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Key Points:

- New zircon trace element data record thin crust at 200–80 Ma and major phases of crustal thickening at 80–55 and 35 Ma to present
- Thickening at 55–35 Ma is obscured by a magmatic lull due to shallow subduction, during which thickening continued via crustal shortening
- Results highlight the complex spatiotemporal relationship of late Cenozoic Central Andean crustal thickening, surface uplift, and climate

Supporting Information:

Supporting Information may be found in the online version of this article.

Correspondence to:

K. E. Sundell, kurtsundell@isu.edu

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Author Contributions:

Conceptualization: Kurt E. Sundell, Sarah W.M. George, George E. Gehrels, Mihai N. Ducea, Joel E. Saylor, Martin Pepper

Data curation: Kurt E. Sundell Formal analysis: Kurt E. Sundell Methodology: Kurt E. Sundell, Sarah W.M. George, George E. Gehrels, Mihai N. Ducea, Joel E. Saylor, Martin Pepper Writing – original draft: Kurt E. Sundell, Sarah W.M. George, Barbara Carrapa, Mihai N. Ducea, Joel E. Saylor

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Crustal Thickening of the Northern Central Andean Plateau Inferred From Trace Elements in Zircon

Kurt E. Sundell^{1,2}, Sarah W.M. George¹, Barbara Carrapa¹, George E. Gehrels¹, Mihai N. Ducea^{1,3}, Joel E. Saylor⁴, and Martin Pepper¹

¹Department of Geosciences, University of Arizona, Tucson, AZ, USA, ²Now at Department of Geosciences, Idaho State University, Pocatello, ID, USA, ³Faculty of Geology and Geophysics, University of Bucharest, Bucharest, Romania, ⁴Department of Earth, Ocean, and Atmospheric Sciences, University of British Columbia, Vancouver, BC, Canada

Abstract The timing of crustal thickening in the northern Central Andean Plateau (CAP), at 13–20°S, and its relationship to surface uplift is debated. Zircon qualitatively records crustal thickness as its trace element chemistry is controlled by the growth of cogenetic minerals and relative uptake of light and heavy Rare Earth Elements. Jurassic to Neogene zircons from volcanic rocks, sandstones, and river sediments reveal shifts in trace element ratios suggesting major crustal thickening at 80–55 Ma and 35–0 Ma, coincident with high-flux magmatism. An intervening magmatic lull due to shallow subduction obscures the magmatic record from 55 to 35 Ma during which thickening continued via crustal shortening. Protracted thickening since the Late Cretaceous correlates with early elevation gain of the CAP western margin, but contrasts with Miocene establishment of near modern elevation in the northern CAP and the onset of hyperaridity along the Pacific coast, highlighting their complex spatial and temporal relationship.

Plain Language Summary Detailed relationships between crustal thickening, surface uplift, and climate remain unresolved. Although most mountains seem to be in isostatic equilibrium today, there is an imperfect correlation between elevation and crustal thickness in the modern continental lithosphere globally. We report trace element geochemical data of zircons extracted from volcanic rocks, sandstones, and modern river sediments collected in the northern Central Andean Plateau. Our analysis suggests that crustal thickness significantly increased from 80 to 55 Ma and from 35 Ma to present. The magmatic record between these two periods is obscured because the subducting slab shallowed, causing a decrease in magmatic activity; despite this, crustal thickening continued due to crustal shortening. The ~80 Myr history of crustal shortening correlates with an early phase of surface uplift of the western side of the orogenic system, but contrasts with the main phase of plateau uplift and establishment of hyperaridity along the Pacific coast.

