rotation. Likewise, they do not establish the location of the modeled profile relative to the load and forebulge, nor do they evaluate the effect of a bending moment on the plate end, a distributed vs. point load, or non-elastic plate rheology. Thus, in no way does *Shillington et al.* [2008] establish the geometry of the flexural foreland basin along the southern margin of the Caucasus. Second, V18 assumes that the behavior of the Tuapse trough in the Black Sea is representative of that in the Rioni basin, ~300 km to the east, which fails to account for possible along-strike variability in the nature and behavior of the lower plate. Third, V18 implicitly assume that the modern effective elastic thickness and load size and distribution are all the same as they were in the Oligocene, which fails to account for growth of the orogen, changes in the nature of the plate being underthrust, and increases in bending moment over time. Fourth, as noted below, the original locations of the Oligocene sedimentary sections within the basin and relative to the growing Caucasus orogen remain unknown.

b) Oligo-Miocene basinal sediments are lacking within the "suture zone"
The general absence of Oligo-Miocene strata from the interior of the Greater Caucasus orogenic wedge is mostly simply interpreted as resulting from erosion of the shallowest structural levels of the thrust belt. As noted above, we place the

Oligo-Miocene deposits exposed along the Chanis and Tskhenis River sections (and in the Tsageri-Ambrolauri belt to the east of the Tskhenis R.) within the Greater Caucasus Basin, as broadly defined. This contrasts with V16 and V18, who place them outside the basin and to the south (Fig. 1 in V18). We also note that unit ages in the Greater Caucasus region are based almost exclusively on Soviet-era biostratigraphic data with little independent confirmation.

V18 Section 3.3) Data from the Russian Caucasus are key

V18 correlate late Jurassic to Eocene shallow water deposits along the southern flank of the Greater Caucasus range from north and west of the Russian-Abkhazian border to north of the Dzirula massif, and then assert that these deposits define the southern shelf of the Greater Caucasus Basin. They further state that the Shatsky ridge, basement of the Rioni basin, and Dzirula Massif define the southern margin of the Greater Caucasus Basin, thus implying that the Jurassic-Eocene shallow water units along the southern flank of the Greater Caucasus range were deposited on this basement ridge.

However, the structure of this portion of the Greater Caucasus orogen remains uncertain. As noted above, all of the Oligo-Miocene and older deposits exposed along the southern flank of the Greater Caucasus have been detached from the underlying basement [e.g., *Banks et al.*, 1997] and transported unknown distances to the south. Until the structural context of these sections is established, the along-strike correlation of shallow-marine deposits proposed by V16 and V18 should be viewed with caution. The correlated units may have been deposited at different latitudes within the Greater Caucasus Basin and then juxtaposed during subsequent shortening. More significantly, we do not agree with the assertion that shallow water deposition equates with the southern margin of the basin. In contrast to V16 and V18, who assume that the Greater Caucasus Basin is only defined by deep-water facies, we accept that the bathymetry of the basin evolved over time, and was likely shallow for most of the Cenozoic due to sedimentary infilling and structural thickening of the basin fill as it was shortened. Thus, we have no problem placing these deposits within the Greater Caucasus Basin, the ultimate southern margin of which was the Lesser Caucasus. We interpret the Shatsky rise-Dzirula Massif as an intrabasin high (as the Shatsky is in the modern Black Sea), not as a block that defined its southern margin. The paleogeographic evolution of this block and its implications for the sub-basin structure in the transition between the Eastern Black Sea the Greater Caucasus Basin are important outstanding problems that remain to be addressed.

a) There is no geological rationale for excluding data W of 41.5°E.

As we explained in C16, we largely excluded data from the Russian western Greater Caucasus because the relict basin (i.e., Black Sea) is still open to the west and our focus was on the role of relict-basin closure in accommodating post-collisional plate convergence. Although V18 summarize recent work on tectonostratigraphic zonation of this region, the magnitude and timing of shortening within this part of the belt are not addressed.

b) No evidence for basinal facies north of the Variscan basement.

As noted above, we define the Greater Caucasus Basin more broadly than V16 and V18, and include within the basin strata deposited along its northern margin, regardless of depositional environment.

c) Sample WC99/3 is located south of the inferred suture.

