Models for metamorphic core complexes

**Extensional MCC (rolling hinge):** Regional extension removes hanging wall rocks, driving isostatic footwall uplift

**‘Gneiss dome’ MCC:** Lack of significant extension requires buoyancy to drive advection

**Hybrid MCC:** Buoyant upwelling captured by detachment fault during regional extension; detachment and mylonites may be decoupled

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**Figure 1. Spectrum of metamorphic core complex models:** $\rho_1$ and $\rho_2$ are the density of the upper and lower crust, respectively. MCC—metamorphic core complex.
correlative sweep of potassic volcanism may reflect melting of an enriched mantle source, indicating asthenospheric upwelling (e.g., Manley et al., 2000) following slab removal (Fig. 2A).

After, or potentially overlapping with, the phase of migrating Eocene-Miocene volcanism, MCCs developed across western North America as two distinct sets. In the north, Paleogene MCCs formed in the hinterland of the Sevier thrust front (Armstrong, 1968; Yonkee and Weil, 2015), from British Columbia, Canada, down to southern Nevada (Fig. 2). Conversely, in the south, Oligocene-Miocene MCCs formed in Arizona, eastern California, and southern Nevada, in the foreland region of the Sevier thrust front (Fig. 2). The hinterland MCCs have characteristic spacing of ~200 km, whereas the foreland MCCs are spaced ~50 km (Fig. 2A), defining an ~4:1 spacing ratio. Hinterland MCCs initiated while the Juan de Fuca plate was still subducting beneath North America, which was not yet in an extensional state (e.g., Stevens et al., 2017), whereas foreland MCCs developed during triple-junction migration and regional extension (Atwater and Stock, 1998; Jepson et al., 2022) (Figs. 2B and 2C). Based on their locations relative to the Sevier thrust front, hinterland MCCs likely developed in thicker crust than the foreland MCCs. The timing and spacing represent a distinct dichotomy between the hinterland and foreland MCCs, which has not been satisfactorily explained with existing tectonic models.

RAYLEIGH-TAYLOR INSTABILITY MODEL FOR METAMORPHIC CORE COMPLEXES

To explain the MCC dichotomy, we propose a simple model that links MCC formation with the thermal state and thickness of crust. Our model is reminiscent of buoyant diapirism (i.e., Rayleigh-Taylor instability, RT) in a two-layer medium with a denser upper layer. Dimensional analysis (Selig, 1965) and analog models (Marsh, 1979) show characteristic diapir spacing, $\lambda$, is related to the viscosity contrast $R = \frac{\mu_2}{\mu_1}$ between the upper ($\mu_1$) and lower layers ($\mu_2$) (where $\mu_2 \leq \mu_1$), and the thicknesses of the lower density lower crust layer, $H_c$. For a range of $R$, $\lambda$ was plotted against $H_c$ using analytical solutions to show the positive

Figure 2. (A) Metamorphic core complexes (MCCs) in the hinterland and foreland of the Sevier thrust front in the North American Cordillera. Right graph shows NAVDAT volcanic (black) and potassic rocks (red), MCC and regional extension timing constraints [Supplemental Material [see text footnote 1]], and reconstructed triple-junction formation (square) and migration (arrows). (B, C) Plate reconstructions of Oligocene western North America (Clennett et al., 2020). Hinterland MCCs initiate prior to triple-junction formation and migration, decoupled from these tectonic events, whereas the foreland MCCs develop during triple-junction migration and slab-window development. ARG—Albion–Raft River–Grouse Creek; bt—biotite; ms—muscovite.
correlation between diapir spacing and $H_m$ (Fig. 3A).

To further support the analytical solutions, we conducted two-layer numerical models (Fig. 4) using the MVEP2 thermo-mechanical modeling package (Kaus, 2010; Thielmann and Kaus, 2012). The models used fixed boundaries, $R = 1–100$, and a constant density difference ($\Delta \rho = 0.1 \text{ g/cm}^3$) between the two layers (see Methods in the Supplemental Material) (Fig. 4). These models reproduced the analytical curves with similar $\lambda$ versus $H_m$ correlations (Fig. 3A). The spacing dependence on $R$ paralleled the analytical solutions of Selig (1965).