1. Introduction

The Central Andes Mountains are home to the Central Andean Plateau (CAP), which including the Altiplano and Puna regions, makes up the largest modern orogenic plateau on Earth formed by Cordilleran orogenesis (Jordan et al., 1983; Horton, 2018; Figures 1a and 1b). This impressive topographic feature developed in the absence of continental collisional orogenesis (Isacks, 1988). Despite protracted east-directed subduction of oceanic lithosphere beneath the South American continent since the early Paleozoic, most estimates of paleoelevation suggest high topography and development of the northern CAP (13-20°S) occurred in the Cenozoic (e.g., Ghosh et al., 2007; Ramos, 2009). However, given that the arc was active prior to the Cenozoic, some topography and elevation must have existed along the western side of the orogenic system as recorded by the deformation and exhumation history of the western CAP margin (Boschman, 2021; Henriquez et al., 2019; Juez-Larré et al., 2010). Accompanying the high topography of the CAP is some of the thickest continental crust on Earth, varying from 50 to >70 km (Ward et al., 2016; Figure 1c); how this thick crust formed in the absence of collisional orogenesis remains particularly enigmatic. Specifically, the temporal relationship between thickening of the crust and the development of regional plateau topography is an outstanding question. Globally, the modern Earth has a remarkable correlation between crustal thickness and elevation in regions outside of cratons (Farner & Lee, 2017; Laske et al., 2013). In the CAP, while shortening has been documented as early as \sim 80 Ma in the northern and southern CAP (Horton, 2018; Kley & Monaldi, 1998; McQuarrie et al., 2005), surface uplift to extreme elevation (i.e., >3 km) is speculated to have occurred mostly within in the last 20 Myr (C. N. Garzione et al., 2017), highlighting an apparent temporal offset between shortening and surface uplift. This is in contrast to the Puna Plateau to the







Figure 1. (a) Digital elevation model (DEM) of western South America generated using HYDRO1K data. Dashed box is the location igneous whole rock data were extracted from the Central Andes Geochemical Database (CAnDB, Mamani et al., 2010). (b) DEM generated from the Shuttle Radar Topography Mission 90 m elevation data of the northern Central Andean Plateau and sample locations. (c) Depth to Moho estimates in km from Ward et al. (2016).

south which instead shows a coupling between shortening, crustal thickening and uplift for the last \sim 40 My (P. G. DeCelles et al., 2015).

The temporal relationship between crustal thickening and surface uplift is central to the debate on what process(es) best explain the generation of high topography and development of orogenic plateaus in Cordilleran orogens. The greater Andean orogen is the archetype Cordilleran orogenic system on Earth in the modern, and determining the crustal growth history is valuable to testing wholistic models of plateau development and explaining the relationships between crustal shortening, magmatism, basin development, surface uplift, and climate. Although earlier studies suggest a shortening deficit at most latitudes of the Central Andes (Hindle et al., 2005; Kley & Monaldi, 1998), more recent studies show that shortening can account for most crustal thickening (e.g., Hernandez et al., 2021). Still, the timing of crustal thickening remains an outstanding question in the northern CAP because the region is mantled by Cenozoic volcanic rocks, thus precluding detailed structural analysis as has been done farther to the south (e.g., Anderson et al., 2017).

Trace element geochemistry provides information on the history of continental crustal growth (and decay) by exploiting the relative uptake of light and heavy Rare Earth Elements (LREE and HREE) in cogenetic mineral phases. Specifically, trace elemental ratios can be used to track relative changes in crustal thickness because they reflect the presence or absence of specific mineral phases formed from residual magmatic melt (Chiaradia, 2015; Ducea et al., 2015; Heaman et al., 1990; Mantle & Collins, 2008; Profeta et al., 2015; Turner & Langmuir, 2015). Similarly, zircon $(ZrSiO_4)$ can be used to infer the chemistry of SiO₂ rich melts because it is refractory and resistant to post-crystallization diffusion (Cherniak et al., 1997), and it forms over a wide range of conditions (Watson & Harrison, 1983). Zircon preferentially incorporates HREEs via xenotime coupled substitution wherein $REE^{3+} + P^{5+} = Zr^{4+} + Si^{4+}$ (Finch et al., 2001), and is thus sensitive to other cogenetic minerals that incorporate HREEs such as amphibole and garnet. Zircon with low LREE/HREE can be interpreted to have formed in thin crust where there is an abundance of HREE and limited LREE in the residual melt, as the latter are preferentially incorporated into plagioclase and the former are generally incompatible. Conversely, zircon with elevated LREE/ HREE can be interpreted to have formed in thick crust where HREE are preferentially incorporated into garnet and amphibole and there is an abundance of incompatible LREE. Zircon Eu/Eu* $(Eu/\sqrt{Sm \times Gd})$ is also a function of crustal thickness due to the preferential incorporation of Eu²⁺ into plagioclase, and due to elevated Fe³⁺/ Σ Fe by garnet sequestration resulting in increased oxidation of Eu²⁺ to more compatible Eu³⁺ (Tang et al., 2021). Collectively, the combination of trace element proxies with U-Th-Pb geochronology (e.g., Gehrels et al., 2008; Schoene, 2014) makes zircon is a useful petrochronometer (Kylander-Clark, 2017) that can be applied to produce time-resolved records of geochemical change in igneous melt chemistry and used to infer qualitative changes in crustal thickness (e.g., Balica et al., 2020; McKenzie et al., 2018; Tang et al., 2021).