The location of this sample relative to the suture is unclear because its structural context remains to be established. In C16 we place this sample in the northern part of the basin, based on its location on the north side of the Black Sea and west of the collision between the Greater and Lesser Caucasus. The phrasing on p. 2931 was an error of omission, and should have read "north of the suture or in the northern part of the basin." All other references to the paleogeography of this sample in C16 refer to it as being in the northern part of the basin. The uncertainty of its position is illustrated in Fig. 6 of C16 by the termination of the area delineating the buried suture zone east of the sample.

d) The Main Caucasus Thrust (MCT) and relict-basin strata are mislocated east of Mt. Kazbek

V18 cite Figure 3 of *Mosar et al.* [2010] for the location of the MCT. However, that study provided no primary data bearing upon the location of this important structure. As noted in *Forte et al.* [2015], the location of the MCT is disputed, and we follow that study in placing the MCT along the Zangi thrust. Near Lagich, Azerbaijan, the Zangi thrust juxtaposes markedly different sections of Aptian-Albian strata, with turbidite facies deposits of the relict basin in the hanging wall thrust over volcaniclastic deposits of Lesser Caucasus affinity in the footwall to the south. The two Cretaceous sections are separated in places by Eocene-Oligocene anoxic deep-water shale deposits of the Maikop Formation [*Kopp*, 1985]. Furthermore, V18 incorrectly cite *Egan et al.* [2009] as indicating dominantly thick-skinned deformation within the Greater Caucasus via reactivation of deep-rooted normal faults. That study investigated the subsidence history of the South Caspian Basin offshore of Azerbaijan, and found that a model incorporating subduction of oceanic-type crust beneath the Apsheron sill best matches observed subsidence along the northern margin of South Caspian

Basin. They clearly show that reconstructing fault slip alone fails to reproduce the subsidence history.

e) The Racha-Lechkumi fault marks the southern margin of the basin.

V18 cite *Yakovlev* [2012] as indicating that the southern structural margin of the Greater Caucasus is defined by the Racha-Lechkumi fault (see Fig. 1 in V18 for location), a partly inverted normal fault across which Upper Jurassic to Cretaceous deposits thin from 8 to 1 km. *Yakovlev* [2012] presented and applied a nontraditional method of strain analysis to develop a 3D structural model of the northwestern Greater Caucasus. It provided no primary data documenting the regional significance of the Racha-Lechkumi fault or sedimentary thickness variations across it. Such thickness variations are shown in a theoretical sketch of their model, in which the Racha-Lechkumi fault formed as a half stretching fault [e.g., *Means*, 1989], with reverse separation at shallow levels and normal separation at deeper levels due to bulk shortening and thickening in the hanging wall relative to the footwall. Thus, it is not clear if *Yakovlev* [2012] envisioned the thickness juxtaposition across the Racha-Lechkumi fault as a primary depositional feature or the result of Cenozoic shortening. If the juxtaposition in stratal thickness is real, we note that it could also reflect large-magnitude thrust displacement along the Racha-Lechkumi fault, unless hanging wall and footwall cutoffs with different thicknesses in the same unit can be documented.

f) Oligo-Miocene sections along the Chanis and Tskhenis were deposited on a shelf that defined the southern margin of the Greater Caucasus Basin.

We do not accept the assumed paleobathymetry or placement of these sections along the southern margin of the basin. We view these section as having been deposited in a successor flexural foreland basin that was superimposed on the Greater Caucasus Basin and which migrated across the basin as the underlying crust was underthrust/subducted beneath the advancing Greater Caucasus orogen.

g) A “crustal segment” separates the inverted western Greater Caucasus Basin from the Eastern Black Sea.

We interpret this crustal segment as one or more intrabasinal highs. Basin deposits overlap this high (e.g., Shatsky ridge) from at least mid-Miocene to present [e.g., *Shillington et al.*, 2008]. The paleogeographic maps of *Popov et al.* [2004] show a continuous marine basin extending from the Eastern Black Sea across this intrabasin high and into the Tuapse, Rioni, Kura, and South Caspian basins from the Eocene up to mid-late Miocene time.

V18 Section 3.4) The context of Oligo-Miocene provenance samples is misinterpreted.