Our RT upwelling model suggests that distinct differences in thermal state and rheology between the hinterland and foreland regions of the North American Cordillera explain the observed MCC dichotomy (Fig. 2). Specifically, the model predicts that the wider-spaced hinterland MCCs developed with thicker $H_m$ values, greater $R$ values, or a combination of factors (Fig. 3A).

**BOUYANT DOMING IN VARIABLY THICK CRUST**

To test the RT model, we examined how $H_m$ may have varied across the Cordillera, assuming $H_m$ scales with the thickness of crust that might undergo partial melting above the solidus, say when $T > 700 \, ^\circ\text{C}$ (e.g., Rey et al., 2009). In this framework, there are two parameters that affect $H_m$ thickness: crustal thickness and the temperature at the base of the crust. Assuming similar thermal parameters in the crust, a thicker $H_m$ will result from thicker crust or a hotter Moho.

The spatial location of the different MCCs in either the hinterland or foreland of the Sevier thrust front (Fig. 2A) implies that they developed in crust with variable thickness. Support for the Late Cretaceous Nevadaplano orogenic plateau (DeCelles, 2004) with relatively thick crust (~60+ km) in the Sevier hinterland includes observed deeply incised paleovalleys (Henry et al., 2012), geochemical thickness proxies (Chapman et al., 2015), moderate-to-high magnitudes of Mesozoic crustal shortening in the Sevier thrust belt and its hinterland (e.g., Long et al., 2014; Yonkee and Weil, 2015; Zuza et al., 2021), Late Cretaceous deep burial (~7–8 kbar) of supracrustal rocks in exhumed MCCs that supports substantial crustal thickening (Lewis et al., 1999; Hallett and Spear, 2014), reconstructions of Cenozoic extension that imply thickened pre-Cenozoic crust (Coney and Harms, 1984), and stable-isotope paleoaltimetry (e.g., Snell et al., 2014).

Conversely, direct evidence for substantial Mesozoic-Cenozoic crustal thickening in the foreland region is lacking. The region is southeast of the Sevier thrust-front and northeast of the Maria fold-thrust belt (e.g., Knapp and Heizler, 1990) (Fig. 2A). Structural reconstructions of Cretaceous–early Cenozoic contractional deformation do not suggest substantially thickened crust (e.g., Davis, 1979; Clinkscales and Lawton, 2018). Geochemical proxies suggest thickened crust across Arizona in the late Cretaceous (~60 km) but relatively thinner crust (~40 km) at 40–30 Ma (Jepson et al., 2022). Therefore, prior to the initiation of Oligocene-Miocene MCCs, we assume the foreland region was relatively thin at ~40 km.

Assuming a thicker hinterland (~60 km) and thinner foreland (~40 km) at the time of Cenozoic MCC generation, steady-state geotherms were plotted to examine the thickness of $H_m$ above ~700 °C (Fig. 3B). We used an 800 °C Moho temperature to represent the hot lower crust heated via mantle upwelling after slab rollback. A set of numerical models simulating partial melting (Supplemental Materials [see footnote 1]) also support that the hinterland-type crust would have a thicker $H_m$ than the foreland (right panel in Fig. 3B), $H_m$ versus $H_{\text{mb}}$ respectively. We estimate that the foreland lower crustal layer ($H_{\text{mb}}$) was ~7 km and the hinterland lower crustal layer ($H_{\text{mb}}$) was ~20 km, a ratio of ~3:1. With these estimates, a plot of MCC spacing versus $H_m$ fits well on analytical curves, demonstrating a predictable positive correlation (Fig. 3A).

