

Dehydration and deformation: the sources and impacts of metamorphic fluids from oceanic crust
in subduction zone forearcs

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Abstract

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Subduction zones are host to major seismic and volcanic hazards, drive plate tectonics, and are an critical force in global geochemical cycling. The mechanical properties of the interface between the subducting and overriding plates are important controls on how these processes progress, but those properties and how they are moderated by chemical processes like metamorphism and metasomatism are complex and poorly understood. This thesis focus on providing a better understanding of the environment in which deep slow slip and tremor occurs in subduction zones and the interacting chemical and mechanical processes that control rock deformation. Specifically, this works is motivated by wanting to understand the sources of metamorphic fluids in this region and the impacts that they can have. I apply a range of

thermodynamic modeling, field geology, petrographic, and geochemical analyses tools to explore this question in modern and ancient subduction zones.

Phase equilibrium modeling of oceanic crust and mantle wedge lithologies in the subduction zone at Mexico combined with geophysically-constrained thermal models and slow slip and tremor distributions give us insights into the metasomatic impacts of fluids and their possible relation to facilitating slow slip and tremor. We show how and where talc can be produced via metasomatism of mantle wedge serpentinites by local advection of Si-bearing fluids from the subducting Cocos plate. Major dehydration reactions in subducting metabasalts have the potential to produce fluid fluxes that result in the production of rheologically significant volumes of talc at the plate interface.

In the geologic record, we search for evidence of dehydration reactions similar to the ones predicted by our phase equilibrium modeling to occur beneath Mexico. We apply petrologic and geochemical analyses to metabasalts of the Epidote Amphibolite Unit in the Catalina Schist to constrain their metamorphic evolution, and use phase equilibrium models to estimate what that entails for their dehydration history. Epidote-rich zones in these metabasalts have geochemical signatures and textures that suggest that they could represent channel-forming reactive porosity networks that developed as a result of pumpellyite breakdown at $\sim 300^{\circ}\text{C}$. Epidote textures and compositions further record prograde, peak, and retrograde metamorphism—the latter stage associated with underplating and refrigeration in a cooling subduction zone. This work demonstrates the utility of epidote as a tracer of metamorphic processes during low- and medium-grade metamorphism.

We have also reconstructed the metamorphic evolution of the Epidote Blueschist Unit of the Catalina Schist, which followed a similar P - T path to the Epidote Amphibolite Unit, but

exhibits much stronger deformation fabrics. We therefore pair petrologic and microstructural analyses using electron backscattered diffraction to place the deformation of the Epidote Blueschist Unit in the context of its subduction and underplating metamorphic story. We show that these rocks deformed early in subduction while still at seismogenic zone conditions ($<300^{\circ}\text{C}$), but that the dominant deformation fabrics in these rocks likely represent underplating. Amphibole and quartz deformed by dissolution precipitation creep and dislocation creep, respectively, during underplating, and quartz recrystallized grain sizes allow us to estimate differential stresses of $\sim 50\text{--}180\text{ MPa}$. This work highlights the importance of contextualizing structural observations within a metamorphic framework, and provides additional context for the environment at the base of the seismogenic zone where slow slip and tremor may occur.

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Chapter 1: Introduction

Subduction zones are critical tectonic environments that generate significant geohazards and drive our planet's geochemical cycles. Subduction zones host the largest earthquakes on Earth and produce arc volcanism, presenting myriad geohazards that impact society. They also enact a first order control on geochemical cycling of volatiles between deep earth and the surface, and thus are an important component of global geodynamic and climate processes over geologic time (e.g., Bebout, 1995; Stern, 2002; Kelemen and Manning, 2015). Both deformation and element cycling are affected in large part by the activity of fluids, which impact the mechanical and chemical properties of the rocks in subduction zones.

Fluids are significant in all regions of subduction zones and mediate processes like deformation, mass transfer, and melting. In subduction settings we commonly conceptualize of fluids as being dominantly H₂O, but other volatiles and dissolved elements are likely present in the fluid phase (e.g., Manning, 2004). Fluids are transported into subduction zones by the downgoing plate either as pore fluids, which are squeezed out of rocks at depths <15 km, or as mineralogically bound volatiles which eventually will be released through metamorphic dehydration reactions (e.g., Saffer and Tobin, 2011; Hacker, 2008). Fluids produced through dehydration reactions in the subducting oceanic crust can travel through the slab and plate interface via channelized or pervasive flow (e.g., Bebout and Penniston-Dorland, 2016). When encountering rocks with which they are not in chemical equilibrium, fluids can exchange elements with the host rock and alter composition of those rocks. This leads to the development of metasomatic and hybrid lithologies that can have both unique mechanical properties and unique chemical properties that can impact how mass is transferred through the subduction zone (e.g., Bebout and Penniston-Dorland, 2016; Codillo et al., 2018; Hoover et al., 2025).

Fluids can also affect the mechanical behavior of rocks in several key ways. If fluid pressures are high enough, it can promote fracturing and/or frictional slip by lowering the effective stress within the rock. Fluid pressures greater than the minimum compressive stress can open fractures through hydraulic fracturing. High fluid pressures near but not necessarily at lithostatic pressures can also promote frictional sliding on preexisting shear surfaces by reducing the effective normal stress on those surfaces (e.g., French and Condit, 2019; Gao and Wang, 2017). The common occurrence of crack-seal veins in exhumed subduction complexes is one piece of evidence for at least transiently high fluid pressures in and around the subduction plate interface (e.g., Behr and Bürgmann, 2021; Condit et al., 2022). Fluids also influence deformation by activating fluid-mediated viscous mechanisms such as pressure solution creep (e.g., Gratz, 1998; Wassmann and Stöckhert, 2013; Condit et al., 2022; Behr and Platt 2013, Fisher and Hirth, 2024). Fluids can therefore change the deformation mechanisms within the viscous regime from power law relations to linearly viscous (Condit et al., 2025), which profoundly influences the relationships between stress and strain rate in these rocks, and modifies the overall viscous strength significantly.

Slow slip events, a mode of deformation within subduction zone deformation that has only recently been discovered, are potentially facilitated by the presence of fluids (e.g., Bürgmann, 2018). Slow slip events involve geodetically-measured slip in a plate interface shear zone at rates of mm per day that can release the energy of a Mw 6–7 earthquake over the course of months to years (e.g., Wech and Bartlow, 2014; Schwartz and Rokosky, 2007; Gao and Wang, 2017). These typically occur in warm modern subduction zones such as Cascadia, Mexico, and Nankai (e.g., Peacock, 2009), and may occur at regular intervals of months, which is termed episodic tremor and slow slip (ETS). Where it occurs, deep slow slip is thought to be a mechanism for transferring stresses updip to the seismogenic zone (e.g., Wech and Creager, 2011; Obara and Kato, 2016). The physical processes by which slow slip events occur is still debated, but they likely occur in a fluid-rich environment based on geophysical observations. These include high P- to S-wave velocity ratios (e.g., Audet and Bürgmann 2011; Delph et al., 2018) and magnetotelluric studies showing high conductivity zones (e.g., Wannamaker et al., 2014). These zones are often inferred to reflect the presence of pore fluids at near-lithostatic pressures (Audet and Kim, 2016; Delph et al., 2018). These high fluid pressures may also contribute to the weakness of the plate interface in this zone by reducing effective normal stresses and enabling the sensitivity of slow slip and tremor events to tides (e.g., Peng and Gombert, 2010; Houston, 2015). Geologic evidence include the prevalence of fluids include veining and dissolution precipitation creep microstructures (Behr and Bürgmann, 2021) in rocks from the pressure-temperature conditions at which slow slip likely occurs.

Modeling work has correlated the depths at which slow slip and tremor occur with the conditions at which subducting oceanic crustal rocks are expected to dehydrate, suggesting an important connection between metamorphic dehydration and slow slip (Condit et al., 2020). Not only may these fluids promote the aforementioned high fluid pressures, but metasomatism alters the strength of rocks prior to and during subduction by promoting the growth of weak mineral phases like chlorite and talc that, in turn, can localize deformation (Hernández-Urbe et al., 2020; Gyomlai et al., 2021; Bebout and Barton, 2002; Hirauchi et al., 2013). Metasomatized rocks are common in exhumed paleo-subduction complexes worldwide and therefore a key consideration for understanding deformation at the subduction interface (e.g., Peacock and Hyndman, 1999; French and Condit, 2019; Codillo et al., 2022, Hoover et al., 2022; Easthouse et al., 2025). These rocks that are subducted and then exhumed provide an important perspective on the conditions at the plate interface that complement geophysical and modeling perspectives. In particular, through the record of alteration, hybridization, and fluid-rock interactions, they provide insight into the chemical processes that are relevant.

The work presented here is motivated primarily by a desire to better understand the environment in which deep slow slip and tremor occur. In particular, how do interacting chemical and mechanical processes facilitate these phenomena? Whereas “fast” earthquakes occurring in the shallower crust at lower temperature conditions where chemical (i.e., metamorphic) changes to rocks are minimal and do not significantly influence the frictional

behavior governing earthquake nucleation, below the seismogenic zone, the impact of chemical changes cannot be ignored.

The metamorphic changes of interest here include on one hand the evolution of the mineral assemblage of a rock such that its rheology also evolves. On the other hand, metamorphism is integral to the production of fluids that both impact mechanical processes and facilitate further chemical changes. As introduced above, on the mechanical front, fluids can impact both the frictional and viscous deformation of rocks by reducing effective normal stresses and by influencing the efficiency of diffusive mass transfer deformation mechanisms, respectively. On the chemical front, fluids metasomatize rocks with which they are in chemical disequilibrium, changing their elemental composition and promoting the growth of new minerals with different mechanical properties. To better understand the environment of deep slow slip, I have focused primarily on the question: **what are the sources of fluids in the region that hosts slow slip and how might they impact the occurrence of slow slip?** In my work I explore both the sources of fluids in dehydrating subducting rocks that may promote fluid-rich environments and high pore fluid pressures, the potential effects of these fluids through their ability to produce rheologically significant minerals like talc through metasomatism, and the potential effects of these fluids of viscous deformation mechanisms.

I approach this question from different angles within each of my three chapters. In **chapter one**, I explore where and how chemical changes might occur as a result of silica metasomatism in the mantle wedge in the modern Middle American subduction zone near Guerrero, Mexico. At this subduction zone, a flat subducting plate geometry provides an ideal setup for a suite of thermodynamic models to test the potential of dehydration fluids to alter serpentinites at the plate interface to produce talc-rich rocks, which have been suggested to occur here based on geophysical observations (Song et al., 2009). Such talc-rich rocks are thought to be potential hosts for slow slip events because of the frictional properties of talc (e.g., Shabtian and Hirth, 2025; French and Condit, 2019; Hoover et al., 2022), and the subduction zone at Guerrero hosts regular episodic tremor and slow slip (ETS) events that occur in specific zones of the flat slab region. A technique I apply here that I also apply in subsequent chapters is phase equilibrium modeling using the Gibbs free energy minimization software *Perple_X* (Connolly, 2005; Connolly, 2009). In this case, I combine thermodynamic modeling with a thermal model of the subduction zone, which allows us to determine where water may be liberated from dehydrating rocks in the subducting Cocos plate, how much water this might be, and finally, how much SiO_2 that water could advect into the mantle wedge and subsequently how much talc could be produced as a result of metasomatism. This chapter was published in *Geochemistry, Geophysics, Geosystems* in 2023 and featured in *Eos* that same year (Lindquist et al., 2023)

Chapters two and three are based in field observations of the subduction-related rocks of the Catalina Schist from southern California, USA. These rocks represent slices of the oceanic crust that were underplated in a progressively cooling subduction zone, providing snapshots of subduction processes over a range of pressure and temperature conditions. **Chapter two** explores the geologic record for the potential sources of fluids that could effect the chemical changes like

those proposed in chapter one and produce the high pore fluid pressures that are predicted in the regions that host slow slip and tremor (e.g., Peacock, 2009; Fagereng and Deiner, 2011; Condit et al., 2020). In this chapter I focus on epidote amphibolite metabasalts from the Catalina Schist as a candidate fluid source. I again apply phase equilibrium modeling to estimate pressure-temperature conditions of metamorphism, but now pair these models with petrographic observations and geochemical analyses including major- and trace-element analyses of epidote and other minerals and bulk-rock Sr isotope ratios. The latter provide an indication of how hydrated the metabasalt protoliths may have been prior to subduction, and the epidote compositional information provides a wealth of information about the metamorphic history of these rocks. We propose that epidote in these rocks record multiple stages of growth that relate to a major dehydration event, prograde subduction metamorphism, and retrograde underplating and cooling. This chapter therefore also demonstrates that epidote is a powerful petrologic tracer of metamorphic processes at low- to medium-grade metamorphic conditions, where classical petrologic tools developed for minerals like garnet are not applicable. This chapter was published in *Geochemistry, Geophysics, Geosystems* in 2026 (Lindquist et al., 2026).

Similar to chapter two, **chapter three** explores the metamorphic history of another unit in the Catalina Schist, the Epidote Blueschist Unit, but this time pairs petrologic and structural analyses to place the deformation processes recorded in these rocks within a metamorphic framework. Unlike the Epidote Amphibolite Unit of chapter two, the Epidote Blueschists exhibit well-developed deformation fabrics. Again applying phase equilibrium modeling, petrographic observations of mineral textures and zoning, and geochemical analyses of mineral compositions, I reconstructed a pressure-temperature history for a suite of metabasaltic and metasedimentary rocks from this unit. This reveals a subduction history similar to the Epidote Amphibolite Unit and provides a framework for microstructural analyses including electron backscattered diffraction. These analyses reveal the deformation mechanisms of amphibole and quartz in these rocks, which, based on the textures and composition of amphibole, appear to be associated with underplating and early retrograde metamorphism. Our analyses of quartz further provide constraints on the stresses at which these rocks deformed, providing an important data point in understanding the environment near the plate interface at the conditions of slow slip and tremor. This chapter highlights the need for combining structural analyses with careful petrologic analyses when working in subduction metamorphic rocks in order to tease apart the various aspects of a rock's entire metamorphic history.

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Chapter 2: Metasomatism and slow slip: talc production along the flat subduction plate interface beneath Mexico (Guerrero)

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Key Points:

- We explore how the silica metasomatism of serpentinized peridotite beneath Guerrero, Mexico can produce talc-rich rocks.
- Talc is predicted to grow in serpentinite where subducting basalts experience metamorphic dehydration.
- Local advection of silica cannot produce geophysically-observable talc volumes, but will still produce thin zones of talc-rich rocks.

Abstract

Talc-rich rocks are common in exhumed subduction zone terranes and may explain geophysical observations of the subduction zone interface, particularly beneath Guerrero, Mexico where the Cocos plate subducts horizontally beneath North America and episodic tremor and slow slip (ETS) occurs. We present petrologic models exploring (1) the degree of silica metasomatism required to produce talc in serpentinized peridotites at the pressure-temperature conditions of the plate interface beneath Guerrero, and (2) the amount of silica-bearing water produced by rocks from the subducting Cocos plate and the location of fluid pulses. We estimate volumes of talc produced by the advection of silica-rich fluids into serpentinized peridotites at the plate interface over the history of the flat-slab system. In the ETS-hosting region, serpentinites must achieve ~43 wt. % SiO_2 to stabilize talc, but minor additions of silica beyond this produce large volumes of talc. Our models of Cocos plate dehydration predict that water flux into the interface averages $3.9 \times 10^4 \text{ kg m}^{-2} \text{ Myr}^{-1}$, but suggest that only where subducting basalts undergo major dehydration reactions will sufficient amounts of silica-rich fluids be produced to drive significant metasomatism. We suggest that talc produced by advective transport of aqueous silica alone cannot account for geophysical interpretations of km-thick zones of talc-rich rocks beneath Guerrero, although silica-bearing fluids that migrate along the plate interface may promote broader metasomatism. Regions of predicted talc production do, however, overlap with the spatial occurrence of ETS, consistent with models of slow slip based on the frictional deformation of metasomatic lithologies.

Plain Language Summary

The subduction zone at Guerrero, Mexico hosts poorly understood fault slip phenomena called episodic tremor and slow slip (ETS), and seismic studies of this subduction zone suggest the deep interface between the Cocos plate and overriding North American plate contains rocks that are rich in the mineral talc. Talc is a weak mineral that can preferentially host deformation and may play a role in facilitating ETS in subduction zones. The presence of talc, however, requires chemical alteration of rocks at the plate interface, so we use Guerrero as a natural lab for exploring how talc might be produced in order to host ETS and produced the observed seismic signatures. We can use the thermodynamic properties of minerals to predict which minerals will occur in rocks that make up the mantle above the subducting Cocos plate, as well as when minerals in subducting rocks will break down to release water. This allows us to predict how much water will be released by the Cocos plate and subsequently how much talc can grow in the plate interface. Although the model for alteration we explore does not explain the geophysical observations at Guerrero, talc is still expected to be produced within regions that experience ETS.

1 Introduction

The interface between subducting and overriding plates in a subduction zone deforms by a complex suite of slip behaviors that include earthquakes, slow slip events, tectonic tremor, and aseismic creeping. The mode of deformation in these fault zones is a consequence of a wide range of factors, including the composition of the plates (e.g., lithology and mineralogy), pressure, temperature, stress, and the presence and movement of fluids (e.g., Agard et al., 2018; Nishiyama et al., 2023; Tarling et al., 2019; Ujiie et al., 2022). However, the chemical processes that affect the composition and thus rheology of the plate interface in subduction zones remain

understudied. For example, metasomatism can alter the strength of rocks prior to and during subduction by promoting the growth of weak mineral phases like chlorite and talc that, in turn, can localize deformation (Hernández-Urbe et al., 2022; Gyomlai et al., 2021; Bebout et al., 2002; Hirauchi et al., 2013). Such metasomatized rocks are common in exhumed paleo-subduction complexes worldwide (e.g., Peacock and Hyndman, 1999; French and Condit, 2019; Codillo et al., 2022a, Hoover et al., 2022).

Talc ($\text{Mg}_3\text{Si}_4\text{O}_{10}(\text{OH})_2$) is a hydrous sheet silicate that occurs in metasomatic and metamorphic ultramafic and carbonate rocks, and is hypothesized to be rheologically significant in subduction zones (e.g., Peacock and Wang, 2020; French and Condit, 2019). During isochemical metamorphism, talc is not predicted to be stable at typical slab-top fluid saturated subduction zone conditions within major subducted lithologies (e.g., metabasalts, metasedimentary rocks), so talc is thought to be produced through the metasomatism—often hypothesized to form through the interaction of silica-rich fluids with rocks in the subduction zone (e.g., Manning, 1997). Experiments and modeling of the simple $\text{MgO-SiO}_2\text{-H}_2\text{O}$ system indicate that talc is stable from the seafloor to a maximum temperature of $\sim 800^\circ\text{C}$ at 1–2 GPa, and a maximum pressure of 5 GPa at $\sim 700^\circ\text{C}$ (Pawley and Wood, 1995 and references therein). This wide range of stability of talc and the expected ubiquity of silica-rich fluids in the lithosphere is reflected by the common occurrence of talc in oceanic and continental fault zones (e.g., Boschi et al., 2006; Moore and Rymer, 2007; Collettini et al., 2009) and exhumed suites of subduction-related rocks (e.g., Bebout & Barton, 1993; Spandler et al., 2008; Hirauchi et al., 2020; French & Condit, 2019, Hoover et al., 2022). Subduction zones in which silica-saturated fluids released by the subducting slab interact with peridotite in the mantle wedge, are therefore a likely setting for the production of talc in rocks near the plate interface up to high pressures and temperatures.

Mechanically, talc has very low frictional strength (e.g., Moore and Lockner, 2008) and given its potential ubiquity along the plate interface in the forearc suggests it could accommodate deformation over a wide range of pressure and temperature conditions. Even at the highest temperatures at which deformation experiments on talc have been performed (875°C), regions of talc will develop narrow zones of localized shearing and slip in which sheets of talc delaminate and slide past each other (Escartín et al., 2008). At high pressures (0.5–1.5 GPa) as well, talc continues to exhibit frictional behavior, with thermal weakening and an increasing tendency to localize deformation with increasing temperature (Boneh et al., 2023). The experimentally-determined rate-and-state frictional parameters of talc also suggest that talc could be a material that is capable of producing slow slip events. Deformation experiments on samples of mixed antigorite and quartz gouge at $300\text{--}500^\circ\text{C}$ showed that talc growth occurred during deformation and localized strain in the samples, even with low (<10 vol. %) quantities of talc (Hirauchi et al., 2013). Talc exhibits velocity-strengthening behavior, but tends towards velocity-neutral behavior at low slip rates (Hirauchi et al., 2020) and increasing bulk shear strain (Misra et al., 2014). Rheological modeling by French and Condit (2019) indicates that at the high pore-fluid pressures predicted to occur at the plate interface, talc schists should preferentially accommodate slow slip strain rates, corroborated by microstructural observations of natural talc-bearing subduction interface rocks (e.g., Hoover et al., 2022).

Here, we model the production of talc along the flat slab subducting beneath Mexico, where observed slip behaviors include earthquakes, slow slip events (SSEs), and tectonic tremor, occurring episodically in the form of episodic tremor and slip (ETS) events. Through a

combination of thermodynamic models of both (1) subducting Cocos plate oceanic lithosphere metamorphism and (2) metasomatic alteration of hydrated mantle peridotites, we show that silica-rich fluids produced by dehydration reactions in MORB may contribute to the occurrence of talc along the plate interface in the Guerrero, Mexico subduction zone. We show that over a duration of 10 million years, greater than 10 vol. % talc can be produced through silica advection at the plate interface near where large volumes of fluids are predicted to be produced by metamorphic dehydration reactions in the subducting Cocos plate. These results demonstrate the importance of incorporating metasomatism into our models of plate boundary slip processes and provide a possible chemical mechanism to produce some of the common occurrences of talc observed in exhumed terranes.

2 Geologic Setting

2.1 The Middle American subduction zone near Guerrero

The subducting Cocos plate in the Middle American subduction zone near the state of Guerrero, Mexico has a unique flat slab geometry (Fig. 1). The Cocos plate subducts beneath the North American plate at the Middle American trench at rates of ~60 mm/yr near Acapulco, Mexico (DeMets et al., 2010). Here, the plate descends beneath Mexico at a typical subduction dip angle before flattening and horizontally subducting beneath the continental lithosphere of North America (Pérez-Campos et al., 2008). Between 120 and 290 km from the trench, the top of the flat slab lies at ~40 km depth (Fig. 1; Kim et al., 2010). Although the map extent of the flat slab segment of the Cocos plate extends beyond the borders of state of Guerrero, we refer to this as the Guerrero segment to remain consistent with previous literature.

Flat slab subduction initiated here around 16 Ma, and the flat portion reached its farthest landward extent around 10 Ma before beginning to roll back (Ferrari et al., 2012; Manea et al., 2013). This is evidenced by the migration of the volcanic front of the eastern Trans-Mexican Volcanic Belt between 16 and 9 Ma, and its continuing trenchward migration since 9 Ma. Between 20 Ma and 10 Ma, Cocos–North American plate convergence rates near Guerrero were >100 mm/yr, with a significant component of margin-parallel motion. Since ~10 Ma, the convergence rate decreased to the modern, nearly margin-perpendicular rate of ~60 mm/yr for the last ~5 Myr (Sdrolias and Müller, 2006).

The Cocos plate is produced by fast spreading rates at the East Pacific Rise. Spreading rates at the East Pacific Rise where it produces the Cocos plate are relatively fast, ranging from 72 to 128 mm/yr (DeMets et al., 2010). The age of subducting oceanic lithosphere at the Middle American trench near Guerrero is ~15 Ma (Manea et al., 2013). The composition of the Cocos plate has been investigated through limited seafloor drilling. Off the coast of Acapulco and 10 km seaward of the trench, hole 487 from leg 66 of the Deep Sea Drilling Project drilled through 171 m of sediment before striking basalt, and an additional 7 meters of basalt were recovered (Watkins et al., 1982). The sediments themselves consisted of 105.5 m of muddy silt atop 65.5 m of brown pelagic clay.

The thermal model of Manea and Manea (2011) provides an estimate of the present-day steady-state, 2D thermal structure of the subduction zone at Guerrero, using the slab geometry determined by the Meso-American Subduction Experiment. In their model, the top of the slab in the flat segment experiences isobaric heating from 350 to 650°C between 130 and 290 km from the trench at a depth of 40 km. We used this thermal model for investigating thermally mediated

processes at Guerrero because it uses precise plate geometries and age parameters for this this location, compared to the thermal models developed by Syracuse et al. (2010) and van Keken et al. (2018), which average the subducting plate geometry over several hundred kilometers. We use the Manea and Manea (2011) model to determine the pressure-temperature (P - T) parameters to input into our petrologic models, and to determine the temperatures at which the various slip phenomena occur beneath Guerrero by pairing the thermal model with reported spatial distributions of slow slip and tremor events (e.g., Fig. 2).

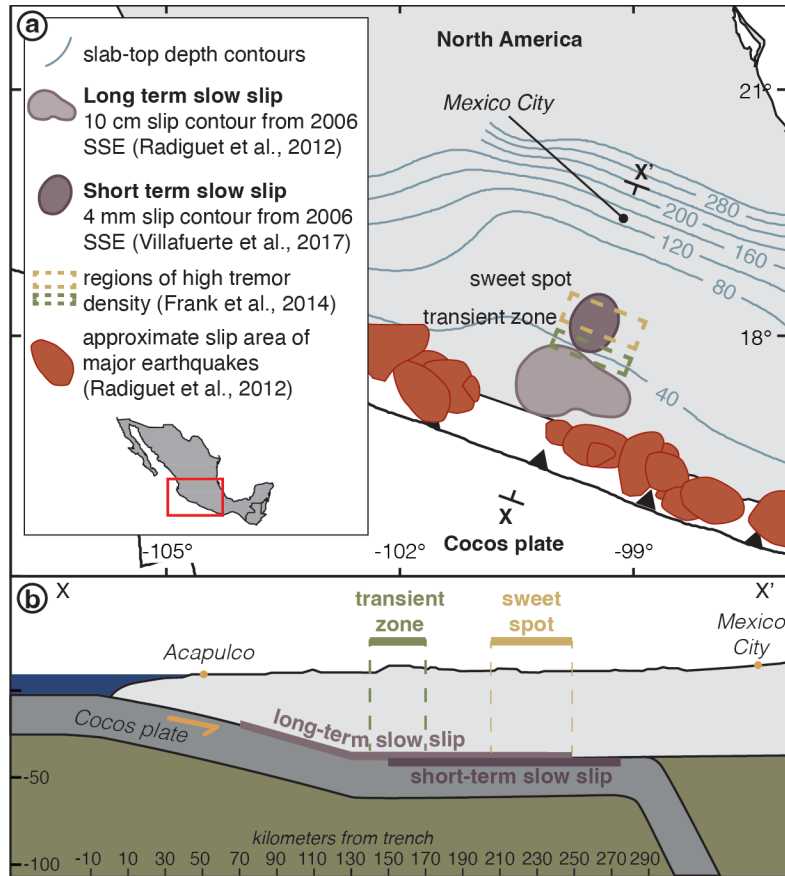


Figure 1. (a) Map of the region around Guerrero as shown by inset map of Mexico. Slab-top depth contours from the Slab2 model (Hayes et al., 2018) show the extent of the flat slab region beneath Guerrero, Mexico, and locations of seismic events are outlined. (b) Schematic cross section of the flat slab subduction showing the distribution of tremor and slow slip on the flat plate interface.

2.2 Geophysical observations of the plate interface and slip behaviors

At the modern Guerrero subduction zone, seismic data suggest the presence of a 3–5 km thick low seismic velocity layer above the oceanic crust of the subducting Cocos plate, separating it from the continental crust of the overriding North America plate (Song et al., 2009; Kim et al., 2010). Song et al. (2009) report that an ultra-low velocity (S-wave velocities ~ 2 – 2.7 km/s) layer is present at the plate interface in the shallowly-dipping to up-dip portion of the flat segment, which transitions to an uneven low velocity layer that extends from approximately 200

km from the trench to the down-dip end of the flat slab at 290 km. These low-velocity zones at the plate interface are suspected to represent mantle wedge material that was entrained during slab-flattening and has been serpentinized (Manea and Manea, 2011). Low V_p/V_s ratios and low V_s values suggest that this zone may contain significant quantities of talc (Kim et al., 2010), an interpretation supported by seismic anisotropy data which has been further interpreted to represent an asymmetric shear fabric defined by deformed and aligned talc (Song and Kim, 2012).

Episodic tremor and slow slip has been regularly observed geodetically in Guerrero since 1998 (e.g., Lowry et al., 2001). The occurrence of ETS in this region is of particular interest because of the lack of $M_w > 7.0$ subduction earthquakes in the ~100 km-wide northwestern part of the Guerrero seismic gap since 1911 (Fig. 1a) (Kostoglodov et al., 2003). SSEs accommodate some of the relative motion between the Cocos and North American plates, likely affecting the earthquake cycle and increasing the recurrence interval for large earthquakes relative to adjacent regions on the Middle American subduction zone (Radiguet et al., 2012). SSEs have also been related to large earthquakes in the Guerrero and Oaxaca regions, potentially serving as a dynamic trigger for several $M_w > 7.0$ earthquakes to the southeast of the Guerrero seismic gap by transferring stresses along the plate interface (Cruz-Atienza et al., 2021).

At Guerrero, the character of SSEs and associated tremor varies down-dip along the subduction interface (Fig. 1). Two distinct regions of tectonic tremor activity have been described, known as the sweet spot and the transient zone, respectively (Kostoglodov et al., 2010; Husker et al., 2012; Frank et al., 2014). The sweet spot is located 205–250 km from the trench and is characterized by near-continuous tremor activity, whereas the transient zone is located 140–170 km from the trench and exhibits tremor activity primarily during SSEs (Husker et al., 2012; Frank et al., 2014). Two types of SSEs also occur along the subduction interface and overlap with these tremor-producing regions: short-term and long-term SSEs. Short-term SSEs occur exclusively on the flat segment of the subduction interface, between 150 and 275 km from the trench, with a ~3-month periodicity and equivalent moment magnitude of ~6.4 (e.g., Frank et al., 2015; Villafuerte and Cruz-Atienza, 2017). Long-term SSEs occur along the up-dip portions of the flat and shallowly-dipping subduction interface between 70 and 250 km from the trench, recur every ~4 years, and have equivalent moment magnitudes of ~7.5 (e.g., Radiguet et al., 2012; Perez-Silva et al., 2021). Long-term SSEs are also spatially correlated with the observed low seismic velocity layer and local seismic anisotropy, suggesting that talc, which may be responsible for this region's unique seismic structure, may also facilitate SSEs (Song et al., 2009; Song and Kim, 2012). However, unless the rocks at the plate interface are chemically altered through silica metasomatism, it is unlikely that volumetrically significant talc will occur in this zone. We investigate the potential role of silica metasomatism in producing talc in this zone.

3 Mechanisms of talc production

The presence of talc beneath Guerrero was suggested by Kim et al. (2010) to be the result of silica metasomatism of mantle wedge rocks that were entrained during the flattening of the Cocos plate before 10 Ma. Silica metasomatism is one of several mechanisms for producing talc at the slab–mantle interface. This particular pathway involves the reaction of Mg-rich minerals—serpentine in the case of hydrated ultramafic rocks and dolomite in the case of carbonate rocks—with silica-bearing fluids (e.g., Manning, 1995; Collettini et al., 2009). These processes produce talc by effectively increasing the amount of SiO_2 relative to MgO in the rock, and so

metasomatic processes that also remove Mg from the rock may also promote talc production by passive Si enrichment. Other ways to produce talc include the addition of CO₂-rich fluids to serpentinite (e.g., Sieber et al., 2022; Okamoto et al., 2021; Spandler et al., 2008). The physical mechanism for the mass transfer involved in each of these may also be achieved through advection of elements in a fluid or diffusion between adjacent lithologies. New hybrid lithologies may also be produced through the physical mixing of various (potentially silica-rich) lithologies that are juxtaposed at the subduction interface (e.g., Penniston-Dorland et al., 2014; Hoover et al., 2022). Other studies of serpentinite metasomatism have employed reactive fluid flow models to predict the production of metasomatic minerals (e.g., Beinlich et al., 2020; Codillo et al., 2022b). Codillo et al. (2022b) suggest that talc will not be a significant metasomatic product in subduction zones based on results of reactive transport modelling, but this is in contrast with geologic studies that indicate talc formation by silica metasomatism (e.g., Hirauchi et al., 2020; Bebout & Barton, 1993). Rather than focusing on fluid compositions, our models simulate the alteration of rock compositions through advective silica transport specifically by focusing on the equilibrium mineralogy of a varying bulk-rock compositions.