C16 indicated that the provenance samples were deposited in both a back-arc basin and successor foreland basins within the Greater Caucasus Basin (as we broadly define it both in C16 and here). V18 presume that we follow their definition of the Greater Caucasus Basin as limited to deep-water facies, which we do not.

a) Oligo-Miocene sediments along the southern margin of the Greater Caucasus were deposited in a series of foreland basins

We broadly agree, although we place these foreland basins within the broader Greater Caucasus Basin. All of the observations they cite (paleocurrent data, plant fragments, reworked Jurassic-Eocene nanofossils, in situ montane palynomorphs, and sandstone compositions with Variscan and East European Craton provenance) are equally compatible with deposition within a still-closing Greater Caucasus Basin, south of a growing orogenic wedge to the north. As noted above, the crux of this disagreement is on the paleogeographic position of these samples. While V18 places them on the southern margin of Greater Caucasus Basin, we place them in the interior of the basin. Deposition of these strata in a foreland-basin setting is compatible with our model because we expect successor foreland basins to have migrated through the central and southern parts of the Greater Caucasus Basin during its progressive closure.

b) Jurassic-Eocene sediments in the Russian Greater Caucasus are dominated by East European Craton zircons in unpublished data, so V18 interpret the East European Craton material in WC99/3 and WG66c/2 as recycled from these sources during inversion.

Because the data are unpublished, we are unable to review and fully evaluate them. However, recycling makes sense and this is a point upon which we agree. C16 noted the possibility of recycling from Mesozoic sediments for samples WC99/3 (Sochi) and WC139/1 (Taman) to explain the ~170 Ma peak in these samples, but it applies to older grain populations as well. We agree that these samples reflect derivation from the northern provenance domain as defined in C16. C16 noted that sample WG66c/2 is dominated by Variscan and East European Craton provenance peaks and concluded that this sample occupied a position within a partially closed basin at the time of deposition, not on the southern margin as V18 argues. We reiterate C16 in noting that WG66c/2 shows only a few (3-4) single-grain peaks of south-domain provenance affinity, and stand by our inference that this indicates deposition in the interior of the basin with a small component of depositional exchange as the basin had partially closed by this time. The lack of strong south-domain peaks in WG66c/2 conflicts with the proposition of V18 that it was deposited on the S margin of the basin. Dominance of East European Craton and Variscan peaks in WC99/3 makes sense with our model because that sample is still located on the north flank of the relict basin, which here comprises the Black Sea.

c) Detrital zircon characteristics of sediments in the western Greater Caucasus are poorly constrained.

We agree that more work is needed. V18 focus on our single sandstone sample NW-GC but we note that the sample of modern sand from the Enguri river also provides insight, as do the samples from the eastern Greater Caucasus. We agree that material sourced from the East European Craton is present, and in the modern Enguri sediment this signature may reflect recycling of older primary material into younger basin deposits during basin shortening. The general absence of such material in sandstone sample SW-GC is significant, because it indicates a lack of sediment sourced from the north-domain at the time of deposition, but this point is overlooked by V18.

d) C16 interpreted plutonic and metamorphic rock fragments in samples WG28c/1, WG22/5, and WG15/5 as sourced from the northern provenance domain and deposited south of the suture zone. It is unclear how material made its way up onto the southern shelf. Deposition in a flexural foreland basin is preferred.

We do not accept that these samples were deposited on a shelf along the southern margin of a deep basin, and dispute the hypothetical bathymetry for the basin at the time the sampled deposits were emplaced. We agree that deposition was likely in a flexural foreland basin, which we view as a successor to the back arc basin. This successor basin was south of the main locus of shortening at the time of deposition, and the strata were subsequently incorporated into the Greater Caucasus thrust belt prior to collision between the Lesser and Greater Caucasus.

V18 Section 3.5) Provenance data provide no insights into the width of the Greater Caucasus Basin.

a) Oligocene and younger sediments should be excluded from the analysis due to their inappropriate geological context.

We fundamentally disagree with V18 on the paleogeographic context of these samples: we place them within the now-shortened back-arc basin and successor foreland basins that developed after shortening initiated. As such, these samples provide insight regarding the progressive shortening of the basin during closure.

b) The basin was probably highly segmented and turbidite systems were likely bathymetrically confined.

