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The role of crustal magmatism in the formation and the evolution of orogenic plateaus

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ABSTRACT

Here we review regional geological, petrological, geochemical, and geophysical data from the Central Andean plateau (comprising two distinct areas, the Altiplano and Puna) and its defunct North American sibling, the Nevadaplano, as well as to a lesser extent Tibet, to show that magmatism in the middle crust of orogenic plateaus plays an important role in the development of relief and high elevation and drives the mechanical behavior of these features. We show that in situ, mostly S-type melting is an intrinsic feature of plateaus and that these melts can reside in the crust for millions to tens of millions of years without freezing just below a critical melt fraction. Some of these in situ partial melt masses escape their source regions and erupt at the surface. The chemical evolution of volcanic masses produced by partial escape from these migmatites can be predicted by simple forward petrologic calculations. Extensive migmatite in sub-plateau middle crust also contributes to the mechanical weakness of plateaus and their buoyancy. In addition, magmatism from the underlying mantle wedge can further sustain the life of these extensive partial melting zones. Additional, much deeper melt accumulation zones exist close to the bottom of these thick-crusted domains, but we have much less information about their evolution and chemistry; they are probably similar to the deepest crustal root zones of the frontal arcs.

We also suggest that there is a link between the ability of Li, B, Cs, and other elements to be concentrated in S-type leucogranites of the extensive migmatite blanket of the middle crust and the concentration of Li (and other metals) in brines and ultimately in the salars of the Central Andean plateau. Hydrothermal activity directly related to the present day Altiplano-Puna magma body or the recent equivalents on South America's plateau may have leaked these incompatible elements into the former lakes that are now hosting the largest concentrations of Li on the planet.

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1. INTRODUCTION

Low-relief, high-elevation orogenic plateaus are some of the most intriguing morphotectonic features in orogenic belts (Isacks, 1988; Lamb and Hoke, 1997; Oncken et al., 2006; Strecker et al., 2009; Kapp and DeCelles, 2019). Not all major modern orogens have plateaus, but when they do form (e.g., in the Andes and Tibet) they contain the highest average elevations on the planet and are underlain by the thickest continental crust, as much as 85 km in places. There are also oceanic plateaus, which could be sites of former hot spots; these are not discussed here. The big continental plateaus, those formed in the same general areas where rugged topography forms nearby, are the subject of this article. What makes a giant plateau such as the Central Andean plateau form at convergent margins, how long do they last, and why do they take a low-relief shape, often accompanied by internal drainage? Can we identify major plateaus of past orogenic belts long after the topography is no longer there? Did such features exist in the earlier tectonic evolution of our planet? Is plateau formation in any way constrained by regional long-term climate patterns such as aridification behind topographic barriers? Do specific mineral resources form while an orogenic area is under plateau-like condition? A large body of literature exists on the topic, and these and other questions can be addressed from a multidisciplinary perspective.

In order to answer these questions and others, we investigate the underside of modern plateaus to understand the metamorphic and magmatic response of plateau formation in conjunction with surface features. A good complementary view of surface versus "at depth" processes can be gained from studying the modern central Andes in conjunction with its defunct geologic equivalent in the North American Cordillera. This is precisely what we present in this paper: We compare surface geologic data and at-depth geophysical data from the modern Central Andean plateau with the ancient Nevadaplano (DeCelles, 2004) an ancient (Cretaceous to latest Oligocene) equivalent in North America, which is now collapsed into the Basin and Range and exposed in places to mid-crustal depths. We will review the surface magmatic products as well as the location of partial melting and melt ponding in the Altiplano-Puna plateau. While Tibet is not the central subject of this paper, we augment our review with some data from modern Tibet (Molnar et al., 1993; DeCelles et al., 2007; Quade et al., 2011). The Anatolian plateau (e.g., Çetiner et al., 2025), a fourth important major orogenic plateau today (or recent past), is not further discussed in this paper since we have enough information from that region to strengthen our points. We then address buoyancy and strength of the middle crust in plateaus and show how partial melting that we can geologically docu-

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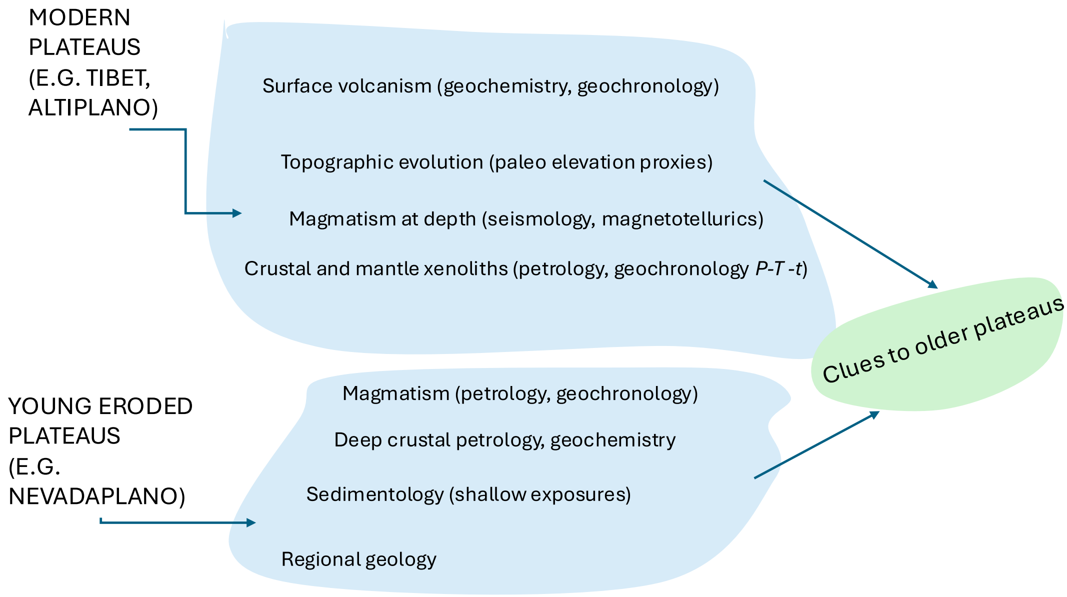

Figure 1. Body of evidence for plateau geology; different types of geological and geophysical data investigated here to better understand magmatism under orogenic plateaus. MT—magnetotellurics; *P-T-t*—pressure-temperature-time.

ment for the Nevadaplano matches well the surface magmatic products of the modern Central Andean plateau. We speculatively make a connection between magmatism under plateaus and the accumulation of Li and other metals in the shallow crust of these domains. A summary of the body of evidence used in this review is given in Figure 1 with the most important references listed. Finally, we compile a set of magmatic and metamorphic characteristics to be expected to form under long-lived plateaus; they are features that one could use in ancient orogens to look for the existence of paleo-plateaus. The great majority of information gathered in this paper is from previously published work. We only report some new simple forward thermal models to characterize the advancement of a plateau's thermal state over time, as well as some thermodynamic models that predict compositions of melts, residues, their densities, and their viscosities.

2. NEVADAPLANO—THE ANCIENT PLATEAU OF THE NORTH AMERICAN WEST

2.1. Introduction to the Nevadaplano

DeCelles (2004) proposed that the Cordilleran interior—between the frontal continental arc range to the west and the Sevier retroarc fold-thrust belt to the east—became a high orogenic plateau and coined the term Nevadaplano because of the possible similarities to the Altiplano. The Nevadaplano extended from British Columbia, Canada, in the north to Sonora, Mexico, and had a surface expression similar to the modern Altiplano-Puna in the Andes (Fig. 2). Evidence for the existence of this plateau has been subsequently added in numerous papers (e.g., Cassel et al., 2012; Henry et al., 2012; Long, 2012; Chapman et al., 2015b; Yonkee and Weil, 2015; Bahadori et al., 2018; Lipman, 2021). To a first order, the existence of the Nevadaplano coincides with the timing of thickened crust in the Cordilleran hinterland (Coney and Harms, 1984), which was confined to the latest Cretaceous to Miocene (Chapman et al., 2015b, 2020). The exact lifetime of the plateau remains imprecisely constrained in detail, although it clearly existed after the Late Cretaceous and most likely persisted until the collapse into the Basin and Range. Collapse took place diachronously as the former subduction margin was replaced by a transform boundary in the western United States (Atwater and Stock, 1998; Dickinson, 2006), but for the most part, this took place in the latest Oligocene to early Miocene. Chapman et al. (2015b) used chemical mohometry (Luffi and Ducea, 2022) to show crustal thickness in excess of 70 km between 85 Ma and 25 Ma. The modern crustal thicknesses close to an average of 30–35 km throughout the Basin and Range (Potter et al., 1986) combined with geologic evidence that indicates ~100% extension on average (McQuarrie and Wernicke, 2005) bolster the notion that before the onset of extension, the continental crust of the Cordilleran interior had to have been in excess of 60 km over large areas. A double-thick crust during Late Cretaceous–Paleogene time is also consistent with estimates of >300 km of upper crustal shortening in the Sevier fold-thrust belt, which must have sent a slab of North American lower crust and lithosphere beneath the west-

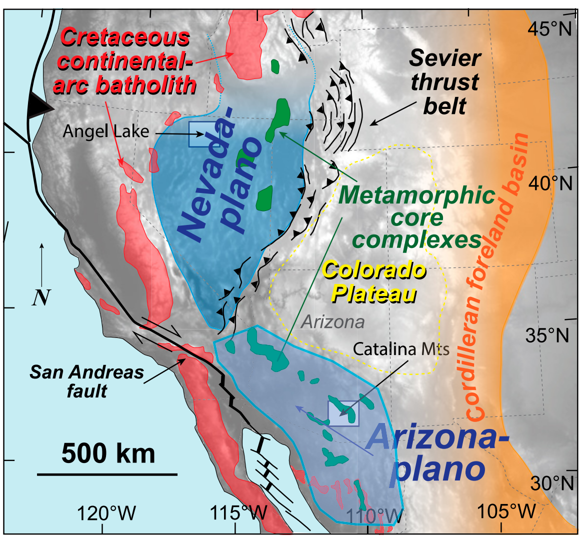

Figure 2. Simplified map of the Nevadaplano and its southern continuation, the Arizonaplano (modified from Chapman et al., 2021). Major tectono-magmatic elements of the North American west during the Cretaceous and younger are shown on the map. Areas discussed in the paper in detail (Angel Lake and Catalina Mountains) are also outlined.

ern Utah and Nevada, USA, hinterland region (DeCelles and Coogan, 2006; Giallorenzo et al., 2018; Long, 2019; Yonkee et al., 2019; Di Fiori et al., 2021). Parts of the plateau correspond to the footprint of the well-known Cordilleran anatectic belt (Chapman et al., 2021, and references therein), which is an extensive zone of in situ partial melting of the local crust (much of which is Precambrian) between the late Cretaceous and the Oligocene

The plateau may or may not be related to the flat slab subduction corridor during the so-called "Laramide orogeny" (Saleeby, 2003), which began diachronously along the strike of the Cordillera between 95 Ma and 75 Ma. An episode of ultra-shallow subduction (Ducea and Chapman, 2018) possibly triggered by the subduction of the Hess-Shatsky conjugate oceanic plateau (Liu et al., 2010) may have led to more coupling between plates, to the widening of the orogenic deformation in the upper plate, and to development of the plateau. Nevertheless, the Nevadaplano remained a high-elevation, low-relief feature (Long, 2012) until the end of major shortening in the Cordillera in the latest Oligocene to Miocene, when the plateau collapsed into the modern, highly extended Basin and Range (Wernicke, 1985). With crustal thicknesses averaging 55–60 km at the height of the plateau existence (Coney and Harms, 1984; Chapman et al., 2015b) the plateau probably had elevations >2 km in the Mesozoic and in excess of 4000 m before basin and range extension. The relatively shallow depth of erosion on the Nevadaplano (Permian–Triassic) suggests that erosion was limited, except along the frontal fold-thrust belt (Gans and Miller, 1983; Van Buer et al., 2009; Konstantinou et al., 2012; Long, 2012; Painter et al., 2014), similar to the Central Andean plateau. Whereas the Central Andean plateau is internally drained, evidence for internal drainage in the Nevadaplano is lacking, in part perhaps due to limited preservation of Cretaceous–Eocene sediment (e.g., Lund Snee et al., 2016; Tye and Niemi, 2024).

The two major plateaus of today, the Central Andean plateau and Tibet are, for the most part, internally drained. It makes sense that a plateau can best survive over time as a relatively flat surface if internally drained. Otherwise, unless it is in an exceptionally arid region, river erosion would carve a hydrographic network that, over time, would alter the stability of a plateau and ultimately destroy it. Despite that, the modern Colorado Plateau, which is not internally drained, shows that in a relatively arid climate, more than 6 m.y. (Flowers et al., 2008) of an extensive erosional network established by the Colorado River and its tributaries have not been able to destroy the integrity of this relatively small plateau. Thus, it is not absolutely necessary to have internal drainage for a plateau to survive over geologic time scales, although an internally drained boundary condition will certainly help stabilize a low-relief surface. Geomorphologic and other geologic evidence (Henry et al., 2012) suggest that the Nevadaplano was, at least throughout the Cenozoic and after the demise of the frontal Sierra Nevada magmatic arc ca. 80 Ma, a high standing region that was higher than the coastal range. This is well documented through extensively studied auriferous gravels brought by Eocene rivers from areas in modern Nevada across the Sierra Nevada toward the Pacific (Cecil et al., 2010; Cassel and Graham, 2011). Additional *P-T-t* paths recorded by the Angel Lake section (see below) show that unroofing was taking place, albeit at a slow pace, in the Nevadaplano. Therefore, slow erosion characterized the Nevadaplano from the Late Cretaceous to at least the Eocene–early Oligocene.

2.2. Magmatism Throughout the History of the Nevadaplano

Data from the compilation effort by the NAVDAT project in the early 2000s (Walker et al., 2006; Ducea and Barton, 2007) showed patterns of volcanism on the Nevadaplano better than any previous synthesis. More geochemical and geochronologic data of volcanic rocks from the Cordilleran interior have been published since, but no updated compilation postdates NAVDAT. Earlier ideas pertaining to a regional inboard sweep followed by retreat of magmatism presumably due to the shallowing and later steepening of the Farallon slab during the Laramide orogeny (see the classic Dickinson and Hatherton, 1967) relationship applied to the North American west) do not hold when more recent age data are compiled (Ducea and Barton, 2007). Volcanism is concentrated in rather distinctive domains that then migrate over time without much connection to the distance from the trench. Cenozoic Nevadaplano volcanism was in no way distributed linearly and/or parallel to the trench. Overall, the pattern of Cenozoic volcanism on the Nevadaplano appears very much like the uneven spatial distribution and composition seen in the more recent Central Andean plateau (see Section 6).