We analyzed Jurassic to Neogene zircons extracted from a suite of volcanic rocks, sandstones, and modern river sediments in the northern CAP (Figure 1). Identifying crustal growth trends works best for igneous rocks and

minerals that formed in intermediate melts in order to avoid deep mafic melts that undergo major diversification during homogenization in the melting–assimilation–storage–homogenization (MASH) zone and highly fractionated melts or melting of metasedimentary rocks in the upper crust (J. B. Chapman et al., 2016; Profeta et al., 2015). We interpret geochemical results using a multivariate predictive model to infer zircon melt chemistry following the methods and training data outlined in Belousova et al. (2002). Temporal trends in zircon trace element data are then placed in the context of published regional igneous whole rock data, surface uplift estimates, basin stratigraphy, and paleoclimate records from the Late Cretaceous to Cenozoic. Finally, data are synthesized into a regional tectonic model for the northern CAP applicable to Cordilleran orogens in Earth's deep past.

2. Methods

Trace element geochemistry and U-Th-Pb geochronology of zircon was conducted using laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS) at the Arizona LaserChron Center (ALC) following methods outlined in Balica et al. (2020). We analyzed the trace element concentrations of zircons with published U-Pb ages from late Oligocene to Pleistocene volcanic rocks across southern Peru (K. E. Sundell, Saylor, Lapen, et al. (2019)), detrital zircons from late Eocene to late Miocene sandstones in the Peruvian Altiplano (K. E. Sundell et al., 2018), and detrital zircons from modern river sediments in southern Peru and northern Chile (Pepper et al., 2016; Figure 1b). A small subset of previously unanalyzed zircons from the Altiplano samples were analyzed for both geochronology and geochemistry. In total 760 zircons were analyzed. Outliers were detected and removed if the individual element or $\sum LREE/\sum$ HREE results were outside median $\pm 2\sigma$ standard deviation for the complete data set, and/or if chondrite-normalized Lu was less than 1,000 ppm, the latter as a proxy for analysis of non-zircon (Belousova et al., 2002). Data filtering reduced the data set size to n = 710, including n = 326 zircons from N = 58 volcanic rock samples, and n = 384 detrital zircons from N = 7 sandstone samples and N = 5 modern river sediment samples.

We used a model-based approach to infer igneous melt chemistry from zircon trace element concentrations. Although early work suggested zircon trace element chemistry is not useful for determining rock type (Hoskin & Ireland, 2000). Belousova et al. (2002) demonstrated that application of multivariate analysis can be used as a robust predictor of igneous rock type. We apply the classification and regression tree (CART) model developed in Belousova et al. (2002). CART models fall under the broader umbrella of multivariate predictive models in supervised machine learning. Here, multivariate data with known responses are used to "train" a predictive model which can then be applied to new data to make predictions (Kuhn & Johnson, 2013). Belousova et al. (2002) developed a CART based on igneous rock types with known zircon trace element concentrations based on n = 330zircon analyses from N = 18 different rock types to train a predictive CART (Figure 2a). Passing zircon trace element data through this model produces a predicted rock type at confidence levels above 75% (Belousova et al., 2002). We first evaluate results in comparison to Ti-in-zircon crystallization temperature calculated following methods of Watson and Harrison (2005), then temporally. CART analysis, as applied here, effectively acts as a filter for interpreting temporal trends in zircon melt chemistry by excluding zircons inferred to be from highly fractionated, high SiO₂ melts and zircons formed in deep melts formed during MASH zone diversification. Results were also compared to data compiled in the Central Andes Geochemical Database (Mamani et al., 2010) between 13°S and 20°S to compare zircon geochemistry to regional whole rock geochemical records (Figure 1a).