We agree that the Greater Caucasus Basin was likely segmented and note that additional structural segmentation was likely during closure and basin shortening. Our estimate of basin width was approximate, and was not primarily derived from the provenance data. We also considered the modern-day widths of the Black Sea and South Caspian Basin, which have previously been cited as analogs for the Greater Caucasus Basin [*Zonenshain and Le Pichon*, 1986], paleomagnetic data that provide a maximum bound on basin width of 1000 km [*Meijers et al.*, 2015a], and an oroclinal deflection of Eocene magmatic belt south of the basin [*Meijers et al.*, 2015b]. The size of the basin inferred by C16 has subsequently been shown to be compatible with a new minimum value of 200-280 km of orogen-perpendicular convergence in the Greater Caucasus since 35 Ma [*van der Boon et al.*, 2018]. Accounting for northward motion of the Alborz and Eastern Pontides relative to Eurasia would increase this value, bringing it closer to the 350-400 km estimate of C16.

V18 Section 3.6) Implicit reference to oceanic spreading is not supported.

a) C16 are careful to not specify crustal type

Yes, because it remains unknown. We did this intentionally, and were careful to note in the caption to Fig. 8 in C16 that we represent the basement of the basin as thinned continental crust and/or transitional oceanic crust.

b) No robust geological data indicate ocean lithosphere floored the Greater Caucasus Basin.

Using geologic observations to determine the various types of crust that formed the basement to the Greater Caucasus Basin is likely to be challenging, considering that this crust has largely been underthrust/subducted beneath the Greater Caucasus or remains buried within the collision. The model proposed in C16 works regardless of whether the basin was floored by thinned continental or island arc lithosphere, transitional oceanic lithosphere, or true ocean lithosphere. The central point is that its buoyancy was such that it could be underthrust or subducted without significantly slowing plate convergence.

c) Jurassic tholeiitic basalts indicative of oceanic lithosphere are missing.

We emphasize that the Greater Caucasus Basin opened as a back-arc basin along the Eurasian continental margin. As such, we do not expect the lithosphere to have the same characteristics as that produced at a mid-ocean ridge in a large and long-lived ocean basin. As discussed by *Saunders and Tarney* [1984], back-arc basins show a variety of crustal compositions transitional between N-type MORB, island arc, and calc-alkaline basalts, with calc-alkaline lavas characteristic of back-arc basins developed from rifting of continental crust. Back-arc basin crust often differs from MORB in that the basins open episodically, contain remnant-arc fragments as submarine ridges, and typically lack well-defined patterns of magnetic anomalies that may suggest diffuse spreading, with low-grade alteration of lavas by seawater leading to enrichment of K, Rb, and Sr and clouding the original geochemical signature of the lavas [*Saunders and Tarney*, 1984]. Most prior petrologic and geochemical studies lack clear geological and structural context for the suites of samples studied, and isotopic data are largely lacking in the literature for volcanic deposits in the Greater Caucasus. The best study reporting such data [*McCann et al.*, 2010] is from the Tuapse region, northwest of Sochi, and it is difficult to know how to extrapolate those results to the Greater Caucasus in Georgia and Azerbaijan.

d) In C16, most basin opening was post-Bajocian to Paleocene, but there is little record of this magmatic history.

We reiterate that the basement of the Greater Caucasus Basin was largely underthrust/subducted, or has yet to be exhumed within the orogen. It appears that the basal decollement to the Greater Caucasus orogen detached the sedimentary cover of the basin and incorporated that material into the thrust belt, but may not have stepped down into the underlying basement. We also note that phases of the basin extension may have been largely amagmatic, and that the ages of most volcanic/volcaniclastic units within the Greater (and Lesser) Caucasus remain to be confirmed.

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References:

Abbey, A. L., N. A. Niemi, J. W. Geissman, I. Z. Winkelstern, and M. Heizler (2018), Early Cenozoic exhumation and paleotopography in the Arkansas River valley, southern Rocky Mountains, Colorado, *Lithosphere,* in press, published on line Dec. 20, 2018.

Ault, A. K., M. Frenzel, P. W. Reiners, N. H. Woodcock, and S. N. Thomson (2016), Record of paleofluid circulation in faults revealed by hematite (U-Th)/He and apatite fission-track dating: An example from Gower Peninsula fault fissures, Wales, Lithosphere, 8(4), 379-385.

Avdeev, B., and N. A. Niemi (2011), Rapid Pliocene exhumation of the central Greater Caucasus constrained by low-temperature thermochronometry, *Tectonics*, *30*.