Every exposure of footwall rocks in metamorphic core complexes of the Basin and Range that has been exhumed to mid-crustal levels contains abundant granitoids (I type and S type) with ages ranging from 70 Ma to 20 Ma. Intrusive magmatism was active for much of the lifetime of the Nevadaplano. The latest Cretaceous to early Cenozoic age plutons located in the Cordilleran interior, some of which are part of the Laramide copper belt, were intruded into the Nevadaplano.

At certain times, such as during the Paleocene and Eocene, volcanism was scarce throughout the region, but ample evidence exists for intrusive magmatism in the form of S-type leucogranites in southern Arizona, USA (Fornash et al., 2013; Scoggin et al., 2021; Ducea et al., 2020; Chapman et al., 2023) and Nevada (Batum, 2000). Although rhyolites were erupted in the Cordilleran interior while it was a plateau, it is pretty clear that many of the early Cenozoic in situ or slightly migrated migmatitic partial melts that are present in various basement units of the Basin and Range were not able to erupt at the surface, and while they were partially molten bodies for extended periods of time, they provided a significant role in weakening the crust and sustaining a plateau morphology (see discussion in Section 4.2.).

One of the most distinctive magmatic episodes took place throughout the plateau at the end of the Nevadaplano as a high plateau, just as it was transitioning to the extended terrain of today. A flare-up event tentatively started ca. 35 Ma, became well established by 30 Ma, lasted until 22 Ma, and was particularly voluminous between 28 Ma and 25 Ma (Gans et al., 1989). It consisted of a mix of intermediate products (dacites, andesites) and high-silica rhyolites, often erupted in the same region, most of them as extensive ignimbrite fields. This classic ignimbrite flare-up of the Basin and Range is similar to the large eruptions of the past 10 m.y. in the Altiplano-Puna volcanic complex (de Silva, 1989). Over one million km^3 of ignimbrites were formed in the waning stages of the Nevadaplano's existence (Best et al., 2009), making it, together with the Altiplano-Puna volcanic complex, one the largest volcanic provinces of all continental regions during the Phanerozoic.

2.3. Exposure of the Nevadaplano After Basin and Range Extension—The Mid-Crustal View

2.3.1. Angel Lake Section of the East Humboldt Range

The East Humboldt Range in the central Basin and Range of Nevada is one of the few areas where the middle crust of the Nevadaplano is exposed, mainly owing to Cenozoic extension (Fig. 2). The geologic relationships in this area are shown on the Welcome and Wells Quadrangle maps (McGrew and Snoke, 2015). The Angel Lake region is crucial to this discussion. In that region, recent glaciation has created steep relief and great exposures in the footwall of a classic Basin and Range normal fault. The metamorphosed rocks in the footwall have been extensively studied because of their high grade of metamorphism and associated magmatism (Dallmeyer et al., 1986; Hodges et al., 1992; McGrew et al., 2000). Rocks ranging in age from Neo-Archaean metasedimentary rocks (basement of the North American Cordillera) to a variety of igneous rocks of many different ages and variable degrees of metamorphism are known to exist in this section. Metamorphic thermobarometry (Hodges et al., 1992; McGrew et al., 2000) indicates that the section exposed at Angel Lake was subjected to Barrovian metamorphism at pressures as high as 0.9 GPa and temperatures in excess of 750 °C during the latest Cretaceous and early Cenozoic. This peak metamorphism was then followed by relatively slow unroofing consistent with a sub-plateau origin, although surprisingly, no previous studies have identified it as such.

The Angel Lake section core exposes Neo-Archean to Neoproterozoic ortho- and paragneisses followed by Neoproterozoic meta-conglomerates and meta-quartzites, Cambrian to Devonian metamorphosed siliciclastic rocks and marbles, an unmetamorphosed upper Paleozoic passive margin suite, and younger intrusive rocks (Fig. 3). The intrusive rocks comprise Cretaceous to early Miocene leucogranites and pegmatitic granites, making up as much as 60% of most outcrops (Fig. 4), middle Eocene amphibole and biotite dioritic orthogneisses, and Eocene to lower Oligocene monzogranitic orthogneiss. It is not always clear whether the leucocratic veins, sills, and dikes are in situ melts (for which there is evidence in most places; Hallett and Spear, 2014) or whether they migrated into this crustal position from adjacent or subjacent regions. These are classic two-mica and garnet-bearing granitoids with perthitic structures and large K-feldspars. There are three distinct generations of S-type melts, although all show some sign of ductile deformation. Many are true granites, but on average they are actually granodioritic (Batum, 2000; Drew, 2013). The entire package is heavily deformed into large nappe-like recumbent folds demonstrating that ductile deformation and metamorphism took place as late as the latest Oligocene. Middle Miocene (ca. 18–17 Ma) unmetamorphosed and undeformed basaltic dikes crosscut the sequence described above. These basaltic dikes were synchronous with or slightly postdated regional “Basin and Range” extension that made it such that all the units described above are exposed in the footwall of a detachment fault system.

Figure 5 shows the Cretaceous to Miocene pressure-temperature-time (*P-T-t*) path of various rocks in the Angel Lake section. Prior to extension-related unroofing at ca. 20 Ma and after, the section was as deep as $\sim$35 km in the middle crust of the Nevadaplano. Between ca. 85 Ma and 45 Ma, the temperatures stayed between 600 °C and 800 °C, while the section underwent some 15 km of unroofing. This is consistent with the observation that the Nevadaplano, unlike the modern Central Andean plateau, was not entirely an internally drained plateau. (Eocene rivers were flowing from present-day Nevada west toward the Pacific, as documented by the auriferous gravels of California, USA, e.g., Cecil et al. [2010, and references therein].) However, this should not be taken to indicate that the unroofing was due to surficial erosion, because unmetamorphosed Pennsylvanian and Permian strata are well preserved in the hanging wall of the steep Basin and Range normal fault just east of Angel Lake (Fig. 3).

The lessons learned from the Angel Lake mid-crustal section through the Nevadaplano are that the plateau crust was hot $>$30 km beneath the surface. This hot crust was prone to partial melting and indeed contains large proportions of S-type partial melt that resided in the section for as long as 50 m.y., extending back to the time of high temperature on the *P-T-t* path. These S-type melts are not always granitic ($>$70% SiO_2), but are more often than not granodioritic (dacitic) in major elemental compositions. Additionally, more mafic bodies intruded the section from below (quartz-diorites and monzodiorites). Ductile deformation persisted to the end of the plateau’s life, which in this case coincided with extensional core complex formation leading up to the onset of Basin and Range brittle normal faulting. Additionally, *P-T-t* data and regional geology suggest that in this case, the plateau may not have been internally drained. However, the mere existence of unmetamorphosed Permian strata in the hanging wall of the eastern range-bounding normal fault requires limited erosion even during the heyday of the plateau and the great thickening that is inferred in the Angel Lake section.

2.3.2. Southern Arizona Metamorphic Core Complexes

The North American orogenic plateau extended into Arizona and continued to the south in modern Mexico (a geographic part of the plateau sometimes referred to as “Arizonaplano”; Jepson et al., 2022). The mylonitic fabrics in the metamorphic core complexes of southern Arizona and southeastern California are generally considered to be the result of ductile deformation during a high-magnitude extensional event that took place during the Oligocene to Miocene (Davis and Coney, 1979; Spencer et al., 1995; Dickinson, 1991). For example, many of the primary penetrative mylonitic ductile fabrics in the Catalina Mountains metamorphic core complex north and east of Tucson, Arizona, were generated during the Eocene, 48–46 Ma (Ducea et al., 2020). Structural restoration of Oligo-Miocene block rotations suggests that these mylonitic fabrics (foliation) were originally sub-horizontal to shallowly dipping (Spencer et al., 2022; Wong et al., 2023). The older fabrics in these core complexes generally formed concurrently with the intrusion of spatially extensive anatectic leucogranite, and the development of the fabrics may have been related to crustal weakening due to emplacement of magmatic material (Ducea et al., 2020). For example, in the Catalina Mountains, the older fabrics formed during intrusion of the Eocene-age Wilderness Suite and in the Harcuvar Mountains, the fabrics formed concurrently with the Late Cretaceous to Paleogene Tank Pass granite and associated leucogranite (Fornash et al., 2013; Ducea et. al., 2020; Wong et al., 2023). These anatectic leucogranites are volumetrically significant and commonly intrude as dike and sill networks (Fig. 6; Chapman et al., 2021). In the Catalina Mountains, there are at least eight major leucogranitic sills and innumerable smaller dikes and sills, which have a cumulative thickness $>$4 km (Scoggin et al., 2021). The Wilderness suite intrusive bodies crop out in various forms of concordance with the primary ductile fabric (Fornash et al., 2013; Fig. 7). The anatectic leucogranites exposed in southern Arizona did not form in situ as migmatites, but rather are interpreted to be derived from migmatitic sources deeper in the crust (Chapman et al., 2021). The leucogranites are presumed to have intruded at a level of neutral buoyancy in the middle crust and, as in the East Humboldt Range, are exposed at the surface today primarily as a result of late Cenozoic tectonic exhumation related to core complex formation and Basin and Range normal faulting. Thermometry and textural data indicate that the older mylonitic fabrics formed at subsolidus conditions (Ducea et al., 2020; Wong et al., 2023) and limited thermobarometry data from the anatectic leucogranites suggest they intruded at $\leq$15 km depth in the crust (Anderson et al., 1988). This range of depths is consistent with the non-migmatitic nature of the host rocks; the depths were simply too shallow and the temperatures too low for the local crust to be partially molten at the current levels of exposure. The origin of the material that underwent partial melting to form the anatectic leucogranites in southern Arizona is not entirely resolved but highly Sr radiogenic isotopic data and inheritance of ancient zircons indicate that the primary source was Proterozoic basement, including Proterozoic metasedimentary rocks (Keith et al., 1980; Farmer and Depaolo, 1984; Fornash et al., 2013; Chapman et al., 2021, 2023; Scoggin et al., 2021). For example, highly

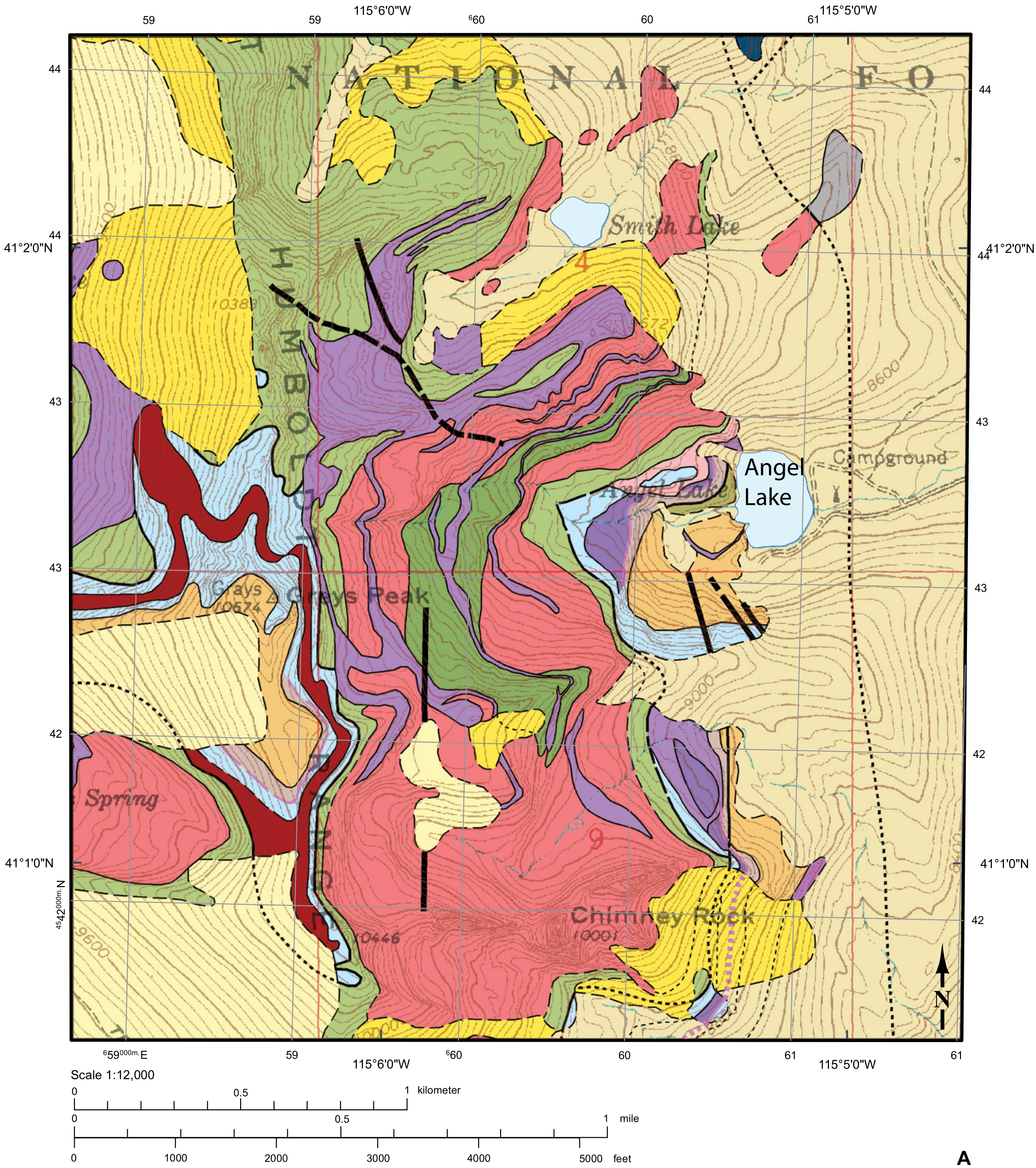

Figure 3. (A) Geologic map of the Angel Lake section of the East Humbold Mountains (simplified from McGrew and Snoke, 2015). (B) Legend to map.