Because intermediate magmatic compositions span such a wide range of depths within the crustal column, and as zircon saturates over a wide range of depths, the trace element concentrations are not necessarily representative of the true crustal thickness. Rather, we consider the range of trace element concentrations to be representative of various (and inherently unknown) depths of crystallization, with the highest LREE/HREE and Eu/Eu* ratios approaching the base of the crust. With this in mind, we seek temporal trends by binning the results into 4 Myr intervals and reporting the maximum values for each bin. We chose a 4-Myr interval because this is the smallest possible range given the limited data density of samples older than 40 Ma.

3. Interpretation of Zircon Trace Element Geochemistry

In zircon, log(Th/U) with values less than -1 can be interpreted to have formed in the presence of high-U metamorphic fluids (Rubatto, 2002). Results show that n = 699 yield log(Th/U) values greater than -1 and only n = 11 analyses yield log(Th/U) values < -1, which suggests that the majority of zircons in this study are igneous





Figure 2. (a) Classification and regression tree (CART) diagram from Belousova et al. (2002) (see Section 2 for details). (b) $\log(Th/U)$ as a proxy to discriminate between metamorphic and igneous zircons. (c) Results from passing trace element data through the CART shown in (a). (d) Zircon $\sum LREE / \sum HREE$ plotted with Ti-in-zircon temperatures. Data are separated into groups of zircons inferred to be from 65% to 70%, 70%–75%, and >75% SiO, melts.

in origin (Figure 2b). Application of the CART model from Belousova et al. (2002) suggests that 76% of analyzed zircons came from granitic melt with 65%–70% SiO₂, 10% from granitic melt with 70%–75% SiO₂, 1% from granitic melt with >75% SiO₂, 1% from basaltic melt, 10% from doleritic melt, 1% from Ne-syenitic melt, and 0% from carbonatitic, syenitic, and kimberlitic melts (Figure 2c).

Plotting $\sum LREE/\sum HREE$ against Ti-in-zircon crystallization temperature (Figure 2d) shows that analyzed zircons from high SiO₂ melt (70%–75% and >75%) generally formed at lower temperatures and lower $\sum LREE/\sum HREE$, whereas zircons from 65% to 70% SiO₂ melt formed over a wider range and generally higher temperatures and $\sum LREE/\sum HREE$. Low $\sum LREE/\sum HREE$ is consistent with melt chemistry in thin crust where there is an abundance of HREE and a deficit of LREE due to its preferential uptake in plagioclase feldspar (Heaman et al., 1990). High $\sum LREE/\sum HREE$ results point to residual melts where there is a depletion of HREEs due to competition with increasing abundances of cogenetic, high lithostatic pressure (>1 GPa) minerals such as garnet and amphibole (e.g., Mantle & Collins, 2008).

High SiO₂ melts undergo fractionation at shallower crustal levels, and thus do not provide reliable proxies of crustal thickness. Temporally, this can be seen in $\sum LREE / \sum HREE$ and Eu/Eu* where there is a clear separation from zircons derived from 65% to 70% SiO₂ melt after 35 Ma; the latter likely reflects the combined effect of decreased uptake of Eu²⁺ in plagioclase and oxidation of Eu²⁺ to Eu³⁺ (Figures 3a and 3b). This same trend can be seen in Ti-in-zircon crystallization temperature and Th/U, both of which show an increase in range after 35 Ma corresponding to increased range in crustal thicknesses and a diversification of magma. For the latter, lower Th/U likely corresponds to Th removal by accessory minerals that saturate before zircon such as monazite (Bea, 1996), rather than the presence of metamorphic fluids as only few zircon analyses show $\log(Th/U)$ values < -1 (Figure 2b), further supporting the idea that the crustal column became extremely thick after 35 Ma. In the absence of an assumed geothermal gradient, Ti-in-zircon temperatures can be interpreted as qualitative proxies of melt depth; results show a steady increase post 80 Ma, with a pronounced decrease in melt depths between 55 and 35 Ma (Figure 3c). Additionally, La/Yb, Sm/Yb, Ce/Y, and Dy/Yb zircon ratios are useful proxies for crustal thickening, with higher ratios indicating thicker crust, where garnet and amphibole are stable. Zircon data inferred to be from 65% to 70% SiO₂ granitoid melt compositions show an increase in all of these ratios between 80 and 55 Ma and 35 Ma to present (Figures 3e-3h). Available igneous whole rock data earlier than Neogene are limited; however, similar trends of increased zircon and whole rock trace element ratios post 35 Ma are interpreted as the result of increased crustal thickness, which appears to have continued to present day (Figures 3e-3h).