In this framework, observed MCC spacing overlap curves for diapirism with reasonable viscosity contrasts of 2–3 orders of magnitude

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Figure 3. (A) Analytical and numerical predicted diapir spacing vs. the thickness of the lower density lower layer ($H_m$) for different viscosity contrasts ($R$), plotted with hinterland and foreland metamorphic core complex spacing (~10) and estimated $H_{\text{mb}}$ (±5 km) from B. (B) Estimates of lower-layer thickness (above ~700 °C), assuming ~800 °C Moho, for the hinterland ($H_{\text{mb}}$), thin foreland ($H_{\text{mb}}$), or thick, cold (~750 °C Moho) foreland ($H_{\text{mb}}$). See text for explanation. Numerical models of partial melting confirm relative thickness differences (Supplemental Fig. 2 [see text footnote 1]). (C) Tradeoff between $R$ and $H_m$ for the Selig (1965) curve, with observed spacing contours emphasized.

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1Supplemental Material. A synthesis of timing constraints for the North American Cordillera metamorphic core complexes, a brief discussion of the conjugate Shatsky Rise, details of zircon HF compilation, and methods and results of numerical simulations. Go to https://doi.org/10.1130/GSAT.S.21253911 to access the supplemental material; contact editing@geosociety.org with any questions.
between the partially melted lower crust and colder, more viscous upper crust (Fig. 3A) (Whitney et al., 2004; Rey et al., 2009). There is a tradeoff between viscosity contrast ($R$) and the thickness of the lower crustal layer ($H_m$), which we explored for Selig’s (1965) solution (Fig. 3C): observed spacing dichotomy may result from (1) nearly constant $R$ in both the hinterland and foreland, which implies variable $H_m$ (~3:1 ratio); (2) generally similar $H_m$, which implies substantial $R$ variations between hinterland and foreland (~100:1 ratio); or (3) some intermediate scenario. We argue that variable $H_m$, modulated by thermal state or thickness discussed above (Fig. 3B), may be most responsible for spacing variations, which permits similar $R$ values within each setting.

A potential caveat is that it has been postulated that Laramide flat-slab subduction could have refrigerated the upper plate to cool the Moho and thermal structure of the overlying crust (Dumitru et al., 1991). Reconstructions of the subducted CSR (Fig. 2A) show that it would have projected directly beneath the foreland MCCs but not the hinterland MCCs (Livaccari et al., 1981; Axen et al., 2018). This predicts that the foreland crust may have been colder than the hinterland, and therefore MCC diapirism in the colder foreland region would have emanated from an even thinner $H_m$ layer than the hinterland. Although more complex, this scenario still satisfies our spacing arguments (Fig. 3). Furthermore, it is possible that Laramide thickening (Bird, 1984) of the Arizona region was more pronounced than we previously assumed (e.g., >45-km-thick crust), possibly driven by alternative thickening mechanisms beside crustal shortening that are hard to track in the geologic record, such as channel flow (Bird, 1991) or magmatic inflation (e.g., Chen et al., 2018). A potentially thicker foreland region would impact the MCC dichotomy model, but Laramide slab refrigeration may counteract this effect. That is, if the foreland was thick but relatively colder due to these combined impacts, a thinner $H_m$ layer is predicted (Fig. 3A) to explain closer MCC spacing. Despite some uncertainties, thermal state through crustal thickness or basal temperature boundary conditions impact $H_m$ (Fig. 3B) and thus diapir spacing (Fig. 3C).

### FARALLON SLAB DYNAMICS DRIVE LOWER CRUSTAL HEATING

Buoyant MCC doming is driven by vertical density differences in the crust, rather than plate-boundary forces, regional extension, hanging wall removal, and isostasy (Fig. 1). Heating of the lower crust reduces its density and viscosity, for example as shown by numerical simulations and tectonic models for some of the MCCs in southwest Canada (Vanderhaeghe et al., 1999; Rey et al., 2009; Whitney et al., 2013). We envision the RT instabilities initiated with an increase of Moho temperature caused by post-Laramide slab rollback, potentially coupled with slab-window development, that allowed influx of hot asthenosphere that intensely heated the crust (Babeyko et al., 2002; Axen, 2020; Lund Snee and Miller, 2022). Thus, the timing of MCC generation should be strongly coupled with the timing of volcanism and crustal heating, and not necessarily correlated with kinematic shifts in plate-boundary conditions and the initiation of regional extension.