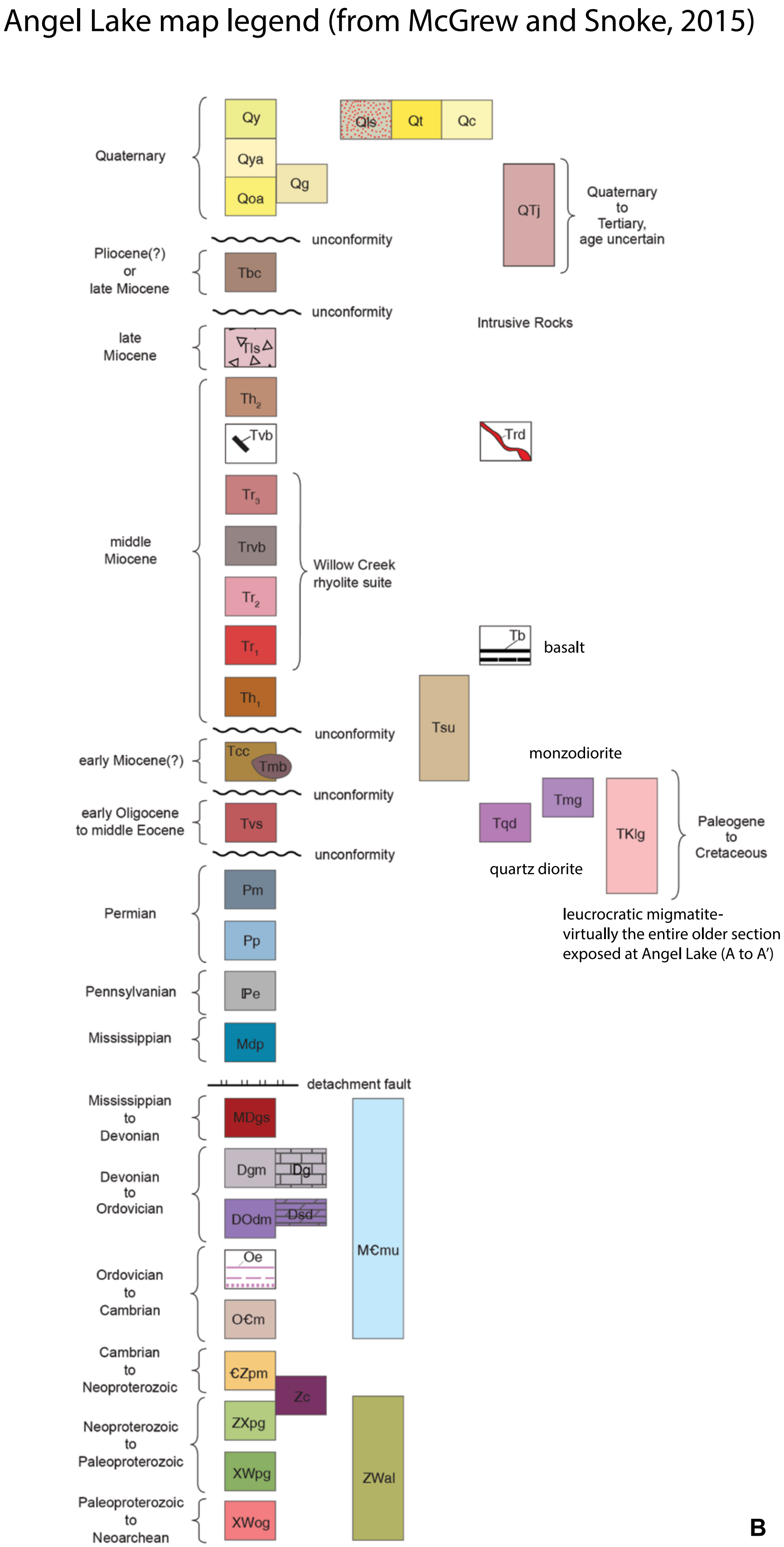

Figure 3. (***Continued***)

radiogenic $^{87}Sr/^{86}Sr$ isotopic ratios from the Wilderness Suite have been interpreted to be produced by partial melting of the Mesoproterozoic Oracle granite and Paleoproterozoic Pinal Schist (Keith et al., 1980; Fornash et al., 2013; Scoggin et al., 2021).

Another feature of the anatectic leucogranites in southern Arizona is that they contain zircon populations indicating crystallization was occurring over protracted periods of 10–20 m.y. (Fornash et al., 2013; Ducea et al., 2020; Scoggin et al., 2021; Chapman et al., 2023; Wong et al., 2023). Zircon U-Pb crystallization dates from the Wilderness Suite range from 55 Ma to 34 Ma (Biegel et al., 2023). Thermochronological and petrochronological data show minor cooling prior to ca. 40 Ma concomitant with ∼10 km of crustal thinning, which has been interpreted to represent crustal thinning of a proto plateau (the Arizonaplano) via ductile mid- to lower-crustal flow, and limited upper crustal extension prior to detachment faulting and Basin and Range extension; significant exhumation occurred at ca. 26–19 Ma during core complex formation and at 17–11 Ma during Basin and Range extension (Jepson et al., 2022). The elevated temperatures and protracted crystallization ages support the hypothesis that southern Arizona was under an orogenic plateau with thick crust for much of latest Cretaceous to Paleogene time (Chapman et al., 2020, 2021, 2023; Jepson et al., 2022). For example, steadily increasing Ti-in-zircon temperatures (650–850 °C) from a Paleocene to Eocene age anatectic leucogranite in the Coyote metamorphic core complex in southern Arizona were interpreted by Chapman et al. (2023) to represent regional prograde metamorphism in response to radiogenic heating and isothermal relaxation associated with crustal thickening and construction of an orogenic plateau.

Like elsewhere in the Nevadaplano, there is evidence from southern Arizona and southeast California that limited erosional unroofing was taking place (Abbott and Smith, 1989), suggesting that the plateau was not completely internally drained. Limited igneous and metamorphic barometry and structural reconstructions suggest that the current exposure levels in the Catalina, Harcuvar, and Whipple core complexes were at 10–20 km depth during the Late Cretaceous to Eocene and at 5–10 km depth at the end of the Oligocene (Anderson et al., 1988; Ducea et al., 2020; Wong et al., 2023), corresponding to relatively low exhumation rates of 0.2–0.3 km/m.y. before extension. Relatively low exhumation rates are consistent with the widespread presence of unmetamorphosed Paleozoic to Mesozoic sedimentary and volcanic rocks in southern Arizona outside of the metamorphic core complexes, indicating limited surficial erosion

Figure 4. (A) Venitic migmatite and (B) diatexitic migmatite of the Angel Lake region. Hammer for scale.

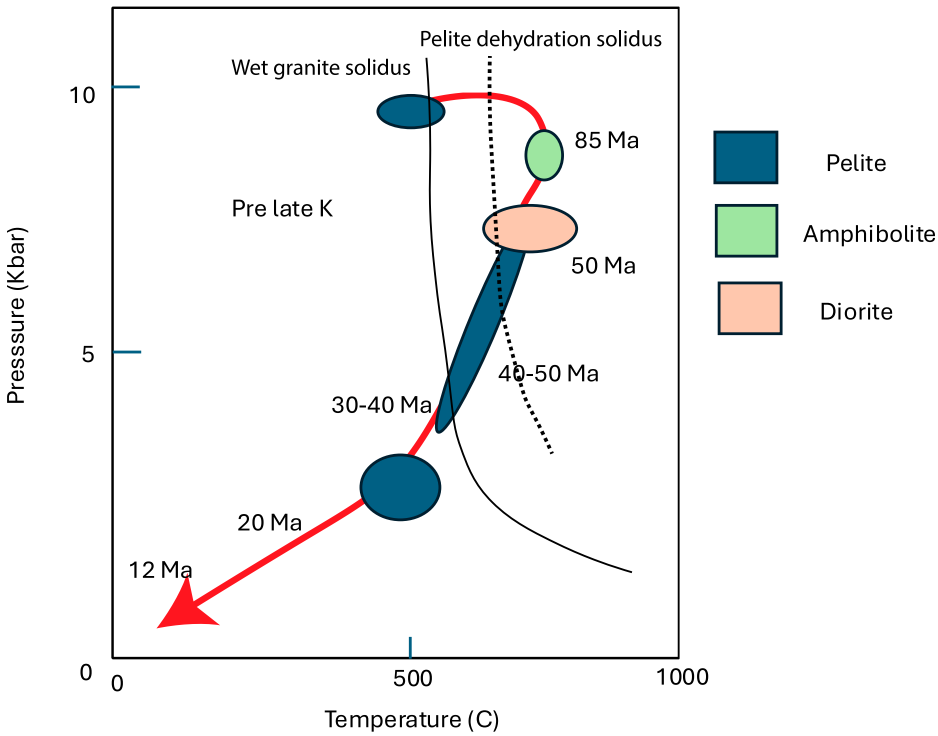

Figure 5. Pressure-temperature-time (*P-T-t*) timeline of magmatic products at Angel Lake. Cretaceous to Miocene *P-T-t* path for the Angel Lake section modified after McGrew and Snoke (2015); metapelite solidus after Pavan et al. (2021); wet granite solidus taken from Philpotts and Ague (2022), with specific references to several experimental determinations therein.

(Chapman et al., 2023). The sub-plateau picture that emerges from the Arizonaplano is that long-lived partial melting in the deep crust produced anatectic leucogranites that migrated into the middle crust, where they existed at depths of 10–20 km and at elevated temperatures for tens of millions of years until they were tectonically exhumed and cooled during late Cenozoic extensional deformation. The presence of long-lived partial melting in the deep crust and extensive anatectic intrusive complexes in the middle crust suggests that the crustal column was significantly weakened, prone to development of a sub-horizontal ductile fabric, and unable to support high topographic relief. Similar arguments have been used to explain the plateau morphologies of Tibet and the central Andes (Jamieson et al., 2011); however, in those two cases the evidence for long-term mid-crustal anatexis is largely inferential because of the very shallow levels of exposure.

3. THE CENTRAL ANDEAN PLATEAU (ALTIPLANO-PUNA)

3.1. The Aggregation of the Modern Central Andean Plateau

The making of the Central Andean plateau (Fig. 8) primarily as a response to crustal thickening is relatively well understood within the overall evolution of the central Andes (Lamb and Hoke, 1997; McQuarrie et al., 2005; Oncken et al., 2006; DeCelles et al., 2011; Garzione et al., 2017). The plateau is internally drained and bound by the frontal magmatic arc to the west (the Western Cordillera) and the mostly amagmatic retroarc fold-thrust belt to the east (the Eastern Cordillera), both of which have higher average elevations. The northern part of the Central Andean plateau, the Altiplano is believed by some to be older than the Puna (Allmendinger et al., 1997). That said, the Puna has thinner crust and higher elevation, presumably because it experienced more mantle lithosphere and/or crustal loss via foundering during the Cenozoic (Kay et al., 1994; Wang et al., 2021). The northern Central Andean plateau too is suspected to have lost its lower crust to foundering during the Cenozoic (Ghosh et al., 2006; Garzione et al., 2008, 2017). However, the Altiplano has a lower than predicted elevation for its crustal thickness and density, presumably supported by a dense mantle root that has not yet experienced a lower crustal foundering event equivalent to the rest of the Central Andean plateau (Ward et al., 2016). Delamination or foundering (Currie et al., 2015) is almost a requirement for the evolution of plateaus because lithospheric mantle is supposed to double (Houseman et al., 1981) to values never

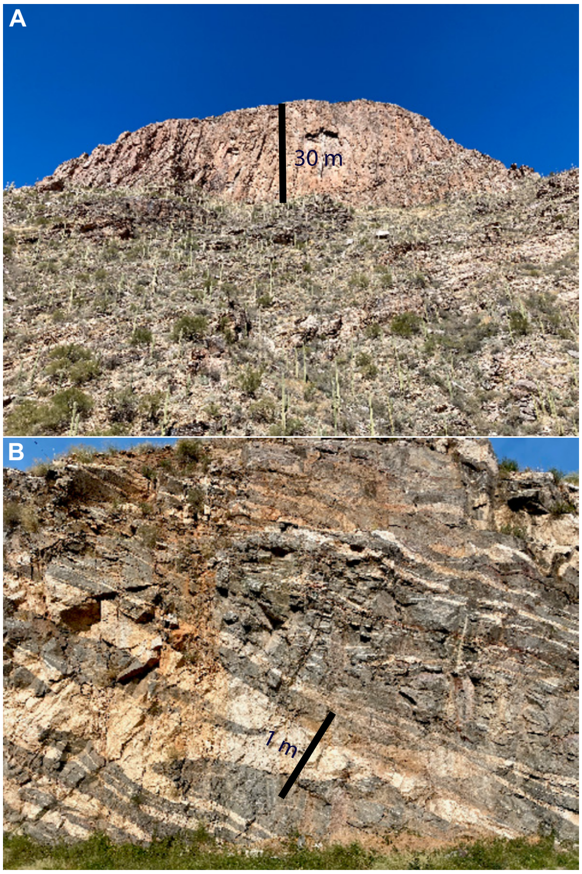

Figure 6. Paleocene–Eocene leucogranite sills of the Catalina Mountains area—from innumerable smaller dikes and veins to concordant large-scale (up to 100 m thick) sills. (A) Large sill in Ventana Canyon. (B) Smaller sills and veins along the Catalina highway both in the Catalina Mountains. Scale shown in both images near or on sills.

seen under plateaus. Thick crust also can lead to eclogitization or arclogitization of the lower crust, which makes the crust unstable. But the size of drips and frequency of dripping are still highly debated, as some suggest that individual drips are not larger than the size of a volcanic field or a large salar (Ducea et al., 2013; DeCelles et al., 2015b; Currie et al., 2015; Wang et al., 2021).

The Central Andean plateau has been associated with crustal thickening for >50 yr (James, 1971; Isacks, 1988; Allmendinger et al., 1997. The crust is as much as 75 km thick under the plateau (Beck and Zandt, 2002; Yuan et al., 2002;). In addition to local shortening of plateau crust, high magnitude shortening from the back-arc side, in the Eastern Cordillera and Subandean zone in the case of the Altiplano-Puna (Kley et al., 1997; McQuarrie, 2002; Oncken et al., 2006; Anderson et al., 2018; Henriquez et al., 2023; Ferroni et al., 2024) feeds lower crust and lithosphere beneath the Central Andean plateau, doubling its crustal thickness. Additionally, up to ∼10 km of sediment accumulation in internally drained areas contributes to thickening of crust and accumulation of upper crustal materials with high heat production (Sobolev and Babeyko, 2005). Magmatic additions from the mantle wedge complete the mass balance of thickening.

The thickening of the Central Andean plateau is most likely progressive and commenced during the Cretaceous, as documented by shortening events in the Chilean Cordillera (Henriquez et al., 2023) as well as whole rock (Garzione et al., 2017) and zircon (Sundell et al., 2022) chemical mohometry. Crustal xenoliths from the Altiplano show that the crust was already quite thick and hot by Eocene time (Chapman et al., 2015a). It is also possible that in the future the plateau will grow more southward at the latitudes of the modern Sierras Pampeanas. Shortening and aggregation of these basement cored uplifts may lead to the southward expansion of the Puna in the future.