Banks, C. J., A. G. Robinson, and M. P. Williams (1997), Structure and regional tectonics of the Achara-Trialet fold belt and the adjacent Rioni and Karli foreland basins, Republic of Georgia, in *Regional and petroleum geology of the Black Sea and surrounding region*, edited by A. G. Robinson, pp. 331-346.

Cook, S. J., J. R. Bowman, and C. B. Forster (1997), Contact metamorphism surrounding the alta stock: Finite element model simulation of heat-and $^{18}O/^{16}O$ mass-transport during prograde metamorphism, American Journal of Science, 297(1), 1-55.

Cowgill, E., A. M. Forte, N. Niemi, B. Avdeev, A. Tye, C. Trexler, Z. Javakhishvili, M. Elashvili, and T. Godoladze (2016), Relict basin closure and crustal shortening budgets during continental collision: An example from Caucasus sediment provenance, *Tectonics*, *35*(12), 2918-2947.

Cui, X., P. I. Nabelek, and M. Liu (2001), Heat and fluid flow in contact metamorphic aureoles with layered and transient permeability, with application to the Notch Peak aureole, Utah, Journal of Geophysical Research, 106(B4), 6477-6491.

DeCelles, P. G. (2011), Foreland Basin Systems Revisited: Variations in Response to Tectonic Settings, in *Tectonics of Sedimentary Basins*, edited, pp. 405-426, John Wiley & Sons, Ltd.

DeCelles, P. G., and P. C. DeCelles (2001), Rates of shortening, propagation, underthrusting, and flexural wave migration in continental orogenic systems, *Geology*, *29*(2), 135-138.

Dewey, J. F., and K. Burke (1974), Hot Spots and Continental Break-up: Implications for Collisional Orogeny, *Geology*, *2*(2), 57-60.

Dimalanta, C., A. Taira, G. P. Yumul, H. Tokuyama, and K. Mochizuki (2002), New rates of western Pacific island arc magmatism from seismic and gravity data, *Earth and Planetary Science Letters*, *202*(1), 105-115.

Dzhanelidze, A. I., and N. A. Kandelaki (1957), K-38-XIII, Ministry of Geology and Mineral Protection USSR, Moscow.

Egan, S. S., J. Mosar, M.-F. Brunet, and T. Kangarli (2009), Subsidence and uplift mechanisms within the South Caspian Basin: insights from the onshore and offshore Azerbaijan region, *Geological Society, London, Special Publications*, *312*(1), 219-240.

Ehlers, T. A. (2005), Crustal thermal processes and the interpretation of thermochronometer data, Reviews in Mineralogy and Geochemistry, 58(1), 315-350.

Forte, A. M., and E. Cowgill (2013), Late Cenozoic base-level variations of the Caspian Sea: A review of its history and proposed driving mechanisms, *Palaeogeography, Palaeoclimatology, Palaeoecology*, *386*(0), 392-407.

Forte, A. M., E. Cowgill, and K. X. Whipple (2014), Transition from a singly vergent to doubly vergent wedge in a young orogen: The Greater Caucasus, Tectonics, 33(11), 2077-2101.

Forte, A. M., K. X. Whipple, and E. Cowgill (2015), Drainage network reveals patterns and history of active deformation in the eastern Greater Caucasus, *Geosphere*, *11*(5).

Forte, A. M., K. X. Whipple, B. Bookhagen, and M. W. Rossi (2016), Decoupling of modern shortening rates, climate, and topography in the Caucasus, Earth and Planetary Science Letters, 449, 282-294.

Gamkrelidze, P. D., and I. R. Kakhazdze (1959), K-38-VII, Ministry of Geology and Mineral Protection USSR, Moscow.

Gazis, C., H. P. Taylor, K. Hon, and A. Tsvetkov (1996), Oxygen isotopic and geochemical evidence for a short-lived, high-temperature hydrothermal event in the Chegem caldera, Caucasus Mountains, Russia, Journal of Volcanology and Geothermal Research, 73(3-4), 213-244.

Gudjabidze, G. E. (2003), Geological Map of Georgia, Georgian State Department of Geology and National Oil Company "SAQNAVTOBI".

Harris, R., M. W. Vorkink, C. Prasetyadi, E. Zobell, N. Roosmawati, and M. Apthorpe (2009), Transition from subduction to arc-continent collision: Geologic and neotectonic evolution of Savu Island, Indonesia, *Geosphere*, *5*(3), 152-171.