4 Petrologic Modeling

To model chemical alteration of hydrated oceanic lithosphere entrained between the two plates along the flat slab, we created two coupled thermodynamic models of silica metasomatism, one focused on the amount of silica addition needed to create talc in hydrated mantle peridotites, and a second exploring the fluid and silica fluxes that are predicted from the subducting Cocos plate sediments and the hydrated top of the oceanic slab.

4.1 Modeling Approach

We model the equilibrium mineral assemblages of metasomatized serpentinite and subducting metasedimentary and basaltic rocks using the Gibbs free energy minimization software *Perple_X* 6.9.0 (Connolly and Petrini, 2002). To explore the silica metasomatism of hydrated peridotites, we created temperature-composition (*T*-*X* models) pseudosections at a constant pressure of 1.3 GPa—equivalent to ~40 km depth of the flat slab in the thermal model of Manea and Manea (2011)—and temperatures varying from 350°C to 650°C, encompassing the range of temperatures estimated for the plate interface at Guerrero over this flat segment. We also simulate the isochemical metamorphism of metasedimentary and metabasalts over this same pressure-temperature (*P*-*T*) path. Modeled pressures and temperatures are based on estimates of conditions at the plate interface that are extracted from the thermal model of Manea and Manea (2011). From the phase-equilibrium models, we can also extract physical properties of the modeled rocks, such as density, based on the predicted mineral assemblages.

Serpentinized peridotites are modeled in the chemical system CaO–FeO–MgO–Al₂O₃–SiO₂–H₂O–O₂ (CFMASHO) using the ds55 thermodynamic database of Holland and Powell (2004) and fluid equation of state model of Holland and Powell (1991). Metasedimentary and basaltic rocks are modeled using the MnO–Na₂O–CaO–K₂O–FeO–MgO–Al₂O₃–SiO₂–H₂O–TiO₂ (MnNCKFMASHT) and Na₂O–CaO–FeO–MgO–Al₂O₃–SiO₂–H₂O–TiO₂ (NCFMASHT) systems, respectively, and the ds62 thermodynamic database of Holland and Powell (2011) and fluid equation of state model of Pitzer and Sterner (1994). The phase solution models used for each lithology are reported in

Tables S1–3. All models were run at water-saturated conditions to simulate a minimum H₂O content considering contribution of fluids from the top-most portion of the subducting Cocos plate (the volcanic component and thin sedimentary cover), consistent with the expected fluid-rich environment at the plate interface and the presence of free water (e.g., Condit et al., 2020; Hernández-Urbe and Palin, 2019; Bebout and Penniston-Dorland, 2016).

4.2 Bulk compositions

4.2.1 Silica metasomatism of hydrated peridotite

We vary the silica content of the serpentinite bulk compositions to simulate the enrichment of these rocks in silica via metasomatism. Across a bulk composition based on an average serpentinitized harzburgite from the mantle wedge (Deschamps et al., 2013), we only vary the normalized silica content from that equivalent to hydrated peridotite (39 wt. % SiO₂) to that of MORB (49 wt. % SiO₂). We use the serpentinitized harzburgite composition to represent a refractory peridotite composition from the mantle wedge that has been hydrated by slab-derived aqueous fluids. Because the oxidation state of the forearc mantle wedge is poorly constrained, we run the serpentinite model with oxygen fugacity defined by variously reduced and oxidized redox buffers, ranging from $\Delta\log_{10}f\text{O}_2$ (FMQ) values of -2 to +2 to encompass a range of reported oxygen fugacity estimates for mantle wedge serpentinites (Deschamps et al., 2013). Complete initial (C₀—hydrated harzburgite) and final (C₁—silica-metasomatized harzburgite with 49 wt. % SiO₂) bulk rock compositions for both models of silica metasomatism are reported in Table 1.

Table 1. Bulk rock compositions in oxide wt. %

Oxide	Serpentinized Harzburgite C ₀	Serpentinized Harzburgite C ₁	Cocos sediment	Cocos basalt
Na ₂ O	—	—	2.5	2.05
MgO	39.84	33.34	2.87	8.53
Al ₂ O ₃	0.54	0.45	13.13	16.58
SiO ₂	39.5	49.37	54.24	49.18
K ₂ O	—	—	1.23	0.14
CaO	0.29	0.24	1.73	13.34
TiO ₂	—	—	0.57	0.85
MnO	—	—	2.33	—
FeO	6.83	5.72	8.5	7.26
Fe ₂ O ₃	—	—	2.82	1.54
H ₂ O	13	10.88	—	—

4.2.2 Silica and fluid production from the Cocos plate

We also model fluid production from subducted lithologies to examine the sources of fluids and silica. Bulk compositions for sedimentary and basaltic rocks are based on samples recovered from boreholes in the modern Cocos plate. Approximately 150 meters of sediment and seven meters of basalt were recovered from Hole 487 off the coast of Acapulco during Leg 66 of the Deep Sea Drilling Project (Watkins et al., 1982). The sedimentary rock bulk composition used here (Table 1) is based on the average bulk composition of the sediment column from Hole 487 reported in Plank and Langmuir (1998), with Fe_2O_3 content estimated from reported total Fe using the average $\text{Fe}^{3+}/\text{Fe}_{\text{Total}}$ ratio of 0.23 for metapelites reported in Pattison and Forshaw (2021). The sedimentary rocks recovered from Hole 487 were hemipelagic–pelagic muds and clays with very little carbonate content (Watkins et al., 1982). The basaltic bulk composition used here (Table 1) is based on an average of three bulk compositions reported for the basalt from Hole 487 (Joron et al., 1982), with Fe_2O_3 content estimated from reported total Fe using the average $\text{Fe}^{3+}/\text{Fe}_{\text{Total}}$ ratio of 0.16 for global average MORB (Cottrell and Kelley, 2021). This MORB composition is comparable to global average MORB compositions (cf. Gale et al., 2013).

4.3 Results of silica metasomatism modeling

Figure 2 shows the results of T-X modeling of silica metasomatism of serpentinized harzburgites at end-member redox conditions. The major minerals (antigorite, olivine, talc) behave comparably in both the reduced (Fig. 2a) and oxidized (Fig 2b) models. Antigorite is stable and abundant over almost all of the model space, but antigorite volume begins to decrease with increasing temperature starting around 500°C, with a concomitant increase in olivine abundance until antigorite completely breaks down around 650°C. Petrologic models were also constructed for intermediate redox conditions, but these do not show significant variations from the end-member models. Modeling results of other redox conditions can be found in Fig. S1.

In both oxidized and reduced settings, the addition of silica is required for talc growth in serpentinized harzburgites at the lower temperature conditions of the flat slab beneath Guerrero (Fig. 2). From the starting point of ~39 wt. %, the silica content must increase to ~43 wt. % in both oxidized and reduced settings in order for talc to stabilize. Adding additional silica to the system at these lower temperatures—350–500°C in the reduced model and 350–600°C in the oxidized models—increases the modal abundance of talc in the rock. Plots of the modal abundance of each mineral across the TX model at 450°C for each model indicates that the dominant talc-forming reaction is antigorite + $\text{SiO}_2 = \text{talc} + \text{H}_2\text{O}$ (Fig 2e-f). From low to high values of silica wt. %, the predicted modal abundance of antigorite decreases at the expense of talc formation (Fig. 2e-f). To achieve a rheologically significant ~10 vol. % talc in the rock, silica contents must be greater than 44–45 wt. % (Fig 2c-f). At higher temperatures of 580–620°C and greater, 15–20 vol. % talc can grow in serpentinized harzburgites with no additional silica, and greater modal abundance is reached with increasing silica metasomatism compared to lower temperatures.

The stability of minor phases besides antigorite and talc differs between the reduced and oxidized serpentinite models. Minor phases are typically present at less than 5 vol. % (Fig. 2e-f), such as amphibole (dominantly actinolite) and clinopyroxene (dominantly diopside). In the

reduced model (Fig. 2a), the increased stability of amphiboles between 515°C and 585°C results in a relative lower abundance of talc. In particular, between silica contents of 43 and 46 wt. %, talc disappears completely from the system. In the rest of the modeled space, a clinoamphibole phase is predicted to be stable with < 3 vol. %, but in this temperature range at $\text{SiO}_2 > 43$ wt. %, amphiboles are predicted to constitute up to ~25 vol. % of the rock near 585°C. In the oxidized model (Fig. 2b), a single amphibole phase is only ever present at ≤ 3 vol. % over the model space. Although more talc is produced for a given silica content at lower temperatures in the more oxidized model, at high temperatures talc is more abundant in the reduced model for low silica content (Fig. 2). At 600°C and 43 wt. % SiO_2 , for example, the reduced model stabilizes 15 vol. % talc, compared to only ~7 vol. % talc in the oxidized model (e.g., Fig. 2c vs 2d).

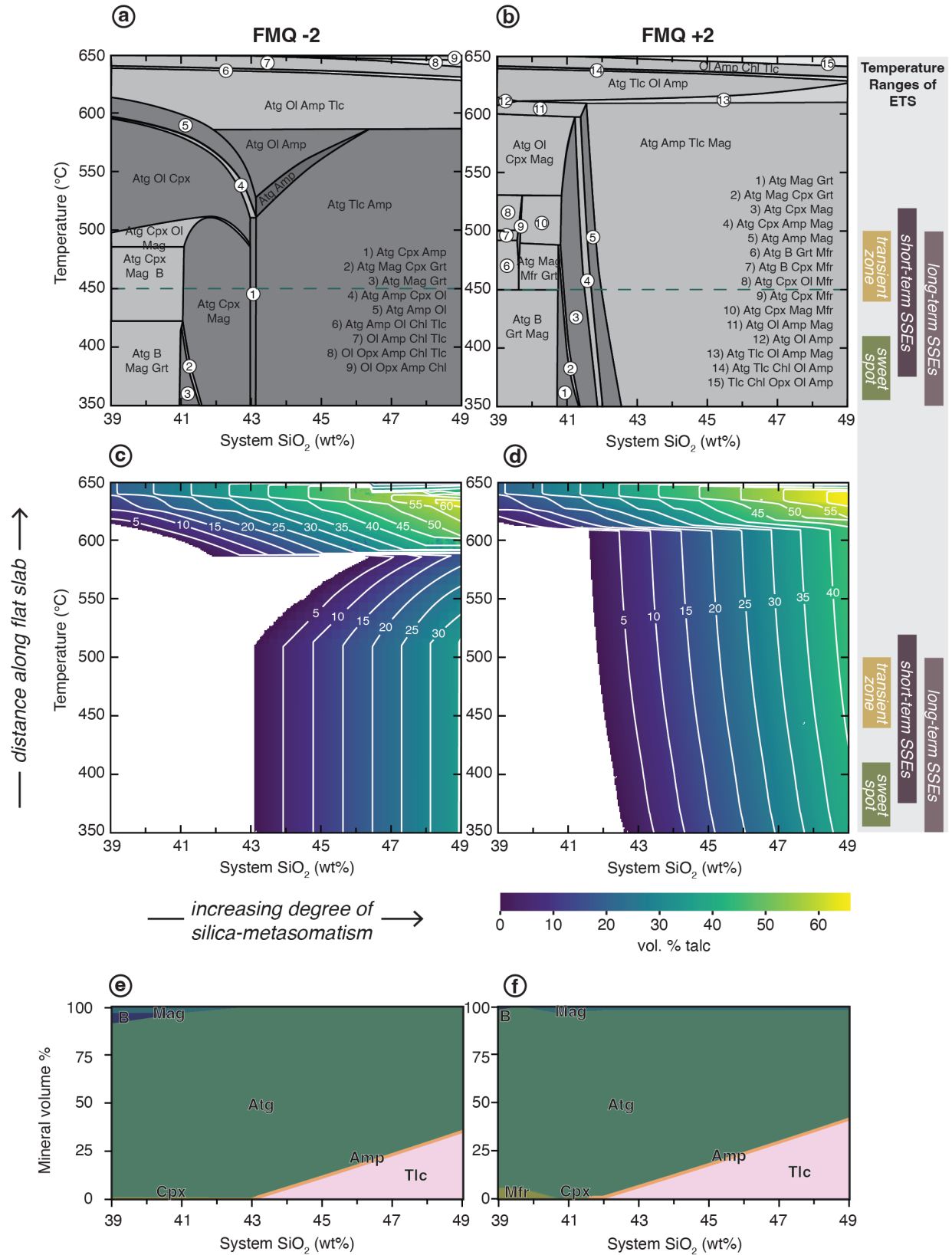


Figure 2. Silica metasomatism of serpentinized harzburgites at end-member redox conditions and 1.3 GPa and nominally water-saturated conditions. (a) and (b) show equilibrium mineral

assemblages at reduced and oxidized conditions, respectively. (c) and (d) show the modal abundance of talc predicted by these reduced and oxidized models, respectively. Both oxidized and reduced models show that silica concentrations must reach ~43 wt. % to stabilize talc over the temperature range at which slow slip and tremor occur beneath Guerrero. (e) and (f) show mineral modal abundance with increasing wt. % SiO₂ at 450°C—dashed line in (a) and (b)—in the reduced and oxidized models, respectively. Mineral abbreviations after Whitney and Evans (2011).

4.4 Results of modeling subducted lithologies

We also calculated thermodynamic models of metasedimentary and metabasaltic rocks from the subducting Cocos plate using a pressure-temperature path at 1.3 GPa over the range of flat slab temperatures to eventually estimate the volumes of metamorphic fluid production. Figure 3 shows the predicted modal abundance of mineral phases in the basaltic and metasedimentary rocks along this *P-T* path and the associated amount of mineral-bound water hosted in those phases. Decreases in the wt. % mineral-bound H₂O with increasing temperature in both models is inferred to correspond to fluid release from the subducting Cocos plate. Water-saturated metasedimentary rocks start with less than 4 wt. % H₂O and are predicted to dehydrate continuously along the flat slab segment (Fig. 3b), containing ~1 wt. % H₂O at 290 km from the trench. In contrast, the metabasalt model starts with 6 wt. % H₂O and exhibit three pulses of dehydration (Fig. 3a): (1) at ~425°C and 190–200 km from the trench, the breakdown of lawsonite releases >2 wt %H₂O, (2) at ~ 460°C and 220 km from the trench and (3) 520°C and 270 km from the trench of chlorite breakdown produce >0.5 wt% H₂O (Fig. 3a) at each location.

These dehydration reactions occur across very narrow temperature ranges and therefore short distances along the flat slab.

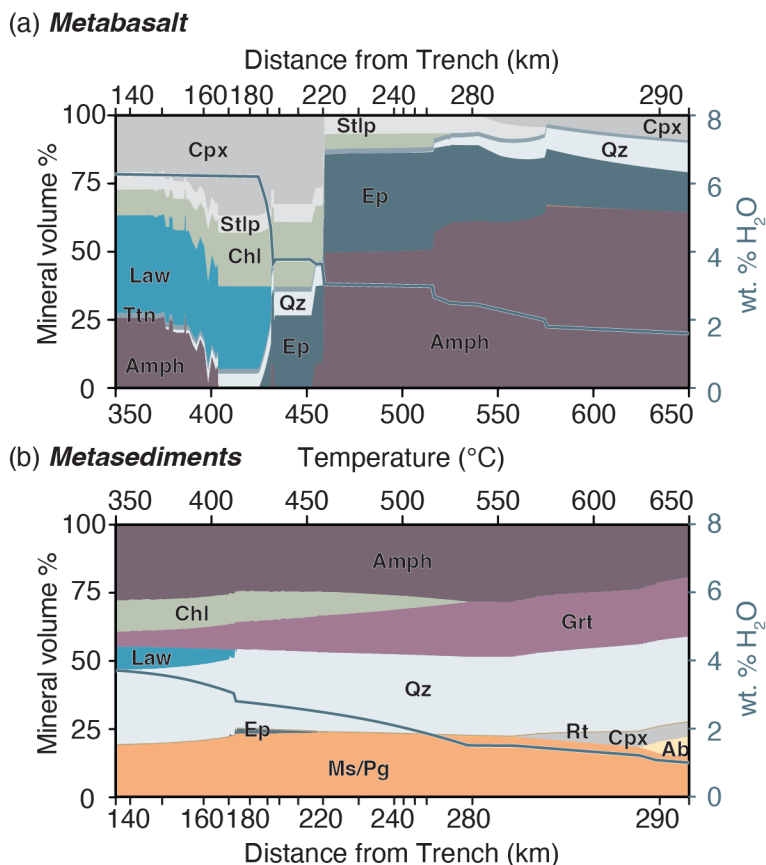


Figure 3. Plots of mineral modal abundance and the weight percentage of mineral-bound water (blue lines) for (a) metabasaltic and (b) metasedimentary rocks from the modern Cocos plate at 1.3 GPa and over the range of temperatures along the flat slab top beneath Guerrero.

5. Water and silica flux and time-dependent talc production

Using the predicted changes in the quantity of mineral-bound water and assumptions about the geometry and composition of the subducting Cocos plate, we estimate the volume of water released over the flat slab segment beneath Guerrero and the amount of silica that could be advected in solution into the overlying mantle wedge. We also compare this to the silica and water flux required to produce serpentinites with a rheologically significant 10 vol. % talc based on our T - X models of metasomatism.

We construct a water and silica flux model based on modern plate velocities and thickness estimates of the rocks that comprise the Cocos plate and the overlying mantle wedge (sect. 2.1). Using the thermal model of Manea and Manea (2011), we extract the temperature conditions of the flat plate interface at 40 km depth between 120 and 290 km from the trench.

We binned the flat plate interface into 5 km segments, and using the temperature data from the thermal model, we calculate the corresponding water content and rock density from the MORB and metasedimentary Perple_X models. For each segment, the amount of water lost by subducting rocks is calculated as the difference between the maximum and minimum water content (in kg/m^3) of the modeled lithology over the temperature range of that 5 km segment. Using the modern convergence rate of 60 km Myr^{-1} and the thickness of the water-producing rock (150 m of metasediments, 600 m extrusive basalt), we calculate the flux of water across the plate interface in $\text{kg m}^{-2} \text{ Myr}^{-1}$. This model assumes that all fluid released during metamorphism in a given 5 km-long region of the subducting Cocos plate is transported vertically into the overlying 5 km-long region of the subduction interface.

The structure of the subducting Cocos crust—i.e., the thicknesses of the layers of basalt and sedimentary rocks—is assumed to be comparable to what is presently outboard of the trench. We assume that the thickness of the sedimentary rock is 150 m—approximately the thickness of the sediment in hole 487 from the Cocos plate off the coast of Acapulco (Watkins et al., 1982). We use a thickness of hydrated MORB of 600 m, which is assumed to be a representative thickness of the layer of extrusive material based on seafloor drilling (e.g., Jarrard, 2003).

5.1 Water fluxes

Our water flux estimates show constant low levels of fluid release by the downgoing slab across the plate interface, interspersed with large pulses of water release that are localized where the subducting basalt undergoes metamorphic dehydration. Figure 4a shows the calculated flux of water across the plate interface averaged for each 5 km-long segment of the flat slab, and the relative contributions of metabasalts and metasedimentary rocks. Total water fluxes range from 10^3 to $3.5 \times 10^5 \text{ kg m}^{-2} \text{ Myr}^{-1}$ and average $3.9 \times 10^4 \text{ kg m}^{-2} \text{ Myr}^{-1}$ over the whole flat slab. Metabasalts contribute an average of 7.7 times higher flux than metasediments, or 89% of the total water flux over the flat slab as a result of both the greater volume of metabasalt than metasediments in our model crustal structure and the larger amount of mineral-bound water that can be carried by basaltic lithologies.

As expected from our modeling of subducted Cocos slab lithologies and their water content (Fig. 3), water is released from the Cocos plate in three main pulses, one at 180–200 km from the trench, one at 215 km from the trench, and another at 270–290 km. These regions of high fluid flux are associated with the three main pulses of metabasalt dehydration shown in figure 3. The first pulse of dehydration results in the greatest water fluxes up to $3.5 \times 10^5 \text{ kg m}^{-2} \text{ Myr}^{-1}$, the second pulse up to $1.6 \times 10^5 \text{ kg m}^{-2} \text{ Myr}^{-1}$, and the final pulse up to $1.5 \times 10^5 \text{ kg m}^{-2} \text{ Myr}^{-1}$. In all cases, our models suggest the fluid released from metabasalts is the dominant contributor to the pulse, demonstrating that the dehydration of metabasalts provides a much more significant

flux of water into the mantle wedge and along the plate interface than that of metasedimentary rocks, as has been observed in previous modelling studies (e.g., Condit et al., 2020).

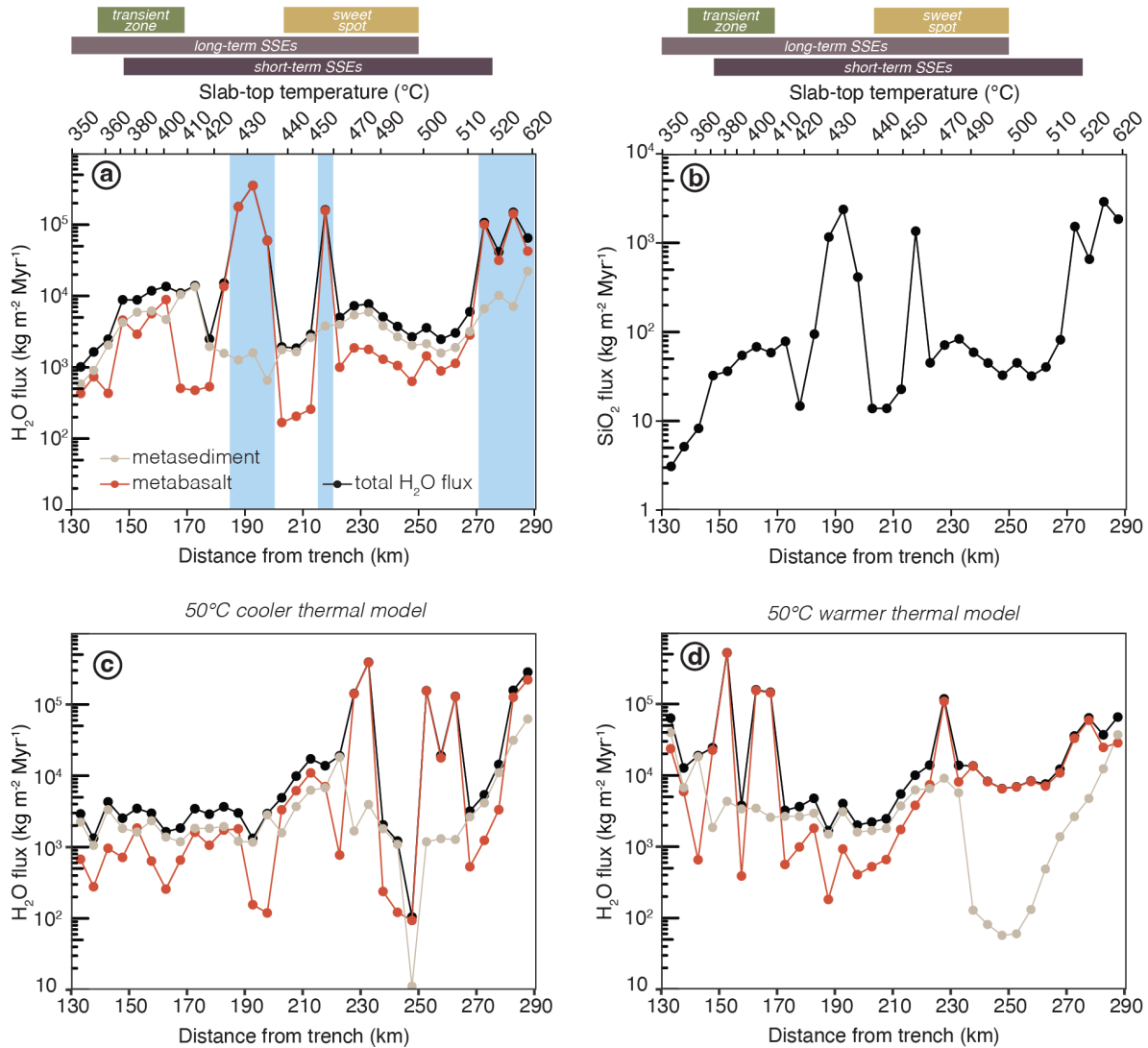


Figure 4. Estimates of (a) water and (b) silica flux into across the plate interface assuming vertical transport of all fluids released by the slab into the overlying interface. (a) shows the relative contributions to water flux of metasedimentary and metabasaltic rocks in the Cocos plate. Regions highlighted in blue are the main areas of high water flux into the plate interface. (c) and (d) show the effects of 50°C cooler and warmer thermal models, respectively, on the locations of water flux pulses, shifting the locations of large water and silica flux into the interface by up to 50 km from the original model in (a).

5.2 Silica fluxes

We combine the estimates of the water flux into the mantle wedge over the flat slab segments with an empirical model of silica solubility in water (Manning, 1994) to calculate the

amount of silica that could be advected from the subducting Cocos plate into the plate interface serpentinites. Assuming that all the silica that could be advected is added to the serpentinized harzburgites in the plate interface—which is reasonable given the silica undersaturation of peridotites—we calculate a silica mass flux across each 5 km segment of the flat plate interface (Fig. 4b). These calculations assume silica saturation of a pure H₂O fluid that is transported vertically into the plate interface.

Our model only considers the advection of silica in aqueous fluids, so the pattern of silica flux (Fig. 4b) generally follows the pattern of water flux (Fig. 4a), with low levels of silica flux along the entire flat slab punctuated by pulses of high silica flux where dehydration occurs in the subducting metabasalts. This silica flux is also dependent on temperature, which has a strong influence on the solubility of silica in aqueous fluids (e.g., Manning, 1994). In both models, only eight of the 5-km wide bins exhibit higher than average silica fluxes. As with the water flux, there are three primary regions of silica flux around 180–200 km, 215 km, and 270–290 km from the trench, and because silica solubility increases with temperature, silica flux is greater farther from the trench as the slab top continues to heat up. Silica flux across the plate interface ranges from 3.1 to 2900 kg m⁻² Myr⁻¹ and averages 410 kg m⁻² Myr⁻¹ across the entire flat slab. The first high-flux pulse of fluids around 190 km from the trench results in silica fluxes up to 2400 kg m⁻² Myr⁻¹. The second pulse at 215 km reaches 1400 kg m⁻² Myr⁻¹, and the final pulse from 270–290 km contributes up to 2900 kg m⁻² Myr⁻¹ (Fig 4b).

Additionally, recognizing that thermal models at these depths are associated with uncertainties of ± 50 °C (Peacock, 2009), we also calculate water flux from the subducting slab in a 50 °C cooler (Fig. 4c) and 50 °C warmer (Fig. 4d) thermal structure. These models provide an indication of the sensitivity of the predicted locations of dehydration pulses to the thermal structure of the subduction zone and uncertainties in thermodynamic modelling approach and the compositional variability of subducting rocks. As figures 4c and 4d show, a hotter or colder thermal model for Guerrero may shift the locus of dehydration—and therefore fluid and silica flux—by tens of kilometers in some cases. We further consider the implications of this uncertainty and sensitivity in the discussion.

5.3 Time dependent talc production by pervasive fluid flow

The estimated silica fluxes from the subducting slab then allow us to build a simple time-dependent model for talc production at the plate interface. We use this to predict the growth of talc along the flat slab over its minimum lifetime of 10 Myr (Ferrari et al., 2012). Pairing the silica flux estimates with our models of silica metasomatism, we can estimate the silica content of the serpentinite along the plate interface in million-year increments, and determine the modal abundance of talc for the corresponding T-X conditions. When considering the effects of these fluids on the bulk rock composition of serpentinites, we use two end-member models for the thickness of the overlying mantle material: a 4 km-thick zone of serpentinite based on the thickness of the observed low-velocity layer (Song et al., 2009; Kim et al., 2010); and a 100 m-thick zone. We use this range of thicknesses to simulate silica-bearing fluids reacting with different volumes of hydrated peridotite. This allows us to explore the effect of different interactive volumes at the plate interface on the ability of the zone to grow significant quantities of rheologically important minerals. The thinner 100 m-thick zone endmember may be a more geologically realistic interaction volume and reflect the possible confinement and channelization

of fluids at the plate interface (e.g., Bebout, 1991; Ganzhorn et al., 2019). However, to compare to the geophysical interpretations of Kim et al. (2010), we include the 4 km-thick model endmember as well.

Figure 5 displays the results of this time-dependent metasomatism model for each endmember oxygen fugacity and interface thickness parameter by plotting the initial wt. % SiO₂ in serpentinites for each 5 km segment of the flat slab and the final SiO₂ content after 10 million years, as well as the corresponding talc abundance. For reference, we also extract from each T-X model of silica metasomatism the values of SiO₂ content at which the serpentinite achieves 10 vol. % talc at the conditions of each 5 km segment (Fig. 5). As with the silica flux models, these calculations are based on the assumption that silica is advected in a pure H₂O fluid that is transported vertically into the plate interface. We further assume that the silica is homogeneously mixed with the serpentinite at the interface, where it instantaneously reacts to increase the silica content of the serpentinite. These models therefore only inform us of the potential for talc production by advective silica metasomatism through pervasive fluid flow.

For both reduced and oxidized serpentinites at the plate interface, the flux of silica into the 4 km-thick interface is not high enough anywhere along the flat slab to produce any appreciable talc in the plate interface after 10 million years. Assuming pervasive metasomatism of the 4 km-thick interface, the added silica would be too dilute to drive the bulk composition of the serpentinite to greater wt. % SiO₂ (Fig. 5a-b). In the 100 m-thick interface model, however, the magnitude of silica flux relative to the volume of interface rock is high enough drive up the silica content of the serpentinite (Fig. 5c-d). In this case, talc is produced in the serpentinite at the interface in the regions that experience large fluid fluxes at 180–200 km, 215 km, and 270–290 km from the trench.

Larger volumes of talc are stabilized by the oxidized serpentinites than the reduced serpentinites. In the first fluid pulse around 190 km, 27 vol % talc is produced in the reduced serpentinite compared to 33 vol. % talc in the oxidized serpentinite, and similarly at 280 km from the trench, up to 33 vol. % talc is produced compared to up to 43 vol. % talc in the oxidized serpentinite model (Fig. 5c-d). At the fluid pulse near 215 km from the trench, only the oxidized model achieves 10 vol % talc.

It should be noted that the wt. % SiO₂ values reported in Figure 5 are the amount of silica in the model rock, which is different from the wt. % SiO₂ values on the reported in Figure 2 that are the total amount of silica in our modeled system. The difference between these is a function of the amount of mineral-bound water, as the total amount of water in the system is fixed. Temperature-dependent differences in the amount of water that is mineral-bound in serpentinites accounts for the variations in the starting SiO₂ content of serpentinites in both reduced and oxidized models of metasomatism (Fig. 5).

Our simple time-dependent models show that talc production over 10 million years of flat slab subduction beneath Guerrero is limited by generally low silica fluxes due to low water production and silica solubility systematics along the length of the flat slab segment. Appreciable talc growth through 10 million years of pervasive silica metasomatism is only predicted to occur if the interface is thin so that there is a low interaction volume for the advected silica to metasomatize. Even so, areas of the interface that can achieve ≥ 10 vol. % talc in 10 Myr are

localized to regions where the subducting Cocos plate undergoes large pulses of dehydration that can produce sufficient volumes of silica-rich fluids (Fig. 5c-d).