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Figure 3. Zircon trace element results plotted as (a) $\log(\sum LREE/\sum HREE)$, (b) $\log(Eu/Eu^*)$, (c) Ti-in-zircon temperature, (d) $\log(Th/U)$, (e) $\log(La/Yb)$, (f) $\log(Sm/Yb)$, (g) $\log(Ce/Y)$, and (h) $\log(Dy/Yb)$. Data are separated into groups of zircons inferred to be from 65% to 70%, 70%–75%, and >75% SiO₂ melts based on CART analysis (see Figure 2a). Parts e-h include igneous whole rock data (yellow squares) from the Central Andes Geochemical Database (see Figure 1 for spatial range; Mamani et al., 2010).

4. Crustal Thickening, Surface Uplift, and Paleoclimate

Zircon trace element data provide a nearly continuous record of crustal thickness change through time (Figures 3 and 4e). Although data are limited between 200 and 80 Ma, they provide a baseline of relatively thin crust. The overlap in zircons inferred to be from 65% to 70% and 70%–75% SiO₂ melts between 200 and 80 Ma followed by a separation after ~80 Ma supports a later diversification in the crustal column (Figures 3a–3d). Results support contrasting hypotheses based on two apparent pulses of crustal thickening at 80–55 Ma and 35–0 Ma (Figures 3 and 4e). Either the crust thickened in two separate stages, with an intervening period of crustal thinning (removal), or alternatively, the crust thickened continuously, albeit with subdued magmatism during shallow subduction, thus obscuring any thickening trends that would otherwise be seen in zircon trace element data. We prefer the latter hypothesis based on (a) an inboard stepping of the magmatic arc after ~45 Ma (Figure 4f; Mamani et al., 2010); (b) largely decreased Ti-in-zircon temperatures at 55-35 Ma, consistent with limited high-T (i.e., deeper) melt production; (c) nearly continuous advancement of the retroarc thrust front since ~80 Ma (Figure 4g;



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Figure 4. (a–d) Tectonic interpretation for the northern Central Andean Plateau modified from Ramos (2009). (e) Binned (at 4 Myr intervals) maximum log scale trace element ratios for zircon data inferred to be from 65% to 70% SiO₂ melt in Figure 3. (f) Compilation of igneous whole rock ages from the Central Andes Geochemical Database (Mamani et al., 2010) and detrital zircon ages (K. E. Sundell, Saylor, & Pecha, 2019 and references therein) shown as a kernel density estimate constructed using a 1 Myr kernel bandwidth; arc width and position from Mamani et al. (2010). (g) Crustal shortening estimates and position from Horton (2018). AP, Altiplano; EC, Eastern Cordillera; FB, Foreland basin; PC, Precordillera. (h) Maximum surface uplift estimates from (1) Sempere et al. (1997); (2) Picard et al. (2008); (3) Gregory-Wodzicki (2000); (4) C. N. Garzione et al. (2008); (5) Kar et al. (2016); (6) K. E. Sundell, Saylor, Lapen, et al. (2019); (7) Jeffery et al. (2013); (8) Boschman (2021), AP = dashed line, WC = dotted line. (i) Sediment accumulation in the Altiplano retroarc foreland basin from K. E. Sundell et al. (2018) and simplified regional basin stratigraphy from Carlotto (2013).

Horton, 2018); (d) Permo-Triassic rocks of the Eastern Cordillera that reached granulite facies at ~45 Ma (A. D. Chapman et al., 2015); and (e) increased retroarc foreland basin subsidence rates at 55–35 Ma, suggesting a coupled plate margin (Figure 4i; Horton, 2018); all of which is consistent with proposed timing and location of shallow subduction and tectono-magmatic model for the northern CAP of Sandeman et al. (1995). This interpretation suggests that magmatism, in addition to crustal shortening, is a dominant driver of crustal thickening (Haschke & Günther, 2003) during normal (i.e., not shallow) subduction.