Hess, J., H. Lippolt, A. Gurbanov, and I. Michalski (1993), The cooling history of the late Pliocene Eldzhurtinskiy granite (Caucasus, Russia) and the thermochronological potential of grain-size/age relationships, Earth and Planetary Science Letters, 117(3-4), 393-406.

Heusser, L., and W. L. Balsam (1977), Pollen distribution in the northeast Pacific Ocean, *Quaternary Research*, *7*(1), 45-62.

Hooghiemstra, H., A.-M. Lézine, S. A. G. Leroy, L. Dupont, and F. Marret (2006), Late Quaternary palynology in marine sediments: A synthesis of the understanding of pollen distribution patterns in the NW African setting, *Quaternary International*, *148*(1), 29-44.

Jicha, B. R., and O. Jagoutz (2015), Magma production rates for intraoceanic arcs, *Elements*, *11*, 105-112.

Jones, R. W., and M. Simmons (1996), A review of the stratigraphy of Eastern Paratethys (Oligocene-Holocene), *Bulletin of the Natural History Museum of London*, *52*(1), 25-49.

Kandelaki, N. A., and I. R. Kakhadze (1957), K-38-XIV, Ministry of Geology and Mineral Protection USSR, Moscow.

Karig, D. E., M. B. Lawrence, G. F. Moore, and J. R. Curray (1978), Structural frame work of the fore-arc basin, NW Sumatra, *J. Geol. Soc.*, *137*(1), 77.

Keating, G. N. (2005), The role of water in cooling ignimbrites, Journal of Volcanology and Geothermal Research, 142(1), 145-171.

Kopp, M. L. (1985), Age and nature of deformations of sediments comprising the Lagich Syncline, *Moscow University Geology Bulletin*, *40*(1), 23-32.

Král, J., and A. G. Gurbanov (1996), Apatite fission track data from the Great Caucasus pre-Alpine basement, *Chemie der Erde*, *56*(2), 177-192.

McCann, T., F. Chalot-Prat, and A. Saintot (2010), The Early Mesozoic evolution of the Western Greater Caucasus (Russia): Triassic–Jurassic sedimentary and magmatic history, *Geological Society, London, Special Publications*, *340*(1), 181-238.

Means, W. D. (1989), Stretching faults, *Geology*, *17*(10), 893-896.

Meijers, M. J. M., et al. (2015a), A paleolatitude reconstruction of the South Armenian Block (Lesser Caucasus) for the Late Cretaceous: Constraints on the Tethyan realm, *Tectonophysics*, *644–645*(0), 197-219.

Meijers, M. J. M., et al. (2015b), Progressive orocline formation in the Eastern Pontides–Lesser Caucasus, in *Tectonic Evolution of the Eastern Black Sea and Caucasus*, edited by M. Sosson, R. A. Stephenson and S. A. Adamia, Geological Society, London, Special Publications, 428, First published online October 27, 2015, https://doi.org/10.1144/SP428.8.

Moore, G. F., and D. E. Karig (1980), Strutural geology of Nias island, Indonesia: implications for subduction zone tectonics, *American Journal of Science*, *280*, 193-223.

Moores, E. M. (1981), Ancient Suture Zones within Continents, *Science*, *213*(4503), 41-46.

Mosar, J., T. Kangarli, M. Bochud, U. A. Glasmacher, A. Rast, M.-F. Brunet, and M. Sosson (2010), Cenozoic-Recent tectonics and uplift in the Greater Caucasus: a perspective from Azerbaijan, in *Sedimentary Basin Tectonics* *from the Black Sea and Caucasus to the Arabian Platform*, edited by M. Sosson, N. Kaymakci, R. A. Stephenson, F. Bergerat and V. I. Starostenko, pp. 261-280, Geological Society, London.

Murray, K., P. Reiners, and S. Thomson (2016), Rapid Pliocene–Pleistocene erosion of the central Colorado Plateau documented by apatite thermochronology from the Henry Mountains, Geology, 44(6), 483-486.

Paterson, S. R., and M. N. Ducea (2015), Arc magmatic tempos: gathering the evidence, *Elements*, *11*, 91-98.