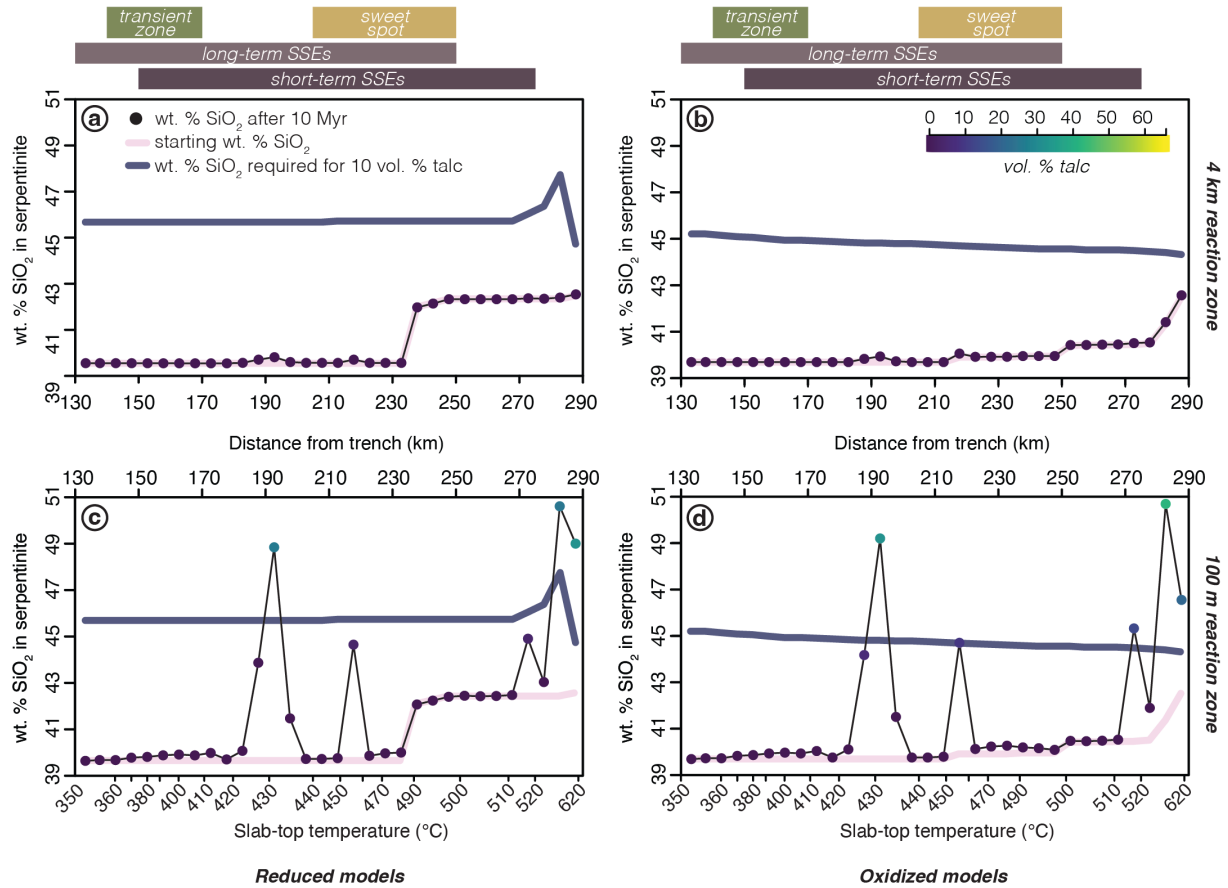


Figure 5. Models of the amount of talc in serpentinites at the plate interface based on flux estimates into 4 km (top panels) and 100 m thick (bottom panels) plate interface zones from Fig. 4, and reduced and oxidized serpentinite metasomatism models in Fig. 2, assuming 10 Ma of flat slab subduction and advective silica flux into the plate interface. Light pink lines show the amount of silica that serpentinites at the plate interface start with before the addition of silica-rich fluids, circles show the amount of silica predicted after 10 Ma of subduction and fluid flow and are colored according to the modal abundance of talc on the same scale from Fig. 2, and dark blue lines shows the threshold of silica required to produce a rheologically significant 10 vol. % talc for that part of the flat slab.

5.3.1 Calculating fluxes required for significant talc volume

Given the values of silica content at which our modeled serpentinites achieve 10 vol. % talc, we also perform a back-calculation to estimate the silica and water fluxes required at each 5 km segment of the flat slab to achieve this rheologically significant abundance of talc within 10 million years. We take the difference between the silica content of the serpentinites with 10 vol. % talc and their initial, unaltered silica contents to determine the silica flux necessary to produce $\geq 10\%$ vol. talc in 10 million years for both reduced and oxidized models (Fig. 6a). Again using

the silica solubility model of Manning (1994), we calculate the amount of water necessary to advect this target quantity of silica every million years (Fig. 6b). Figure 6 compares both target and predicted silica and water fluxes for the flat plate interface.

As expected, over the whole flat slab, the fluxes predicted by our mass balance calculations are much lower than these target silica and water fluxes for a 4 km-thick serpentinite interface. For example, even with the high silica fluxes at 190 km from the trench, the needed silica fluxes to the 4 km-thick interface are ~ 27 times larger than those predicted (Fig 6a). These silica fluxes would then require corresponding water fluxes into a 4 km-thick interface on the order of 10^6 – 10^7 kg m⁻² Myr⁻¹ (Fig 6b), which are ~ 10 – 10^4 times greater than the water fluxes actually predicted (Fig. 4a). In contrast, needed silica fluxes to produce significant talc in a 100 m-thick interface average around 1600 kg m⁻² Myr⁻² (Fig 6a), which corresponds to target water fluxes of $\sim 10^3$ – 10^4 kg m⁻² Myr⁻². As demonstrated in the previous section, our predicted fluxes into a 100 m-thick interface match or exceed these values in three locations (Fig 6), corresponding again to the locations of dehydration pulses in subducting metabasalt.

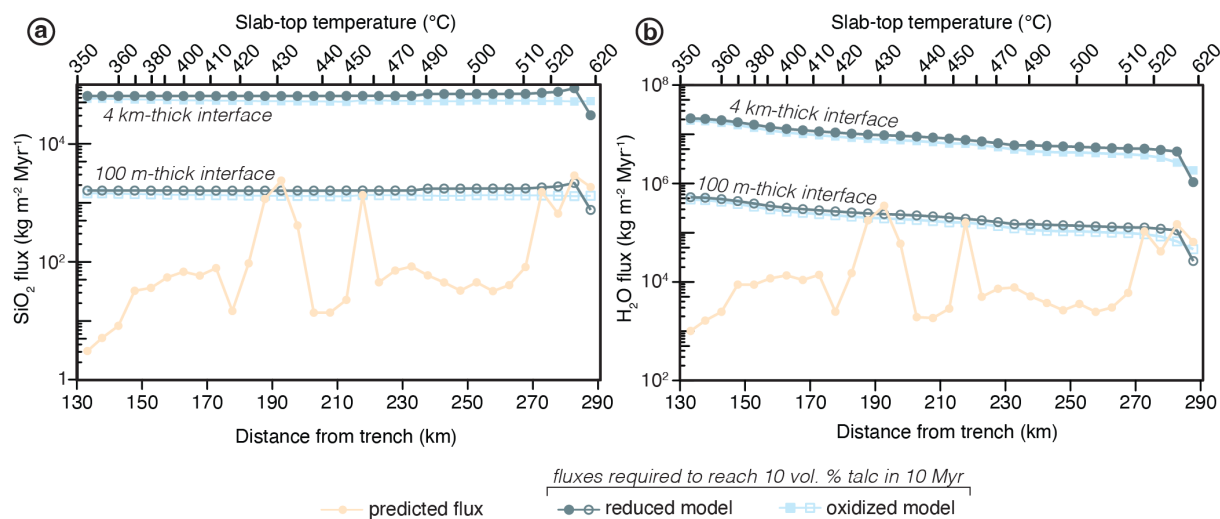


Figure 6. Estimates of the necessary (a) silica and (b) water fluxes into the plate interface to stabilize 10 vol. % talc in serpentinites along the flat slab beneath Mexico in 10 million years. Dark blue circles and light blue squares plot the necessary fluxes for reduced and oxidized models, respectively, and closed and open shapes correspond to the target fluxes for a 4 km-thick and 100 m-thick interface, respectively.

5.4 Time-dependent talc production by localized metasomatism at the plate interface

Whereas the previous section considered the effects of metasomatism by pervasive fluid flow resulting in the homogeneous distribution of silica through a prescribed thickness of serpentinite, we may also consider the case of localized metasomatism at the plate interface. In this case, all silica in the fluid released by the Cocos plate reacts with the serpentinite immediately at the plate interface and is scrubbed from the infiltrating fluid to produce a nearly pure talc schist with a bulk composition with ~ 61 wt. % SiO₂. Using the predicted silica flux

from the Cocos plate, we estimate what the thickness of a talc-rich layer of rock at the plate interface would be as a result of this type of metasomatism.

Again using serpentinite TX models, we calculate the difference between the silica content (kg m^{-3}) of an ultra-silicified serpentinite with the maximum amount of talc it could have (96–98 vol. % talc at ~61 wt. % SiO_2) and the initial silica content of serpentinites along the flat slab beneath Mexico (Fig S2). Dividing the predicted silica flux from the Cocos plate by this difference in silica content yields a value for the rate at which a zone of talc schist would vertically thicken at the plate interface (Fig. 7). These results indicate talc zone growth rates ranging from $4.4 \times 10^{-3} \text{ m Myr}^{-1}$ where silica fluxes are low closer to the trench, up to 5.4 m Myr^{-1} where dehydration reactions produce large pulses of silica-rich fluid. Because we calculate this vertical growth rate of a layer of nearly pure talc, this represents a minimum growth rate for a talc-rich layer at the plate interface, as layers containing less talc would require less silica, allowing for thicker talc-rich zones. Over 10 Myr, this would result in the growth of ~50 m-thick layers of talc schist at the high-flux regions of the plate interface, or on the order of several meters on average over the entire flat slab top.

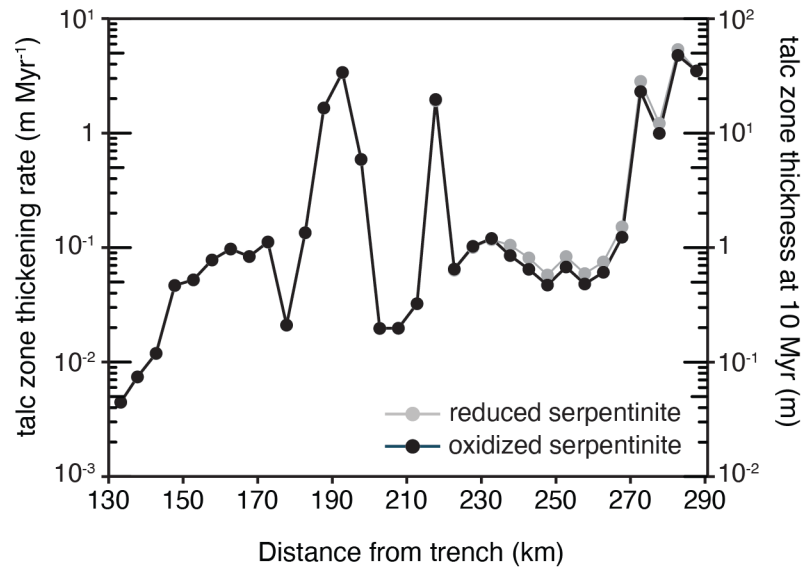


Figure 7. Calculated rate at which a zone of talc schist (96–98 vol % talc) at the plate interface would thicken for reduced (grey) and oxidized (black) serpentinites based on the predicted silica fluxes in figure 4b. Full thicknesses of the zones of talc schist after 10 Myr are given by the scale on the right.

6 Discussion

6.1 Silica metasomatism of serpentinites

Based on geophysical observations of the subduction zone beneath Guerrero, Mexico, a layer of talc has been inferred at the flat plate interface that hosts slow slip and tremor events (Kim et al., 2010). Our T-X models of serpentinitized harzburgites at the conditions of the flat slab top beneath Mexico indicate the necessary extent of advective silica metasomatism to producing significant quantities of talc at this plate interface through advective silica metasomatism (Fig. 2). For any talc to be present on the flat plate interface where slow slip and tremor occur beneath Guerrero—up to 520 °C at 1.3 GPa—our calculations indicate that the silica content of hydrated serpentinites must increase from ~39 wt. % to ~43 wt. %. The initial production of talc in these rocks is associated with the breakdown of clinopyroxene in the reduced model and the breakdown of antigorite in the oxidized model. Subsequent increases in talc in both sets of calculations are associated with decreasing abundance of antigorite. To stabilize 10 vol. % talc at the plate interface, serpentinites must achieve ~45 wt. % SiO₂. This amount of talc would be rheologically significant and would likely localize deformation at the plate interface (e.g., Hirauchi et al., 2013; Peacock and Wang, 2021). Over the temperature range of slow slip and tremor phenomena on the flat slab (~350–520°C), each additional 1 wt. % SiO₂ in the serpentinitized harzburgites will stabilize an additional 5 vol. % talc, and increasing the silica content of these to that of MORB (~49 wt. % SiO₂) stabilizes up to 35–40 vol. % talc (Fig. 2c, 2d). Chlorite, another weak phyllosilicate mineral that may influence the rheology of the plate interface, is not predicted to stabilize in our serpentinite models until >625°C, which is only reached at the farthest end of the flat slab from the trench.

Our results give us a lower bound for the amount of silica metasomatism needed to produce rheologically significant talc within hydrated entrained mantle wedge peridotites along the flat slab. These results are not significantly affected by redox state: our two end-member models for the redox state of serpentinite at the flat plate interface both show that the stability of talc in T-X space and the amount of silica necessary to produce talc only vary slightly between these two models (Fig. 2). In evaluating the ability of silica-rich water produced by the predicted dehydration reactions in our models in the next section, we keep both oxidation state end-member models in mind to understand the range of conditions that may allow the stabilization of large volumes of talc near the plate interface through silica-metasomatism.

6.2 Water and silica flux at Guerrero

6.2.1 Fluid production and slip behaviors

The production of metamorphic fluid through dehydration reactions is influenced by both the bulk rock compositions from the subducting slab in Guerrero and the *P-T* path these rocks take as they subduct. Both metasedimentary and metabasaltic rocks of the Cocos plate near the flat plate interface will undergo dehydration reactions and produce water in three pulses along the flat slab (Fig. 3, 4). The first pulse of fluid flux into the interface is the result of lawsonite and clinopyroxene breakdown into epidote and chlorite at ~425°C. The second pulse is associated with chlorite breakdown at ~460°C, and the third with chlorite, epidote, ±stilpnomelane breakdown over 520–575°C. Although this last pulse occurs over a wider temperature range, these temperatures occur at the end of the flat slab region near the mantle wedge where the flat slab top heats up rapidly through these temperatures so that they occur

within a 20 km region along the slab top. The minerals that are predicted to carry water in both the sedimentary and basaltic rocks of the Cocos plate are consistent with those in other petrologic studies of slab dehydration (e.g., Kerrick and Connolly, 2001; Condit et al., 2020.; Hernández-Urbe and Palin, 2019).

The locations of the two pulses of fluid production nearest the trench lie within the spatial extent of both long-term and short-term slow slip events. These fluids will carry silica that will likely metasomatize rocks at the plate interface, as discussed in the next section, and will contribute to the elevation of pore fluid pressures at the plate interface. Two of the main fluid pulses, at 180–200, and 270–290 km from the trench, are anti-correlated with the location of the primary regions of tremor activity: the transient zone and sweet spot 140–170 and 205–250 km from the trench, respectively (Fig. 4). However, the middle pulse that our models predict at 215 km from the trench lies in the range of the tremor sweet spot. Both short- and long-term SSEs overlap the predicted locations of the two main fluid pulses nearer the trench from our models (Fig. 4). However, both types of SSEs occur over much wider regions than the extent of the water and silica pulses. While considering uncertainties in thermal structure may shift the location of these pulses within the zones of SSEs (Fig. 4c, d), the spatial extents of local water and silica flux pulses into the plate interface are much narrower than that of SSEs.

Our results also show that the water production by metasedimentary rocks is significantly lower than that of metabasaltic rocks. In particular, each of the major pulses of water are dominated by fluids produced from the dehydration of metabasalts (Fig. 4a, b). This is consistent with the results of other authors (Manea and Manea, 2011; Perry et al., 2016), although the degree of dehydration for the modeled metabasalts considered here is greater because of the geologically realistic assumption of water-saturated conditions near the plate interface. Although this means that our calculations consider the maximum amount of water that could be carried by sedimentary and basaltic rocks in the Cocos plate, we only consider the contribution of water from the upper 600 m of volcanic rocks that make up the oceanic crust. Deeper portions of the oceanic crust are likely hydrated heterogeneously along fault zones that may penetrate into the mantle lithosphere as observed at other subduction zones (e.g., Naif et al., 2015). The release of these fluids from the oceanic lower crust and suboceanic mantle may also contribute to the water flux into the plate interface (e.g., Gutiérrez-Aguilar et al., 2022).

6.2.2 Silica flux and magnitude of talc metasomatism

Given the predicted fluid and silica flux from the Cocos plate, our models of time-dependent silica enrichment of serpentinites beneath Guerrero for end-member interface thicknesses of 100m to 4 km (sect. 5.3-4) predict the locations and potential magnitudes of talc production along the flat plate interface. With the silica fluxes that we calculated from the dehydration potential of the subducting slab, we find that our simple advective model of silica metasomatism does not predict pervasive talc production along the entire flat slab interface, which should be analogous to the low velocity layer of Kim et al. (2010).

The extent of silica enrichment depends both on the thickness of the serpentinite layer (i.e., the volume of rock with which fluids interact) and the oxidation state of these serpentinites (Fig. 5). The results of our time-dependent models show that after 10 million years of flat-slab subduction, talc production by this pervasive fluid flow can only occur in thin (~100 m) interface

zones, and is localized to portions of the flat slab where dehydration reactions in the subducting slab liberate large volumes of water. At these regions of major fluid production, serpentinitized harzburgites in a 100 m-thick zone may experience increases in silica content from 39 wt. % up to ~50 wt. % SiO_2 after 10 million years of fluid and silica flux from the subducting Cocos plate (Fig. 5). Oxidation state then influences the amount of talc that may be produced by similar changes in the silica content of reduced and oxidized serpentinites, with the latter generally stabilizing greater volumes of talc.

Importantly, it is difficult to produce rheologically significant quantities of talc in the 4 km thick interface models or enough talc to explain the seismic observations. In neither reduced nor oxidized models of pervasive silica metasomatism of a 4 km-thick zone does significant talc stabilize because the amount of silica introduced into the interface is not high enough to drive the silicification of such a large volume of serpentinite (Fig 6). These results suggest that, by itself, talc production by advection of silica from the subducting slab into the plate interface cannot explain the low V_p/V_s ratios and low V_s values in the ~4 km-thick layer imaged by Kim et al. (2010).

Considering our model of localized silica metasomatism at the plate interface (sect. 5.4), however, it is possible that meter-scale layers of talc-schist can grow immediately at the plate interface as silica-rich fluids react with “fresh” serpentinites and are scrubbed of silica (Fig. 7a). Unlike the other models based on pervasive fluid flow and homogeneous metasomatism of prescribed thicknesses of serpentinite (Fig. 5, 6), localized metasomatism at the plate interface would produce a layer of talc schist along the entire flat plate interface beneath Mexico. With an average thickness after 10 Myr of ~7 meters (Fig. 7b), such a zone of talc-rich rock is consistent with the geologic record of talc schists in exhumed subduction zone rocks, which typically occur in layers tens of centimeters to meters thick (e.g., Spandler et al., 2008; Hirauchi et al., 2020; Hoover et al., 2022).

In addition to the amount of silica introduced to the plate interface, oxidation state of the rocks at the plate interface may influence the effects of metasomatism (and vice versa). Oxidized serpentinites generally can stabilize larger volumes of talc than reduced serpentinites with the same silica content (Fig 2), and in our time-dependent models oxidized serpentinites do indeed achieve more talc than reduced serpentinites (Fig 5). The redox conditions of serpentinites in the mantle wedge are not well constrained (Deschamps et al., 2013), but more serpentinitized peridotites are likely to be more oxidized with higher $\text{Fe}^{3+}/\text{Fe}_{\text{total}}$ than fresh peridotites (Evans, 2008). Our oxidized T - X models may thus more closely represent the mineralogy of serpentinites at the plate interface in the Mexican subduction zone. However, the control of redox conditions on the abundance of talc at the plate interface mainly becomes significant at high temperatures, and the difference between oxidized and reduced models is thus not as relevant to the majority of the area of the flat slab that host ETS.

6.2.3 Sensitivity of fluid production to thermal structure

As Figs. 4c and 4d show, our predicted locations of dehydration are extremely sensitive to changes in the thermal model as shown in other subduction-related petrologic models (e.g., Hernández-Urbe & Palin, 2019; Kerrick and Connolly, 2001). In particular, the low thermal gradient down-dip along the top of the flat slab beneath Guerrero may result in large spatial

uncertainties in the location of water and silica fluxes. For the water flux pulses centered at 190 km and 215 km from the trench, this 50°C uncertainty has the potential to shift these loci of dehydration by ± 40 –50 km from the trench. The large pulse of fluids and silica that we predict occurs at the northeast end of the flat slab could be shifted 60 km trenchward if temperatures at the flat slab are 50°C higher than predicted in the thermal model that we use. This would place this large fluid pulse in the region of the tremor sweet spot. It is therefore difficult to definitively correlate the predicted locations of dehydration pulses with the distribution of slow slip phenomena at the plate interface. Even when considering the uncertainties in the exact location of dehydration reactions in the downgoing Cocos plate, however, the location of the water flux pulse nearest to the trench is always encompassed by the slip areas of both the short- and long-term SSEs (Fig. 4).

Previous petrologic modeling has also shown that the dehydration of metabasalts and metasedimentary rocks should provide significant pulses of fluid to the plate interface beneath Guerrero (Manea et al., 2004; Manea and Manea, 2011, Perry et al., 2016, Condit et al., 2020), although the predicted locations of dehydration pulses differs from the results of the models presented here. Manea and Manea (2011) and Perry et al. (2016) construct their own models for the thermal structure of the subduction zone at Guerrero. Perry et al. (2016) consider the added effect of fluid circulation in the subducting oceanic lithosphere in their models, which accounts for at least some of the difference between our predictions of fluid flux from the dehydrating oceanic lithosphere. The preferred model of Perry et al. (2016) predicts pulses of fluid flux near 140, 225–275, and >300 km from the trench. Manea and Manea (2011) predict that the majority of dehydration in the metasedimentary and metabasaltic rocks occurs between 230 and 250 km from the trench. This range overlaps with landward-end of the tremor sweet spot, from 205–250 km from the trench (Frank et al., 2014), however neither of these two studies pairing petrologic and thermal models at Guerrero have considered the effects of uncertainty in the thermal models that clearly can have a large impact on the predicted locations of dehydration reactions (Fig. 4e, f). In this work, we use the thermal model of Manea and Manea (2011) to establish the P - T parameters of our petrologic modeling, so the differences between our models and those of Manea and Manea (2011) may be due to differences in bulk compositions and phase-equilibrium model parameters. Manea and Manea (2011), for example, use phase diagrams from Rüpke (2004), who used bulk compositions of global average sediment compositions for metasedimentary models, and average carbonate-altered MORB from the Atlantic Ocean for metabasalt models, and constructed these models using an earlier generation thermodynamic database. These differ slightly in bulk composition from the local sedimentary and basaltic rocks of the Cocos plate near Guerrero that we use here, and this compositional variation likely contributes to the differences between our petrologic predictions.

6.3 Assumptions and Caveats

Both our petrologic models and mass balance calculations rely on several assumptions that are important to consider in interpreting these results. The petrologic models that form the foundation of this work assume that rocks are in thermodynamic equilibrium and thus do not account for the potential effects of reaction overstepping. However, as considered by Condit et al. (2020), rocks near the plate interface are likely to experience metamorphic reactions close to equilibrium P - T conditions because of the abundance of fluids, additional strain energy contributed through deformation, and, when considering the devolatilization of the subducting

plate, dehydration reactions that are associated with large entropy changes which reduce the magnitude of reaction overstepping. Nonetheless, in phase equilibria modelling, 2σ uncertainties of $\pm 50^\circ\text{C}$ and ± 0.1 GPa in the location of phase assemblage field boundaries are also expected as a result of uncertainties associated with end-member thermodynamic properties and the development of activity-composition models (e.g. Powell and Holland, 2008; Palin et al. 2016; Gervais and Trapy, 2021).

Our modeling of the silica metasomatism of serpentinites at the plate interface is also necessarily simplified in several respects. In our first set of analyses, we take a bulk-rock approach by assuming that silica enrichment occurs homogeneously throughout a body of rock. This effectively assumes that fluid flow is pervasive and occurs along grain boundaries, but does not account for the possibility of anisotropic fluid movement through highly deformed rocks (e.g., fracture networks, shear zones), which would be especially likely in the case of our localized silica metasomatism models producing meter-scale zones of talc schist (e.g. Fig. 7). The channelization and transport of fluids over 10s of kilometers along the plate interface has been documented in various exhumed subduction records (e.g., Angiboust et al., 2014; Bebout and Penniston-Dorland, 2016). As demonstrated in our modeling of water and silica flux into the plate interface, if the fluids that we predict to be produced by the subducting plate interact with smaller volumes of serpentinite, it is easier to drastically change the bulk composition of these rocks and produce talc. However, this bulk-rock approach to modeling the mineralogy of the plate interface under Guerrero is consistent with the uncertainties of geophysical observations, which are only able to resolve low-velocity zones with thicknesses on the order of kilometers. Producing a 3–5 km-thick zone of talc-rich serpentinite as predicted by Kim et al. (2010), requires extensive alteration that is appropriately modeled by varying the bulk-rock composition of rocks at the plate interface.

The silica metasomatism models also assume that the only chemical change in the rocks is the addition of silica to the composition of the serpentinites and that the fluid is pure H_2O . Talc may also be produced by the reaction of CO_2 with serpentinites (e.g., Okamoto et al., 2021), however, because of the lack of carbonate materials in sediments recovered from the Cocos plate (Watkins et al., 1982), we suggest that this is unlikely to be a significant process beneath Guerrero. Water is also expected to be the most abundant volatile comprising fluids released by the subducting slab, but solutes like Al and Ca, in addition to Si, will also be present in these fluids (Manning, 2004; Codillo et al., 2022b). The involvement of these elements could promote the growth of other minerals like chlorite, which was predicted to grow during some types of metasomatism based on reactive transport modeling (Codillo et al., 2022b), and is a common phase in metasomatic rocks exhumed from subduction zones that may be involved in the localization of deformation at the conditions of ETS (e.g., Ujiie et al., 2022; Hoover et al., 2022; Nishiyama et al., 2023). However, rheological modeling (French and Condit, 2019) and geologic evidence (Hoover et al., 2022) suggest that talc schists may preferentially localize deformation at slow slip strain rates over chlorite schists.

The mass-balance-based models of water and silica flux (Fig. 4, 5, 6, 7) assume that all the fluid being produced in each 5 km segment of the subducting slab is going directly into the corresponding 5-km segment of the overlying mantle. This approach ignores the possibility of fluid transport along the plate interface. Furthermore, because this model is based on the solubility of silica in pure water, and subduction zone fluids likely carry multiple solutes

(Manning, 2004), this calculation likely provides a lower bound on the quantity of silica that would realistically be advected in solution by fluids released from the dehydrating slab. Using the modern convergence rate of 60 mm/yr in estimating water and silica flux is also conservative because Cocos–North American plate convergence rates have been faster in the last 20 Ma (Sdrolias and Müller, 2006).

6.4 Implications for Guerrero and beyond: Talc metasomatism and slip behaviors

The simple model of silica advection beneath Guerrero that we present here suggests that silica-rich fluids produced by local dehydration in the upper <1 km of the subducting Cocos plate can alter the composition of overlying serpentinites enough to produce significant volumes of talc in these rocks. These predicted areas of talc production, however, are localized to the regions where increasing temperature causes the metamorphic dehydration of subducting metabasalts (Fig. 8). If talc is necessary to produce the observed low seismic velocity regions at Guerrero, local fluid and silica fluxes alone cannot explain the hypothesized alteration because silica fluxes are not high enough to increase the silica content of serpentinites to ≥ 43 wt. % along the entire region of the flat slab where the low velocity zone is imaged (Fig. 5). This suggests that either other mechanical and chemical processes contribute to the transfer of silica from the subducting Cocos plate into the overlying plate interface, or that the low V_p/V_s ratios and low V_s values imaged at the plate interface beneath Guerrero (Kim et al., 2010) may be produced by some combination of talc, other minerals, fluids, and rock structures. The 4 km-thick zone may also not reflect the size of the actively deforming and metasomatism region, but rather may be the result of progressive underplating of material, including thinner talc-rich zones, through time.

Although our models predict that the local advection of silica-rich fluids from the subducting slab into the plate interface is insufficient to produce thick talc-rich zones along the entire plate interface beneath Guerrero, it is possible that the migration of fluids within the plate interface may help to distribute silica and promote metasomatism in regions that are not directly adjacent to rocks in the subducting plate that undergo large dehydration reactions (Fig. 8). This process is consistent with geophysical observations at Guerrero, where variations in the rate of low-frequency earthquakes (LFEs) during slow slip events may be explained by the migration of pulses of pore fluids trenchward along the plate interface from the region of the sweet spot (Frank et al., 2015). This migration of fluids along the plate interface may provide a mechanism for distributing silica-rich fluids from regions of high local fluid flux to regions of the flat slab with low levels of local fluid flux, and similar fluid flow patterns have been suggested by stable isotope studies from the rock record (e.g., Jaeckle et al., 2018; Condit et al., 2022; Epstein et al., 2021). The creation of talc and other phyllosilicate minerals may contribute to the production of anisotropic permeability that can help to confine fluids to the plate interface region. Such impermeable barriers have been suggested as a explanation for geophysically inferred high pore-fluid pressures (e.g., Hyndman, 1988; Peacock et al., 2011), and have been suggested to channelize fluids and concentrate silica (Hyndman et al., 2015). Experiments on antigorite, on the other hand, suggest that deformed serpentinites may provide enough permeability along grain boundaries to allow fluid movement (Ganzhorn et al., 2019). The development of impermeable seals of talc schist, which are predicted by our localized metasomatism model (Fig. 7), may contribute to the redistribution of fluid along the plate interface (Fig. 8e) which might diminish

the amount of silica introduced into the regions that experience high degrees of local silica flux and increase the silica delivery to up-dip parts of the interface that have low local silica flux.

The contrast between our 4 km-thick, 100 m-thick, and localized metasomatism models also highlight the importance of the physical pathways of fluid flow at the subduction interface, as the same volume of water released by subducting rocks may cause different degrees of metasomatism depending on the volume of serpentinite with which it interacts (e.g., Fig. 8d vs Fig. 8e; Fig 5a,b vs. Fig 5c,d). A thin zone of talc produced by the metasomatism of serpentinite immediately adjacent to the plate interface, as predicted in our localized metasomatism models (Fig. 7), is likely rheologically significant, as suggested by the experiments of Hirauchi et al. (2013) in which less than 10 vol. % talc was sufficient to localize frictional slip. As shown in Fig. 4, the locations of pulses of high water and silica flux into the plate interface overlap with the spatial extent of slow slip events, even when considering possible uncertainties in the thermal models. Talc-rich rocks are therefore likely to be present at the plate interface within some of the region that hosts slow slip—if not the entire flat plate interface—even if lateral fluid movement does not lead to a wider distribution of metasomatized rocks over the flat plate interface. No matter which process of metasomatism is dominant beneath Mexico, it is likely that a small volume of talc-bearing rocks are present along some or all of the flat plate interface and influence the slip phenomena that occur there.

The model for silica flux and talc production presented here only considers advective silica metasomatism, but other mechanisms of metasomatism may contribute to talc production in subduction zones including mechanical mixing and diffusion. However, field based studies of exhumed rocks from accretionary wedges and deeper in subduction zones suggest that advective silica transport can be important and externally-derived fluids can transport large quantities of silica through subduction zones. For example, Breeding and Ague (2002), document extensive quartz veining in low metamorphic grade accretionary wedge rocks indicating up to 10% increases in rock mass through quartz precipitation. Externally derived fluids have also been invoked to explain some of the vein formation and metasomatism in higher-grade *mélange*, metasedimentary, and metabasaltic rocks of the Catalina Schist on the basis of trace element and oxygen isotope data (Bebout, 1991; Bebout and Barton, 1993). Our results therefore represent

the effects of one, likely important, end-member mechanism for silica metasomatism and talc production at Guerrero.