The absence of proxy data precludes robust estimation of paleoelevation in the late Cretaceous-early Cenozoic. However, there are clues to this early paleoelevation history that can be reconciled with new qualitative estimates of crustal thickening (Figures 4e and 4h). The Altiplano and western Andean margin appear to have different surface uplift histories (Figure 4h; Boschman, 2021). Before 80 Ma, the northern CAP was likely a neutral plate margin undergoing orogenic stasis and sediment bypass with low elevation, as evidenced by the lack of retroarc foreland basin records (Figures 4a and 4i; Horton, 2018). Localized surface uplift of the CAP western margin starts at ~80 Ma (Figure 4h), consistent with 80–55 Ma zircon trace element data which show an increased range of crustal depths associated with different parts of the crustal magmatic column (Figures 3 and 4b). Although marine rocks now in the orogenic hinterland require that at least parts of the northern CAP Altiplano be at sea level in the latest Cretaceous (Figure 4h; Sempere et al., 1997), retroarc foreland basin stratigraphy suggests the development of a Western Cordilleran crustal load of thickened crust and moderate elevation between 60 and 25 Ma resulting in increased sediment accumulation rates associated with the development of a Paleogene retroarc foreland basin system (Figure 4i; P. G. DeCelles and Giles, 1996; Horton et al., 2002; K. E. Sundell et al., 2018).

Late Cenozoic surface uplift in the northern CAP Altiplano may have been partly decoupled from crustal thickening, in contrast to the earlier elevation gain in the western CAP margin (Figure 4h) and Puna region of the southern CAP (Canavan et al., 2014; Quade et al., 2015). For example, phylochronologic analysis of potato parasite nematodes suggests the maximum paleoelevation of the northern CAP between 25 and 18 Ma was no more than half of its current elevation (Picard et al., 2008). Gregory-Wodzicki (2000) similarly required a ~2 km elevation maximum by early Miocene time based on leaf physiognomy, although this study disregards climatic effects. C. N. Garzione et al. (2008) presented δ^{18} O isotopic data that allowed for maximum elevations of 2.2 km between 25 and 10 Ma, which was bolstered by later paleoelevation estimates documented by δ^{2} H analysis of volcanic glass, δ^{18} O and Δ 47 clumped isotope analysis of soil carbonates, pollen assemblages, and leaf wax lipids which collectively produced a maximum paleoelevation of 2.8 km between 18 and 7 Ma (Kar et al., 2016). Regional δ^{2} H stable isotopic analysis of volcanic glass yield a maximum elevation of ~2 km elevation between 25 and 19 Ma (K. E. Sundell, Saylor, Lapen, et al. (2019)). Finally, Monte Carlo based river incision modeling of the Rio Ocoña in southern Peru provide a maximum paleoelevation estimate of ~3 km at 20 Ma (Jeffery et al., 2013).

The transition to hyperarid conditions occurred long after the establishment of moderate to high elevations along the western CAP margin and the attainment of near modern crustal thickness (Figure 4e). Miocene palustrine and lacustrine carbonates and fossil soils place the onset of extreme aridity in the Atacama Desert at 12–10 Ma (Rech et al., 2019), after the attainment of extreme elevation in the Altiplano (Figure 4h). Similar evidence is seen slightly earlier in hinterland basin stratigraphy of the Peruvian Western Cordillera where a nearly bimodal switch from dominantly fluvial to gypsum-bearing lacustrine-evaporitic deposition takes place at ~18 Ma (Figure 4i; Carlotto, 2013). This is in contrast to paleoclimate models which predict that the opposite took place: that the CAP became wetter, not drier, due to surface uplift (Chandan & Peltier, 2017; Ehlers & Poulsen, 2009), which are inconsistent with earlier climate modeling work (e.g., Garreaud et al., 2010) and recently documented fossil records of pollen, leaves, and wood (Martínez et al., 2020).

Despite the protracted history of crustal thickening since \sim 80 Ma and maximum zircon trace element ratios reaching near modern values by \sim 30 Ma (Figures 4e and 4g), the northern CAP did not reach near modern elevations until the middle to late Miocene (Figure 4h), implying that these processes were in part decoupled in the late Cenozoic. Alternatively, if our preferred hypothesis is incorrect and results indeed suggest that the crust thickened in two separate stages, with a period of crustal thinning in between, results might indicate several periods of uplift and subsidence. In either scenario, the apparent inconsistences in paleoelevation we observe may be related to the complex effect of global climate change, at least in Miocene, and warrants further research on stable isotopes used as paleoelevation proxies (e.g., Carrapa et al., 2019).

Data Availability Statement

The geochemical data used in the study are available from the authors upon request and at Zenodo via https://doi.org/10.5281/zenodo.5800271.

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