A common theme in plateau geology is understanding the role (if any) of flat-slab subduction in making the plateau. Slab dip is variable across the sub-plateau area but generally shallower than ∼30° (Syracuse and Abers, 2006). All areas under the plateau have space for a convective mantle wedge of at least 30 km thickness. To a first order, most researchers do not see an obvious connection between the development of the Central Andean plateau and the steepness of the slab (Isacks, 1988; Kay et al., 2010; DeCelles et al., 2015a).

But the making of the plateau and its specific internal low-relief morphology has to be at least in part, if not exclusively, a result of crustal pro-

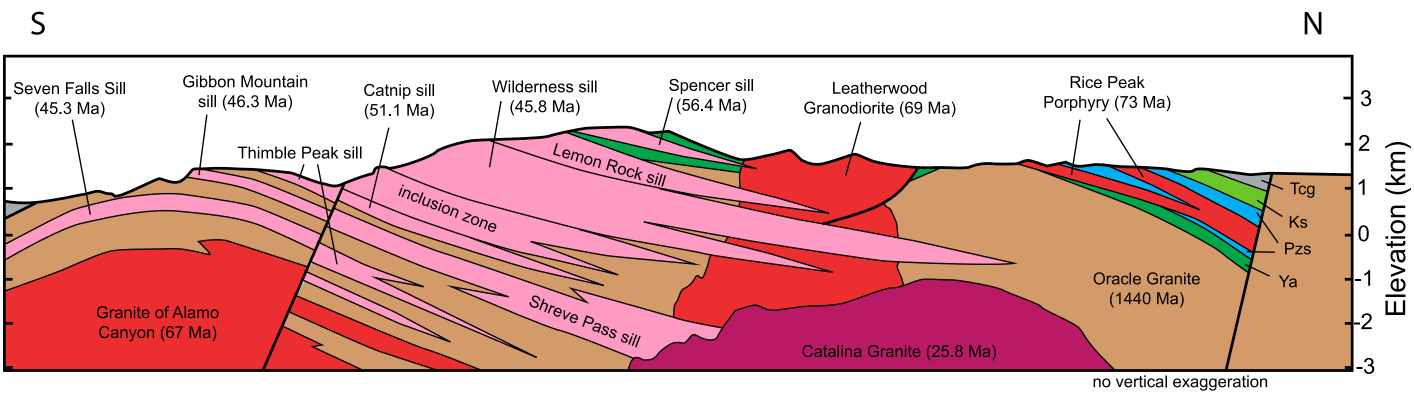

Figure 7. Sills of the Catalina Mountains (from Fornash et al., 2013). This S to N cross section through the Catalina Mountains of the Arizonaplano shows the overall sill-like distribution of magmatic bodies of the latest Cretaceous to latest Oligocene and particularly the sill-like distribution of S-type magmatic bodies in the upper crust of the former plateau. Frontal arc magmatism is dominated by stock-like bodies intruded into each other's and progressively building a vertically textured crust, not a sill-like paleo horizontal fabric.

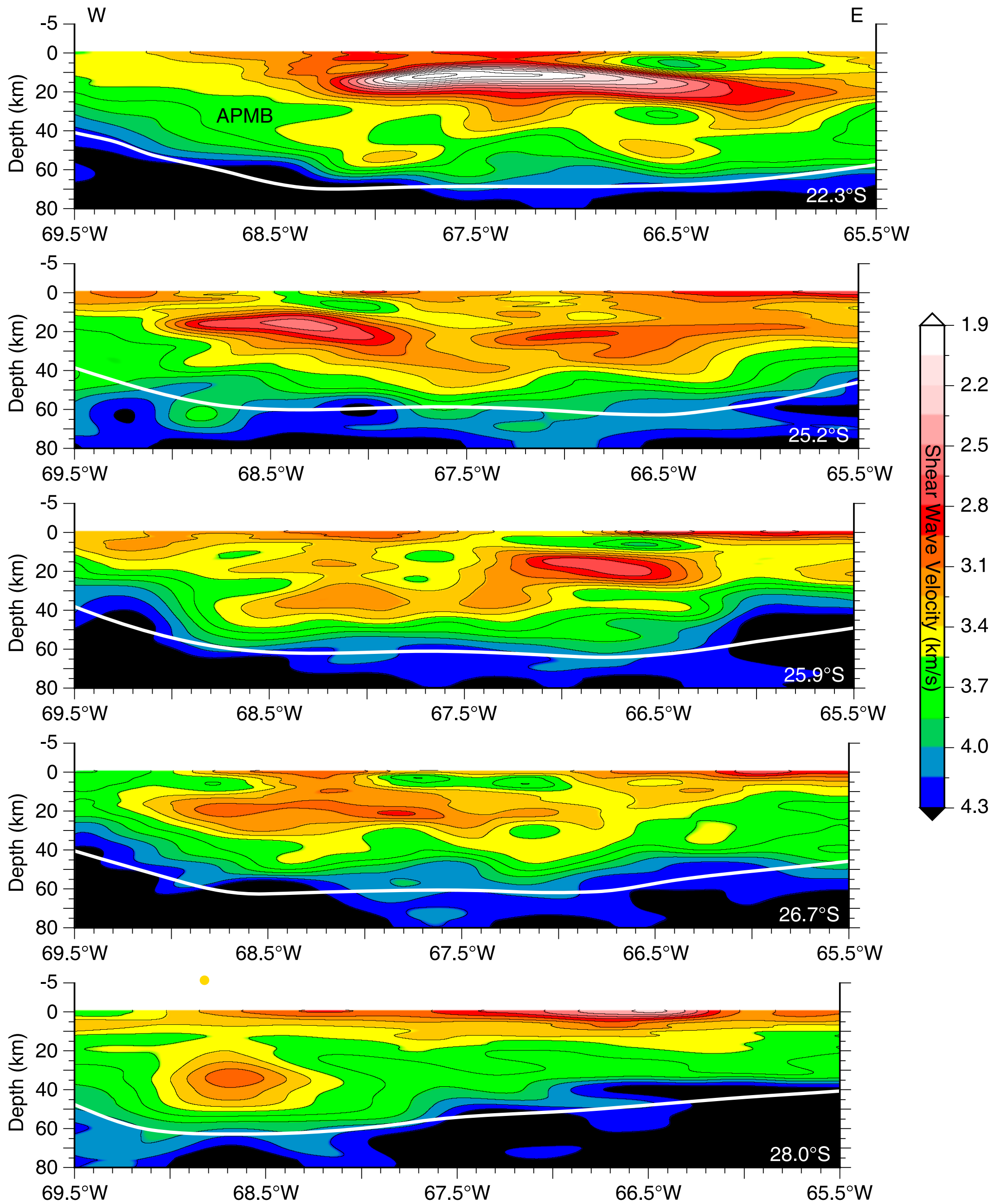

Figure 8. Cross sections through several W-E sections within the Altiplano Puna from 21°S to 28°S showing the extent and depth of various magma bodies (typically where shear velocity is below 2.9 km/s). These are either shallow magma bodies (as shallow as 5 km) or mid-crustal (20–35 km) sill-like bodies with large diameters (simplified from Ward et al., 2017). APMB—Altiplano-Puna magma body.

cesses, and in particular of a weak middle crust (Clark and Royden, 2000), which is postulated in some cases to be partially molten (Jamieson et al., 2011). There is no question that a felsic upper crust will turn into a weak material subject to ductile deformation in a region with elevated heat flow, and this could be reflected in seismic images of low velocity zones or high radial anisotropy below the Altiplano (Yuan et al., 2002); It takes millions to tens of millions of years for radiogenic heat production to raise the thermal gradient (see Section 4.1.).

We also know that the plateau has a number of along-strike high angle faults that were structurally inverted from reverse- to normal-sense multiple times (Schonenbohn and Strecker, 2009). This is most likely because once a certain elevation has been reached, there is a tendency of silicate rocks typical of the continental crust to begin to "collapse" (Dewey, 1988). Continuous shortening imposed by the boundary condition represented by the subduction margin of South America creates the need to constantly raise the elevation. These two opposing factors led to the multitude of fault reactivations along the southern Puna and may turn many of the local depocenters into bobber or delamination-related basins (e.g., Arizaro-type) subject to local removal of crustal/lithospheric drips (Drew et al., 2009; DeCelles et al., 2015b; Andersen et al., 2022). Collapse and local extension and widening of the plateau can be accommodated by the migration of the fold-thrust belt eastward into the foreland. If this is what controls the relatively uniform average elevation of 4000 m of the plateau, it can also explain why plateaus become wider over time (Molnar and Lyon-Caen, 1988). The plateau is making space for itself to stay at 4 km elevation as shortening accumulates to larger values over time.

The western boundary of the plateau is the frontal magmatic arc of the central Andes. An interesting feature at the latitudes of 23°–25°S is that the modern frontal arc, which is identifiable as a line of young volcanoes, appears to have stepped eastward into the Altiplano plateau, as many magmatic arcs do where the subduction margin is erosive (Ducea and Chapman, 2018). By doing so, the arc essentially moved a fragment of the former Altiplano, the Salar de Atacama basin of Chile, into a forearc position along the subduction margin. Part of the plateau has thus been moved out of the internally drained and high heat-flow setting and transferred into the cold forearc region.

The plateau must be underlain by a variety of basement rocks, and it is well known from its surface exposure that the Arequipa-Antofalla terrane cuts through it as well as other smaller basement domains (Ramos, 2008). Volcanic and intrusive rocks of the Famatinian and Pampean arcs are also found at the surface (Ducea et al., 2015c). The most spatially extensive sedimentary unit in the region is the Puncoviscana Formation, an Ediacaran to Cambrian siliciclastic sequence that crops out over large areas in central south America (Pearson et al., 2012). The Puncoviscana was a peripheral Pampean basin made of a variety of first-cycle sedimentary rocks, of which perhaps the most dominant is a hydrous (meta)pelite that is highly susceptible to melting at mid-crustal levels (Zimmermann, 2005). It was found as deep as 25 km in the Sierra Valle Fertil (as migmatitic granulite; Otamendi et al., 2012) without evidence of thickening in that section at any time in the past (Cristofolini et al., 2012). Without a doubt, rocks of the retroarc fold-thrust belt to the east are present under the plateau. Chapman et al. (2015a) show that in the northern plateau, metasedimentary rocks equivalent to the Triassic Mitu Formation of the Eastern Cordillera are found at 30 km under the plateau (sampled as crustal xenoliths in Quaternary volcanics). Precambrian basement of the South American craton (Ramos, 2008) is also inferred to exist but is not unambiguously found in xenoliths. Andean age (that is, as recently as Quaternary) mafic and intermediate igneous rocks from the mantle complete the most likely rocks to be found at depth.

Seismic surveys performed regionally within the Central Andean plateau over the past few decades cannot distinguish specific lithologic units in their interpretation, despite significant recent improvements (e.g., Beck et al., 2023) compared to a couple decades ago (Yuan et al., 2002; Beck and Zandt, 2002). These studies, however, do point out a rather felsic composition for at least the upper part of the crust, which is similar to what we know from the Nevadaplano from local geology. This is also consistent with interpretations of modern Tibet (Owens and Zandt, 1997; Klemperer, 2006; Hetényi et al., 2011). All these findings bolster the idea of in situ melting in the middle of plateau crusts.

3.2. Melting Within the Altiplano-Puna

Melting within the Central Andean plateau region is extensive (Fig. 9). Melt appears to be found at 3 major levels in the lithosphere:

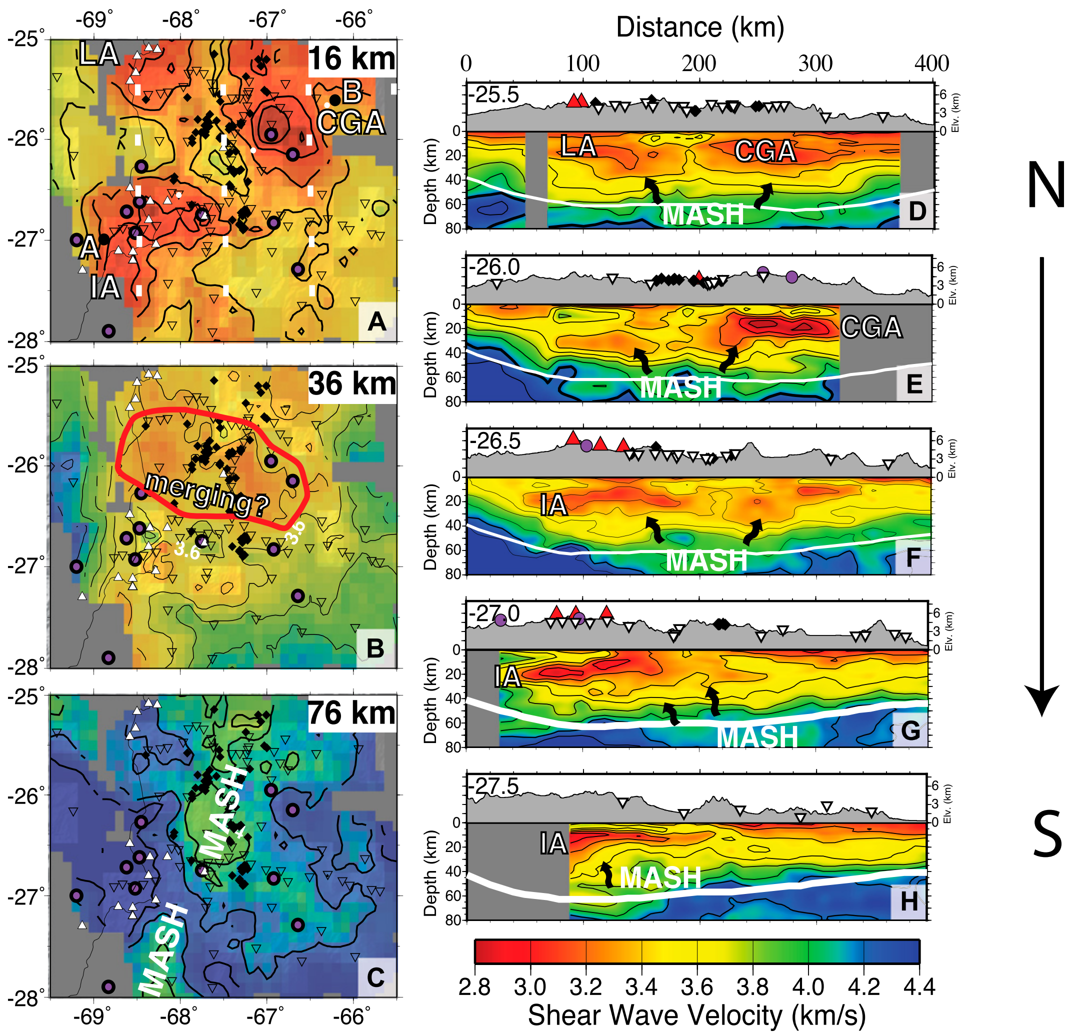

Figure 9. S wave velocity anomalies in the crust at different depth (from Delph et al., 2017). This figure shows how plausible magmatic bodies interact at different depths and also shows the deeper crustal magmatic reservoirs ("MASH zones" probably) close to the crust mantle boundary (shown in bold white line). CGA—Cerro Galan Anomaly; IA—Incahousi Anomaly; LA—Lazufre Anomaly; MASH—melting, assimilation, storage, and homogenization.

within the "MASH" zone (melting-assimilation-storage-homogenization) near the Moho where melts comprise additions from the mantle mixed with local lower crust, at mid-crustal levels (~10–30 km) where in situ melting of the crust dominates, and locally in the upper crust at <10 km below volcanic centers (Kay et al., 2010; Risse et al., 2013; Comeau et al., 2015; Ward et al., 2017; de Silva and Kay, 2018; Delph et al., 2021; Unsworth et al., 2023). Surface volcanism, particularly in ignimbrite fields, is massive and represents a major late Miocene to Quaternary flare-up (Fig. 8). Some of these crustal magmatic reservoirs escape to the surface as volcanic products. As soon as the middle crust partially melts, the strength of the plateau crust diminishes abruptly and remains weak for the entire time the in situ partial melting (migmatite) continues to exist in the source zone. However, if and when that melt erupts, the residue of melting will be a relatively dry, hot assemblage that may or may not deform ductily depending on the composition of the residue (Kidder and Ducea, 2006; Bowman et al., 2021). As long as some melt is retained in the middle crust, the strength of the crust will be very low (Jamieson et al., 2011).