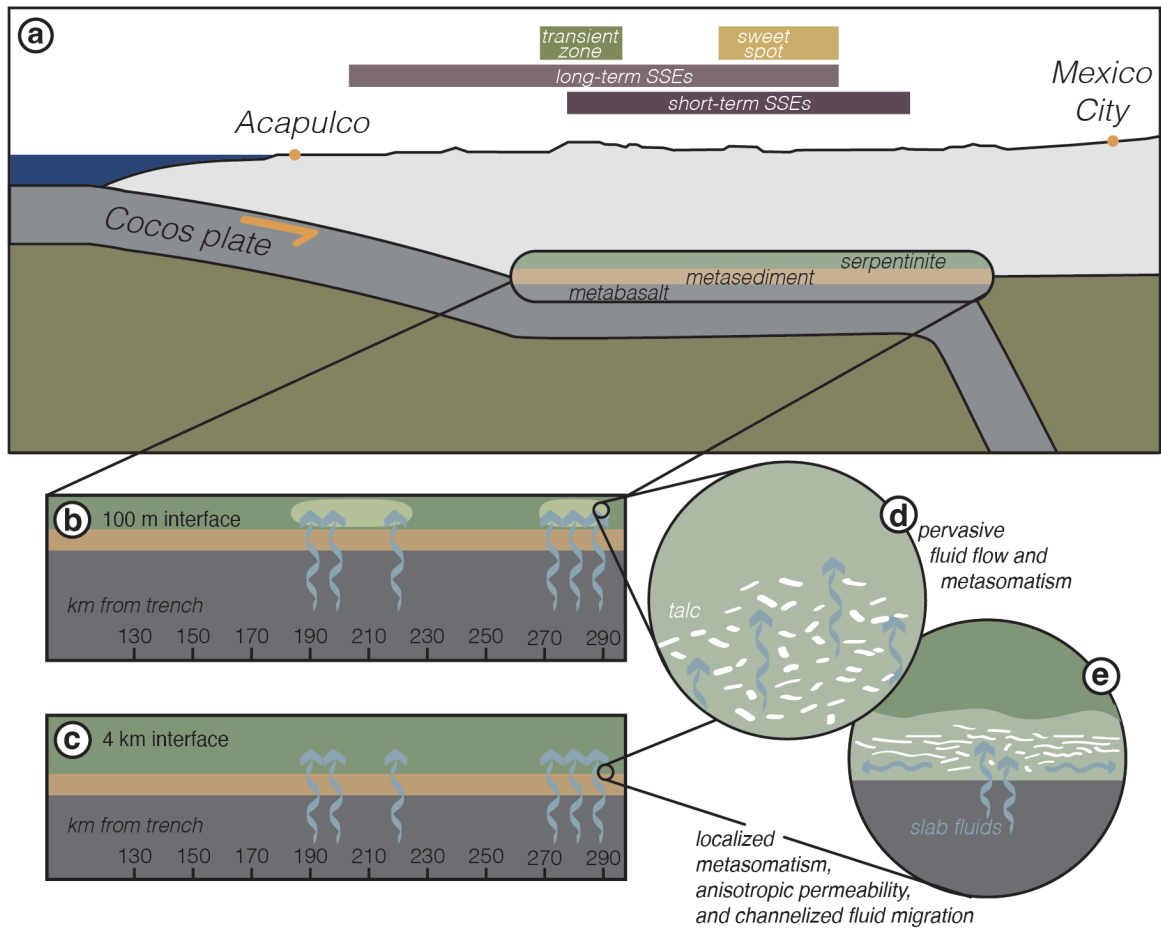


Figure 8. Schematic cross section of flat slab subduction beneath Guerrero (a) demonstrating the possible mechanisms for silica metasomatism of serpentinites at the plate interface that are explored here and comparing talc production in the (b) 100 m-thick interface and (c) 4 km-thick interface models. Blue arrows show the locations of major fluid pulses, and light green regions experience significant talc production as a result of silica metasomatism over 10 million years. The simple models of silica metasomatism considered here assume pervasive fluid flow and reaction of silica-bearing fluids with serpentinite, as in (d), but anisotropic permeability that develops in deformed talc-rich rocks as in (e) as a result of localized metasomatism that will likely occur even if pervasive metasomatism of the whole interface is unlikely, may act to channelize fluids at the plate interface and cause the migration of silica-rich fluids to the rest of

the interface where local fluid fluxes may not be high enough to promote significant talc growth, or produce pervasive talc growth as in the 4 km-thick interface.

7 Conclusion

We use petrologic models to explore the effects of silica metasomatism on serpentinites at the plate interface and the ability of silica advection to produce talc in these rocks above the flat slab beneath Guerrero. At the conditions of ETS beneath Guerrero, serpentinites must achieve ~43 wt. % SiO₂ to begin to stabilize talc, and ~45 wt. % SiO₂ to stabilize a rheologically significant 10 vol. % talc. Estimates of water production along the flat slab based on our petrologic models of the rocks that make up the subducting Cocos plate suggest that water and silica flux will only lead to silica metasomatism of the entire plate interface near sites of significant dehydration of metabasalts. However, localized metasomatism of serpentinite directly adjacent to the plate interface can easily produce thin zones of talc-rich rocks given the silica fluxes that we predict to occur along the entire flat slab. We also suggest that fluids are likely to migrate along the plate interface, distributing silica farther from the loci of dehydration, although silica advection by saturated aqueous fluids alone is still likely not enough to cause extensive metasomatism of the plate interface. We note, however, that other mechanisms such as mechanical mixing of more silica-rich rocks with serpentinites near the plate interface, and diffusion of silica may still lead to metasomatism and talc production. Although our estimates of talc production are not large enough to explain previous seismic velocity models of the flat plate interface beneath Mexico, our results indicate that rheologically significant volumes of talc are nonetheless likely to occur.

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Open Research

Archives of the whole-rock data used for the petrologic modeling as well as modeling results are available in a Zenodo repository (<https://doi.org/10.5281/zenodo.8034024>): <https://zenodo.org/record/8034024> (Lindquist et al., 2023).

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Chapter 3: Epidote records (de)hydration from the seafloor through subduction and underplating: implications for subduction zone fluid budgets and slow slip

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Key Points:

- Studied metabasalts record dehydration at ~300°C from a pumpellyite–epidote reaction, near the base of the subduction seismogenic zone.
- Networks of vein-like epidote structures may represent channel-forming reactive porosity developed during dehydration.
- Epidote is a powerful recorder of subduction and petrologic processes and can track pre-subduction, prograde, and retrograde metamorphism.

Abstract

When, where, and how much fluid is released from subducting oceanic crust control deformation, element cycling, and magmatism in subduction zones. However, the palimpsest of tectonometamorphic processes in the exhumed rock record complicates relations of geologic observations to geodynamic models of subduction fluid sources and transport. We present textural observations and major- and trace-element geochemical measurements of epidote combined with phase equilibrium modeling to decipher the pressure-temperature and dehydration history of metabasalts from the Catalina Schist Epidote Amphibolite Unit (California, USA). We distinguish four generations of epidote, representing (1) hydrothermal alteration of pillow basalts, (2) prograde epidote formation and fluid release during pumpellyite breakdown at 290–310°C and 0.4–0.7 GPa, (3) equilibration at peak *P-T* conditions of 550–600°C and 0.6–0.9 GPa, and (4) retrograde epidote (re)growth during underplating and refrigeration. Geochemical data suggest that basalts were hydrated at the seafloor, and during subduction, metabasalts experienced a large pulse of fluid release during pumpellyite breakdown, recorded by epidote-rich self-organizing porosity structures which helped to drain fluid from the subducting slab. This water was released from the subducting slab at depths corresponding to the base of the seismogenic zone and the shallowest depths at which deep slow slip and tremor occur in modern subduction zones, providing a geologic example of local metamorphic dehydration contributing to the maintenance of high pore fluid pressures at the plate interface. We show that epidote can serve as a recorder of progressive metamorphic processes when careful consideration of textures and mineral chemistry are applied.

Plain Language Summary

At subduction zones, water bound in the crystal structure of minerals is carried down into the mantle by oceanic plates. This water is released during chemical reactions that occur as rocks in the downgoing plate heat up and their constituent minerals dehydrate. The depths at which this water is released depends on the composition of the rocks and the specific minerals present. The amount and location of this water release controls things like the faulting or flow of rocks at the subduction interface, the cycling of elements through the Earth, and the source of magma for volcanic arcs. This work explores the history of these dehydration reactions in rocks of the Catalina Schist, which were subducted beneath North America during the Cretaceous and subsequently returned to the surface. We use the composition of epidote minerals and the textures that it forms to tease out conditions at which it grew, which often coincide with the breakdown of more water-rich minerals. In this case, we suggest that initial subduction-related epidote grew from the mineral pumpellyite at ~300°C and depths at which it is thought that water plays an important role in how subducting and overriding plates slip past each other.

1 Introduction

The composition and water content of the subducting oceanic crust play important roles in modulating deformation in subduction zones, melt generation at magmatic arcs, and element cycling on Earth (e.g., van Keken et al., 2011; Hacker, 2008; Peacock, 2009; Abers et al., 2013). The fluid inventory of the subducting slab, and where and how it releases fluids through pore collapse and metamorphic devolatilization reactions are especially critical for the mode of deformation at the subduction plate interface, including megathrust seismic slip, slow slip phenomena, and intermediate depth earthquakes (e.g., Saffer and Tobin, 2011; Condit et al., 2020; Abers et al., 2013; Hacker et al., 2003). High pore-fluid pressures can activate frictional

behavior at depth in subduction zones (e.g., Bürgmann, 2018; French and Condit, 2019; Peacock and Wang, 2021), and the presence of fluids within viscously-deforming rocks can enhance the activity of dissolution-precipitation creep (e.g., Wassmann and Stöckhert, 2013), facilitate the growth of weak hydrous phases (e.g., Bebout and Penniston-Dorland, 2016), and otherwise promote chemical changes at the plate interface (e.g., Rüpke et al., 2004; Hoover et al., 2022a). In addition, seismic velocity models of the plate interface in modern warm subduction zones reveal that slow slip and tremor are hosted in a fluid-rich environment (e.g., Audet et al., 2009; Shelly et al., 2006; Song et al., 2009; Wallace and Eberhart-Phillips, 2013). The sensitivity of slow slip and tremor to tidal modulation further suggests a weak plate interface, likely as a result of near-lithostatic pore fluid pressures (e.g., Rubinstein et al., 2008; Houston, 2015). Thermodynamic modeling of subducting lithologies has demonstrated a correlation between the depths at which dehydration reactions are predicted to occur and those at which deep slow slip and tremor events occur, suggesting a likely link between local metamorphic reactions and high pore-fluid pressures at the plate interface (e.g., Peacock, 2009; Fagereng and Diener, 2011; Condit et al., 2020).

Beyond the depths where free fluids are forced out of pore spaces by compaction, subducting oceanic crust carries water and other volatiles bound in minerals into subduction zones. These can be released by devolatilization reactions at forearc to subarc depths, or retained within the subducting slab as it descends further into the mantle, processes commonly explored through phase equilibrium modeling. Models predict that cold subduction zones may experience a single large release of fluid associated with the breakdown of lawsonite marking the blueschist to eclogite transition (e.g., Peacock, 1993), while warmer subduction zones may exhibit continuous or episodic fluid release (e.g., Schmidt and Poli, 1998). In these warmer subduction zones, it is a series of reactions at lower pressures than those of the blueschist-eclogite transition that produce fluids in the forearc region, often corresponding to the conditions at which slow slip and tremor occur (Peacock, 2009; Condit et al., 2020). The composition of the subducting oceanic lithosphere and the path it takes through pressure-temperature (P - T) space, however, are primary controls on which specific metamorphic reactions occur. This in turn controls the location and magnitude of the fluids that are released from the subducting slab and the elements that can be carried with them (e.g., Lindquist et al., 2023). The reactions that occur in subducting oceanic crust are therefore critical to understanding the water and mass budgets of subduction zones, and geological constraints on the behavior of fluids are necessary to understand the production and ultimate fate of fluids during subduction.

The mineral products of dehydration reactions can serve as recorders of these reactions. For example, at blueschist to eclogite conditions, garnet growth occurs due to the breakdown of hydrous minerals (e.g., Baxter and Caddick, 2013), and chemical zoning (major-, trace-element, and isotopic) of garnets can help constrain the timing and duration of fluid production (e.g., Dragovic et al., 2012; Broadwell et al., 2019; Hoover et al., 2022b). At low- to medium-grade metamorphic conditions where garnet is not stable, epidote group minerals may serve as a similar recorder of dehydration processes. With the general formula $\text{Ca}_2\text{Al}_2(\text{Fe}^{3+}, \text{Al})\text{Si}_3\text{O}_{12}(\text{OH})$, epidote group minerals are predicted by thermodynamic models to be a prograde reaction product following the breakdown of more hydrous phases (e.g., lawsonite, chlorite, pumpellyite; Condit et al., 2020). Epidote group minerals are stable across a wide range of P - T conditions—from prehnite-pumpellyite facies through epidote-blueschist, amphibolite and into eclogite facies (Franz and Liebscher, 2004). Epidote can also grow during seafloor hydrothermal alteration of oceanic crust, such as in epidote alteration zones or epidote veins as observed in both ophiolites

(e.g., Richardson et al., 1987; Gilgen et al., 2016) and modern oceanic settings (e.g., Banerjee et al., 2000; Chen et al., 2023). These minerals also act as significant hosts of trace elements, particularly Sr, Pb, and light rare-earth elements (LREEs; e.g., Frei et al., 2004; Spandler et al., 2003). Epidote may therefore be useful for tracking hydrothermal and metamorphic processes from the seafloor into subduction zones, perhaps especially as a proxy for dehydration reactions. In this work, we analyze the geochemistry of epidote minerals to look for evidence of epidote growth during (de)hydration processes and shed light on the reactions and processes that can affect subducting oceanic crust.

We use metabasalts from the Catalina Schist (Pimu'nga/Santa Catalina Island, CA) which represent an exhumed sliver of a subducted slab to explore the geologic record of (de)hydration of oceanic crust before and during subduction. From these exhumed metabasalts, we analyze epidote textures, major- and trace-element mineral compositions, and whole rock elemental compositions and Sr isotope ratios. We pair these textural and chemical observations with phase equilibria modelling to constrain a likely *P-T* path for the metamorphic history of these rocks. We further use these models to identify the likely reactions that liberated water during subduction and the conditions at which these reactions occurred, their relevance for deformation processes at the plate interface and slow slip and tremor phenomena, and demonstrate the utility of epidote group minerals as a tracer of subduction zone processes.

2 The Catalina Schist

The Catalina Schist on Pimu'nga/Santa Catalina Island (California, USA) is an exhumed paleosubduction terrane (Platt, 1975; Grove and Bebout, 1995). It was assembled via the incremental underplating of suites of metamafic and metasedimentary rocks from the downgoing oceanic crust during the Cretaceous. The oldest and structurally highest of these packages include amphibolite facies rocks (metamorphic conditions of ~700°C and 1.34–1.44 GPa) while the youngest and structurally lowest are lawsonite-albite to lawsonite-blueschist facies rocks (~300°C and ~1 GPa) (Sorensen and Barton, 1987; Penniston-Dorland et al., 2018; Harvey et al., 2021a; Sorensen, 1986; Platt and Schmidt, 2024). These higher- and lower-grade units are separated by an “intermediate” sequence of epidote-blueschist and epidote-amphibolite facies rocks (Platt, 1975; Grove and Bebout, 1995; Grove et al., 2008; Harvey et al., 2021a; Platt and Schmidt, 2024). The overall sequence of younging with decreasing metamorphic grade of tectonometamorphic units is inferred to reflect the progressive underplating of oceanic crust in a cooling subduction zone. Each of these units often include and may be bounded by *mélange* zones that contain metasedimentary, metamafic, metaultramafic, and metasomatized and hybridized lithologies thereof (e.g., Platt, 1975; Sorensen, 1988; Penniston-Dorland and Harvey, 2023; Condit et al., 2025).

The Epidote Amphibolite Unit includes a coherent suite of metabasalts, metasediments, and a *mélange* zone that is interpreted as a paleo-plate interface shear zone (Fig 1c; Hoover et al., 2022a; Hoover et al., 2025). Detrital zircon U-Pb ages from metasedimentary rocks within the Epidote Amphibolite Unit yield a maximum depositional age of 110–116 Ma, and hornblende ⁴⁰Ar/³⁹Ar ages yield metamorphic ages of 106–108 Ma and white mica ⁴⁰Ar/³⁹Ar ages yield ages of 97–101 Ma, the latter suggesting cooling of the unit through 400°C by that time (Suppe and Armstrong, 1972; Grove and Bebout 1995; Grove et al., 2008). Peak metamorphic temperatures have been estimated to be 546 ± 50°C and 534–603°C using Raman spectroscopy of carbonaceous material (RSCM; Platt and Schmidt, 2024) and Zr-in-rutile thermometry (Harvey

et al., 2021a), respectively. Platt and Schmidt (2024) also estimate peak pressures of 1.1–1.5 GPa based on the mineral assemblage of metacherts. Hoover et al. (2025) similarly estimate peak temperatures in the epidote amphibolite facies mélangé zone of ~550°C on the basis of Zr-in-rutile and RSCM, and pressures of ~1 GPa on the basis of phase equilibrium modeling. The exposed Epidote Amphibolite Unit is folded and overturned and thus preserves an inverted seafloor stratigraphy: metabasalts overlie metacherts and micaceous schists (Platt, 1976), which in turn overlies a mélangé zone that is interpreted as a paleo-subduction interface (Hoover et al., 2022a; Easthouse et al., 2025; Hoover et al., 2025). Hoover et al. (2025) used rutile and titanite U-Pb geochronology to estimate the timing of mélangé development and suggest that rutile ages of 100–106 Ma provide a minimum age for the metasomatism of metasedimentary lithologies, and that titanite, with ages of 89–104 Ma and commonly present as rims on rutile, records the infiltration of Ca-rich fluids during metamorphism and deformation in the mélangé. Hoover et al., (2025) suggest this Ca-rich fluid is sourced from the downgoing metabasalts—the rocks we focus on in this work.

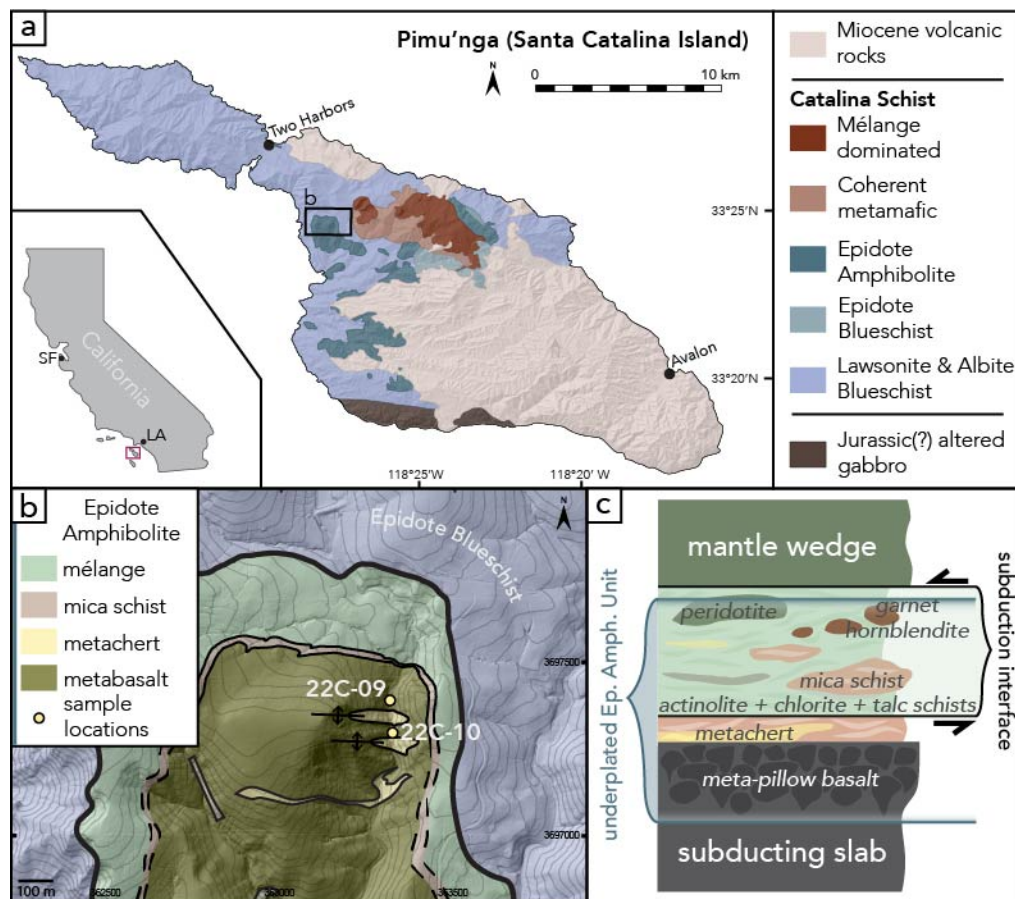


Figure 1. (a) Geologic map of the Catalina Schist exposed on Pimu'nga, after Platt et al. (2020). (b) detailed map of the Little Springs Canyon field area after Hoover et al. (2022a) and Platt et al. (2020). The Epidote Amphibolite Unit structurally overlies rocks of the Lawsonite Blueschist unit, with the epidote-amphibolite facies mélangé zone separating the two (Hoover et al., 2022a). (c) schematic lithostratigraphic column of the inferred tectonic setting of the Epidote Amphibolite Unit; adapted from Hoover et al. (2025).

We analyzed metabasalts from two ~100 m² outcrops exposed at the bottom of Little Springs Canyon (Fig. 1b). Although deformation fabrics are rare in the metabasalts, the outcrops of interest are located within what are inferred to be two synformal anticlines of metabasalt based on the presence of adjacent antiforms of metachert (Platt, 1976). The epidote-amphibolite metabasalts are dominantly fine-grained, with visible epidote and plagioclase porphyroblasts within an amphibole-dominated matrix. The metabasalts are interlaced with networks of epidote-rich veins. For the purposes of structuring our petrographic and geochemical results below, we distinguish here between epidote occurring in interpillow networks and occurrences of epidote in textural settings likely related to subduction metamorphism, which define two distinct sets of textures.

2.1 Interpillow networks

We observe pillow lava structures in several outcrops within this unit, as identified primarily by shapes outlined by epidote that commonly join together in “nodes” that we interpret as having originated as interpillow hyaloclastite or void-filling material (Fig. 2). Breccia textures that appear similar to seafloor basalt emplacement breccias are also present elsewhere in the Epidote Amphibolite Unit (Fig. 2e), though not in the outcrops that we will describe here.

2.2 Subduction-related textures

In addition to the interpillow epidote, these metabasalts are commonly crisscrossed by anastomosing networks of epidote-rich, vein-like zones that we interpret as forming during subduction metamorphism because of the presence of inclusions of metamorphic amphibole, titanite, and pyroxene in these zones (Fig. 2, 3, 4). These networks lack any apparent preferred orientation and can have wispy branches that terminate within centimeters. These vein-like epidote-rich zones (ERZs) are composed of ~mm to ~cm thick zones of coarse epidote minerals ± fine-grained amphiboles. We refrain from using “vein” here because, texturally, these epidote-rich bands appear distinct from the coupled brittle fracture and precipitation related features that are typically called veins. Boundaries between ERZs and the host metabasalt are not sharp, but rather gradational (Fig. 3). Epidote porphyroblasts (~mm-scale) are also visible within the fine-grained metabasalt that hosts the interpillow structures and epidote-rich zones. Although the majority of metabasalts in Little Springs Canyon lack any clear deformation fabric, ~2 m² of one of the outcrops exhibits a weak foliation defined by the preferred orientation of fine-grained amphibole crystals and a lineation defined by the preferred orientation of single epidote crystals and aggregates. These deformation fabrics appear to postdate the earliest development of metamorphic epidote-rich networks as these structures appear deformed and dismembered within the fabric. These fabrics may be related to deformation during or after underplating of the Epidote Amphibolite Unit.

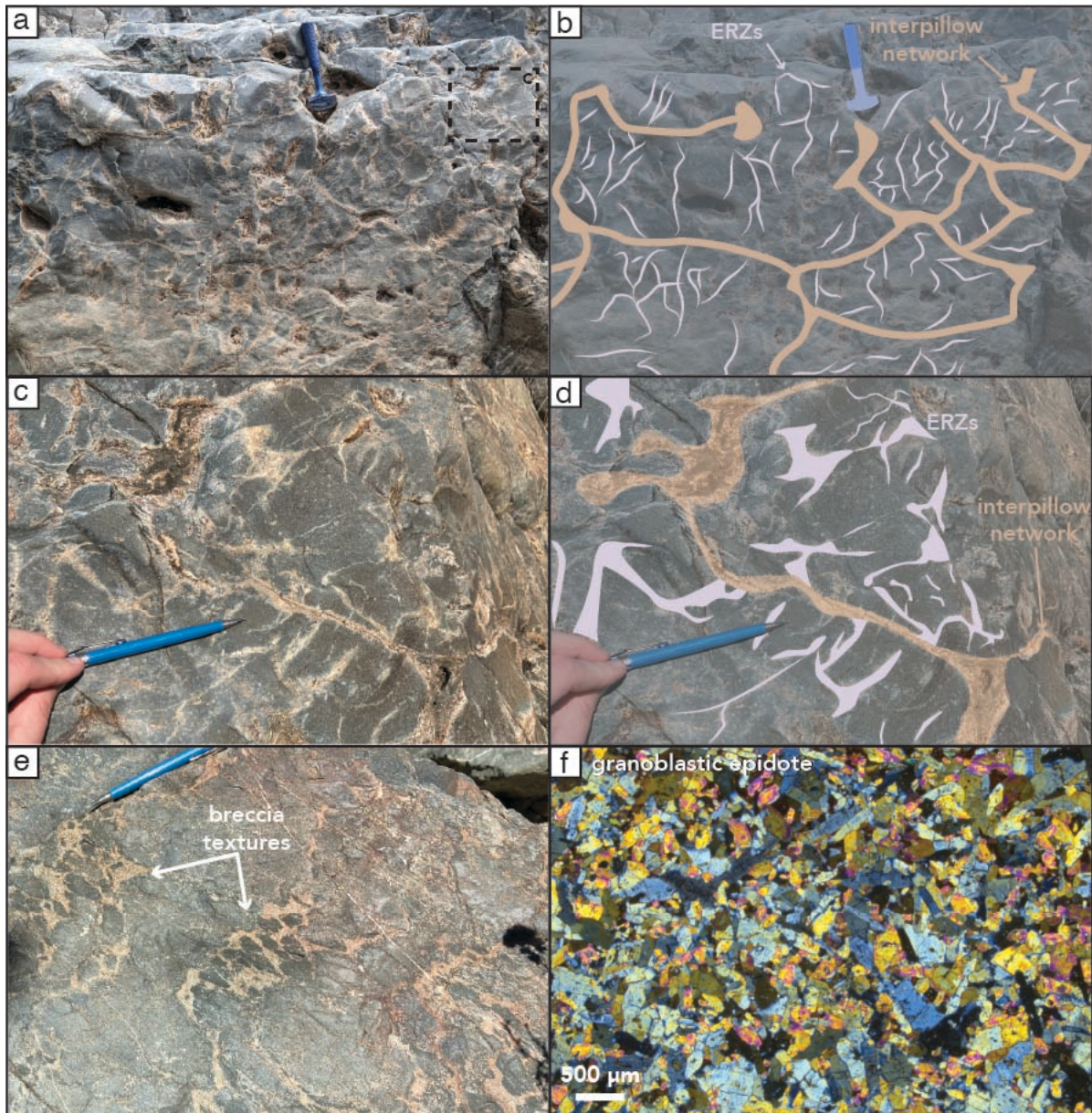


Figure 2. Examples of interpillow and other epidote textures in the field and in thin section. (a-b) Outcrop scale view of meta-pillow basalts showing the network of interpillow epidosite and smaller, less connected networks of epidote-rich zones. (c-d) A closer view of ERZs interacting with interpillow networks of epidosite. (e) Breccia textures we interpret as having originated as hyaloclastite breccias preserved in Little Springs Canyon. (f) XPL photomicrograph view of interpillow epidosite in which granoblastic epidote exhibiting a range of optical properties because of zoning, intergranular compositional differences, and orientation.

3 Methods

3.1 *In-situ* analyses

3.1.1 EPMA and SEM

Major element analyses of amphibole, epidote, and pyroxene were conducted at the University of Nevada, Reno with a JEOL JXA-iHP200F electron probe micro-analyzer using a focused beam with a 15 kV accelerating voltage and 20 nA beam current for spot analyses. For epidote-group minerals, we assume that all iron and manganese measured via EPMA are present as Fe^{3+} and Mn^{3+} , respectively, and distinguish the composition of epidotes based on the epidote fraction, $X_{\text{Ep}} = \text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Al} - 2)$, where $X_{\text{Ep}} = 1$ represents endmember epidote, and $X_{\text{Ep}} = 0$ is endmember clinozoisite (e.g., Franz and Liebscher, 2004). Although many of our analyses here indicate compositions with more clinozoisite-rich compositions ($X_{\text{Ep}} < 0.5$), we use “epidote” here for simplicity to refer to the whole spectrum of epidote-group minerals. For mineral recalculations of amphiboles, we follow the method of Locock (2014), and base amphibole nomenclature on A-site occupancy, Si apfu, and Mg-number after Leake et al. (1997).

Epidote, amphibole, and plagioclase were also analyzed on a Cameca SX100 at the University of Massachusetts, Amherst, where the full-thin section map in Figure 3 was also collected. Spot analyses were collected at 15 kV accelerating voltage and 120 nA beam current, whereas the full-thin section map used a 100 nA beam current. The full-thin section map in Figure 4a–c was collected by EDS in the Molecular Analysis Facility at the University of Washington on an Thermo-Fisher Scientific Apreo SEM with an Oxford Instruments Ultim-Max 100 energy dispersive spectroscopy detector using a 20 kV accelerating voltage and 3.2 nA current.

3.1.2 LA-ICP-MS

In situ trace element analyses on epidote minerals were collected using a Thermo Scientific iCAP Q Quadrupole inductively coupled plasma mass spectrometer (ICP-MS) in the Keck Collaboratory for Plasma Spectroscopy at Oregon State University. Analyses were conducted using a constant 5.5 mJ beam energy on 30 μm spots, and a 10 Hz laser pulse rate. Timing of analyses included 25 s of background, 35 s of ablation, and 10 s of washout. Three standards were analyzed after blocks of 20 sample analyses. Analytes included: ^{29}Si , ^{45}Sc , ^{47}Ti , ^{51}V , ^{52}Cr , ^{59}Co , ^{60}Ni , ^{85}Rb , ^{88}Sr , ^{89}Y , ^{90}Zr , ^{93}Nb , ^{133}Cs , ^{137}Ba , ^{139}La , ^{140}Ce , ^{141}Pr , ^{146}Nd , ^{147}Sm , ^{153}Eu , ^{157}Gd , ^{159}Tb , ^{163}Dy , ^{165}Ho , ^{166}Er , ^{169}Tm , ^{172}Yb , ^{175}Lu , ^{178}Hf , ^{181}Ta , ^{208}Pb , ^{232}Th , and ^{238}U . Data from individual analyses were processed and reduced using the LaserTram-DB software (Lubbers et al., 2025) with GSD-1G as the calibration standard, and ^{29}Si as the internal standard, and BCR-2G and ATHO-G as secondary standards.

3.2 Whole-rock analyses

3.2.1 Bulk rock major elements

Whole-rock major element abundances were measured using a Malvern PANalytical Zetium X-ray fluorescence (XRF) vacuum spectrometer at Franklin and Marshall College, and ferrous iron concentrations were determined by titration using the methods described in Mertzman (2000) and Waterton et al. (2020). Materials for bulk-rock XRF analysis were cut from samples so as to avoid weathered surfaces, and material for the purpose of phase

equilibrium modeling was cut from host metabasalt, avoiding both epidosite and epidote-rich zones.