Peyton, S. L., and B. Carrapa (2013), An Introduction to low-temperature thermochronologic techniques, methodology, and applications, in Application of Structural Methods to Rocky Mountain Hydrocarbon Exploration and Development, edited by C. Knight and J. Cuzella, pp. 15-36, AAPG.

Popov, S. V., F. Rögl, A. Y. Rozanov, F. F. Steininger, I. G. Shcherba, and M. Kovac (Eds.) (2004), *Lithological-paleogeographic maps of Paratethys: 10 maps late Eocene to Pliocene*, Schweizerbart'sche Verlagsbuchhandlung, Stuttgart.

Saunders, A. D., and J. Tarney (1984), Geochemical characteristics of basaltic volcanism within back-arc basins, *Geological Society, London, Special Publications*, *16*(1), 59.

Shillington, D. J., N. White, T. A. Minshull, G. R. H. Edwards, S. M. Jones, R. A. Edwards, and C. L. Scott (2008), Cenozoic evolution of the eastern Black Sea: A test of depth-dependent stretching models, *Earth and Planetary Science Letters*, *265*(3), 360-378.

Somin, M. L. (2011), Pre-Jurassic Basement of the Greater Caucasus: Brief Overview, Turkish Journal of Earth Sciences, 20(5), 545-610.

Spotila, J. A. (2005), Applications of low-temperature thermochronometry to quantification of recent exhumation in mountain belts, Reviews in Mineralogy and Geochemistry, 58(1), 449-466.

van der Boon, A., D. J. J. van Hinsbergen, M. Rezaeian, D. Gürer, M. Honarmand, D. Pastor-Galán, W. Krijgsman, and C. G. Langereis (2018), Quantifying Arabia–Eurasia convergence accommodated in the Greater Caucasus by paleomagnetic reconstruction, *Earth and Planetary Science Letters*, *482*, 454-469.

Vincent, S. J., A. Carter, V. A. Lavrishchev, S. P. Rice, T. G. Barabadze, and N. Hovius (2011), The exhumation of the western Greater Caucasus: a thermochronometric study, Geological Magazine, 148(1), 1-21.

Vincent, S. J., F. Hyden, and W. Braham (2014), Along-strike variations in the composition of sandstones derived from the uplifting western Greater Caucasus: causes and implications for reservoir quality prediction in the

Eastern Black Sea, *Geological Society, London, Special Publications*, *386*(1), 111-127.
Vincent, S. J., A. C. Morton, F. Hyden, and M. Fanning (2013), Insights from petrography, mineralogy and U–Pb zircon geochronology into the provenance and reservoir potential of Cenozoic siliciclastic depositional systems supplying the northern margin of the Eastern Black Sea, *Marine and Petroleum Geology*, *45*(0), 331-348.
Vincent, S. J., A. C. Morton, A. Carter, S. Gibbs, and T. G. Barabadze (2007), Oligocene uplift of the Western Greater Caucasus: an effect of initial Arabia-Eurasia collision, *Terra Nova*, *19*, 160-166.
Vincent, S. J., W. Braham, V. A. Lavrishchev, J. R. Maynard, and M. Harland (2016), The formation and inversion of the western Greater Caucasus Basin and the uplift of the western Greater Caucasus: Implications for the wider Black Sea region, *Tectonics*, *35*(12), 2948-2962.
Vincent, S. J., A. Saintot, J. Mosar, A. I. Okay, and A. M. Nikishin (2018), Comment on 'Relict basin closure and crustal shortening budgets during continental collision: An example from Caucasus sediment provenance' by Cowgill et al. [2016], *Tectonics, doi:10.1002/2017TC004515, in press*.
von Blanckenburg, F., and J. H. Davies (1995), Slab breakoff: A model for syncollisional magmatism and tectonics in the Alps, *Tectonics*, *14*(1), 120-131.
Whipp, D. M., and T. A. Ehlers (2007), Influence of groundwater flow on thermochronometer-derived exhumation rates in the central Nepalese Himalaya, Geology, 35(9), 851-854.
Yakovlev, F. L. (2012), Methods for detecting formation mechanisms and determining a final strain value for different scales of folded structures, *Comptes Rendus Geoscience*, *344*(3), 125-137.
Zonenshain, L. P., and X. Le Pichon (1986), Deep basins of the Black Sea and Caspian Sea as remnants of Mesozoic back-arc basins, *Tectonophysics*, *123*, 181-211.