To test this hypothesis, we compiled biotite and muscovite $^{40}$Ar/$^{39}$Ar ages (Supplemental Material [see footnote 1]), which track cooling through closure temperatures of ~300 °C and 400 °C, respectively (McDougall and Harrison, 1999). We interpret these dates to broadly constrain the late phases of mylonite development in quartz-rich rocks along the flanks of the evolving MCCs. Lower temperature thermochronometers track brittle normal faulting and related exhumation. Argon dates from hinterland MCCs are younger than the northwest, whereas those from foreland MCCs are older (Fig. 2). MCC doming age patterns parallel volcanic trends (Gans et al., 1989), but only Ar dates from the foreland MCCs show a correlation with the propagation of initial regional Basin and Range extension tracked by plate reconstructions, low-temperature thermochronology, and the extensional basin record (Miller et al., 1999; Colgan et al., 2010; Konstantinou et al., 2013; Lee et al., 2017; Jepson et al., 2022; Supplemental Material [see footnote 1]) (Fig. 2). Our compilation of volcanism, MCC doming, regional extension, and triple-junction migration suggests that MCC development is more strongly correlated with trends of rollback volcanism rather than the propagation of regional extension due to migrating triple junctions (Fig. 2).

The implied causal relationship between magmatism and MCC generation can be further tested by magmatic source characteristics. All MCCs involve pre-/syn-kinematic magmatism (e.g., Gans et al., 1989; Howlett et al., 2021). Available zircon $eH_f$ data from different but adjacent hinterland MCCs broadly overlap with parallel trends (Fig. 5A). $eH_f$ trend toward evolved values ($eH_f < ~20$) during Late Cretaceous anatexis followed by a juvenile excursion ($eH_f < ~10$) during Eocene slab rollback reflecting mantle influx and melting (Howlett et al., 2021). An Oligocene evolution toward more evolved values ($eH_f < ~30$) can be interpreted as protracted crustal heating and melting (Konstantinou et al., 2013) (Fig. 5A). Foreland MCCs show more subdued isotopic trends (Fig. 5B), likely reflecting different melt sources compared to the hinterland region. Within uncertainty, the foreland trend is either flat or there is a juvenile excursion with the arrival of mantle-derived volcanism (Fig. 5B). In the hinterland, there is a pronounced


DECOUPLED MCC DOMING AND DETACHMENT FAULTING

Advances in field and geochronology studies reveal a decoupled two-phase deformation history for the hinterland MCCs. In the Albion–Raft River–Grouse Creek, the primary mylonitic shear zones formed in the Oligocene and Basin and Range extensional faulting started ca. 14 Ma (Konstantinou et al., 2013). In the northern Ruby Mountains–East Humboldt Range, Oligocene mylonites are crosscut by undeformed 17 Ma basalt dikes, which are cut by Miocene detachment faults that were associated with syn-kinematic extensional basin sedimentation (Wright and Snoke, 1993; Zuza et al., 2021, 2022). The Miocene detachment continues south along strike for ~150 km (Colgan et al., 2010), where its footwall is no longer mylonitic or magmatic, thus suggesting the mylonites are not genetically or kinematically linked with detachment faulting. In the Snake Range, the Oligocene mylonitic shear zone was cut by ca. 22 Ma undeformed dikes (Lee et al., 2017), and a later phase of extensional exhumation is recorded by ca. 17 Ma fission track ages (Miller et al., 1999).