A set of whole-rock powders were also analyzed via ICP-MS at the GeoAnalytical laboratory at Washington State University. Trace element data from these analyses were used for Rb-decay corrections in our analysis of whole-rock Sr isotopes.

3.2.2 Strontium (Sr) Isotope Ratios

Rock powders produced for whole-rock XRF analysis were also analyzed for Sr isotopes. Samples of 50–130 mg of powder were weighed out and digested in sequential digestions of HF and HNO₃ solution and HNO₃. Strontium was isolated with column separations using Eichrom Sr-Spec resin, and isotopic ratios were measured using a Nu Plasma multi-collector ICP-MS following procedures described by Brach-Papa et al. (2009). Strontium isotope ratios were corrected for mass fractionation to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$ and normalized to NIST987 ($^{87}\text{Sr}/^{86}\text{Sr} = 0.710240$) by sample-standard bracketing. External precision is ± 30 ppm (2σ), and in-run precision and accuracy were monitored via repeated analyses of the IAPSO seawater standard, which yielded 0.709165 ± 0.000008 (2σ).

3.2.3 Phase equilibrium modeling

To determine the *P-T* path taken by the Epidote Amphibolite Unit metabasalts during subduction based on our petrologic observations, we modeled equilibrium mineral assemblages for one sample using the Gibbs free energy minimization software Perple_X (v. 7.1.5; Connolly, 2005; Connolly, 2009). We chose a representative metabasalt sample that did not include any metamorphic epidote-rich zones or interpillow epidosite material for these analyses, the composition of which (determined by XRF and titration) is reported in Table S1. We calculated models in the chemical system Na₂O–CaO–FeO–Fe₂O₃–MgO–Al₂O₃–SiO₂–H₂O–TiO₂, and we use the internally-consistent thermodynamic database of Holland and Powell (1998) and fluid equation of state model of Holland and Powell (1991). Potassium was excluded from the modeled composition because of its low abundance (0.25 wt. %) in the bulk-rock major element analysis, and white mica in the sample from which the bulk rock composition is derived is primarily observed as an inclusion in other minerals and does not appear to be associated with the equilibrium mineral assemblage. The activity-composition models used in our modelling are reported in Table S2.

4 Epidote textures: Petrography and compositions

Below, we present the results of our petrographic and geochemical analyses on epidote in the context of the two main textural classes introduced in Section 2: interpillow epidosite (Fig. 2), and subduction-related epidote-rich zones (ERZs; Fig. 3) and porphyroblasts (Fig. 4-5). We differentiate between generations of epidote (Epx) on the basis of textural associations, our mineral chemistry data (Fig. 6), and epidote REE data (Fig. 7). Throughout, we use the term ERZ as a description of the textural setting of some subduction-related epidote, while Epx represents an interpretation of the generation of epidote growth as constrained by both texture

and major- and trace-element chemistry. We have also summarized key textural and geochemical characteristics of each textural domain in Table 1.

4.1 Epidosite (Ep_0)

As introduced in Section 2.1, basalt pillows are outlined by interpillow epidosite networks (Fig. 2A). In thin section, the interpillow epidote is dominated by fine-grained, granoblastic epidote (Ep_0 ; Fig 2f). Chondrite-normalized REE patterns from Ep_0 have relatively flat slopes, but exhibit strongly negative Ce anomalies (Fig. 7a), with Ce/Ce* values spanning 0.2–0.7, while Ce/Ce* of non-interpillow epidote range from 0.8 to 1.2 (Fig. S3a). Ep_0 have X_{Ep} values of 0.28–0.47 (Fig. 6a), and have concentric or very fine-scale oscillatory zoning (Fig. S1). Epidosites are difficult to differentiate from the other epidote-rich features described below in the field, but can be distinguished based on petrographic and geochemical analyses, particularly their negative Ce anomalies.

4.2 Subduction-related epidote

Outside of interpillow epidosite networks, epidote occurs as porphyroblasts throughout the matrix of the metabasalts and as epidote-rich zone (ERZ) structures (introduced in Section 2.2). These two groups exhibit textural and geochemical differences, so we discuss these separately below. Both ERZs and porphyroblastic epidote occur in an amphibole-dominated matrix that includes actinolite + hornblende/pargasite/edenite + albite + titanite $\pm$ white mica $\pm$ biotite, with micas only present in the matrix of more K-rich metabasalts than that used to inform phase equilibrium modeling. In some samples, clinopyroxene from peak metamorphic conditions is present as inclusions within masses of amphibole (hornblende + actinolite) and titanite that define clinopyroxene pseudomorphs (Fig. S1). Albite is commonly present in ERZs and as porphyroblasts with inclusions of amphibole and, rarely, micas (Fig S3). While both actinolite and hornblende are present within the albite porphyroblasts, actinolite is the dominant inclusion and hornblende is always rimmed by actinolite (Fig. 6b). Magnesiohornblende, pargasite, and edenite are more common in the matrix, but are always rimmed by actinolite. Representative mineral compositions are shown in Tables S3, S4, S5, and S6, and in Figure 6.

4.2.1 Epidote-rich zones (Ep_1)

Epidote-rich zones (ERZs) are mm- to cm-wide tabular features that extend for ~1–10s cm, typically narrowing towards their terminus, that crisscross the metabasalts in our field area, sometimes merging with interpillow epidosite zones (Fig. 2c,d). Relative to the surrounding host rock, the bulk composition of ERZs is slightly enriched in REEs (Fig. 7f). In thin section, ERZs are dominated by large, typically anhedral epidote crystals occurring with variable amounts of albite, pargasite, titanite, white mica, and rare clinopyroxene (Fig. 3). Epidote is typically rich in inclusions of fine-grained amphiboles (both pargasite and actinolite), titanite, quartz, and white mica (e.g., Fig. 3f). In general, both within individual ERZs and across samples, epidote compositions vary significantly, with X_{Ep} ranging from 0.16 to 0.67, sometimes within the same epidote crystal. Epidote crystals within ERZs show two main types of zoning patterns: patchy, irregular zoning or more coherent concentric zoning, sometimes with oscillatory zoning visible (Fig. 3f). This zoning is apparent optically and in BSE images and EDS maps and defined primarily by the Fe/Al content (Fig. 3, 4). In concentrically zoned epidote, cores are primarily,

but not always, defined by more Fe-rich epidote that grades to more clinozoisitic compositions at the rims, with some growth zones of more Fe-rich epidote.

Elemental X-ray maps show discrete regions with low Ti-abundance (primarily a result of the absence of titanite) that are commonly present within these epidote-rich zones (Fig 3c, 4c). In one ERZ sample (22C10B), the ERZ and associated Ti-poor regions are accompanied by seams of titanite within the ERZ and ~100 μm discrete titanite crystals that outline epidotes in the matrix (Fig. 3). Epidote major and trace element compositions do not vary significantly between Ti-poor zones and the surroundings.

In some ERZs, we observe dark patches (in XPL and BSE images) with tabular shapes that we refer to as ‘ghost’ grains (Ep_{1g}) in the cores of concentrically zoned epidote crystals (Fig 3, 4; cf. Whitney et al., 2020). These ‘ghost’ grains have clinozoisitic compositions with some of the lowest X_{Ep} values we measure, averaging 0.25. The epidote surrounding these ‘ghost’ grains is typically patchily or concentrically zoned with higher X_{Ep} values. Ep_{1g} also show distinct trace element abundances compared to the crystals in which they are found (Fig. 7b). ‘Ghost’ grains, and most epidote in ERZs exhibit relatively flat REE patterns with subtle enrichment in HREE compared to LREE (Fig. 7b,c). This pattern is apparent in chondrite-normalized $\text{La}/\text{Lu}_\text{N}$ values of 0.3–1.3 (mean of 0.73) for Ep_{1g} , 0.04–1.5 (mean of 0.6) for ERZ Ep_1 in sample 22C10B2 (Fig S3c). One exception to this pattern is the ERZ in sample 22C09A2 (Fig. 7d), which shows relatively flat, but mostly negative REE slopes ($\text{La}/\text{Lu}_\text{N} = 0.6\text{--}6.6$; Fig. S3d). Within this ERZ, analyses from the core of the ERZ differ subtly from those made on the margin of the ERZ near the surrounding matrix, with the former exhibiting concave up REE profiles (mean $\text{Dy}/\text{Dy}^* = 0.8$), and the latter exhibiting concave down profiles (mean $\text{Dy}/\text{Dy}^* = 1.3$; Fig. 7d, S3d).

One of the rare samples of deformed metabasalt contains evidence for deformed epidote-rich zones that have been transposed into the amphibole-defined foliation and boudinaged (Fig. 4). These transposed ERZs contain the same Ti-poor regions (Fig. 4c) that are visible within undeformed ERZs (Fig. 3c). Deformed samples also include strongly-zoned epidote crystals within the ERZs and as porphyroblasts with syn-kinematic rims (Fig 4). These rims include both magnesiohornblende and actinolite, suggesting that deformation occurred during peak through early retrograde metamorphism. The cores of some epidote crystals in these deformed epidote rich zones also contain ghost grains (Fig. 4d,e), sometimes situated in Fe-rich cores (X_{Ep} values of 0.59, decreasing to 0.41 towards the rims; Fig. 6a). In deformed samples, the epidote surrounding these ghost grains has REE patterns that differ from undeformed examples of Ep_1 . The surrounding epidote typically has higher concentrations of LREEs compared to the ghost grains, with REE patterns defining a negative slope that decreases to HREE abundances comparable to that in the ghost grains (Fig. 7b), which is also apparent in $\text{La}/\text{Lu}_\text{N}$ values (Fig. S3b). While textural observations suggest that this epidote originated in ERZs, geochemical data from the synkinematic growth zones of these epidotes are highly variable. LREE-enriched, concave down patterns of some bright (in BSE) zones more closely resemble those patterns of porphyroblastic epidotes described in the following section than they do the REE patterns in undeformed epidote-rich zones.

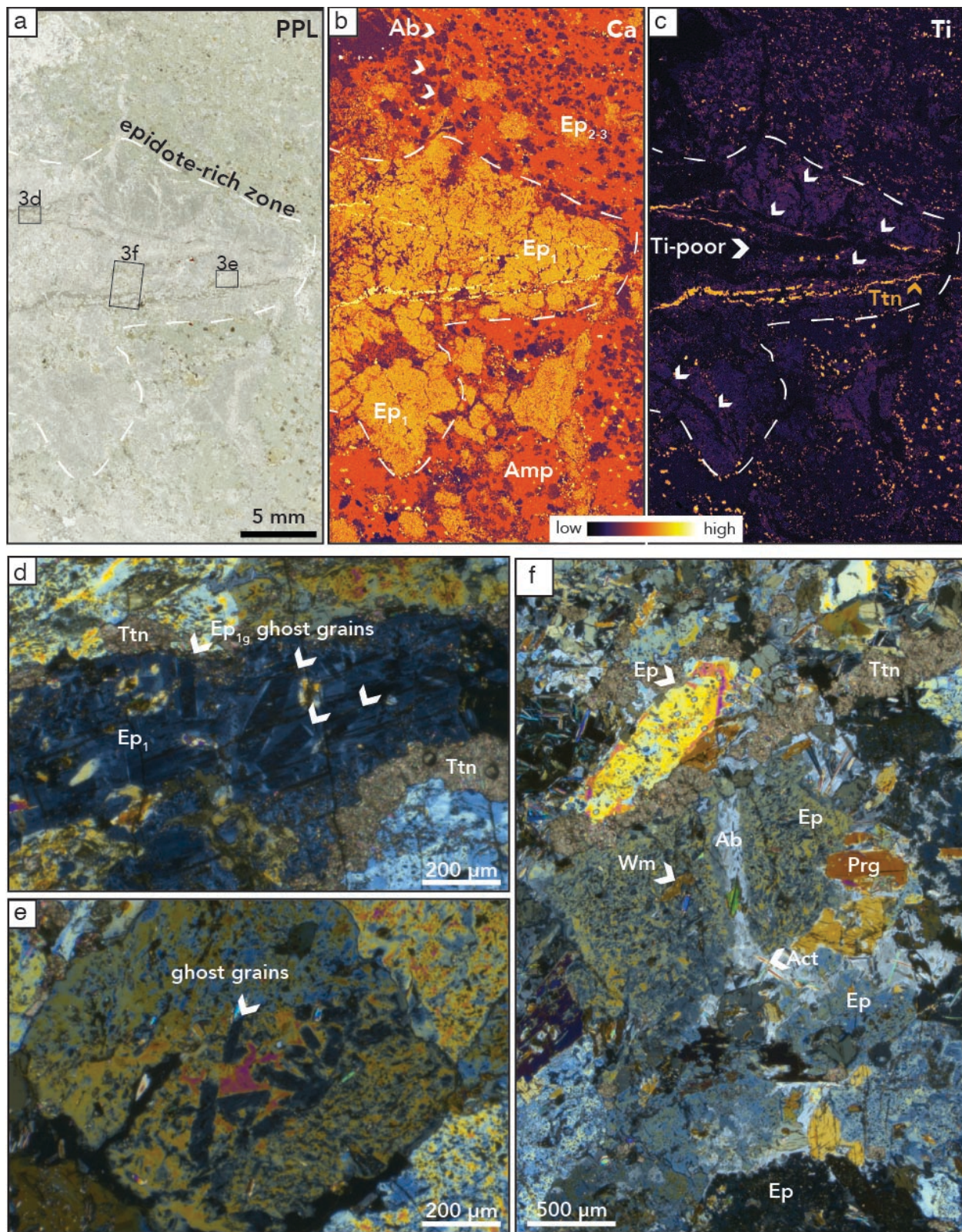


Figure 3. Figure 3. Textures of an undeformed epidote-rich zone. (a-c) show optical (a) and WDS (b-c) scans of full thin sections. White dashed lines outline the epidote-rich zone, which includes discrete Ti-poor zones visible in the Ti WDS map in (c). XPL photomicrographs (d-e)

show examples of ghost grain textures within the epidote-rich zone, and (f) shows examples of mica and amphibole inclusions therein. Mineral abbreviations in this and subsequent figures follow Whitney and Evans (2010).

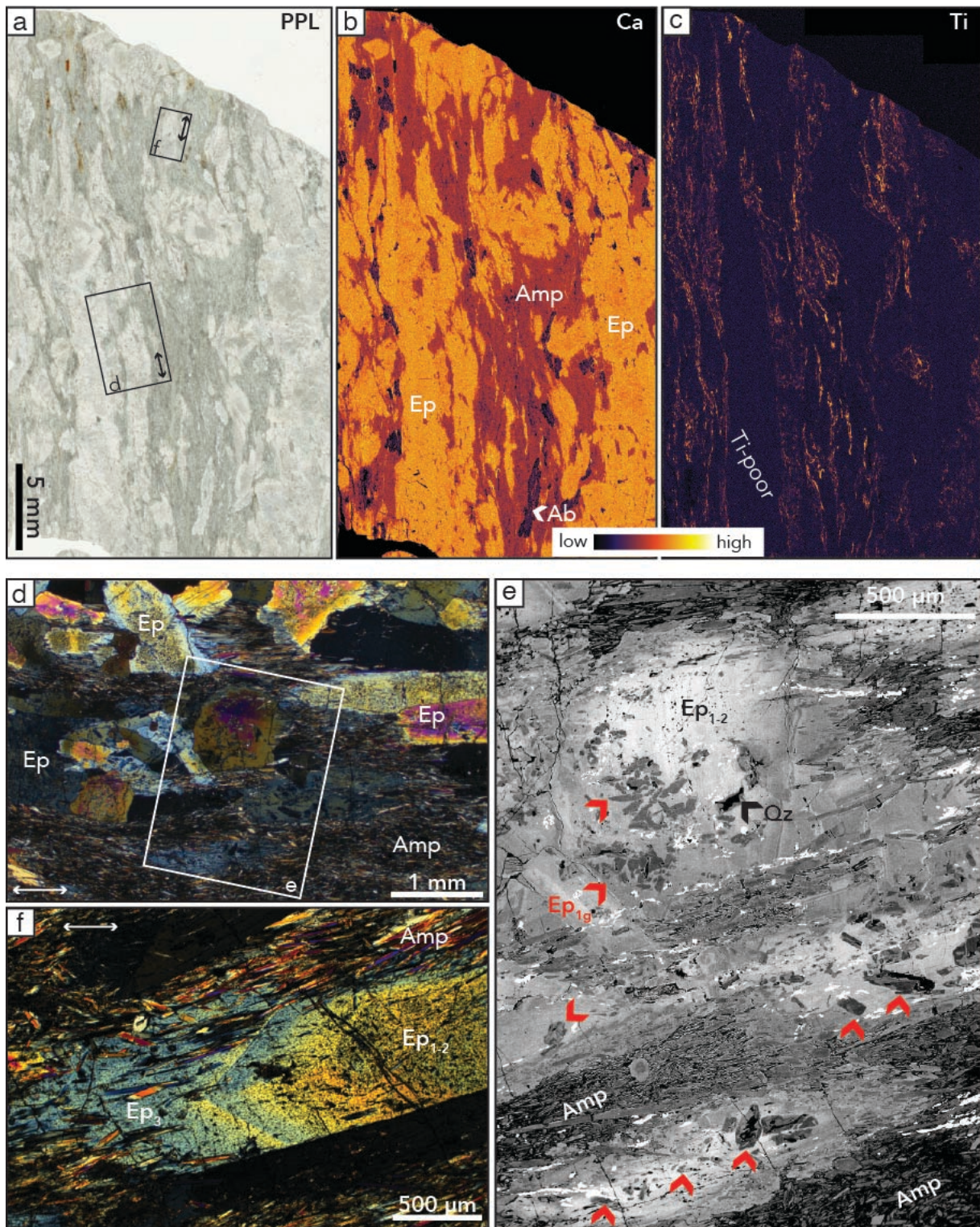


Figure 4. Textures of a deformed epidote-rich zone. (a-c) show optical (a) and EDS (b-c) scans of full thin sections. In (c), Ti-poor zones are visible transposed into the foliation. XPL photomicrographs (d) and (f) show examples of optical zoning in epidotes. (e) BSE image of ghost grains in epidote-rich zones, highlighted with red arrows. Double-ended arrows in (d) and (f) correspond to those in the image outlines in panel (a) to show orientation within the sample.

4.2.2 Porphyroblastic Epidote (Ep_{2-3})

Epidote porphyroblasts are also present throughout these metabasalts, and commonly feature patchy zoning similar to that observed in some ERZs and are rich in amphibole inclusions (Fig. 5). These porphyroblasts are commonly subhedral with roughly tabular shapes. Porphyroblasts typically have a narrower range of compositions than epidote in epidote-rich zones with X_{Ep} values only spanning a range of ~ 0.30 – 0.50 (compared to 0.16 – 0.67 in ERZ (Fig. 6a). Porphyroblastic epidote in all samples also exhibit concave-down chondrite-normalized REE patterns, with greater LREE abundances than observed in epidote-rich networks (Fig. 7d). This pattern that is apparent in values of La/Lu_N (0.1 – 11.3 with a mean of 2.86) and Dy/Dy^* values (0.9 – 2.6 with a mean of 1.5) of porphyroblasts, which indicate LREE abundance over HREEs and concavity, respectively (Fig. S3). In general, porphyroblasts show both positive and negative Eu anomalies while ERZs exhibit dominantly positive Eu anomalies (ERZ mean $Eu/Eu^*=1.7$; Fig 7, S3, S4).

Some porphyroblasts (e.g., in sample 22C09A2) have two distinct compositions of epidote. One (Ep_2) is brighter in BSE images and exhibits X_{Ep} values of 0.42 – 0.50 , while the other, darker epidote (Ep_3) has X_{Ep} values of 0.32 – 0.42 (Fig. 5, 6). The more Fe-rich epidote appears only as isolated, ameboid patches within the more clinozoisitic epidote (Ep_3), with which it shares a sharp boundary (Fig. 6). There is only a subtle difference between the REE patterns of these two groups of epidotes, with both exhibiting slightly concave patterns ($Dy/Dy^* > 1$) and negative slopes ($La/Lu_N \geq 1$), but several analyses of Ep_3 show positive Eu anomalies (mean $Eu/Eu^*=0.9$), compared to consistently negative Eu anomalies in Ep_2 analyses (mean $Eu/Eu^*=1.1$; Fig. 7e, S4). Amphiboles with both actinolite and magnesiohornblende compositions are included in these porphyroblasts and present in the matrix. Actinolite commonly appears as rims on magnesiohornblende cores and as euhedral crystals in the matrix surrounding the porphyroblasts (Fig 6b). The more clinozoisitic epidote (Ep_3) appears in textural equilibrium with the actinolite and sometimes appears to embay amphiboles with magnesio-hornblende–pargasite compositions (Fig. 5e).

In the metabasalt matrix adjacent to interpillow epidosite, epidote porphyroblasts with Ep_3 -like mineral compositions and REE patterns (Fig. 7a) and magnesio-hornblende and pargasite inclusions may also exhibit moderately negative Ce anomalies (mean Ce/Ce^* of 0.8), similar, but not as extreme, as those of Ep_0 found in the interpillow epidosite (mean Ce/Ce^* of 0.2).

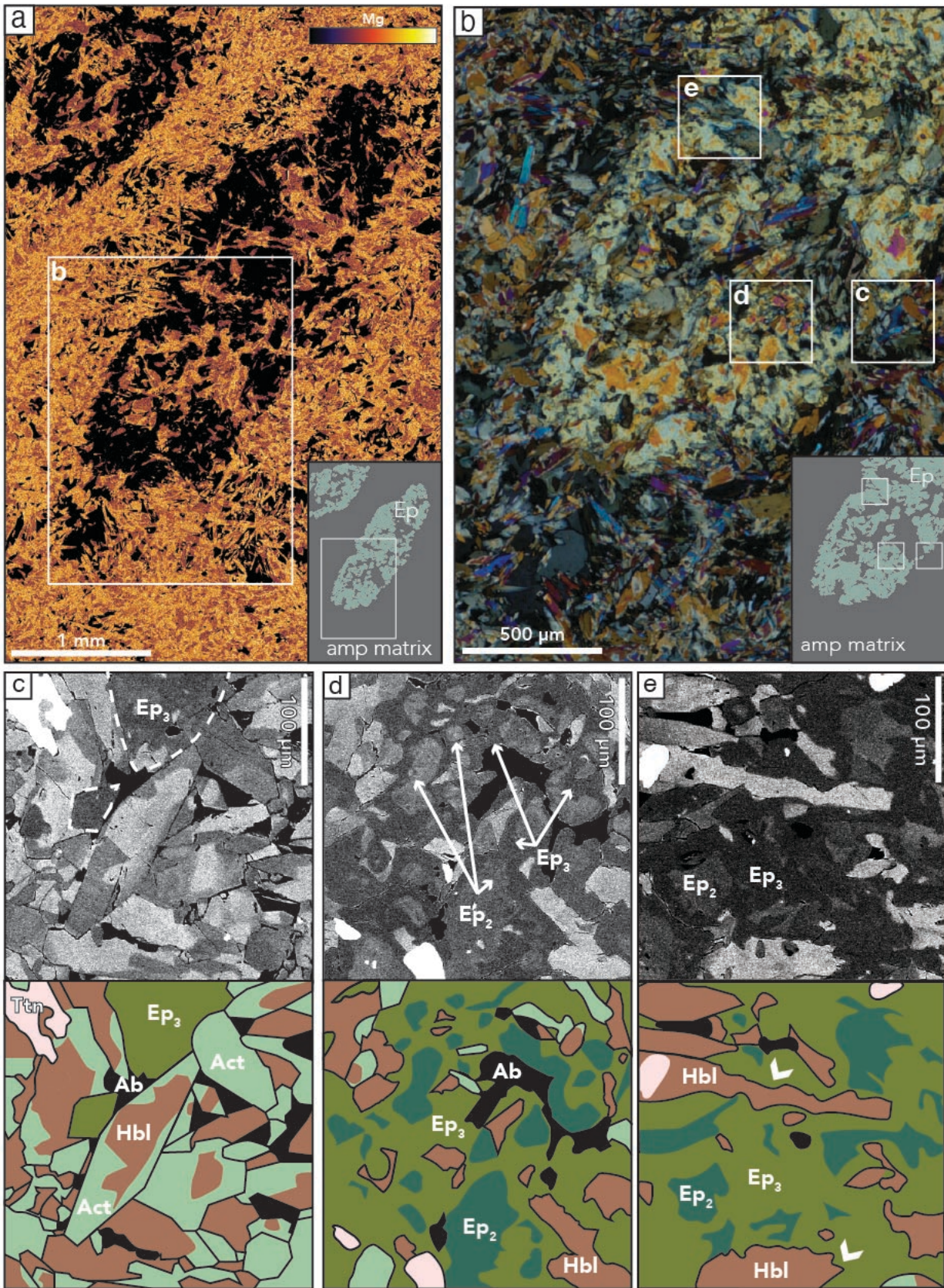


Figure 5. Example of an epidote porphyroblast within the matrix of a metabasalt. (a) Mg EDS map highlighting the morphology of epidote porphyroblast structures (black) within the amphibole-dominated matrix (orange–red). (b) Optical XPL image of the porphyroblast; zoning of the epidote is visible optically with Ep₂ appearing as orange–magenta and Ep₃ appearing as blue–grey. (c–e) BSE images and sketches showing details within the porphyroblast: (c) actinolite rims on hornblende; (d) Ep₂ forming distinct cores surrounded by Ep₃; (e) white arrows highlight where hornblende is embayed and included by Ep₃.

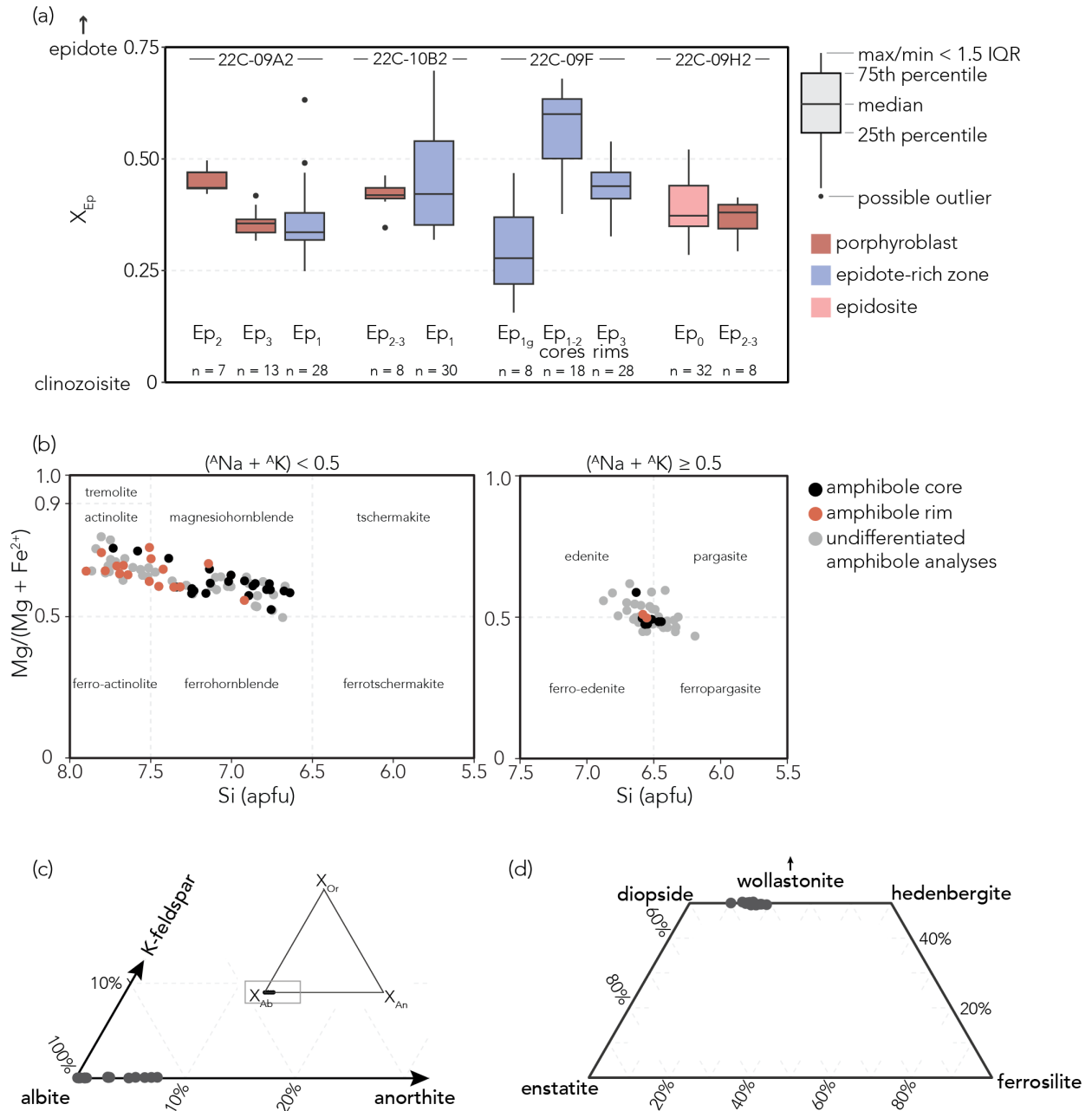


Figure 6. Plots of mineral compositions. (a) Epidote compositions in box-and-whisker plots showing variations between the different generations and textural domains of epidote, (b)

amphibole compositions, with core and rim analyses highlighted, (c) feldspar compositions, and (d) clinopyroxene compositions.

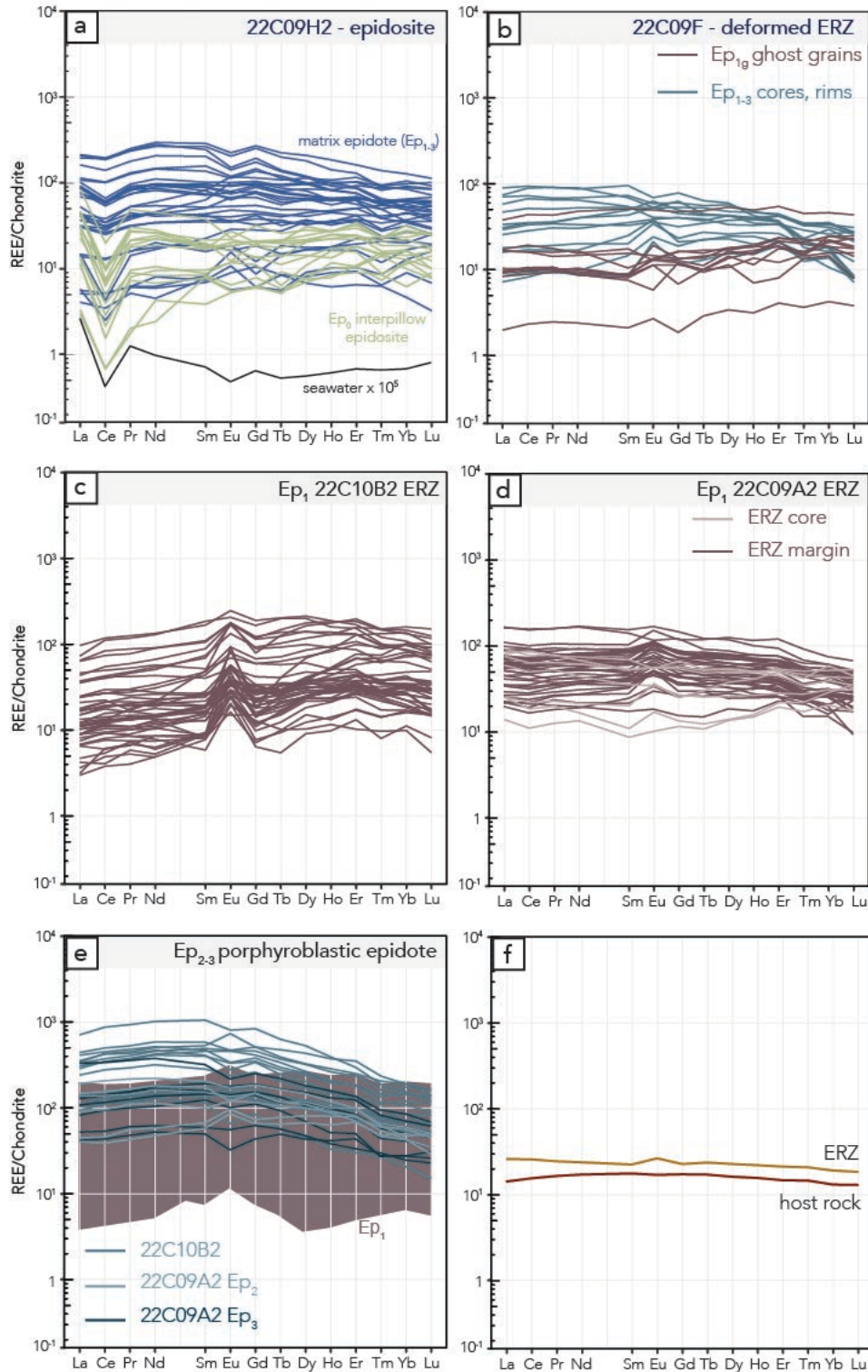


Figure 7. REE patterns from *in-situ* and bulk-rock LA-ICP-MS analyses normalized to chondritic values from McDonough and Sun (1995) (a) Analyses from epidote in an epidosite-

dominated sample showing variations between epidosite and adjacent metamorphic epidote that has inherited a moderate negative Ce anomaly. The plotted modern seawater values are an average of seawater compositions from samples >1000 m depth in the western Pacific (Deng et al., 2017). (b) Deformed ERZ analyses with core and rim analyses showing patterns similar to (e) porphyroblasts, and ghost grains showing patterns more comparable to epidote-rich zones (c). (d) Variation is evident within some ERZs where analyses from the margins of the ERZ differ from those in the core. (f) comparison of bulk-rock REE patterns between host rock and metamorphic epidote-rich zones.