Currently, there are several inferred magma reservoirs under the Central Andean plateau from geophysical studies. Below the Altiplano-Puna volcanic complex, Chmielowski et al. (1999) and Zandt et al. (2003) found a widespread low velocity zone using receiver functions that coincides with previous refraction results and conductivity anomalies (Schwarz et al., 1994; Wigger et al., 1994); they called this anomaly the "Altiplano-Puna magma body." The addition of new data and methods refined the spatial scale of this low velocity zone to be ~200 km in diameter and at least 11 km thick (Fig. 10), with evidence for at least some amount of melt extending from 4 km to 25 km depth (Fig. 10; Ward et al., 2014, 2017).

The oblate shape and spatial extent of these low velocity zones is consistent with a zone of in situ partial melting (migmatite) since there are no 200-km-diameter sills or magmatic bodies (or even 50 km, as seen in the Southern Puna) found in the geologic record. This partial melting zone in the Altiplano-Puna has clear contributions from deeper, more mafic magmas as seen from the geochemical signatures of erupted melts, typically being ~1:1 contribution of crustal versus mantle material (Kay et al., 2010; Freymuth et al., 2015; Delph et al., 2021), similar to what is documented in the Angel Lake section of the Nevadaplano. Melting of clay- and lithic-rich sandstones (a.k.a. "graywackes," described in Section 4.1.), which are the most likely candidates to form the mid-crust under the plateau

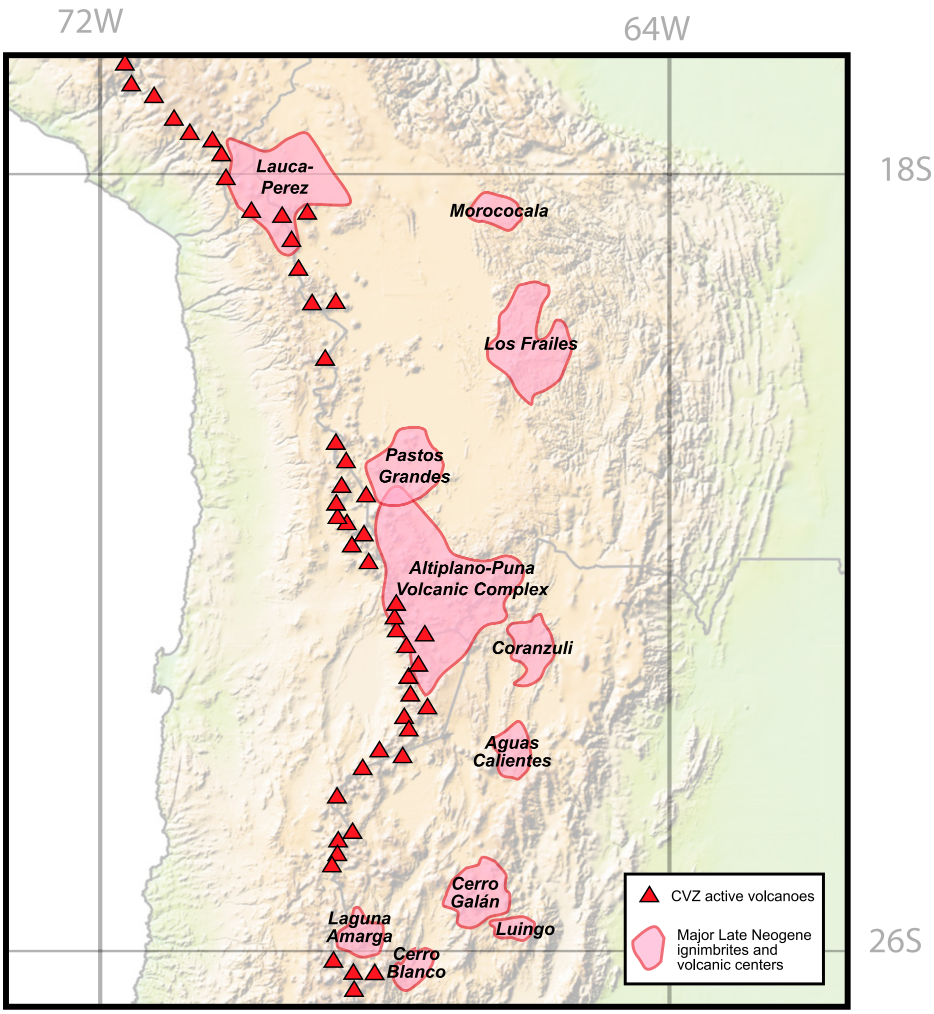

Figure 10. The general position of the central Andean plateau within the central Andes. Map also shows the architectural style of frontal arcs versus plateau volcanism shown on the digital elevation map of the central Andes. While the frontal arc follows an arcuate line of stratovolcanoes that are spaced at variable distances but typically no more than 50 km, plateau magmatism is spread out throughout the plateau and often comprises large ignimbrite fields (shown in figure), although stratovolcanoes exist on the plateau as well. Some of the spread in magmatism is due to the extension of ignimbrite fields far from their source but the occurrence of linear array of frontal volcanoes is distinct from the spatially randomness of the plateau volcanism. CVZ—Central volcanic zone.

(McLeod et al., 2013; Chapman et al., 2015a) will generate dacitic to rhyolitic liquids. Any mafic additions from the mantle or lowermost crustal MASH zone (Hildreth and Moorbath, 1988) will not only provide heat to keep this reservoir partially molten but will shift the composition to more mafic. With little ongoing erosion in the Central Andean plateau (e.g., DeCelles and Carrapa, 2023), this reservoir can retain its partially molten state for millions of years or more (Babeyko et al., 2002; Tierney et al., 2016), which is consistent with the imaged magmatic reservoirs' lateral footprints being similar to the >10 m.y. history of the recent Altiplano-Puna volcanic complex and southern Puna magmatic episodes (Delph et al., 2017; Ward et al., 2017). This is also what we see for the analogous crustal section exposed in the Catalina Mountains in the Arizonaplano (Ducea et al., 2020). Evidently, these reservoirs can, at times, evacuate to the surface. The Central Andean plateau is characterized by one of the largest Neogene–Quaternary provinces of homogeneous ignimbrites (Francis et al., 1983; Davidson and de Silva, 1995; de Silva et al., 2006; Kay et al., 2010), together forming the Altiplano-Puna volcanic complex, which is the erupted equivalent of the Altiplano-Puna magma body.

If the comparison between the Central Andean plateau and its North American sister, the Nevadaplano, is valid, we also speculate that the large magmatic systems of the Altiplano-Puna plateau will continue to exist at mid-crustal levels with some punctuated volcanic events until the convergent plate boundary ends its compressional life cycle. At that point, if extensional collapse ensues, the remainder of magma is more readily evacuated to the surface in a flare-up event similar to that of Oligocene age in the modern Basin and Range (Best et al., 2009). A popular model for mid-Cenozoic flare-up in the Nevadaplano region is the removal of the Farallon slab and upwelling of asthenospheric mantle during the transition from a convergent to transform plate boundary that triggered Basin and Range extension (Atwater and Stock, 1998; Humphreys et al., 2003; Farmer et al., 2008).

3.3. Volcanic Rocks of the Central Andean Plateau

Several seminal papers have been written on the volcanology, petrology, and geochemistry of the Central Andean magmatism, beginning with the early efforts by Francis et al. (1983) and de Silva (1989) when this gigantic volcanic province was just being identified, to the large databases created in more recent years (de Silva et al., 2006; Kay et al., 2010; Freymuth et al., 2015). There are very few trends that one can observe; in no way does plateau magmatism follow the cyclic geochemical and isotopic behavior of long-lived frontal magmatic arcs (e.g., Ducea et al., 2015a). Overall, most researchers identify a 1:1 mantle to crustal mass contribution to the Altiplano-Puna volcanics (e.g., Kay et al., 2010) with a tendency for mantle input to become slightly larger over time.

Most papers studying Altiplano-Puna magmatism casually call upon subduction-related magmas having formed on the plateau (Ward et al., 2017; Pritchard and Gregg, 2016), although the frontal arc (the main subduction-related arc) is found almost always to the west and forms a well-defined linear belt interrupted only by regions that are temporarily amagmatic. The distribution of plateau volcanoes is quite different and at no time follows a linear belt that is in some form parallel to the trench (Fig. 10). The frontal arc is typically not more than 30–35 km wide at any given time, although it can migrate over time (Ducea et al., 2015b). Plateau volcanics are distributed over a wide area, hundreds of kilometers inboard from the trench with no clear spatio-temporal arrangement (de Silva and Kay, 2018). Compositionally, both of these types of magmatic rocks—frontal arc versus plateau—are broadly similar in that they are very diverse, calc-alkaline, and they exhibit a large range of silica content. In some places, truly frontal arc lavas appear to be spatially remarkably close to plateau volcanic fields (e.g., the Antofalla volcano in the Puna region), making them difficult to distinguish from each other, as there is some overlap. However, in most places, the surface distribution of plateau magmas is rather different, with widespread volcanic centers clustering across the surface of the plateau (Fig. 10). While the composition of these plateau magmas is to a first order also calc-alkaline, covering a wide spectrum of silica concentrations, very much like frontal arc volcanic rocks, there are some differences in detail. First, there are some S-type volcanic rocks, and they are crucial to sustaining the plateau topography, as we argue in this paper. Second, the kinds of I-type volcanic rocks typical of plateaus are uniformly dacitic, including shoshonitic and adakitic varieties in much higher proportions than found in the frontal arcs. Overall, plateau magmatism is more silicic than frontal arc magmatism. If average volcanic rocks in continental arcs straddle the boundary between andesite and dacite (Ducea et al., 2015a), at around 63%–64% SiO_2 in the case of Altiplano-Puna plateau rocks, they are on average close to the boundary between dacites and rhyolites at around 68% SiO_2. Some stratovolcanoes exist on the plateau (e.g., Volcan Tuzgle), but most plateau volcanism is associated with large calderas that erupt homogeneous ignimbrites (e.g., Aguas Calientes). Thus, while not so different in the complexity of volcanic products, the silicic (S-type) element is quite easy to demonstrate—from petrography, geochemistry, and isotopes—as significantly more abundant in plateau magmatism, when compared to frontal arc compositions.

3.4. Pressure and Temperature of the Middle Crust Under the Central Andean Plateau

Mid- to lower crustal xenoliths in Quaternary flows discovered from the northern Altiplano in Bolivia (McLeod et al., 2013) and Peru (Chapman et al., 2015a) provide unique windows into the deeper parts of the plateau. These are, together with small crustal xenoliths found in central Tibet (Hacker et al., 2000), the only direct (petrographic and petrologic) evidence for the compositional-thermal regime of sub-plateau crust. Various *P-T* and age data obtained from these xenoliths in the Altiplano show that the plateau had a thick crust certainly by 45 Ma and probably before that, and that Triassic sedimentary rocks (Mitu Group) that crop out today some 35 km to the east in the Eastern Cordillera have been recently present at a depth of 25 km beneath the western part of the plateau in the form of granulite-grade garnet-bearing gneisses (Fig. 11). The pressure-temperature equilibrations of these metasedimentary gneisses at the time of xenolith entrainment (Quaternary) are between 0.7 GPa and 1.0 GPa and 750 °C and 800 °C, also suggesting that they may have lost some melt from those depths (25–35 km). Monazite U-Th-Pb geochronology records ages between 44 Ma and 3 Ma in these metasedimentary rocks, with that range measured in individual samples. This is another indication that the typical metamorphic rock in a sub-plateau environment is exposed to a long-lived high-temperature regime, with temperature actually increasing over time in internally drained plateaus. Similar to Angel Lake in North America, the sub-plateau temperatures at 30 km depths under a mature plateau are >750 °C making it likely that most fertile, first-cycle siliciclastic sedimentary rocks found at those depths will be subject to partial melting.

The other lesson from McLeod et al. (2013) and Chapman et al. (2015a) is that mafic or even ultramafic xenoliths become the only messengers from depths $>\sim35$ km (1.0 GPa), suggesting that dense basement rocks dominate at depths >35 km. This in turn suggests that even if the temperatures continue to increase at depth, it is not absolutely required to have long-lived partial melts in situ in these mafic lithologies without additional heat and mass influx from the mantle (mafic melts in a deeper MASH zone, see next section). The presence of spinel peridotites with mantle mineral chemistry from the Southern Peru xenoliths may additionally indicate that the sub-plateau lithosphere could have crust-mantle imbrications as often suggested by seismic images of the central Andes (e.g., Beck et al., 2023).

3.5. The Deeper Melt Zone of the Altiplano-Puna

Recent seismic studies of the Altiplano-Puna crust and shallow mantle are showing that a second zone of magma generation and/or accumulation exists at greater depths in the crust, near the crust-mantle boundary (Delph et al., 2017, 2021; Ward et al., 2017). This is more like a MASH-zone of the frontal arc except that, as is the case with shallower plateau magmas, these magma-bearing and probably magma-rich reservoirs are laterally widespread, with horizontal diameters of as much as 150 km and vertical dimensions of 10–15 km (Fig. 9). In contrast, the footprint of a MASH zone from the frontal arc rarely exceeds 30 km in diameter (Ducea et al., 2015b).