We posit that for hinterland MCCs, the earlier, temporally decoupled phase of buoyant doming established mechanical or thermal weaknesses that were exploited by Miocene detachment faulting, thus explaining the apparent connectivity between Paleogene doming and Miocene detachment faults (e.g., Konstantinou et al., 2013; Ducea et al., 2020; Zuza et al., 2021) (Fig. 2A). This also explains the perplexing observation that Paleogene MCCs did not generate syn-kinematic basins, whereas Miocene extensional basins were well developed (Colgan and Henry, 2009; Zuza et al., 2021). Domal upwarps in the mid-crust did not generate space for surface sedimentation, but hanging wall removal during detachment faulting allowed for supra-detachment basins (e.g., Friedmann and Burbank, 1995). Foreland MCCs may have similarly involved two phases that occurred on nearly overlapping time scales (Jepson et al., 2022) due to coeval slab-window development, magmatism, and extension initiation (Atwater and Stock, 1998) (Fig. 2).

Extension-related detachment fault models for MCC generation (e.g., Wernicke and Axen, 1988) (Fig. 1A) cannot satisfactorily explain MCC spacing, age trends, and generation prior to plate-boundary conditions switched to initiate regional extension (Fig. 2). MCC spacing has previously been interpreted in the context of elastic buckling (e.g., Yin, 1991), but this type of instantaneous solution does not uniquely constrain observed age trends across the Cordillera (Fig. 2) and diminishes the role of a viscous, partial-melt–rich mid-lower crust. The aforementioned two-phase deformation history of many MCCs complicates models of simple protracted detachment faulting. However, it remains possible that some MCC spacing is partially modulated by corrugations or elastic buckling that overprinted an established first-order buoyantly domed architecture.

A comprehensive summary model in Figure 6 unifies observations from across the Cordillera and provides testable predictions for future investigations. Mesozoic shortening thickened the hinterland region more than the foreland. Laramide flat-slab subduction underplated schists beneath the foreland region, potentially refrigerating the upper-plate lithosphere. In the hinterland, post-Laramide slab rollback drove SW sweeping juvenile magmatism that heated the crust. Thermal incubation over ~10 m.y. resulted in a hot, melt-rich lower crust that rose as buoyant diapirs to form hinterland gneiss dome MCCs with strong shearing along the upwelling margins (Fig. 1B). This style of MCC development resulted in pure-shear attenuation along the flanks and tops of the rising domes (Miller et al., 1983; Zuza et al., 2022). Paleogene doming would have overprinted and incorporated preexisting Mesozoic fabrics and structures, thus creating locally complex domal geometries. Detachment faulting during Miocene-present Basin and Range extension exploited the domal structures to exhumate them in the detachment footwall (Fig. 1C).

The spatial correspondence of the CSR and slab window beneath the foreland region may
imply complex influence on MCC generation, including earlier lithospheric refrigeration and hydration followed by focused heating of the base of the crust. Coupled mantle upwelling through the slab window and a plate configuration conducive to regional extension drove diapiric upwellings that were almost immediately (within several Myr) impacted by regional extension. Detachment faults captured the rising domes in their footwalls, resulting in more traditional kinematic evolution and geometries, with more stratigraphic omission and syn-kinematic basins (Fig. 1C).

The development of the classic North American Cordilleran MCC belt was not uniquely and initially driven by regional extension because the hinterland MCCs developed before mid-Miocene plate-boundary conditions drove regional extension (Colgan and Henry, 2009). Instead, we argue that Farallon slab dynamics and subsequent mantle heating led to buoyant RT upwellings at characteristic spacings. Individual MCCs undoubtedly experienced differing Mesozoic-Cenozoic geologic histories, but as outlined here, the thickness and thermal state of the crust exerted a first-order control on the observed MCC dichotomy. Our model may be transferrable to other similar coupled subduction-intraplate settings. Mesozoic subduction in southeast China involved MCC generation following Jurassic-Cretaceous flat-slab subduction, rollback, and mantle-derived magmatism (Li and Li, 2007), similar to North America. The links between flat-slab events, subsequent rollback, magmatism, partial melting, and heat redistribution during MCC development require further evaluation.

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