4.3 Strontium isotopes

Strontium isotopic analyses can provide an indication of the degree of seafloor alteration of the epidote-amphibolite metabasalts (e.g., Chapman and Spooner 1977; Bickle and Teagle, 1992). We compare our measured and corrected isotopic ratios to estimates of Jurassic seawater $^{87}\text{Sr}/^{86}\text{Sr}$ and compiled $^{87}\text{Sr}/^{86}\text{Sr}$ values for MORB. Jurassic seawater values are reported to range from 0.70685 to 0.70750 (McArthur et al., 2001), while MORB values range from $\sim$ 0.702–0.703 (e.g., Rubin and Macdougall, 1992). Measured whole-rock $^{87}\text{Sr}/^{86}\text{Sr}$ values from our metabasalts have strongly radiogenic values ranging from 0.704974 to 0.706491 (Fig. 8).

To account for ^{87}Rb decay, we calculated a range of possible initial $^{87}\text{Sr}/^{86}\text{Sr}$ values using the whole-rock Rb and Sr concentrations measured via ICP-MS. Assuming the measured Sr isotopic values were established either during initial emplacement of the metabasalts at the oldest or during metamorphism of these rocks at the youngest, we bracket the range of possible initial $^{87}\text{Sr}/^{86}\text{Sr}$ based on constraints on the timing of metamorphism and protolith formation.

Metamorphism of the Epidote Amphibolite Unit has been dated to $\sim$ 105–96 Ma based on $^{40}\text{Ar}/^{39}\text{Ar}$ hornblende and mica ages (Grove and Bebout, 1995) and U-Pb in titanite and rutile ages (Hoover et al., 2025), providing a minimum time since which Rb decay may have affected the $^{87}\text{Sr}/^{86}\text{Sr}$ value. At the other extreme, plate reconstructions suggest that 80–90 million year old oceanic crust was subducting beneath North America at 100 Ma (Müller et al., 2019), so we take 195 Myr as a possible maximum time since initial emplacement. Only the estimated initial $^{87}\text{Sr}/^{86}\text{Sr}$ values of one sample (22C09A1, host rock) show a significant spread between these possible extremes because of a higher Rb/Sr value (0.063). Between texturally discrete parts of one sample (22C09A2 ERZ vs host rock vs epidosite), Sr values are similar, but there is significant inter-sample variability (cf. deformed, epidote-rich sample 22C09I; Fig. 8), despite these samples being collected within $\sim$ 10 m of each other. Additional explanation of calculations of initial $^{87}\text{Sr}/^{86}\text{Sr}$ values is provided in the supplementary information.

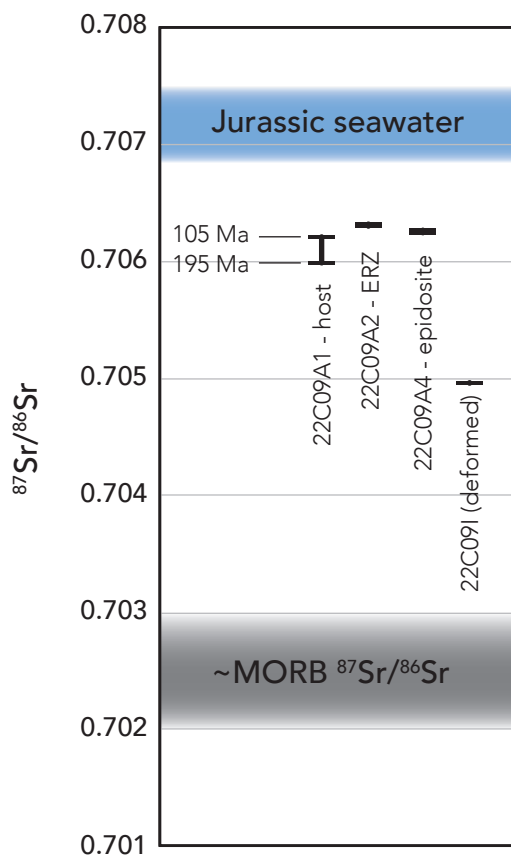


Figure 8. Calculated initial whole-rock strontium isotope ratios for several samples of metabasalts, including a host-rock sample (without ERZs), a dominantly metamorphic ERZ sample, and a dominantly epidosite sample. There is little difference between the Sr isotope ratios of these samples consistent with metamorphic epidote inheriting the isotopic signature of seawater from the host rock. However, Sr-isotopes may vary significantly between samples from different parts of the outcrop suggesting heterogeneous fluid flow within the pillow basalts. Bars represent the range of likely initial $^{87}\text{Sr}/^{86}\text{Sr}$ depending on the age of when the system became closed, but only 22C09A1 shows significant variation between the two endmembers.

Table 1. Summary of textural and geochemical observations of epidote textural domains

Textural Domain	Epidote microtextures	Geochemical characteristics	Epidote generation interpretation	Process interpretation
Interpillow epidotes	fine-grained, granoblastic; fine-scale oscillatory zoning	$X_{Ep} = 0.28\text{--}0.47$; Strongly negative Ce anomalies	Ep ₀	Pre-subduction hydration and hydrothermal alteration
Epidote-rich zones (ERZs)	'ghost' grains within other epidote	$X_{Ep} < 0.5$; flat to positive REE patterns; positive Eu anomalies	Ep _{1g}	Epidote pseudomorphs after pumpellyite
	coarse, anhedral–subhedral epidote crystals; inclusion-rich; patchy to concentric zoning patterns	$X_{Ep} = 0.16\text{--}0.67$; associated with discrete Ti-poor zones; ~flat REE patterns with inter-sample variability in slope	Ep ₁ ; possible Ep ₂ /Ep ₃ overprint	Epidote growing from pumpellyite breakdown to form reactive porosity networks
Porphyroblasts	amoeboid patches within porphyroblasts; comprise optically distinct patches	$X_{Ep} = 0.42\text{--}0.50$; concave down REE patterns with negative slopes	Ep ₂	Epidote equilibrated at peak <i>T</i> conditions during incomplete epidote breakdown
	volumetrically dominant part of subhedral porphyroblasts	$X_{Ep} = 0.32\text{--}0.42$; concave down REE patterns with negative slopes	Ep ₃	Retrograde epidote formed during refrigeration after underplating

4.4 Phase equilibria modeling

We use the bulk composition from one of our Catalina Schist epidote amphibolite-facies metabasalts to model the equilibrium phase assemblages, phase compositions, and phase abundances of a metabasalt over 0.1–1.5 GPa and 200–800°C. Figure 9a shows the results of our phase equilibria modeling, with boundaries of fields containing phases of interest highlighted. Below 1 GPa, metamorphic epidote is predicted to grow in these rocks at ~250–350°C. Epidote growth is associated with the breakdown of prehnite (at pressures <0.2 GPa), pumpellyite (0.2–0.7 GPa), or lawsonite (>0.7 GPa). In all cases involving pumpellyite, the epidote-producing reaction is pumpellyite + quartz + chlorite = epidote/clinozoisite + actinolite + H₂O ± quartz, which is consistent with reactions reported for some low-grade metabasalts (e.g., Nakajima, 1982), from other thermodynamic modeling work (e.g., Liou et al., 1985; Evans, 1990), and from experimental studies (Nitsch, 1971). Albite is stable in our model below 0.9 GPa, and above 500°C plagioclase starts to become more anorthitic.

The greatest epidote volume fractions predicted in these rocks are ~35 vol. % near the lower-temperature limit of epidote stability and intermediate pressures (Fig. 9b). Above 0.5 GPa, the compositions of the first epidote are clinozoisite-rich ($X_{Ep} < 0.5$), but X_{Ep} values increase with increasing temperature while epidote volume fraction decreases (Fig. 9b). With increasing temperature, epidote gradually breaks down to produce amphibole and clinopyroxene. At >1.1 GPa and >700°C, our model predicts coexisting epidote + zoisite. Beyond the higher temperature stability boundary of epidote, equilibrium mineral assemblages consist of clinoamphibole + feldspar + clinopyroxene + quartz + Fe-Ti-oxide ± garnet.

Amphibole is generally stable throughout the modeled P - T space but with varying composition. While sodic amphibole is predicted at low temperatures and high pressures, we observe no sodic amphiboles in the Epidote Amphibolite Unit, so we focus on the composition of the calcic amphiboles predicted in our models. In our samples, we broadly distinguish actinolite-like ($\text{Si} \geq 7.5$ apfu) and hornblende-like Ca-amphiboles ($\text{Si} \leq 7.5$ apfu), and use the results of phase equilibria modeling to explore how Si content and Mg-number ($\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$) are predicted to evolve over P - T to match with those measured via EPMA in the metabasalts. Actinolitic amphibole is predicted to be stable from $<250^\circ\text{C}$ up to $\sim 500^\circ\text{C}$, while amphibole with hornblende-like compositions is predicted to grow at $\sim 400^\circ\text{C}$ at low pressures (near the upper limit of albite stability), resulting in a field where actinolite and hornblende coexist between 400 – 500°C and <0.9 GPa.

Pairing our phase equilibria modelling results with our epidote and amphibole chemical analyses allows us to estimate the P - T conditions at which Ep_2 and Ep_3 in the porphyroblastic epidote grew. Magnesiohornblende, pargasite, and edenite analyses from amphiboles in the matrix of the metabasalts that are rimmed by actinolite have 6.5 – 7.2 Si pfu. We infer that these amphiboles coexisted with Ep_2 , which has X_{Ep} values of 0.42 – 0.50 , at peak T conditions. These constraints, along with the inference from some samples that clinopyroxene stabilized in these rocks at peak metamorphic conditions, allow us to estimate peak metamorphic conditions for Ep_2 growth at ~ 520 – 600°C and 0.6 – 0.9 GPa (Fig. 9b,c). Clinopyroxene compositions predicted by phase equilibria modelling at these conditions are similar ($\text{Mg\#} \sim 0.61$) to our measured diopside compositions ($\text{Mg\#} 0.62$ – 0.80 ; Table S6). Similarly, the actinolite in the matrix has 7.5 – 7.9 Si pfu, and these amphiboles are in textural equilibrium with Ep_3 , which has X_{Ep} values of 0.32 – 0.42 . The presence of albite also significantly constrains the P - T conditions of Ep_3 , which we place at 470 – 510°C (Fig. 9).

The model presented here uses the whole-rock Fe^{3+} concentration as determined via titration, which we assume represents a maximum value for this rock—i.e. it is currently in its most oxidized state since formation. Our $\text{Fe}^{3+}/\text{Fe}^{\text{Total}}$ is 0.28 , which is similar to the measured average $\text{Fe}^{3+}/\text{Fe}^{\text{Total}}$ of 0.16 for modern MORB (Cottrell and Kelley, 2011) and suggests that this is a reasonable range that is representative of the oxidation state the rock may have had during metamorphism. Performing preliminary phase equilibrium modeling with the lower $\text{Fe}^{3+}/\text{Fe}^{\text{Total}}$ ratio does not significantly change the topology of the pseudosection or modal abundance of major phases, other than shifting the stability field of zoisite to lower pressures at the expense of epidote at high T .

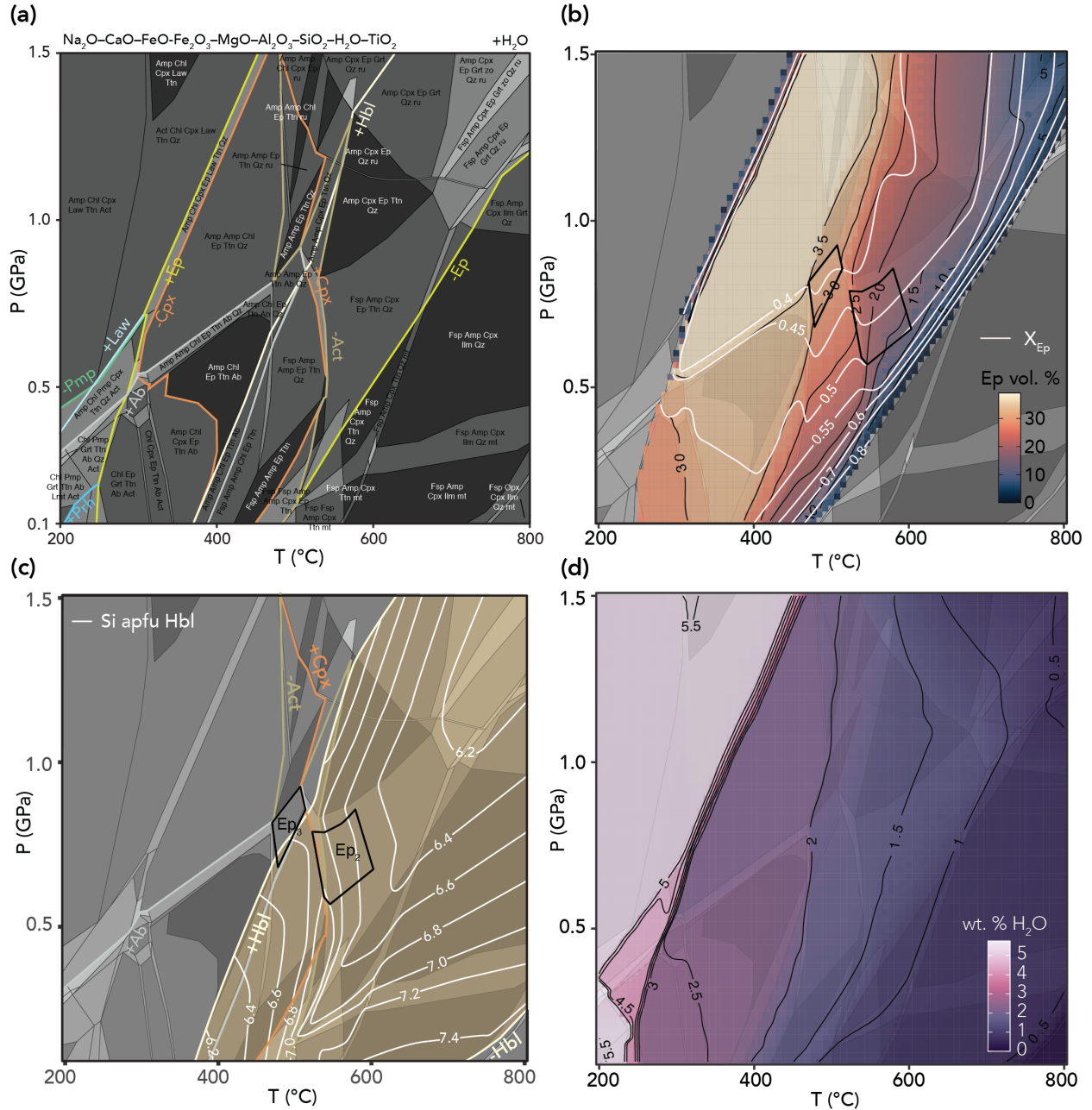


Figure 9. Phase equilibria modelling results. (a) pseudosection showing stable equilibrium mineral assemblage. The stability regions of important phases are highlighted with colored lines. (b) Properties of epidote predicted in our model with contours showing predicted X_{Ep} (white isopleths) and modal abundance (vol. %, black isopleths). (c) White contours show predicted Si content in hornblende (atoms per formula unit), and the stability fields of clinopyroxene and albite, used to help constrain the Ep_2 and Ep_3 fields, are also shown. (d) Pseudosection contoured for predicted abundance of mineral-bound water—note the large reduction in predicted mineral-bound water content associated with the crystallization of epidote.

5 Discussion

5.1 Preserved seafloor alteration signatures

We interpret epidiosites (Ep_0) and the observed macroscopic breccia textures (Fig. 2e) elsewhere in the metabasalt to be relict features from emplacement and subsequent hydrothermal alteration of the oceanic crust. Fine-grained, granoblastic epidote in interpillow networks are consistent with epidiosite textures observed in ophiolites (e.g., Gilgen et al., 2016; Weber et al., 2021) and drill cores from modern oceanic crust (e.g., Banerjee et al., 2000; Chen et al., 2023). The negative Ce anomalies associated with epidote in these zones, and, to a lesser degree, epidote in the adjacent host rock are also diagnostic of seawater interactions (e.g., Menzies et al., 1977; Murray et al., 1991). Seawater typically exhibits strongly negative Ce anomalies as a result of the relative insolubility of Ce^{4+} as compared to Ce^{3+} (e.g., Elderfield and Greaves, 1982), and hydrous minerals forming during seawater-rock interactions will inherit patterns of seawater REE compositions. The epidote porphyroblasts in the matrix adjacent to the epidiosite zones exhibit negative Ce anomalies of a lower magnitude (mean Ce/Ce^* of 0.8) than those of epidotes within the interpillow Ep_0 epidiosite (mean Ce/Ce^* of 0.2; Fig. 9a, S3a). We suggest this signature was imparted on the basalt adjacent to the epidiosite during hydrothermal alteration, and that the porphyroblastic epidote has inherited this signature. Similarly, it is possible that the epidiosite zones themselves, though certainly retaining the signature of seafloor alteration, recrystallized during subduction metamorphism and may exhibit overprinting textures and geochemical signatures. However, phase equilibrium modeling of epidiosites during subduction has also demonstrated that, alone, the phase assemblage of an epidiosite is unlikely to change significantly during subduction to subarc depths because of the broad stability of epidote (Hernández Uribe et al., 2020). This does not preclude textural changes and trace element redistribution, but nonetheless negative Ce-anomalies indicate that these zones experienced significant seafloor hydrothermal alteration that subsequent geochemical changes was unable to overprint.

Epidiosites are commonly associated with high-temperature hydrothermal alteration of basalts. Field studies of ophiolites (e.g., Harper et al., 1988; Richter and Diamond, 2022) as well as modern oceanic crust (Chen et al., 2023) have estimated epidiosite alteration to occur at $\geq 250^\circ\text{C}$. Similar to the seafloor-proximal setting inferred for the protolith of the metabasalts presented here, epidiosites in other ophiolites have been found at shallow structural levels within the extrusive piles of the Josephine (Harper et al., 1988) and Semail ophiolites (e.g., Gilgen et al., 2016). In these settings, epidiosites have also been found as replacements or alterations of interpillow matrix material, indicating hydrothermal circulation concentrated around pillow margins. Canonical models of epidiosite formation, however, based on studies of the Troodos ophiolite suggest the formation of epidiosites often occurs in the deep parts of hydrothermal circulation systems within the sheeted dike complex at mid-ocean ridges (e.g., Richardson et al., 1987; Alt, 1995). Based on the pillow textures observed here, it is likely that these metabasalts represent shallow parts of the subducted slab.

The measured whole-rock $^{87}\text{Sr}/^{86}\text{Sr}$ values also indicate a significant degree of seawater–rock interaction for both epidiosites and host pillow basalts prior to subduction. For simple two end-member mixing of isotopic reservoirs, we can estimate the effective water/rock (W/R) ratios required to shift the $^{87}\text{Sr}/^{86}\text{Sr}$ values of the metabasalts so far towards seawater-like values (e.g., McCulloch et al., 1980; Marks et al., 2015). Using the $^{87}\text{Sr}/^{86}\text{Sr}$ value estimated for the age of the subducting crust and assuming initial Sr abundances in fresh N-MORB of 128 ppm (Gale et al.,

2013) and modern seawater of 8 ppm, and Jurassic seawater $^{87}\text{Sr}/^{86}\text{Sr}$ values ranging from 0.70685 to 0.70750 (McArthur et al., 2001), we calculate water/rock ratios for the whole rock samples analyzed. The range of seawater values represents the range of marine Sr-isotope ratios during the time when these rocks could have experienced seawater-rock interactions. Water/rock ratios calculated using the lower seawater Sr-isotope value are 8.1–14, while the upper value yields 6.9–12. Both ranges are consistent with W/R ratios estimated by Sr isotopes for hydrothermal systems in both ophiolite and modern basaltic settings (e.g., Kawahata et al., 2001; Marks et al., 2015). The equation used to calculate W/R ratios is included in the supplementary information.

These W/R ratio estimates rely on several crucial assumptions including the attainment of equilibrium between the basalt and the hydrating fluid, and the fluid being endmember seawater and not having previously equilibrated with other rocks, neither of which were likely entirely satisfied in the present system. Particularly in the case of the assumption of seawater isotopic values, based on the high temperatures ($>250^\circ\text{C}$) suggested for epidosite alteration, it is likely these rocks were altered by upwelling, hot hydrothermal fluids that had acquired a hybrid, probably more basalt-like Sr-isotope value (e.g., Weber et al., 2021). The W/R ratios reported here therefore should be viewed as minimum values for the amount of water that actually passed through these rocks during hydrothermal alteration at the seafloor.

Epidosite textures, negative Ce anomalies in epidote, and whole-rock Sr-isotopic ratios are clear fingerprints of seawater influence in the protolith of the metabasalts and indicate that these rocks were quite hydrated prior to subduction. The temperatures estimated for epidosite formation ($>250^\circ\text{C}$) in other settings also suggests that these rocks experienced relatively high temperature hydrothermal alteration that would have introduced water not just into pore spaces in the rock, but also reacted with anhydrous minerals to create new water-bearing phases. With effective W/R ratios >6.9 , it is possible that the pillow basalts—both in and out of interpillow zones—were water saturated prior to subduction. This is significant for the water budget of these rocks and their potential for later dehydration, which we discuss below.

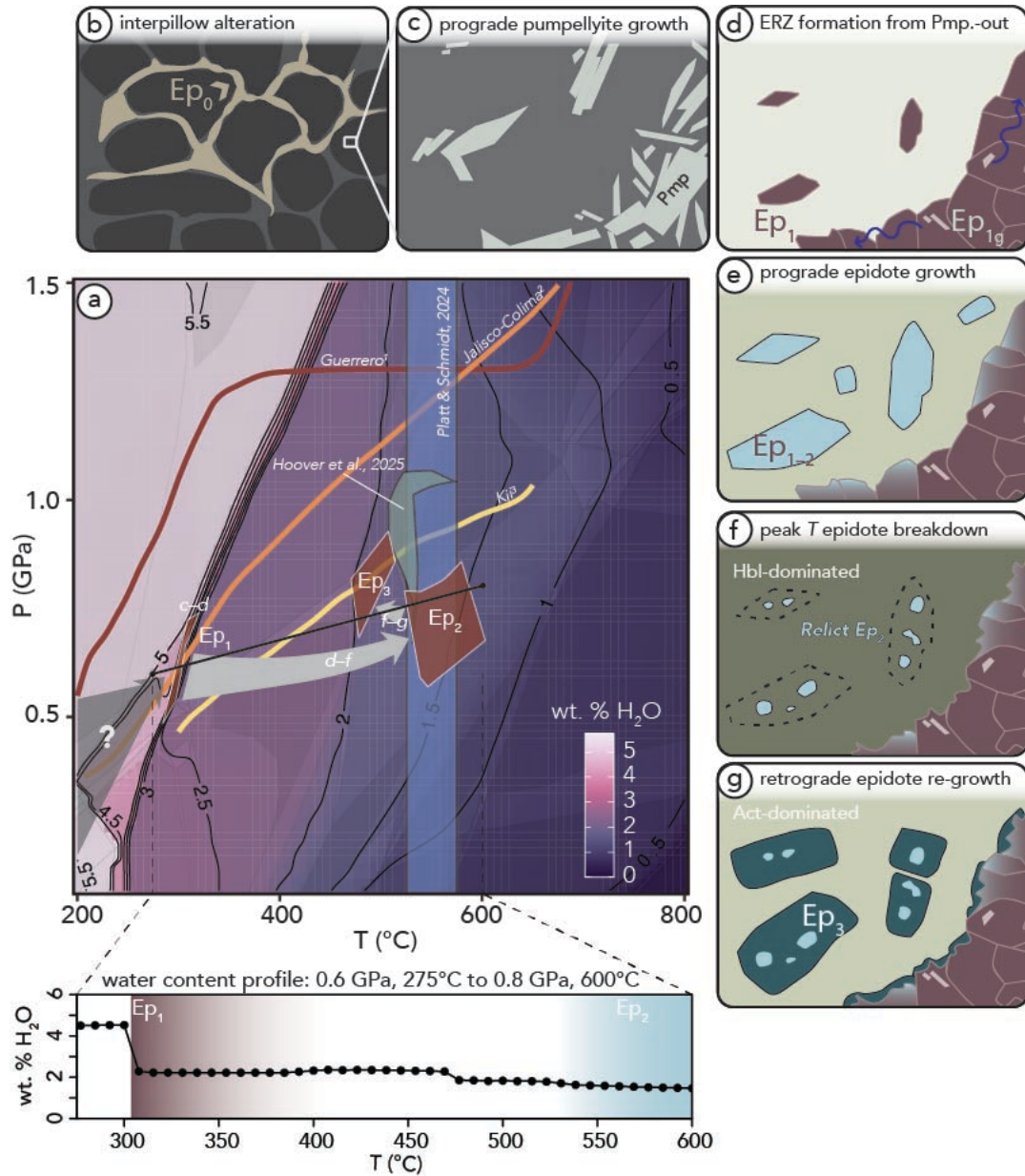


Figure 10. Summary P - T diagram and cartoon showing our interpretation of the evolution of the Epidote Amphibolite Unit metabasalts. (a) same as Fig. 9d, but with labeled stability field of each epidote generation and potential P - T paths connecting them. Letters on the plot correspond to the surrounding panels which summarize the textural evolution of epidote in these rocks. The plot at bottom show a representative evolution of the amount of mineral-bound water in the metabasalts with increasing T from initial metamorphic epidote growth to ~peak P - T conditions. In (a), slab geotherms are from [1] Manea and Manea (2011), [2] Ferrari et al. (2012), and [3] Suenaga et al. (2019), and additional shaded polygons show the peak P and/or T estimates of the Epidote Amphibolite Unit from Platt and Schmidt (2024) and Hoover et al. (2025).

5.2 Metamorphic P - T path of the Epidote Amphibolite Unit

Based on the textures observed in thin section, trace element patterns, and our phase equilibrium modeling, we infer an initial metamorphic epidote-producing reaction and

subsequent prograde and retrograde P - T paths for these metabasalts. This includes epidotes in populations $Ep_{1,2,3}$, with Ep_1 and Ep_2 forming during initial and peak prograde metamorphism, respectively, and Ep_3 forming during early retrograde metamorphism.

An important limitation of the model we present is that the whole-rock major element composition used in our phase equilibrium modeling likely does not perfectly represent the effective bulk composition of the metabasalts at any given point on the P - T path. Because of the presence of zoned minerals and epidote-rich zones, we know that this effective bulk composition will vary during the metamorphic evolution of these rocks and differ from the bulk composition used in our model (Zuluaga et al., 2005; Palin et al., 2016). Despite having measured the current FeO and Fe₂O₃ concentrations in the bulk rock sample we used for phase equilibrium modeling, we also lack constraints on the evolution of oxygen fugacity during subduction, which affects the stability and composition of phases like epidote. Phase equilibrium modeling itself also includes uncertainties in the position of phase assemblage field boundaries that arise from uncertainties in the underlying thermodynamic properties of end-member phases and in activity-composition models. These are commonly estimated as 2σ uncertainties of $\pm 50^\circ\text{C}$ and ± 0.1 GPa (e.g., Powell and Holland, 2008; Hodges and McKenna, 1987). Regardless of these limitations and uncertainties, our estimated peak P - T conditions for the Epidote Amphibolite Unit overlap with existing estimates from this unit (Fig. 10; Platt and Schmidt, 2024; Hoover et al., 2025), suggesting that our constraints on the P - T path are both generally geologically faithful and useful for understanding the geodynamic setting of dehydration reactions.

5.2.1 Prograde metamorphism and dehydration

The integration of our textural observations and mineral chemistry with our phase equilibria modelling show that some of these epidote-rich zones represent a record of the initial metamorphic epidote (Ep_1) that crystallized in these rocks during subduction. The ghost grains (Ep_{lg}) observed in the cores of epidote in these zones have well-developed tabular shapes, with some exhibiting subtle asymmetry that could suggest a monoclinic crystal, though we cannot rule out orthorhombic symmetry (such as one might expect from pseudomorphs after lawsonite; Fig. 3, 4). However, ghost grains also exhibit higher concentrations of HREEs as compared to LREEs, which is common among epidote in ERZs (Fig. 7b,c). We suggest that these REE patterns are likely inherited from preexisting pumpellyite in this rock, which has been shown to have similar REE patterns in other metabasalts, and is consistent with the possibility of the original crystals exhibiting monoclinic symmetry (e.g., Aguirre et al., 1995; Nyström, 1984). Although pumpellyite often exhibits a variety of morphologies in low-grade metabasaltic rocks, including radial, fibrous sprays, it has been reported to occur alone as discrete prismatic crystals or in aggregates thereof, and as a common pseudomorphous phase after primary igneous phases like pyroxene and plagioclase (e.g., Offler et al., 1981; Cortesogno et al., 1984; Metcalfe et al., 1992; El-Shazly, 1994). In the latter case, pumpellyite pseudomorphs after plagioclase laths could also produce some of the observed ghost grain shapes. At higher pressures, epidote may be expected to form by the breakdown of lawsonite (Fig. 9a). However, the tendency of lawsonite to be the dominant host of LREEs in similar rocks (e.g., Martin et al., 2014) are inconsistent measured REE concentrations.

While not definitive, these data are consistent with pumpellyite as a likely precursor to epidote in this rock, which would constrain the initial growth of metamorphic epidote to pressures lower than 0.72 GPa (Fig. 9). The inclusion of quartz in and around these ghost grains further constrains the minimum pressures to 0.48 GPa, as at lower pressure, quartz is entirely

consumed in the epidote-producing reactions. Given the slope of the predicted low-temperature boundary of the epidote stability field, these pressures correspond to temperatures of 290–310°C for the pumpellyite–epidote reaction (Ep₁ field in Fig 10). Based on our pseudosection modeling then, the reaction that produced the first metamorphic epidote during subduction was pumpellyite + quartz + chlorite = epidote/clinozoisite + actinolite + quartz + H₂O. Epidote compositions predicted in our phase equilibria modeling ($X_{\text{Ep}} < 0.5$) at these P - T conditions are also consistent with most Ep₁ compositions (Fig. 6).