What are these deeper melting zones made of and why do they exist under orogenic plateaus?

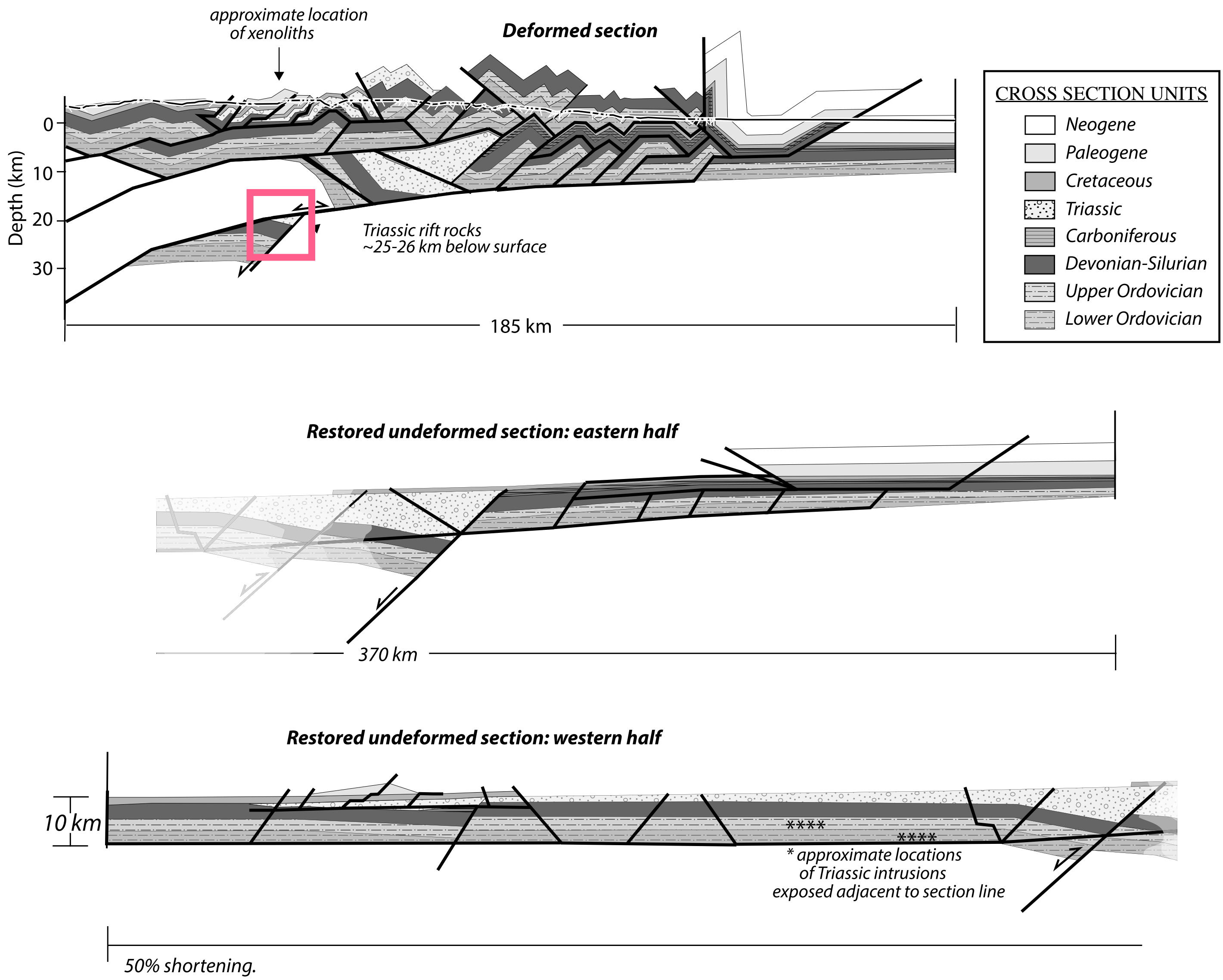

Figure 11. Cross section through the northernmost Altiplano plateau and the eastern Cordillera at the latitude of 14°S showing deformed retroarc fold and thrust belt and undeformed restored section (from Chapman et al., 2015a). The location of northern Altiplano xenoliths in the modern (deformed) section is where supracrustal metasedimentary rocks were found in abundance from the sub-plateau region. Structural restoration of the eastern Cordillera (performed at this latitude by Nadine McQuarrie for Chapman et al., 2015a) is consistent with the existence of such rock in the mid crust of the Altiplano plateau. Moreover, one can make specific predictions as to which rock units of the Eastern Cordillera are present at depth as granulite facies equivalents.

As with the shallower magmatic reservoirs, some in situ partial melts of the lower crust are expected, given the high ambient temperatures (800–900 °C or even more) at 60–80 km depths. In addition, basalts derived by wet melting of the mantle wedge are also expected to contribute to the heat and mass budget (Bowman et al., 2021). Such melts are probably doomed to be captured by the deep crustal partial melt zone, and they probably contribute to the long term persistence of a partially molten domain. Indeed, shear-wave velocities in the lower crust of the southern Puna plateau are consistent with between 1% and 9% of basaltic melt within an ultramafic matrix (arclogite), and P- and S-waves from local seismicity show very high attenuation (Liang et al., 2014; Delph et al., 2021). Very little basaltic melt makes its way to the surface of the Central Andean plateau (Kay et al., 1994; Drew et al., 2009; Ducea et al., 2013); the few examples at the surface appear to be derived from convective removal events locally affecting the deepest crust. This suggests that most mafic material derived from the mantle is being captured and becomes subject to deep crustal MASH zone assimilation and fractional crystallization and mixing (e.g., de Silva and Kay, 2018; Delph et al., 2021). The intermediate I-type volcanic rocks that are common throughout the plateau are probably the result of evacuation events of these MASH zones. The chances of these MASH zone melts to make it to the surface are proportional to the extent to which mid-crustal melting zones do not overlap the deeper crust partial melt areas, as most less-differentiated melts seem to only erupt between or on the edges of imaged mid-crustal magmatic reservoirs despite underlying nearly the entirety of the southern Puna plateau (Delph et al., 2017).

A similar scenario emerges from the Nevadaplano. Eocene and Oligocene I-type "intruders" into the migmatitic and ductile middle crust turn quickly into orthogneisses (McGrew and Snoke, 2015). We do not have a precise view of the magmatic evolution of the Nevadaplano, but it appears that I-type volcanic rocks did form during the Paleocene to Oligocene (navdat.org) but not in areas where a mid-crustal or shallower migmatitic filter existed. For example, in southern Arizona, there is plenty of evidence from core complexes that Eocene–Oligocene melt zones existed (see discussion in Section 2.3.2.) yet neither S-type nor I-type volcanic rocks are known in areas where the shallow crust is exposed today. Instead, the greatest majority of I-type melts migrated toward the surface (to form shallow plutons) or to the surface (to form volcanic rocks) at 28–25 Ma during a well-documented Oligocene flare-up event (Best et al., 2009) that coincided with the collapse of the Nevadaplano as the subduction margin was being replaced by transform faults. At that time and after, I-type melts escaping from the deeper MASH zone were able to cut across the ductile, melt-rich fabric of the shallower migmatite zones as dikes (Ducea et al., 2020). Mafic melts from the mantle maintain the heat budget of the lower crustal MASH zone whereas melts extracted from this MASH zone can further pond, mix, and mingle with the shallower level magmatic systems described above.

4. MID-CRUSTAL MELTING UNDER OROGENIC PLATEAUS—A FORWARD MODELING APPROACH

4.1. Modeling Melting in the Plateau Mid-Crust

The thermal state of a sub-plateau environment is readily predicted by a 1-D model (Turcotte and Schubert, 2014) in which a normal segment of continental crust is thickened to over 70 km and the thermal gradient is allowed to relax over time (Fig. 12). We do this by allowing a thrust system to instantaneously thicken the crust at 40 km below the surface and use the heat flow and heat production data of the Central Andean plateau from the global database of Davies (2013). Surface temperature is 20 °C, surface heat flow is 100 mW/m^2, mantle heat flow is 55 mW/m^2, thermal conductivity is 2.5 W per meter Kelvin, and the length scale for thermal diffusivity is 15 km (the depth at which heat production is a third of what it is at the surface). The bottom boundary is in the mantle, the depth of the model is 120 km, and grid spacing is 2 km. Because there is essentially no erosion for tens of millions of years, the temperatures at any given depth, for example at 25 km and 35 km below the surface (Fig. 12), can be tracked through time. This shows that conditions required for in situ melting in the middle crust of a plateau, even without magmatic heat input from below (magmas migrated from the deeper MASH zone), are met after only 15 m.y. after crustal thickening. The temperature in the depth interval 25–35 km is bound by an interval of ~600–800 °C (Fig. 12). Consequently, all melting forward calculations would need to be conducted within that range of temperatures. Much higher temperatures, in excess of 1000 °C as seen in mid-crustal Tibetan xenoliths (Hacker et al., 2000) require a much longer thermal relaxation time or some unusually high basal (mantle) heat flow.

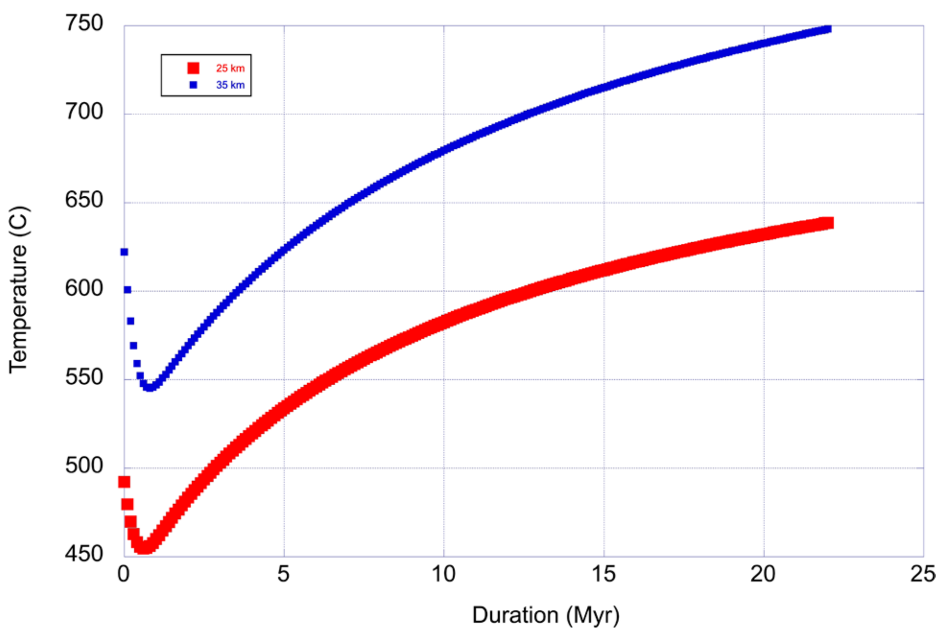

Figure 12. Temperature of a sub-plateau environment at 25 km and 35 km respectively, after thickening of the crust (see text for details) and allowing thermal relaxation for several tens of million years with no erosion. Basal heat flow—55 mW/m^2; surface heat flow—100 mW/m^2; length scale for thermal diffusivity—15 km. The mid crust is predicted to undergo significant crustal melting under plateaus.

One of the most volumetrically important basement units in the central Andes is the Puncoviscana Formation (Zimmermann, 2005) detailed in Section 3.1. It is composed of intercalated quartzite and mudstone at the surface today unmetamorphosed in places (Pearson et al., 2012) but also modified by greenschist facies metamorphism in places as well as amphibolite to granulite facies equivalents in the roots of the Famatinia arc in Sierra de Valle Fertil (Cristofolini et al., 2012). We performed forward melting calculations of the Puncoviscana at the temperatures and depths determined for the sub-Altiplano crust in order to better understand the composition of melts that may form in situ at mid-crustal levels under the Altiplano and to compare these with the intermediate to felsic volcanic rocks of the rest of the plateau (Kay et al., 2010). Results of the forward models carried out with Rhyolite MELTS (Gualda et al., 2012) are shown in Figures 13 and 14. Figure 13 shows major elemental diagrams for melts produced by the average Puncoviscana pelite melted at temperatures of 600–900 °C and pressures of 0.5–1.0 GPa, which are similar to the relatively uniform dacites of the Altiplano-Puna volcanic complex (source: GEOROC, https://georoc.eu/). Residual rocks are dominated by quartz, garnet, biotite, and sanidine. The higher the temperature, the higher the melt fraction, which will result in consumption of sanidine in the melting reaction and a higher concentration of K in the magma (Fig. 14). Overall, in situ melting of a metapelite at mid-crustal depths can alone produce the monotonous ignimbrites of the plateau. Input of mafic magma from the deeper MASH zone may further increase the temperature and decrease the viscosity of these magmas to render them more eruptible.

There is a big difference in the predicted partial melt in the above forward model runs (commonly over 40% at temperatures not exceeding 750 °C) and the much smaller inferred partial melting percentages from ambient noise tomography suggesting no more than 25% partial melt

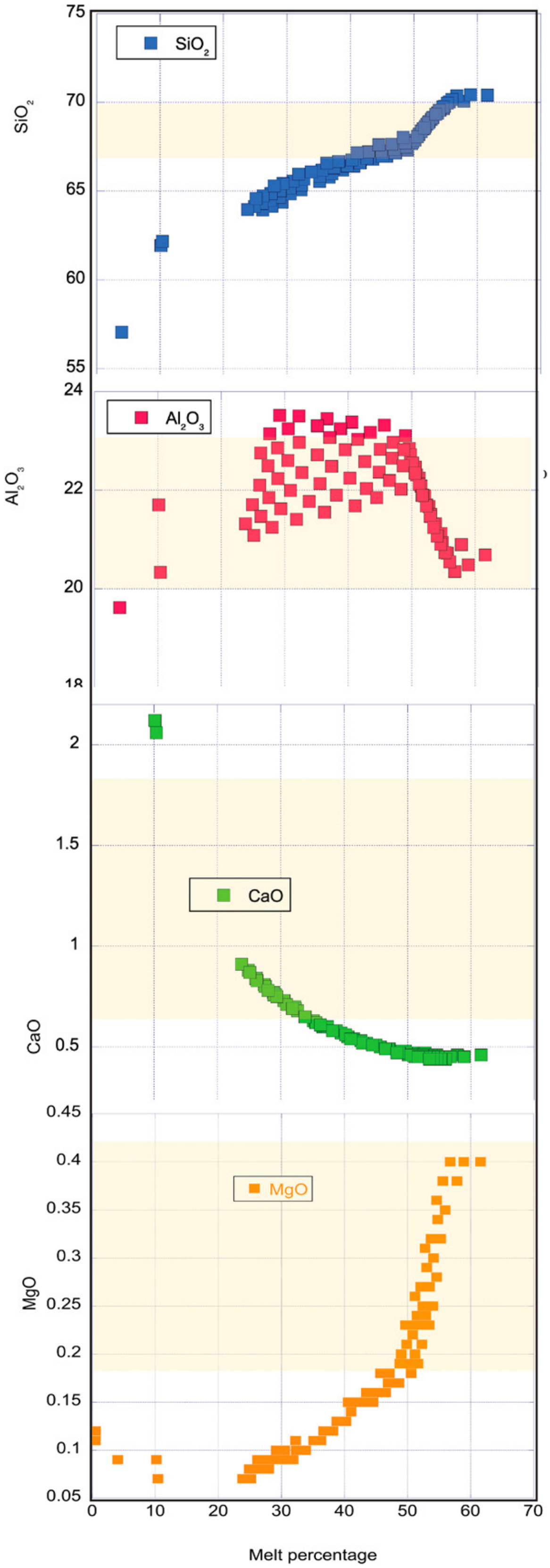

Figure 13. Major elemental compositions of the average Puncoviscana Formation (average of analyses by Zimmermann., 2005) as a function of melt fraction in the temperature range of 600–800 °C and pressures of 5–10 kbar. Compositions were modeled using Rhyolite MELTS. (See text for explanations.)