Thus, the epidote-rich zones record the earliest stages of epidote growth, and we suggest that ERZs represent the coalescence of recrystallizing pumpellyite similar to dehydration veins in serpentinite described by Plümper et al. (2017) or in eclogites by several authors (e.g., Taetz et al., 2016; Hoover et al., 2023). In Plümper et al.'s (2017) work, olivine veins in dehydrating meta-serpentinites are interpreted as self-organizing structures in a developing network of channel-forming reactive porosity that is guided by microscale chemical heterogeneities that localize dehydration reactions and consecutively induce dehydration in adjacent parts of the rock. In the Catalina Schist Epidote Amphibolite Unit metabasalts, the observed discrete Ti-poor zones that are associated with the ERZs (Fig. 3, 4) appear to represent the formation of reaction-induced porosity by the pumpellyite–epidote transition. Our phase equilibria models predict an increase in the density of the bulk rock across this boundary—from 3180 kg m⁻³ to 3240 kg m⁻³—and a corresponding volume change could result in the formation of fractures or pore space. This space, now filled with epidote, would have provided room and connectivity for the release of water liberated in the epidote-producing reaction. Although Ti has been shown to exhibit at least local fluid mobility in metamorphic environments because of its enrichment in metasomatic products and the occurrence of rutile-rich veins (e.g., Van Baalen, 1993; Sorensen and Grossman, 1989; Luvizotto and Zack, 2009), we suggest that the definition of these discrete vein-like regions by a lack of Ti-bearing phases indicates the relative immobility of Ti in this setting. The apparent merging of some ERZs with interpillow networks (Fig. 2) suggests that the preexisting interpillow networks may have helped to drain fluids produced during ERZ formation. In other ophiolites, interpillow networks have been shown to serve as higher permeability pathways than other parts of pillow lava suits (Brett-Adams et al., 2023), so these networks may be destinations for fluids sourced from dehydration reactions.

Interconnecting and branching patterns of ERZs on the outcrop (Fig. 2) and micro-scale (in the pattern of Ti-poor regions; Fig. 3) have a similar spatial pattern to the channel-forming reactive porosity structures represented by olivine vein networks in serpentinites described by Plümper et al. (2017). The epidote-rich zones in our rocks are not as spatially extensive as these olivine vein networks, which are observed at up to the meter scale, while our metamorphic ERZs are wispy, cm- to decimeter long structures that are not as organized or interconnected as the olivine vein networks. The lack of larger-scale interconnectivity could be explained by the relative availability of reactants. In the dehydrating serpentinites of Plümper et al. (2017), antigorite is the volumetrically dominant phase (e.g., Scambelluri et al., 1991), whereas our phase equilibria models predict that the metabasalts would have contained only up to 40 vol. % pumpellyite. In the case of channel-forming reactive porosity in the metabasalts, it is possible that the limited volume of pumpellyite would prevent more extensive, interconnecting networks of ERZs from forming, and the original distribution of pumpellyite would influence the nucleation and pattern of ERZs, with ERZs more likely to form in more pumpellyite-rich domains. The presence of the preexisting networks of interpillow materials, which represent existing higher-permeability zones than pillows themselves, may therefore be an important

structure in draining the subducting crust. At high pressures in subduction zones, pervasive grain-boundary fluid flow is likely limited (e.g., Ague, 2014), so channelized fluid flow is required for effective drainage of the subducting slab (Zack and John, 2007). ERZs linked with preexisting structures—e.g., interpillow material, faults—may operate in tandem as pathways for fluids produced through dehydration to escape the subducting crust. It is possible that fluid flow through the interpillow networks during subduction might overprint the seafloor hydrothermal alteration signatures recorded in epidiosites with the signature of dehydration-derived fluids. However, without additional modeling or analytical constraints, we do not know what the composition of fluids derived from pumpellyite dehydration would be or their effect on epidote. Given the preservation of seafloor signatures in the epidiosite presented here, we suggest that either fluid flow was heterogeneous at the scale of the pillows and seawater signatures are only locally overprinted, or that dehydration fluids did impart some signature of their passage but did not completely overprint the original seawater signatures.

5.2.2 Peak *T* and early retrograde deformation

Additional stages of prograde and early retrograde metamorphism are revealed by the textures and compositions of epidote porphyroblasts and amphiboles in the matrix of the metabasalts. While Ep₁, is recorded only by ERZs in the rock today, Ep₁ likely also developed outside of ERZs in the matrix of the metabasalts, but during prograde metamorphism ultimately developed major- and trace-element characteristics of Ep₂. In the context of the porphyroblasts, the textural relationship of the two epidote compositions, Ep₂ and Ep₃, suggests that Ep₃, the more clinozoisitic epidote, grew after the more Fe-rich Ep₂ (Fig. 5). Further, the inclusions of Ep₂ within Ep₃ have rounded, amoeboid disequilibrium shapes, while Ep₃ may show euhedral grain shapes within the porphyroblast structure (Fig. 5). The textural relationships between the associated amphiboles, with actinolite rimming magnesio-hornblende, are suggestive of typical retrograde actinolite growth. Between epidote and amphibole, we see Ep₃ with straight grain boundaries where in contact with actinolite and sometimes embaying the magnesio-hornblende (Fig. 5e), which suggests that Ep₃ also represents retrograde recrystallization (Fig. 10).

The dominance of Ep₃ in these porphyroblasts and the co-inclusion of both actinolite and magnesio-hornblende can be explained by connecting these textures to our pseudosection (Fig. 9). With increasing temperature, the amount of epidote in these rocks is expected to decrease from its peak abundance following the initial stabilization of epidote, and the epidote that remains is predicted to develop more Fe-rich compositions. This is consistent with Ep₂ in our porphyroblasts representing near-peak temperature conditions, and Ep₃ growing in equilibrium with more actinolitic amphibole during retrogression (Fig. 5, 6, 10). We therefore define fields Ep₂ and Ep₃ representing the regions of *P-T* space in which the measured compositions of Ep₂ + magnesiohornblende/pargasite + clinopyroxene, and Ep₃ + actinolite, respectively, coexist in our pseudosection model. The Ep₃ field is also constrained by the requirement of albite stability, which provides an upper pressure limit of 0.9 GPa.

Put together, the stability fields for Ep₁, Ep₂, and Ep₃ that we identify in *P-T* space based on our pseudosection define a likely path followed by the Epidote Amphibolite Unit during subduction and underplating (Fig. 10). Prograde metamorphism followed a near isobaric heating path from field one to two, and the counterclockwise path from field two to field three suggests cooling of these rocks nearly isobarically. The tectonic significance of this path is discussed below in section 5.4.

The presence of magnesiohornblende, pargasite, and clinopyroxene within ERZs—described above as Ep₁—suggests that these continue to communicate with the surrounding matrix during prograde metamorphism and experience some breakdown of epidote. As discussed in the next section, however, we suggest that these zones may experience less metamorphic recrystallization than epidote porphyroblasts and therefore preserve distinct REE patterns.

5.3 Variations in REE patterns between ERZs and porphyroblasts

A common difference between the trace element composition of ghost grains and epidote-rich zones (Ep₁) as compared to porphyroblastic epidote (Ep₂₋₃) is the shape of the chondrite-normalized REE patterns (Fig. 7b–e). We suggest that epidote rich zones characterized by lower LREE than HREE normalized values (Fig. 7c) preserve inherited REE patterns from pumpellyite that was present in the rock before the growth of metamorphic epidote. However, porphyroblastic epidote (Ep₂₋₃) recrystallizes during peak and retrograde metamorphism, providing the opportunity for it to exchange trace elements with adjacent minerals in the matrix. Some ERZs, especially in epidote along their margins, also have more concave-down REE patterns and negative, but not as negative as porphyroblastic Ep₂₋₃ (Fig. 7d), which suggests that these zones may also have experienced recrystallization and trace element redistribution.

These interpretations are based on the assumption that epidote inherits the REE pattern of the pumpellyite from which it forms, and that epidote is the preferred host of REE in these rocks. Indeed, epidote/amphibole trace element partition coefficients are typically >1 for all but the HFSEs and LILEs except Sr, with LREEs and MREEs generally preferred over HREEs (Frei et al., 2004 and references therein). Alternatively, enrichment of Ep₂₋₃ in LREEs and MREEs relative to Ep₁ may also be explained by the predicted diminishing volume fraction of epidote in these rocks with increasing temperature (Fig. 9b). With some epidote breaking down into more amphibole and clinopyroxene, the remaining epidote would continue to be a preferred host for LREEs in the rock, resulting in the concentration of LREEs in epidote with increasing temperature, similar to the behavior of ferric iron (sect. 4.3).

In deformed ERZs, more of the epidote in these zones have concave-down REE patterns more similar to those of Ep₂₋₃ (cf. Ep₁₋₃ in Fig. 7b and Ep₂₋₃ in Fig. 7e). This includes epidote surrounding apparent ghost grains (Ep_g) with typical Ep₁ REE patterns. As with the occurrence of more Ep₂₋₃-like REE patterns on the margins of ERZs, we suggest that this is related to enhanced recrystallization as a result of the deformation experienced by these rocks. Some epidote exhibits synkinematic growth rims and asymmetric tails which include both hornblende and actinolite, and there is clear evidence for epidote fracturing and (re)crystallization at peak–retrograde metamorphic conditions (e.g., Fig. 4). This deformation may have dismembered epidote rich zones with originally Ep₁-like REE patterns, enhanced chemical communication with the surroundings, and caused the epidote in these rocks to adopt REE patterns more similar to the Ep₂ and Ep₃ porphyroblasts.

5.4 Implications for the tectonic history of the Catalina Schist

The *P-T* path that we estimate for the prograde metamorphism of the Catalina Schist Epidote Amphibolite Unit exhibits heating from ~300°C (Ep₁) to ~550°C (Ep₂) with possibly little increase in pressure (Fig. 10a). The maximum pressure change between fields Ep₁ and Ep₂ is 0.4 GPa as constrained by the lower and upper boundaries of those fields, respectively. This could be explained by a shallow subduction angle of the Farallon plate during subduction and ultimate underplating of the metabasalts (Fig. 11). Flattening of the Farallon slab beneath North

America at around 100 Ma has previously been inferred to explain geochemical observations from plutons of the Sierra Nevada Batholith and Peninsular Ranges Batholiths and late Cretaceous high-flux magmatism (e.g., Chapman et al., 2013; Grove et al., 2008). The varying geometry of the subducting Farallon plate, with flattening of the slab occurring around the time of amphibolite-facies metamorphism, has also been proposed as a model for the inferred P - T path of rocks in the Amphibolite Unit (Dong et al., 2022). The fact that the whole P - T path we predict lies at pressures below 1 GPa would suggest that subduction was occurring beneath relatively thin (≤ 30 km) forearc crust. A 30 km thickness for continental forearc crust has been proposed for the paleo-subduction setting of the Sanbagawa Belt (Aoya et al., 2013), and estimated in modern subduction settings like northern Cascadia (Audet et al., 2025).

The peak P - T conditions that we estimate for the metabasalts and the P - T path that we infer are consistent with conditions in some modern warm subduction zones. The peak epidote amphibolite conditions of 520–600°C at pressures < 1 GPa are consistent with conditions experienced at the top of the subducting Philippine Sea plate beneath the Kii Peninsula (Fig. 10; Suenaga et al., 2019). The shape of the P - T path for the metabasalts, which shows nearly isobaric prograde P - T heating once reaching depths corresponding to pressures of ~ 0.48 – 0.72 GPa, is also similar to that modeled for the Cocos plate subducting beneath North America in Guerrero, Mexico (Manea and Manea, 2011). The subducting plate beneath Guerrero, however, underplates thicker continental lithosphere (~ 40 km) than what the thickness of the North American forearc must have been at ca. 100 Ma. The Kii slab-top P - T path, along with that of Jalisco-Colima (Fig. 10; Ferrari et al., 2012), are some of the warmest modern subduction P - T paths. However, a warm subduction P - T path comparable to that predicted for Kii (Suenaga et al., 2019), is consistent with the peak conditions experienced by the Epidote Amphibolite Unit and Amphibolite Unit of the Catalina Schist, where the latter experienced metamorphism at ~ 1 GPa and 650–730°C (Penniston-Dorland et al., 2018) over ~ 116 – 108 Ma (Anczkiewicz et al., 2004; Harvey et al., 2021b; Fig. 11a). Following underplating of the amphibolite unit, up-dip transport as has been proposed by Platt and Schmidt (2024), would allow for the juxtaposition of the various tectonometamorphic slices of the Catalina Schist (Fig. 11).

Structurally above the section of coherent epidote-amphibolite metabasalts and metasedimentary rocks lies a ~ 100 m thick highly sheared and metasomatized mélange zone (Hoover et al., 2022a; Hoover et al., 2025). Petrologic modeling work done on hybrid lithologies in this zone indicate temperatures of ~ 550 °C and pressures of 1 GPa, which overlap with the peak P - T field associated with Ep_2 here, but extend to higher pressures (Fig. 10). Differences in the P - T conditions between the mélange and the metabasalts may be due to differential movement of materials within the interface shear zone mélange, with up-dip transport of blocks from higher grade conditions within the mélange as has been suggested by Harvey et al. (2021a), or due to the metabasalts not fully recording the peak pressure conditions of the P - T path. The talc-chlorite-actinolite schists that dominate the mélange zone require substantial mechanical and chemical mixing, facilitated in part by infiltrating fluids. The P - T path and model established here for the dehydration history of the slab components adjacent to the mélange zone suggests that these basalts could have provided some of water to the mélange shear zone. While dehydration of epidote and eventually amphiboles continued to peak conditions, our phase equilibria modelling suggests that over 400–600°C at 0.65 GPa, the metabasalts would lose 1 wt.% H_2O (Fig. 9, 10a) along this portion of the P - T path. The metabasalts therefore likely contributed some amount of fluid to the plate interface to facilitate mélange formation. Indeed, the breakdown of epidote into quartz, amphibole, and ultimately clinopyroxene at > 450 °C along

the P - T path we predict could conceivably produce fluids enriched in Ca and LREEs that provide the Ca-rich fluid inferred to infiltrate the mélange zone based on the growth of actinolite and LREE-enriched titanites (Hoover et al., 2025). It is also important to consider that the P - T path and dehydration history model that we construct here is only based on rocks from the upper <100 m of oceanic crust. Hydrated metabasalts, metagabbro, and lithospheric mantle deeper in the subducting slab would experience a different P - T path and undergo a different dehydration evolution as a result of differences in bulk composition and initial hydration, and may also have provided fluids to the plate interface mélange (e.g., Hacker, 2008; Epstein et al., 2024).

The final, retrograde stage of metamorphism of the metabasalts (from Ep_2 to Ep_3) is then consistent with refrigeration of the underplated Epidote Amphibolite Unit in a subduction zone cooling by continued subduction of cool oceanic lithosphere in its new, shallower geometry. This step involves cooling by $\sim 100^\circ\text{C}$ while maintaining the same pressure or slightly increasing in pressure relative to the peak conditions of Ep_2 (Fig. 9). White mica $^{40}\text{Ar}/^{39}\text{Ar}$ dating yields ages of 97–101 Ma, suggesting cooling of the unit through 400°C by that time (Grove and Bebout 1995). These cooling ages overlap with those of the next youngest unit, the Epidote Blueschist Unit, which has white mica $^{40}\text{Ar}/^{39}\text{Ar}$ ages of 95–99 Ma, and we know lawsonite blueschist facies metamorphism at $\sim 330^\circ\text{C}$ and ~ 1.0 GPa occurred in the youngest unit of the Catalina Schist at ca. 95 Ma (Platt and Schmidt, 2024; Grove and Bebout, 1995), roughly 10 Myr following epidote amphibolite facies metamorphism. Near isobaric cooling of the Epidote Amphibolite Unit (from Ep_2 to Ep_3) is therefore consistent with the later underplating of colder rocks. Paired with the structurally-higher and higher-grade amphibolite unit, which was underplated and metamorphosed at ~ 650 – 730°C and 1.3–1.4 GPa (Penniston-Dorland et al., 2018; Harvey et al., 2021a) earlier at ca. 115 Ma, this is all consistent with a model of a cooling subduction zone (e.g., Platt and Schmidt, 2024).

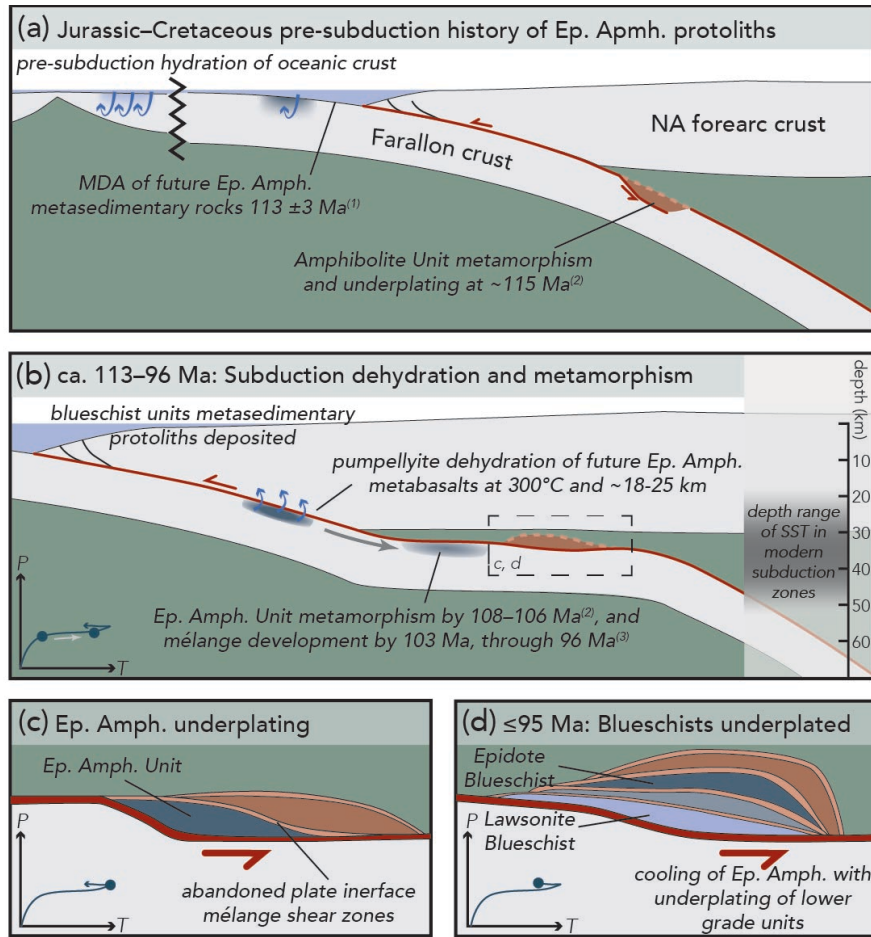


Figure 11. A schematic cartoon illustrating the tectonic evolution of the Catalina Schist. (a) The oceanic crust experiences hydration prior to subduction. MDA = maximum depositional age from detrital zircon data of ¹Grove et al. (2008). (b) Slab flattening allows the rocks of the future Epidote Amphibolite Unit to experience near isobaric heating to epidote amphibolite conditions. Ages are based on hornblende ⁴⁰Ar/³⁹Ar ages of ²Grove and Bebout (1995), and rutile and titanite U-Pb ages from mélangé rocks of ³Hoover et al. (2025). (c) Underplating of the Epidote Amphibolite Unit occurs near peak metamorphic conditions, before (d) retrograde cooling occurs along with underplating of the lower-grade Epidote Blueschist and Lawsonite Blueschist units.

5.5 Implications for subduction dehydration and deformation at the plate interface

Geophysical observations of modern subduction zones suggest that the deep subduction interface is a fluid rich environment, which has significant implications for the possible mechanisms of deformation within this shear zone (e.g., Bürgmann, 2018). In subduction zones including Cascadia, Nankai, Mexico, and Hikurangi, seismological observations of high Vp/Vs ratios suggest high pore fluid pressures below the seismogenic zone where deep slow slip and tremor occur (e.g., Audet et al., 2009; Shelly et al., 2006; Song et al., 2009; Wallace and Eberhart-Phillips, 2013). High pore fluid pressures can promote frictional deformation in rocks at the deep plate interface, which is a possible mechanism for slow slip and tremor events (e.g., Rubin, 2008; Segall et al., 2010; Gao and Wang, 2017; French and Condit, 2019). Metamorphic dehydration reactions at these conditions are commonly proposed as a source for such fluids on

the basis of petrologic modeling (e.g., Peacock, 2009; Fagereng and Diener, 2011; Condit et al., 2020). Reconstructing the P - T path and dehydration history of the Catalina Schist Epidote Amphibolite Unit in this study therefore helps us to constrain the magnitudes of fluid flux from metamorphic reactions and the conditions at which they occurred in ancient subduction zones.

Combining the P - T path we have constructed with the results of our phase equilibria modeling, we provide a model for the dehydration history of the Epidote Amphibolite Unit metabasalts (Fig. 10). The inferred shallow pre-subduction stratigraphic position of the meta-pillow basalts and their Sr-isotope ratios (Fig. 8) and inferred water/rock ratios provide evidence for high water contents prior to subduction. During prograde subduction, the mineral-bound H_2O content of the metabasalts changes from 5.5 wt% to 2.5 wt% across the pumpellyite–epidote transition at 290–310°C and 0.5–0.7 GPa (~18–25 km depth), followed by a gradual decrease to ~0.9 wt% at 550°C (Fig. 10a). The reaction pumpellyite + quartz + chlorite = epidote/clinozoisite + actinolite + quartz + H_2O therefore represents a major dehydration step that could release a large quantity of water along the subduction plate interface. Similarly, in studies of field exposures of sequences of isograds in metabasaltic rocks, authors have inferred that a relatively spatially narrow transition zone (<1 km) between prehnite-pumpellyite facies rocks and actinolite-bearing rocks suggests that the prograde transition between these two parageneses occurs over a narrow temperature range (Starr et al., 2020; Starr and Pattison, 2019). These interpretations support the possibility that the narrow temperature range over which we predict the epidote-in reaction to occur would correspond to a narrow region in the subduction zone into which these metabasalts would have released mineral-bound water through prograde metamorphic reactions.

The P - T path followed by the rocks prior to the initial growth of metamorphic epidote (Ep₁ field, Fig. 10) is also important for the dehydration history because prior to the pumpellyite-out reaction, there is potential for the release of an additional ~1 wt.% H_2O . Depending on the slope of the P - T path of the subducting slab, this additional water may be released in one of two possible pulses. In the case of a warmer P - T path (steeper P/T), this may happen at much lower P - T conditions (<0.3 GPa and <250°C) with the breakdown of laumontite. With a cooler path (shallower P/T), this will occur at similar conditions to the ultimate pumpellyite-out reaction—within 10s of degrees cooler temperatures and a similar range of pressures—with the breakdown of lawsonite (Fig. 9a). While we lack geologic constraints on this early stage of subduction of the metabasalts, this early dehydration history may have implications for fluid abundance and mineral reaction kinetics in the shallow forearc and subduction seismogenic zone.

Our interpretations of the dehydration history of the metabasalts is consistent with the results of previous petrologic modeling of subducting oceanic crust, which suggest that fluids may be released from the slab in pulses at various P - T conditions depending on the subduction geotherm (e.g., Peacock, 2009; Condit et al., 2020). The P - T path that we interpret for the Catalina Schist Epidote Amphibolite Unit is most similar to that modeled for modern subduction zones where flat-slab subduction occurs—i.e., central Mexico, Peru, and central Chile (Manea et al., 2017)—where slab-top temperatures increase at near-isobaric conditions as the subducting plate moves ~horizontally beneath the overriding plate. In our model, however, the Catalina Schist metabasalts experienced lower pressures than those expected for these modern systems (>1 GPa). The pulse of fluid produced by the epidote-in reaction would be delivered to the plate interface by the metabasalts at depths consistent with some of the shallowest depths at which deep slow slip and tremor occur in modern subduction zones, such as at the Nankai Trough where small slow slip patches occur at 10–20 km depth on the plate

interface, potentially overlapping in some cases with the downdip region of the seismogenic zone (Kano and Kato, 2020).

The pulse of fluid release that occurred in our samples also suggests that local dehydration reactions at the depths of the plate interface that host slow slip and tremor may facilitate the high pore fluid pressures that have been geophysically inferred to exist in this region. Long-distance up-dip fluid transport has been suggested along the subduction zone interface as interpreted from isotopic data in the exhumed rock record (e.g., Jaeckel et al., 2018) and from geophysical studies (Frank et al., 2015). However, such advection is not required to produce high pore fluid pressures at shallower conditions like those inferred here for the pumpellyite-epidote reaction. At these shallow conditions, local dehydration of metabasalts as a result of these early dehydration reactions may provide significant quantities of H₂O-rich fluids near the plate interface to produce high pore fluid pressures.

In addition to promoting high pore fluid pressures, these fluids will also effect chemical changes in rocks they encounter, metasomatizing rocks at the plate interface and producing unique, sometimes mechanically-weak lithologies. For example, in the epidote amphibolite mélange zone adjacent to the metabasalts studied here, Hoover et al. (2022a, 2025) and Easthouse et al. (2025) document the mechanical mixing of ultramafic, metamafic, and metasedimentary lithologies within this zone and chemical and isotopic exchange between these lithologies. Periodic influxes of Ca-rich fluids contributed to the formation of actinolite-rich rocks in the mélange, while also increasing pore fluid pressures and enabling frictional slip in talc-rich rocks—a likely mechanism behind the occurrence of deep slow slip at the subduction interface (e.g., French and Condit, 2019). In this way, these fluids that are produced by dehydration reactions in the subducting slab are important not only for maintaining high pore fluid pressures in the plate interface, but also for producing metasomatic rocks that may preferentially localize deformation.

Although the *P-T* path predicted here does not exactly match those modeled for modern subduction zones, we present evidence for dehydration having occurred in these rocks at *P-T* conditions within the range of conditions estimated for the occurrence of deep slow slip. The reactions here are therefore relevant for modern subduction zones and understanding fluid production and movement at the base of the seismogenic zone and below. The approach used here to assess the *P-T* path of these epidote-rich rocks can also be applied to other exhumed suites of underplated units to look for evidence of dehydration at conditions similar to those over the full range of slow slip and tremor depths (Audet and Kim, 2016; Behr and Bürgmann, 2021).

5.6 Epidote as a petrologic tool in sub-amphibolite facies metamorphic rocks

Because of the ability for epidote group minerals to grow over *P-T* conditions as a result of a variety of hydrothermal and metamorphic processes, its common occurrence in low-medium-grade metamorphic rocks, and its propensity for incorporating trace elements into its crystal structure, we argue that epidote minerals present a powerful tool for investigating sub-amphibolite facies metamorphic processes. Whereas garnet geochemistry and geochronology have been well-developed for studying high-grade metamorphic processes in subduction zones (e.g., Skora et al., 2006; Page et al., 2007; Baxter and Caddick, 2013), we demonstrate here that in garnet-free rocks, epidote has the potential to be used as an important recordkeeper of subduction processes. In particular, as shown by phase equilibrium models, metamorphic epidote growth in metabasalts first occurs at ~250–400°C depending on pressure (Fig. 9), which is an important range encompassing the base of the seismogenic zone and regions

hosting slow slip in subduction zones. The first appearance of epidote in metabasalts, as well as subsequent growth zoning and any inclusions that are incorporated can therefore capture important prograde subduction processes and the P - T path of subducting rocks. In other examples, recent work has used epidote compositions from altered basalt cores of modern oceanic crust to differentiate between the signatures of multiple fluid sources based on Sr isotope ratios (Chen et al., 2023). Epidote geochronology also has the potential to provide timing constraints on these processes, such as recently-developed *in-situ* U-Pb dating for vein epidotes (Peverelli et al., 2021; Peverelli et al., 2024). Even lacking direct petrochronological constraints on epidote growth, epidote chemistry and textures can inform P - T path and tectonic reconstructions, as we demonstrate here. In the case of our study, major- and rare earth element compositions of epidote minerals have helped to reveal aspects of the prograde metamorphic history of metabasalts, distinguish between textures produced by seafloor hydrothermal alteration and metamorphism, and track peak and retrograde metamorphism of Catalina Schist metabasalts. The many processes and conditions in which epidote grows can also pose challenges and necessitate careful coupled textural and chemical analyses as the composition of epidote is sensitive to bulk rock chemistry, P - T conditions, and fluid compositions (e.g., Grapes and Hoskin, 2004). There are therefore many opportunities for detailed textural and chemical studies of epidote to extend its utility as a tracer of metamorphic, hydrothermal, and deformational processes.

6 Conclusions

We constrained the P - T path of metabasalts in the Epidote Amphibolite Unit of the Catalina Schist and have used thermodynamic models to estimate a dehydration history for these rocks.

- Metabasalts from the Epidote Amphibolite Unit of the Catalina Schist followed a relatively shallow P - T path across a pumpellyite–epidote reaction at $\sim 300^\circ\text{C}$ and 0.4–0.7 GPa to peak conditions at $\sim 550^\circ\text{C}$ and 0.6–0.9 GPa. Following peak T metamorphism, these rocks were underplated beneath North America and experienced cooling consistent with refrigeration in a cooling subduction zone during underplating of subsequent packages of rock.
- The breakdown of pumpellyite in these rocks led to the development of epidote-rich zones, which form mm–cm scale networks that represent limited channel-forming reactive porosity structures that developed as a result of volume reduction in the formation of epidote. These porosity structures drained to permeable inter-pillow fluid pathways and out of the slab.
- The pulse of water released during pumpellyite breakdown could have contributed to elevated pore fluid pressures at the down-dip limit of an ancient subduction megathrust seismogenic zone, and this and similar reactions may contribute to the high pore fluid pressures present in these regions that could host deep slow slip and tremor.
- Epidote can serve as a powerful record of poly-stage low- to medium-grade metamorphic processes and hydrothermal alteration when pairing textural and geochemical observations.

Inclusion in Global Research Statement

Samples used in this study were collected on the ancestral lands of the Gabrielino-Tongva on Pimu'nga, so-called Santa Catalina Island, California. We affirm their ongoing sovereignty on these lands in the face of settler-colonialism. All sampling was carried out in consultation with Tongva tribal members and with their permission. Permission to collect samples was also granted by the Catalina Island Conservancy.

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Conflict of Interest Statement

The authors have no conflicts of interest to disclose.

Availability Statement

Major element mineral analyses (Table S7), epidote trace element analyses (Table S8), and Sr isotope measurements and W/R calculations (Table S9) are available from the Zenodo repository Lindquist et al. (2026).

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Chapter 4: Metamorphism and deformation during subduction and underplating: the evolution of the Catalina Schist Epidote Blueschist Unit

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Abstract

The metamorphic and rheological evolution of rocks at and near the subduction plate interface are important controls on the slip behavior of the plate interface and the transfer of stresses updip to the seismogenic zone. The environment downdip of the seismogenic zone is likely characterized by lithologic heterogeneity and abundant fluids, but our understanding of this region is complicated by the preservation and exhumation of rocks from these regions. The Epidote Blueschist Unit of the Catalina Schist represents a deformed suite of meta-basaltic and sedimentary rocks that were underplated to North America during the Cretaceous. These rocks provide insight into the metamorphic and structural evolution of rocks near the plate interface during subduction and underplating. Here we present petrologic analyses including mineral compositions and phase equilibrium modeling, as well as microstructural analyses including electron backscattered diffraction of amphibole and quartz. New pressure-temperature constraints for the Epidote Blueschist unit place peak metamorphic conditions at < 0.7 GPa and $470\text{--}520^\circ\text{C}$, which was achieved following near isobaric heating from $< 300^\circ\text{C}$. We demonstrate that these rocks primarily record deformation related to underplating and that stresses during underplating and exhumation were ca. $50\text{--}180$ MPa. During underplating and potentially also during prograde subduction, both Na-amphibole and actinolite in these rocks deformed via dissolution precipitation creep, whereas quartz dislocation creep was activated by higher stresses during underplating. The record of dissolution precipitation creep provides additional evidence to suggest that the subduction interface is a fluid-rich environment at the conditions of slow slip, and the preservation of early foliations in these rocks suggests that subducting metabasalts may be weak and deform viscously even in the seismogenic zone.