(Ward et al., 2014; Delph et al., 2017). Moreover, most experimental petrology data predicts that at these high partial melting values, the liquid will be able to migrate upward. It is possible that much of the melt was in fact extracted from the Altiplano-Puna magma body and that is what the Altiplano-Puna volcanic complex represents. It is also possible that the actual upward melt migration velocities are strongly influenced by the contractional state of the crust and not just interconnectivity bounds. Of course, the seismic images also represent the average structure relatively large spatial scales (kms) and thus may not reflect "locally" higher melt percentages at scales below the seismic resolution.

Regardless, each point of the mid-crust beneath a plateau like the Central Andean plateau increases its temperature due to radiogenic heat production and lack of erosion. That creates a scenario in which additional melting can occur after melt extraction from the middle crust (see, for example, the model by Bowman et al., 2021, for step-extraction after 7% melt increments on a graywacke partial melt). The overall progression from more felsic to slightly more mafic magmatism at each location on the plateau (Kay et al., 2010) is consistent with lesser productivity from crustal rocks over time and a steadier state delivery of mafic magmas from the deeper MASH zone.

4.2. Role of Partial Melt on Weakening and Buoyancy of the Crust

The presence of melt in a mid-crustal migmatite at temperatures of over 650 °C further decreases the strength of a quartzo-felspathic material, which is brought to a minimum due to ductile behavior of both quartz (at >350 °C) and plagioclase feldspars (at >600 °C; Kidder and Ducea, 2006). That said, there is no easy relationship between viscosity and amount of partial melt, as many parameters influence the ability of migmatites to flow. It remains unclear whether viscoelastic flow can transport mid-crustal sections tens or hundreds of kilometers sideways (as in Clark and Royden, 2000, or Beaumont et al., 2004) or whether lateral transport is more restricted spatially even if partial melt is abundant (see Hu et al., 2022, for Tibet), as it will largely depend on the viscosity of the material and lateral pressure gradients within the crust.

The presence of in situ partial melts in mid-crustal areas increases the buoyancy of the crust. If 10 km of the crust has as much as 10%–40% partial melt at any given time, the high degree partial melting decreases the density of the assemblage, whereas the residue has a composition dominated by quartz, sanidine, biotite, and other phases that are slightly denser than

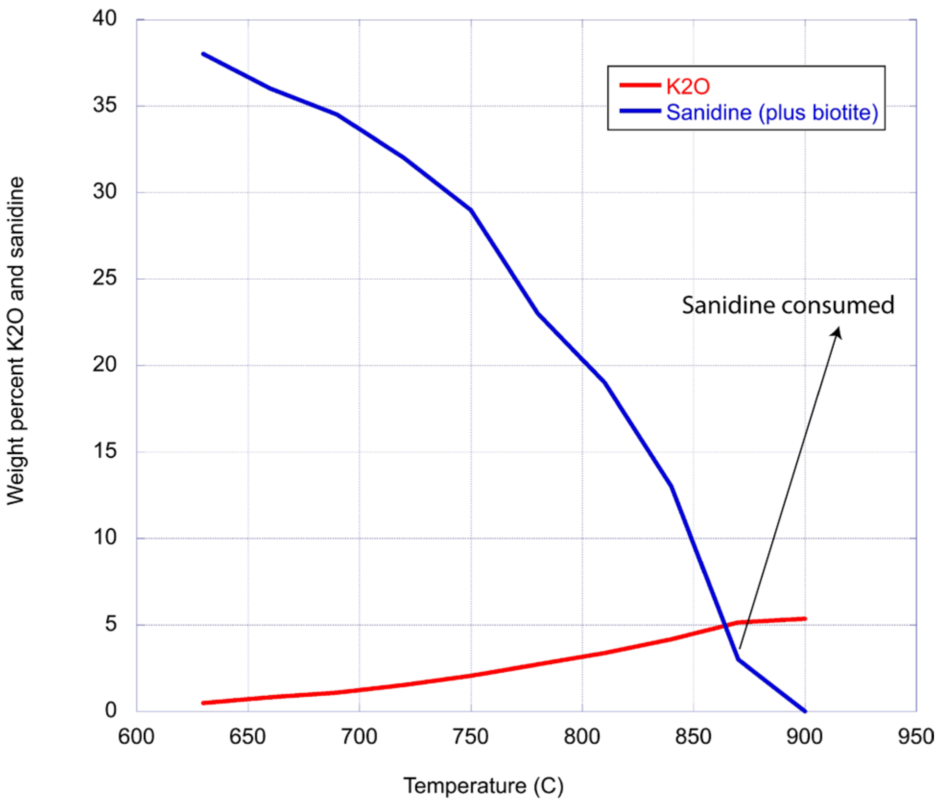

Figure 14. Rhyolite MELTS calculation of K_2O (wt%) and sanidine + biotite (modal percent) versus melt percentages in the temperature range of 650–800 °C at 7 kbar pressures. The figure shows that it is the sanidine and biotite that sequester the potassium concentration in melts, and after their consumption of these phases, melt compositions becomes shoshonitic.

the starting material. Of all felsic minerals that are potentially present in the residue, K-feldspar (sanidine) is the strongest and thus the most likely to give the melt-solid silicate assemblages a higher overall viscosity. Since most plateau ignimbrites are fairly potassic (>3.5 K_2O wt%), high-temperature melts (Kay et al., 2010), we predict that sanidine is consumed from the residue at the time these large volume ignimbrites were able to erupt; thus, the partial melt in the melted region of the mid crust may be large, over 20%–30% depending on how much K is there in the original average source rock (mixed arenitic/pelitic, versus arenite, or other type of rock).

Figure 15 shows the correlation between filtered topography of the Puna plateau and the location of the Altiplano-Puna magma body (from Perkins et al., 2016a). About 0.7 km excess elevation is attributed to the presence of the Altiplano-Puna magma body at depth; this is the most direct demonstration that in situ partial melt can sustain roughly one-sixth of the plateau elevation. We use an approach similar to Bowman et al. (2021) to calculate, using PERPLE-X (Connolly, 2005), the density of residue and melt, as well as the entire partially melted system as a function of partial melt of a metapelite at pressures of 1.0 GPa. The aggregate melt and residue density (assuming they stay within the same volume) is plotted as a function of temperature for each depth range in Figure 16. This shows that in situ partial melting in the mid crust leads to a melt + residue assemblage that is less dense than the original rock. The average negative density anomaly of a 10-km-thick body of metapelite (between 0.7 GPa and 1.0 GPa) undergoing partial melting and holding that in situ relative to the solid body is ~150 kg/m^3, which is well within the range of 300 mgal regional Bouguer anomaly (Perkins et al., 2016b) using GRAVMOD software. At higher pressures, in excess of 1.0 GPa, the melt-solid residue package will have a lesser negative density anomaly due to the increased contribution of garnet in the residue.

5. MINERAL RESOURCES OF PLATEAUS—THE METAL CONNECTION

Plateau mineral resources (Central Andean plateau, Nevadaplano, and Tibet) include various base metal deposits (porphyry copper, etc.) directly related to calc-alkaline magmatism under a thick crust, but the most distinctive types of deposits associated with these tectonic/geomorphic environments are light element enrichments (Li, Cs, B, etc.). In fact, more than 80% of the known readily exploitable resources of the critical metal Li are found on plateaus in the form of accumulations in evaporative salars (Alonso et al., 1991). Salars are dry lakes that are so common in the Central Andean plateau, unlike other evaporites found elsewhere on Earth (e.g., thick Pliocene salt deposits of the Messinian crisis in the Mediterranean, or the salts of the Permian Zechstein Sea in the Alps-Carpathians), which do not have elevated concentrations of these light elements. There appears to be a first-order connection between the concentration of these elements and the formation of orogenic plateaus.

We speculate that the connection is the ability of magmas present in the upper to middle crust of the plateaus to scavenge these elements from their metasedimentary sources in these low-temperature magmas. It is well known that leucogranites are commonly enriched in these elements and often even contain minerals (e.g., spodumene, tourmaline) that host them (Sirbescu et al., 2017). Not all S-type magmas are enriched in Li or B—that is a function of how much they are available in the host rock. In the case of the Central Andean plateau mid-crust, the well-known Puncoviscana metapelites are enriched in Cs (60–100 ppm Cs) compared to the average in the upper continental crust (20 ppm Cs; Rudnick and Gao, 2014). The relative abundances of Li-Cs and B are quite constant in rocks, implying that Li and B (for which there is hardly any data) are probably enriched in the Puncoviscana relative to a background upper crustal value (Meixner et al., 2020). In this case, eutectic melts of the Puncoviscana Formation should be enriched in these highly incompatible elements. These elements, when present in the source regions of migmatites, can also significantly lower the temperatures of granitic eutectics down to as little as 400 °C when large amounts of Li and B "flux" is present (Sirbescu and Nabelek, 2003), although that is unlikely to be common in nature. Whereas we do not have direct samples of the Altiplano-Puna magma body and there is no available study that properly documents the Li concentrations in Miocene to modern volcanic rocks of the Altiplano-Puna volcanic complex, we propose that the link between plateaus and Li (and B, Cs, and possibly other elements) is the very existence of these extensive migmatitic domains in the mid-crust of plateaus (which migrate toward the surface even long after their original formation).

Exactly how Li transfers from magmas into the salars is unclear. Some suggest that rhyolites

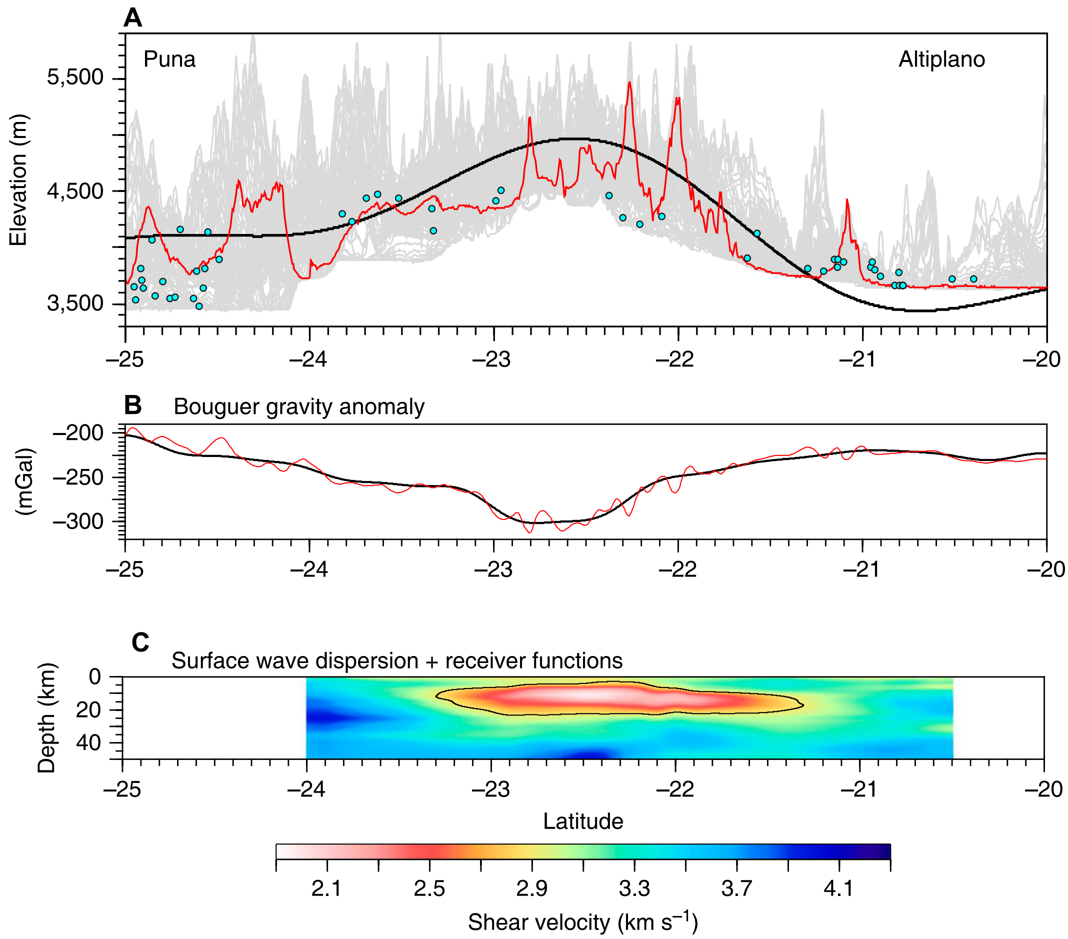

Figure 15. This figure shows the correlation between filtered topography of the Puna plateau and the location of the Altiplano-Puna magma body (APMB; from Perkins et al., 2016a). (A) Elevation vs. latitude along the Puna Altiplano segment where the largest body, the APMB, is located. It shows a topographic swath profile (gray band) along the −67.2 longitude line, an exact topographic profile along the cross-section line (red line), filtered long wavelength topography (black line), and median basement rock outcrop elevations (blue dots). (B) The filtered and unfiltered Bouguer air gravity anomaly obtained from the AnGrav dataset. (C) The S wave velocity anomaly of Ward et al. 2017, where the 2.9 km^{-1} velocity contour corresponds to the boundary of partial melt.

with high Li contents are the direct sources of Li deposits in northern Nevada (the Nevadaplano; Ellis et al., 2022). In other words, Li can be washed from surface lavas and ignimbrites that are enriched in that element. Alternatively, data from Tibet suggest that hydrothermal springs, presumably related to the deeper magmatic bodies of the region, are enriched in Li and other light elements and that the transfer takes place from magma to surface via H_2O-rich hydrothermal fluids (Benson et al., 2023). Much more research needs to be conducted in order to fully grasp the causes of Li, Cs, B, and Rb enrichments in plateau environments. Whether enrichment derives directly from rhyolite lavas or via hydrothermal activity related to them, we propose that the S-type magmatism is a major cause for the ultimate enrichments of those elements in the upper crust of orogenic plateaus (Fig. 17).

6. CLUES FOR OLD PLATEAUS IN THE GEOLOGIC RECORD

Many Barrovian metamorphic terrains covering a range of 0.5–1.0 GPa peak lithostatic pressure conditions may need to be reevaluated as plausible sub-plateau domains. It is fairly well established that these types of metamorphic terrains can be the roots of collisional areas, such as the modern Himalaya. Such areas are subject to crustal thickening by thrust faults and extensive metamorphism along *P-T-t* paths that generate Barrovian metamorphism (England and Thompson, 1984). Crustal shortening along relatively low-angle (thrust) faults rooted in the middle or lower crust and associated Barrovian metamorphism seems to also characterize sub-plateau environments, such as the Nevadaplano or Central Andean plateau, as reviewed above. However, a sub-plateau environment is likely to experience long-term peak metamorphism and associated partial melting, in the order of tens of millions of years because rates of erosion are small.

How will these conditions be fingerprinted in the geologic record? First, the long duration of zircon and/or monazite crystallization in a single rock is a trademark of sub-plateau environments since all other compressional *P-T-t* paths are relatively fast. A sub-Alpine or sub-Himalayan rock goes through a *P-T-t* loop faster and stays at peak conditions typically for not more than a few million years (Philpotts and Ague, 2022). The high surface erosion rates and possible contributions from normal faults in rugged mountain type orogens makes it impossible for any point to linger at or near its metamorphic peak for many millions to tens of million years. The Angel Lake section remained near peak temperature conditions for 45 m.y. (McGrew and Snoke, 2015), the Catalina Rincon for at least 25 m.y. (Ducea et al., 2020), and the Quaternary Altiplano xenoliths contain evidence for ~45 m.y. of high temperatures at mid-crustal depths. This should be a hallmark chronologic signature of ancient plateaus carried by their metamorphic rocks.

Other geochronological and thermochronologic indicators may add to this dimension. Fast unroofing and quick succession of closure of mid-temperature thermochronometers at the time of plateau collapse should be recorded in a sub-plateau domain, not a steady-state cooling/exhumation pattern. This is exemplified by the Arizona exposure presented here, where plenty of geochronometers indicate that the Catalina core complex was hot between 55 Ma and 30 Ma and then experienced fast cooling and unroofing in the late Oligocene–Miocene. While monazite

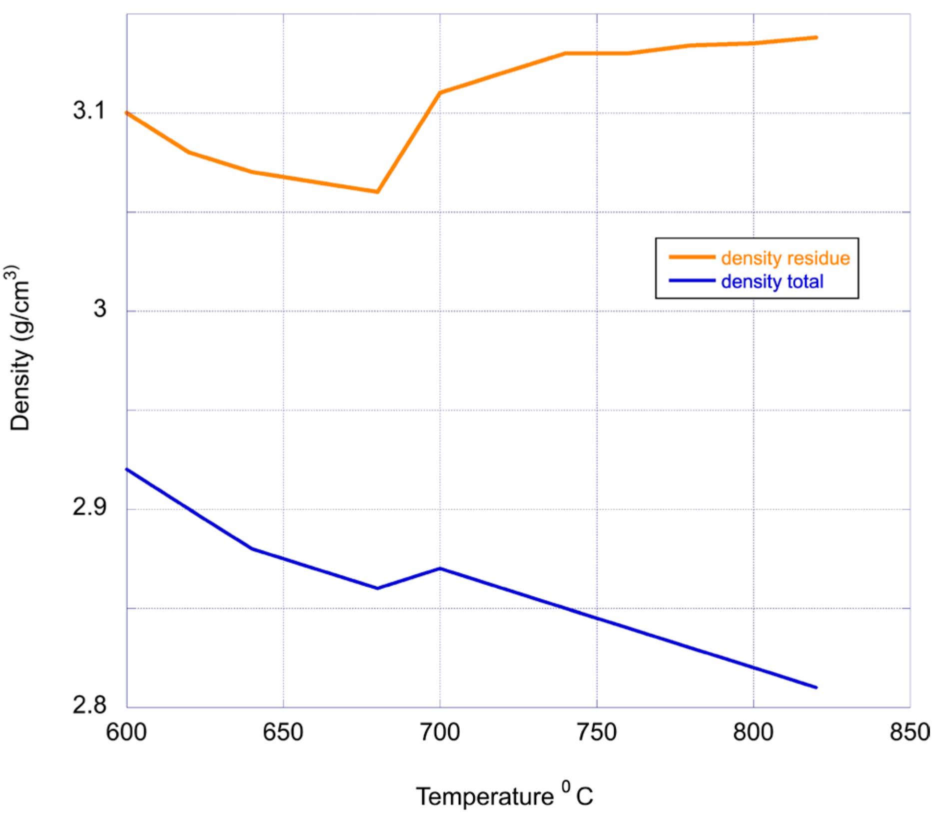

Figure 16. Density of the solid-melt mix determined as a function of temperature (and thus of melt fraction) for the average Puncoviscana Formation (Zimmermann, 2005) at pressures of 7 kbar using PERPLE-X and Rhyolite MELTS packages as well as the classic melt density relationship of Bottinga and Weill (1970). Consumption of sanidine in the residue plays the biggest role in changing the overall density, but the residue never gets dense enough for delamination.

and zircon U-Pb document the length of time when the section was hot, a variety of other chronometers such as mid-temperature Rb-Sr, Ar-Ar systems and then low-temperature chronometers (fission track and U-Th/He) all closed within a few million years at the end of the Oligocene (summary in Ducea et al., 2020).

With respect to thermochronologic signatures of plateau existence, the importance of slow surficial erosion—usually a function of climate (Jepson et al., 2022)—cannot be overemphasized. All of Earth's modern orogenic plateaus are arid. This owes partly to their ability to extract atmospheric moisture along their rugged, high-relief flanks by orographic rain-out (e.g., Strecker et al., 2007;), but it may also be related to latitudinal distribution. The Tibetan, Anatolian-Iranian, and Central Andean plateaus are all located in the subtropics, where long dry seasons alternate with variably strong monsoons. A similar argument may be presented for the Nevadaplano, which was located in a slowly eroding hinterland that was likely sheltered from rapid fluvial erosion by its frontal high-relief fold-thrust belt (Long, 2012), much like the modern Himalaya protects the Tibetan Plateau and the frontal Subandean zone protects the Central Andean plateau (DeCelles and Carrapa, 2023). Without slow erosion, orogenic plateaus would not persist.

Magmatism is rather complex under plateaus, but we suggest that there are some distinctive, albeit non-unique aspects of plateau magmatism (Hu et al., 2025). The coexistence of leucogranitic magmas with more typical subduction-related magmatic rocks is not to be found in Cordilleran frontal arc systems except at the very beginning of subduction magmatism there (Ducea et al., 2015a). Common leucogranitic melts are found in collisional (Himalayan) environments, but there they are not accompanied by intermediate melts. It is difficult to find a single common thread for sub-plateau magmatism, but the consistent association of intermediate classic I-type (sensu Chappell and White, 1992) magmatic rocks, in some cases with adakitic and/or shoshonitic characteristics, with S-type dacites and rhyolites, seems to be a signature magmatic style for these environments. The average composition of plateau volcanism and magmatism in general is close to the boundary between dacite and rhyolite. There are no vertically foliated, stock-dominated large batholiths at depth as defined in Ducea et al. (2015a) and many other papers on frontal arc crustal sections; instead, the dominant fabric of plateau magmatism at depth is sub-horizontal based on both field studies and seismic images. Additionally, the spatial distribution of volcanic rocks is rather random, and does not follow the linear or arcuate distribution of the subduction frontal arc stratovolcanoes.

7. CONCLUDING REMARKS

We review here a number of seemingly disparate features found in the Central Andean plateau with the exposed remnants of the Nevadaplano plateau in western North America in an effort to develop a better view of orogenic plateau evolution and in particular the evolution and role of magmatism. Central to this discussion are the common S-type magmas that are found in plateau regions, either as migmatites in the middle crust or as erupted equivalents. We propose that the existence of long-lived partial melting in the mid-crustal regions of plateaus is an important control on the mechanical elements needed to form and maintain a plateau—gravitational potential to be equalized such that the surface takes on a low-relief shape. Therefore, the evidence for widespread melting under the largest modern plateaus is easily understandable by the internally generated crustal high heat flow of these regions with or without additions of heat and mass from basaltic input. While magmatism in these regions is complex, a blanket of S-type migmatites, which can be hundreds of kilometers in diameter (map view) at any given time, possibly separated as leucogranite sills at shallower levels in the crust, is to be expected. Such partial melting zones can last for millions to tens of millions of years in one place. High viscosity of these melts and the contractional tectonic regime inhibit eruption at the surface; the ratio of plutonic to volcanic material is very high in these environments.

A deeper partial melt generation and ponding zone close to the sub-plateau Moho is also evident in recent geophysical studies. Here, melts from the mantle wedge probably mix with partial melts of the deepest crust in hot zones similar to what is envisioned for frontal arcs. Melts derived from this MASH zone ascend and can be captured by the shallower magmatic systems if they exist. There, they may contribute to the overall heat budget of the mid-crustal migmatites. Direct evidence for this process, including

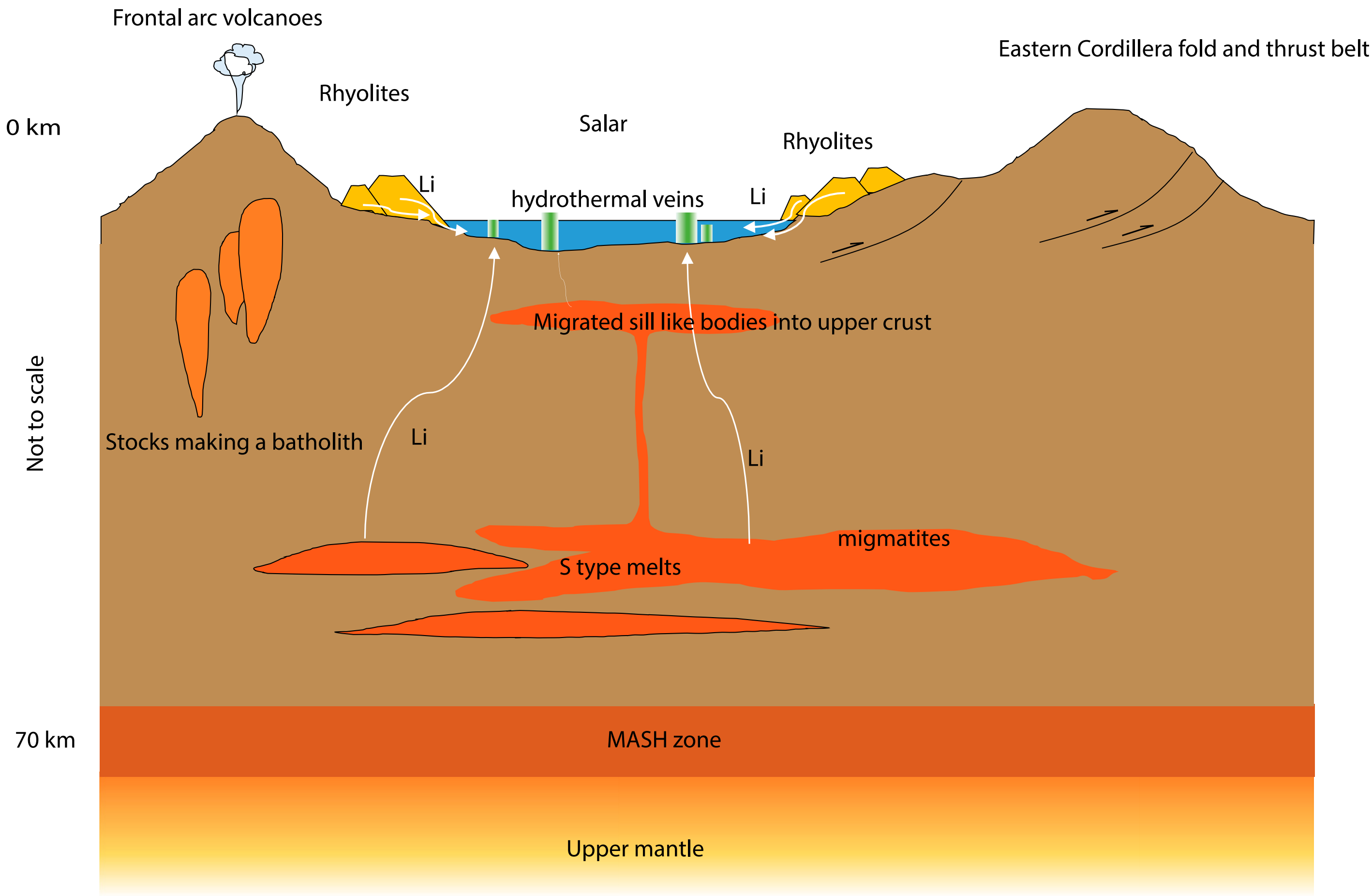

Figure 17. Schematic cross section showing the proposed relationships between Li, Cs, and B in migmatite bodies at depth and their enrichment at the surface of the plateau via hydrothermal veins originating in the shallowest magma chamber or migmatite zone under plateaus or leaching from nearby volcanic rocks.

mixing, exists from exposures of mid-crustal rocks of the Nevadaplano in the East Humboldt Range in Nevada. The addition of I-type intermediate melts from the deeper MASH zone can also explain the fact that most erupted rocks of the Central Andean plateau are relatively high-silica dacitic rocks, not rhyolites.

There may be a link between enrichment of Li, Cs, and other incompatible elements in hydrothermal systems associated with the mid-crustal melting zones of plateaus and the concentration of these elements in salars formed at the surface in these plateau areas. If this is the case, the source of Li from plateau regions is from below and not from alteration of rhyolitic material. We speculate on this potential link and the processes that are responsible for them.

A set of geochemical, petrological, geochronological, and paleoclimatological characteristics of plateau environments are presented as possible tools to be used in the identification of ancient orogenic plateaus in deep time.

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