1 Introduction

Subduction plate interfaces host slip phenomena ranging from aseismic viscous creep, to slow slip and tremor, to earthquakes. At deeper and hotter conditions than the seismogenic zone, the environment of and mechanisms that host deformation remain ambiguous (Behr and Burgmann, 2021; Condit et al., 2022, Lay et al., 2012). This deformation, and the conditions and mechanisms that control it, have important implications for the behavior of the whole subduction interface system. Viscous deformation and slow slip events transmit stresses updip to load the seismogenic zone (Wech and Creager, 2011) and exert control on the overall rate of plate motions (Behr and Becker, 2018). Furthermore, the chemical (metamorphic and fluid-related)

processes that occur during subduction at these same conditions not only influence this deformation, but also play a key role in global element cycling (e.g., Rüpke, 2004; Bebout and Penniston-Dorland, 2016; Hacker, 2008). Therefore, understanding the rocks that exist at the conditions of slow slip and tremor, their structural relations, rheology, and the fluid conditions that exist are important components of a holistic understanding of subduction zones.

In addition to prograde subduction processes, the underplating or basal accretion of slices of the oceanic crust is an important mechanism for exhuming rocks from subduction zones and underplated rocks are the primary sources of insights into subduction deformation and metamorphism from the geologic record (Bebout and Penniston-Dorland, 2016). Domal duplex structures are observed in exhumed accretionary wedge and other subduction terranes in the geologic record (e.g., Hashimoto and Kimura, 1999; Ujiie et al. 2024; Menant et al., 2026) and inferred geophysically in modern subduction zones (e.g., Moore et al., 1991; Ito et al., 2009; Kimura et al., 2010). These structures represent the episodic stepping-down of the megathrust interface into the subducting plate and the resulting detachment of slices of the top of the oceanic crust including sedimentary and basaltic rocks. These slices then have an opportunity to be exhumed and returned to the Earth's surface, providing a record of subduction-related processes in the subducting slab and the plate interface shear zone. It has also been suggested that underplated rocks may continue to host tremor and slow slip phenomena even after underplating (e.g., Littel et al., 2023; Sammis and Bostock, 2025).

This work provides an example of some of the rocks and deformation processes at play near the plate interface at the conditions of underplating and deep slow slip and tremor. Here we present petrologic and microstructural observations from metabasaltic and metasedimentary rocks of the Epidote Blueschist Unit of the Catalina Schist, from which we interpret a metamorphic and deformation history for these rocks. The Epidote Blueschist Unit experienced peak metamorphism at P - T conditions similar to those at which slow slip and tremor likely occur in modern subduction zones. Using phase equilibrium modeling, we are able to reconstruct a P - T path for this unit from subduction through underplating. In addition to petrographic observations, we present electron backscatter diffraction (EBSD) microstructural analyses of amphibole and quartz, which provide constraints on the active micromechanisms of deformation and the stresses driving deformation in these rocks. We find that these rocks record hints of early subduction-related deformation, but are dominated by underplating-related fabrics.

2 Geologic Background

The Catalina Schist on Pimu'nga/Santa Catalina Island (California, USA) was constructed through the incremental underplating of subducted slivers of oceanic crust beneath North America during the Cretaceous. Each unit of the Catalina Schist consists of mafic and sedimentary lithologies as well as *mélange* zones that occur as the tectonic boundaries of most of the units, all of which were metamorphosed at different P - T conditions. The top and bottom of the stack of units are defined by the structurally highest and oldest Amphibolite Unit and structurally lowest and youngest Blueschist Unit, which includes lawsonite-albite and lawsonite-blueschist facies rocks (Fig. 1) (Platt, 1975, 1976; Grove and Bebout, 1995). In between these units lie the Epidote Blueschist Unit and Epidote Amphibolite Unit—the former being structurally lower and younger than the latter. Both these units were originally included as part of the Greenschist Unit of Platt (1975), but subsequently divided by Grove and Bebout (1995) into

the separate units.

The focus of this work is the Epidote Blueschist Unit, which includes the rocks described by Sorensen (1986) as glaucophanic greenschists, consists of metabasalts, metasedimentary rocks, and quartz-rich graphitic schists (e.g., Platt, 1976). Both metabasaltic and metapelitic rocks exhibit epidote-blueschist and epidote-actinolite greenschist mineral assemblages, with variation between the two assemblages sometimes occurring at the cm-scale in layers that are coplanar with the foliation in the rocks. On the basis of mineral assemblage, Sorensen (1986) estimated the glaucophanic greenschists of the Catalina Schist to have peak temperatures of 450–550°C and pressures of 0.6–1.2 GPa. Platt and Schmidt (2024) report RSCM temperatures for the Epidote Blueschist Unit of 390–470°C, and, similar to Sorensen (1986), they use the mineral assemblage (lack of jadeitic pyroxene and presence of glaucophane) to constrain peak pressures to 0.7–1.3 GPa. Timing of the formation of the Epidote Blueschist Unit is constrained by detrital zircon U-Pb ages giving a maximum depositional age of 100 ± 3 Ma, and by white mica $^{40}\text{Ar}/^{39}\text{Ar}$ cooling ages of 95–99 Ma. This places the timing of deposition and metamorphism of the Epidote Blueschist Unit protoliths between that of the older Epidote Amphibolite Unit and the younger Blueschist Unit.

In this work, we focus in particular on a suite of samples collected in Cottonwood Canyon which lie within several hundred meters of the top of the Epidote Blueschist Unit in a semi-coherent section of ocean plate stratigraphy (Fig. 1). The samples we describe here include epidote-blueschist metabasalt, epidote-actinolite greenschist metabasalt, metachert, and metapelitic samples with cm-scale blueschist-greenschist layering (Fig. 2, 3). All rocks contain a pervasive foliation, which, in non-metachert lithologies, is primarily defined by the shape-preferred orientation of amphibole and mica crystals as well as epidote-rich layers that may represent transposed veins.

The foliation of the Epidote Blueschist Unit in this area dips shallowly to the ENE with roughly down-dip plunging mineral lineations defined by amphibole crystal shapes (Fig. 1). Long-wavelength, gentle folding of the foliation around ~ESE-plunging fold axes (Platt, 1976) accounts for the recurrence of metasedimentary lithologies farther downstream to the W of the focus area, suggesting that much of this exposure lies within several hundred meters of what would have been the top of the oceanic crust. The Epidote Blueschist Unit is overlain by coherent mafic amphibolites of the Amphibolite Unit. The contact between the units is the Ollas fault zone, as described by Platt (1975, 1976) and Harvey et al. (2021). The thickness of this fault is not well-constrained due to a lack of outcrop over ~150 m between the base of the Amphibolite Unit—where Harvey et al. (2021) describe a *mélange* matrix of mafic blocks in variably actinolite-dominated and talc + biotite matrix—and the metachert at the top of the Epidote Blueschist Unit.

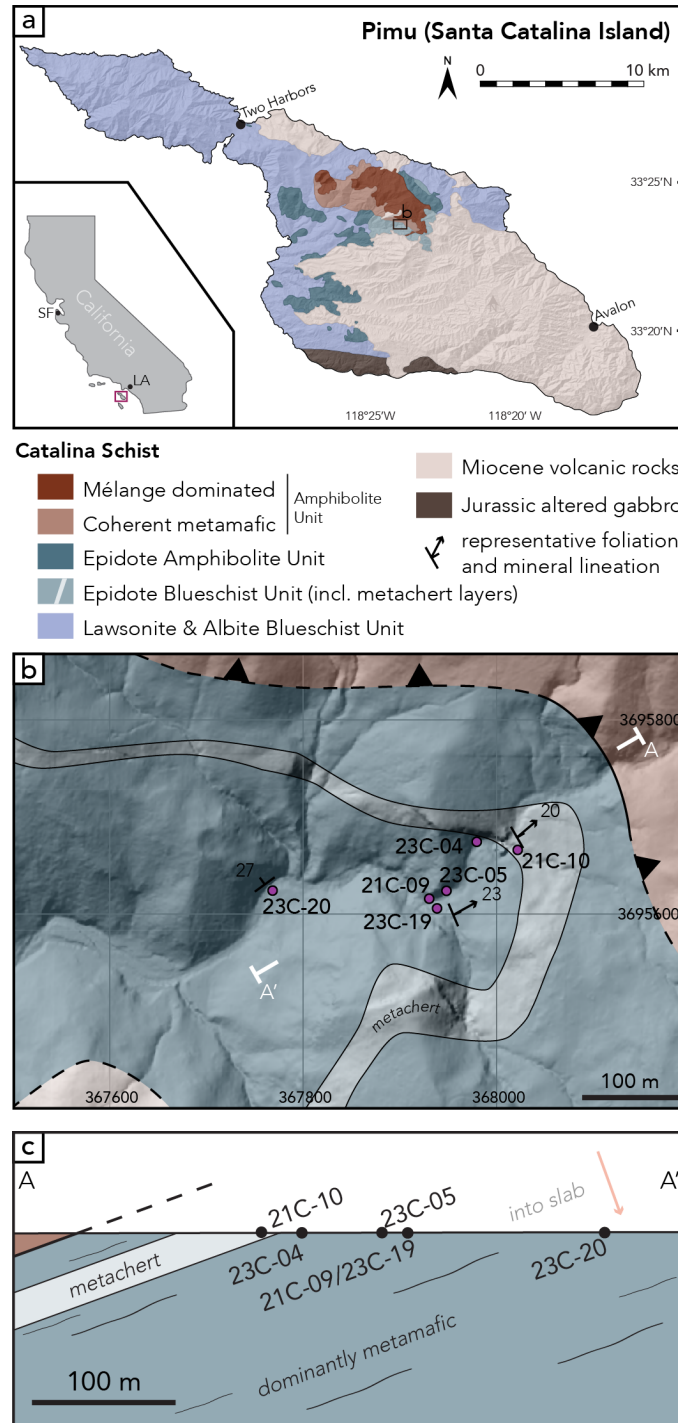


Figure 1. (a) Geologic map of Pimu after Platt (1975, 1976). (b) Detail map of Cottonwood Canyon. Purple dots represent sample locations. Map after Platt (2020). (c) Cross-section A-A' in (b). Topography is not expressed on the cross section.

3 Methods

3.1 EPMA, SEM, and EBSD analyses

Major element analyses of amphibole, epidote, plagioclase, white mica, garnet, and lawsonite were conducted at the University of Nevada, Reno with a JEOL JXA-iHP200F electron probe micro-analyzer using a focused beam with a 15 kV accelerating voltage and 20 nA beam current for spot analyses. For mineral formula recalculations of amphibole analyses, we follow the method of Locock (2014), based on the amphibole nomenclature of Hawthorne et al. (2012), and for recalculations of garnet analyses we use the method of Locock (2008).

EDS maps, EBSD data, and BSE images were also collected in the Molecular Analysis Facility at the University of Washington on an Thermo-Fisher Scientific Apreo SEM with an Oxford Instruments Ultim-Max 100 energy dispersive spectroscopy detector, and an Oxford Instruments Symmetry 2 EBSD detector, using a 20 kV accelerating voltage and 1.6–3.2 nA current.

Crystal orientation data for glaucophane, actinolite, and quartz were collected via EBSD using a sample with 70° tilt and 13 mm working distance. Step size ranged from 1 to 4.9 μm depending on grain size of the analyzed area. Indexing of EBSD data used Oxford HKL and American Mineralogist databases in the Oxford Aztec software. Post-collection data processing was done using the MTEX toolbox for MATLAB (Bachmann et al., 2010), which we used to plot phase and misorientation maps, and pole figures of crystallographic preferred orientations. All analyzed thin sections were cut close to parallel to any apparent mineral lineation and perpendicular to foliation.

3.2 Bulk rock compositions

Whole-rock major element abundances for several samples were measured using a Malvern PANalytical Zetium X-ray fluorescence (XRF) vacuum spectrometer at Franklin and Marshall College, and ferrous iron concentrations were determined by titration using the methods described in Mertzman (2000) and Waterton et al. (2020). Materials for bulk-rock XRF analysis were cut from samples so as to avoid weathered surfaces.

Bulk rock compositions for metasedimentary samples were calculated from the composition (as measured by EPMA) and modal abundances of minerals. The modal abundance of major minerals was estimated by calculating the area fraction of those minerals in thin section by thresholding EDS elemental maps and pixel counting using ImageJ. Bulk rock compositions used in phase equilibrium modeling are reported in Table 1.

3.3 PerpleX phase equilibrium modeling

To estimate the P - T conditions experienced by the Epidote Blueschist Unit and the path followed by these rocks during subduction and underplating, we use the Gibbs free energy minimization software Perple_X (v. 7.1.5; Connolly, 2005; Connolly, 2009) to model the equilibrium mineral assemblages of a suite of metabasaltic and metasedimentary rocks. All models are run over the P - T range 0.1–1.5 GPa and 200–650°C at H_2O -saturated conditions. Mafic bulk-rock compositions were modeled in the chemical system Na_2O – CaO – FeO – MgO – Al_2O_3 – SiO_2 – H_2O – TiO_2 – O_2 , and metasedimentary bulk compositions were modeled in the chemical system Na_2O – CaO – FeO – MgO – MnO – Al_2O_3 – SiO_2 – H_2O – TiO_2 – O_2 . We use the

internally-consistent thermodynamic database of Holland and Powell (1998) and fluid equation of state model of Holland and Powell (1991). The activity-composition models used in this modeling are reported in Table 2. Pseudosections are included as supplemental figures S1–S5.

Table 1. Bulk compositions used for phase equilibria modelling

Oxide (wt.%)	23C04D blueschist metabasalt	21C09B greenschist metabasalt ¹	23C04A blueschist metapelite ²	23C04A greenschist metapelite	21C10A1 metachert micaceous layer ²
SiO ₂	51.29	50.83	66.92	62.19	92.62
TiO ₂	2.65	2.19	0.11	0.45	0.14
Al ₂ O ₃	12.48	13.23	16.24	14.33	3.74
FeO [†]	14.31	12.60	6.02	9.67	0.42
MgO	6.39	9.21	2.93	5.13	0.28
CaO	6.30	8.85	1.05	3.79	0.04
Na ₂ O	3.54	2.68	1.21	3.35	1.16
K ₂ O	1.14	0.06	5.28	1.05	0.69
MnO	—	—	0.17	0.18	0.22
O ₂	0.48	—	—	0.10	—
H ₂ O	sat.*	sat.*	sat.*	sat.*	sat.*

*assumed water saturated for phase equilibrium modeling

[†]all Fe is reported as FeO

¹ imposed μ_{O_2} of -405340 J/mol as extracted from the blueschist metabasalt model

² imposed μ_{O_2} of -424310 J/mol as extracted from the greenschist metapelite model

Table 2. Activity-composition models used for phase equilibria modelling

Mineral	Solution Model	Citation
feldspar	feldspar	Fuhrman & Lindsley (1988)
clinoamphibole	cAmph(DP)	Diener & Powell (2012)
chlorite	Chl(HP)	Holland & Powell (1998)
pumpellyite	Pu(M)	Massonne & Willner (2008)
clinopyroxene	Omph(GHP)	Green et al. (2007)
orthopyroxene	Opx(HP)	Powell and Holland (1999)
olivine	O(HP)	Holland & Powell (1998)
epidote	Ep(HP)	Holland & Powell (1998)
ilmenite	Ilm(WPH)	White et al. (2000, 2014)
brucite	B	
carpholite	Carp(M)	Massonne & Willner (2008)
sudalite	Sud(M)	Massonne & Willner (2008)
antigorite	Atg(PN)	Padrón-Navarta et al. (2013)
chloritoid	Ctd(HP)	Holland & Powell (1998)

garnet	Gt(WPPH)	White, et al. (2005)
biotite	Bio(WPH)	White, et al., (2007)
stilpnomelane	Stlp(M)	Massonne & Willner (2008)
spinel	Sp(WPC)	White et al., (2002)
cordierite	hCrđ	
staurolite	St(HP)	Holland & Powell (1998)

4 Results

4.1 Petrographic observations and mineral chemistry

4.1.1. Metabasaltic Samples

Metabasalts occur in two main forms: one with an epidote-blueschist assemblage of Na-amphibole + white mica + epidote + albite + chlorite + quartz + titanite and minor biotite, and one with an epidote-actinolite greenschist assemblage of actinolite + chlorite + albite + epidote + quartz + titanite. In both types of metabasalts, epidote and albite are present as porphyroblasts to porphyroclasts and contain inclusions of lawsonite, white mica, sodic amphibole, and titanite. We note that this is the only description of lawsonite in the Epidote Blueschist Unit aside from a passing mention by Sorensen (1986). The matrix of both metabasalts is defined by the shape preferred orientation of amphibole and mica, and inclusion trails within porphyroblasts define a folded foliation.

4.1.1.1. Epidote-blueschist metabasalt

Epidote and albite are present as porphyroblasts in the epidote-blueschist metabasalt. Feldspar has end-member albite compositions (Table 3). Epidote exhibits subtle core-rim zoning with more Fe-rich cores and Al-rich rims, but the difference is minor and epidote analyses show a median composition of $\text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Al} - 2)$ near 0.5, intermediate between epidote and clinozoisite (Fig. S6). Epidote also occurs in masses that likely represent relict veins that have been dismembered and transposed (Fig. 2b). The dominant zoning pattern observed in amphibole inclusions is magnesioriebeckite cores with glaucophane rims (Fig. S7b). White mica inclusions in epidote show a broad spread in Si pfu over 3.1–3.7. Lawsonite inclusions occur as anhedral blebs (Fig. S7a).

In the matrix of the epidote-blueschist metabasalt, amphibole commonly exhibits zoning, with dominantly glaucophane cores rimmed by magnesioriebeckite (Fig. 3e,f). More rare are magnesioriebeckite cores and glaucophane mantles, like the pattern observed in inclusions, with additional magnesioriebeckite rims that appear to truncate some earlier core-rim zoning (Fig. 3e,f). Phengite crystals in the matrix have compositions that cluster tightly around 3.5–3.6 Si pfu (Fig. 4b). Minor biotite is present in this sample—appearing mainly to be an alteration product after white mica—primarily in regions associated with small fractures that cross-cut the foliation—and in tails on porphyroblasts.

Extensional C' shear bands are visible in the blueschist metabasalt (Fig 3d). The C' shear band is defined primarily by fine grained (<10 μm) quartz and albite, but is visible also in the warping of adjacent micas. Adjacent to at least one C' shear band, amphibole also shows conspicuous microboudinage textures, with more calcic amphibole apparently growing in boudin necks. While microboudinage textures are observed in amphibole elsewhere in the matrix of the blueschist metabasalt, it appears more obvious and extensively developed adjacent to the shear band.

4.1.1.2. Epidote-actinolite greenschist metabasalt

In the greenschist metabasalt, while no Na-amphibole or white mica are present in the matrix, these phases do occur as dominantly unzoned, anhedral to subhedral glaucophane inclusions in epidote and albite, along with very fine inclusions of tabular, euhedral white micas, subhedral epidote, and amoeboid lawsonite. White mica inclusions in albite are euhedral and have 3.3–3.4 Si pfu. Similar to in the epidote-blueschist metabasalt, lawsonite inclusions in albite occur as anhedral blebs. Both epidote and albite porphyroblasts are dominantly inclusion-rich, but most also feature thin (<50 μm), inclusion-free rims. On epidote, these rims appear as distinct euhedral growths of more Fe-rich epidote interfingering with chlorite in the matrix (Fig S7c). Epidote and albite porphyroblasts also commonly feature quartz-filled fractures or boudin-necks (Fig. 7).

In the matrix of the greenschist metabasalts, actinolite is commonly zoned with rims that appear bright in BSE images, with EPMA analyses suggesting lower Ca and Mg, and higher Na and Fe compared to the cores (Table 3). Incipient microfractures perpendicular to foliation direction in actinolite are possible microboudinage textures. Coarse-grained actinolite and chlorite occur in strain shadows on porphyroblasts, and chlorite in these zones often shows interlocking, randomly-oriented textures in which actinolite crystals also appear randomly oriented. These tails occur in sharp contrast to the strongly-oriented actinolite and mica that define the foliation in the rest of the matrix (Fig. S7d). Fine, euhedral titanite occurs throughout the matrix.

Like in the blueschist metabasalts, the greenschist metabasalts include apparent folded and dismembered segments of epidote veins. In one greenschist sample, we observe a core of Fe-oxides and in an epidote vein that shows decreasing Fe-content of the epidote from core to margin.

Table 3. Representative mineral compositions

plagioclase		amphibole				mica		garnet		epidote			
lithology setting analysis #	blueschist metabasalt	greenschist metabasalt	blueschist metabasalt			blueschist metapelite	blueschist metapelite	metachert	metapelite		greenschist metabasalt	blueschist metabasalt	
		core	rim	core	rim	core	rim			metapelite			
		Act	Act	Gln	Mrb								
	599	261	262	352	353	430	431	633	782	381	411	422	
SiO ₂	67.98	55.36	54.42	56.95	55.94	49.71	49.99	36.10	37.00	SiO ₂	37.65	38.28	38.47
TiO ₂	0.00	0.02	0.03	0.03	0.09	0.09	0.08	0.25	0.20	TiO ₂	0.09	0.06	0.09
Al ₂ O ₃	19.81	1.78	1.79	9.92	4.38	28.56	25.37	19.67	20.51	Al ₂ O ₃	24.67	26.95	26.78
Cr ₂ O ₃	0.00	0.00	0.00	0.00	0.00	0.01	0.00	-	-	Cr ₂ O ₃	0.04	0.02	0.00
MnO	0.01	0.26	0.26	0.13	0.28	0.03	0.05	29.77	17.25	Mn ₂ O ₃	0.93	0.19	0.31
FeO	0.07	12.66	13.85	14.63	19.45	2.79	4.87	7.27	12.48	Fe ₂ O ₃	12.14	9.36	9.36
MgO	0.00	15.61	15.21	8.03	8.59	3.08	3.36	0.25	0.21	MgO	0.01	0.03	0.03
CaO	0.09	11.75	11.26	1.03	1.29	0.00	0.00	4.95	11.67	CaO	22.56	23.19	23.34
Na ₂ O	11.59	0.70	0.76	6.62	6.21	0.43	0.12	0.03	0.02	Na ₂ O	0.01	0.02	0.00
K ₂ O	0.04	0.05	0.05	0.01	0.02	10.60	11.00	0.00	0.00	K ₂ O	0.00	0.00	0.01
F	0.00	0.04	0.00	0.00	0.00	0.04	0.01	-	-	F	0.00	0.05	0.03
Cl	0.00	0.00	0.00	0.00	0.00	0.00	0.00	-	-	Cl	0.00	0.00	0.00
Total	99.58	98.232	97.639	97.351	96.249	95.35	94.83	98.30	99.34	Total	98.117	98.153	98.431
							Spessartine	0.70	0.39				
							Pyrope	0.01	0.01				
							Almandine	0.14	0.26				
							Grossular	0.11	0.31				
							Andradite	0.03	0.02				

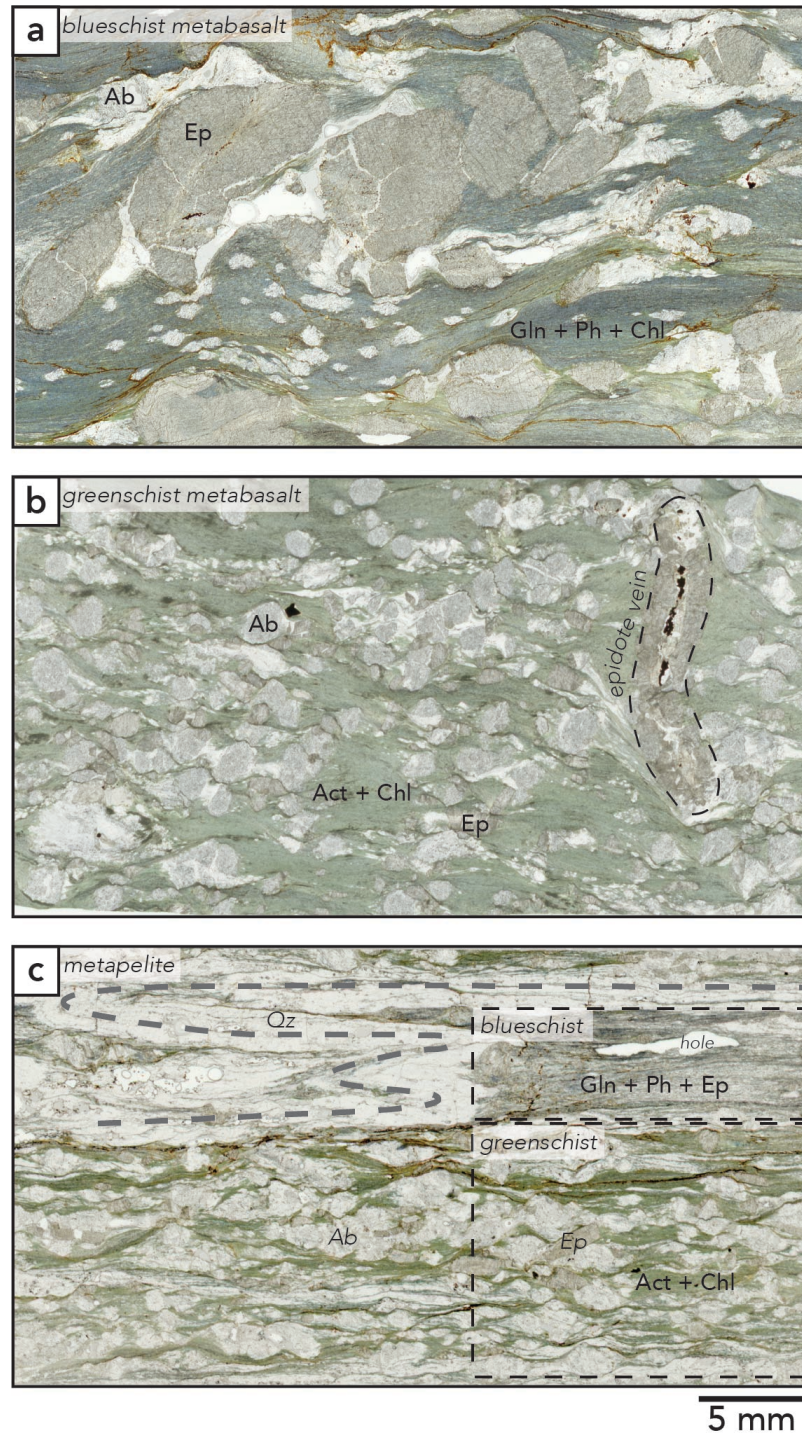


Figure 2. PPL scans of thin sections described in this work demonstrating cm-scale textures and compositional domains. (a) the epidote-blueschist metabasalt; (b) the epidote-actinolite metabasalt with outlined dismembered epidote + Fe-oxide vein; (c) the compositionally-layered metapelite. Dashed grey line highlights folding in the quartz

layer. Dashed boxes outline the fields that were mapped by EDS to estimate bulk compositions for the blueschist and greenschist metapelitic layers. Mineral abbreviations here and in following figures after Whitney and Evans (2010).

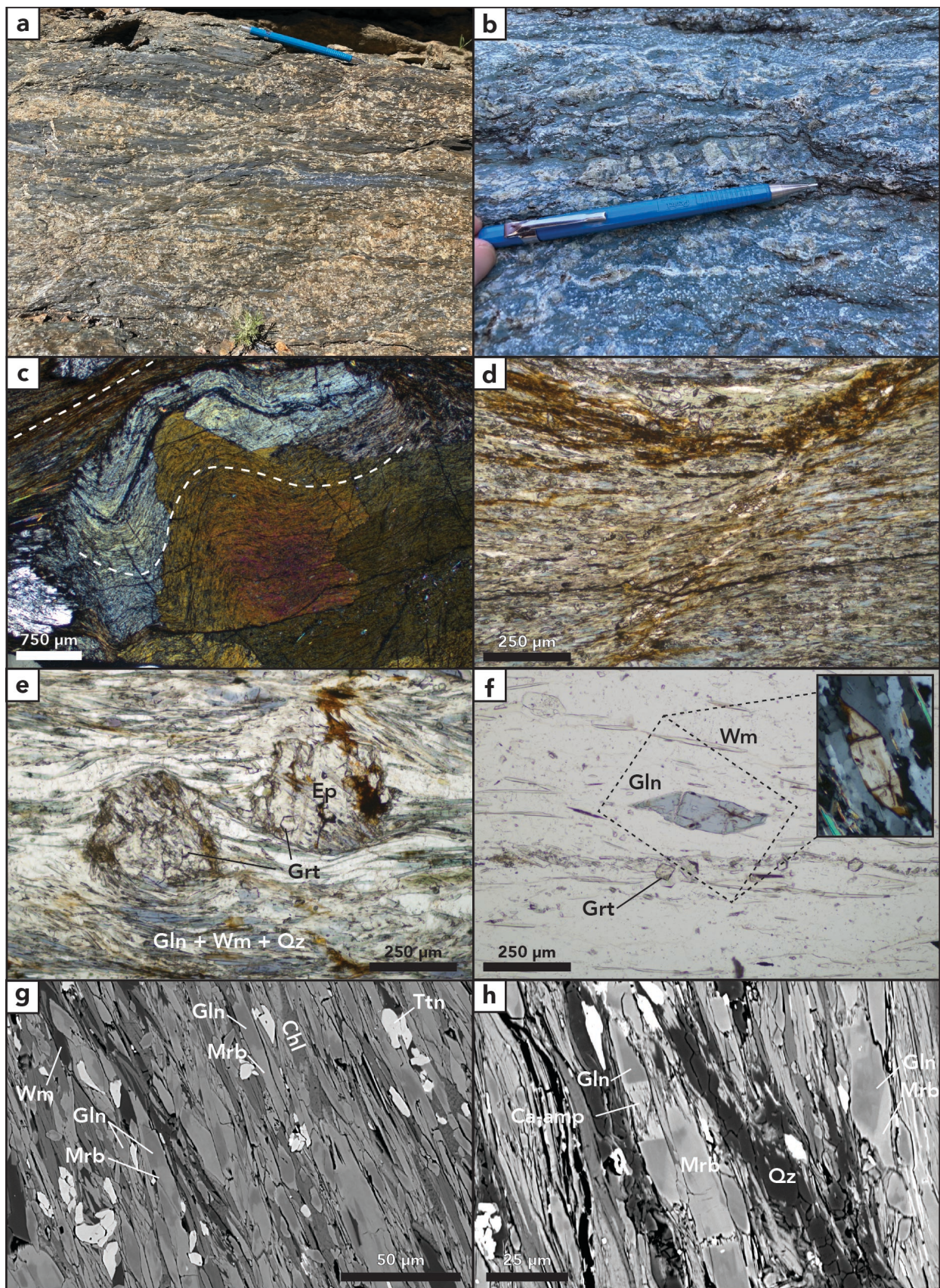


Figure 3. Field photos (a–b), photomicrographs (c–f), and BSE images (g–h) highlighting micro- and macro-structures.

4.1.2. Metasedimentary samples

We analyzed two metasedimentary samples from the Epidote Blueschist Unit: one garnet-bearing metachert (21C10A), and one more metapelitic sample (23C04A) that includes alternating glaucophane- and actinolite-rich layers.

4.1.2.1. Metachert

The metachert is dominantly (>90%) quartz, with Ph + Ab + Na-Amp + Grt, and minor Fe-oxides. Non-quartz phases mostly concentrate in more “pelitic” layers and share a preferred orientation which help define the foliation in the metachert. The appearance of quartz in elongate ribbons with mm-scale long-axes parallel to the foliation also contribute to defining the foliation. Quartz also exhibits undulose extinction and patchy subgrain textures. In phengite-rich layers, the micas define an S-C’ fabric in which quartz exhibits significantly smaller grain size in the C’ shear bands than in the surrounding matrix.

Garnets in the metachert are subhedral to euhedral and exhibit concentric zoning, and are spessartine rich ($X_{\text{Sps}}=0.67\text{--}0.73$, $X_{\text{Alm}}=0.09\text{--}0.15$, $X_{\text{Grs}}=0.09\text{--}0.15$, $X_{\text{Py}}=0\text{--}0.02$, Table 3). These garnets show minor core–rim compositional variation, but a large variability in the trajectory of zoning patterns, with trends of both increasing and decreasing Mn concentrations from core to rim, with some concentric zones reflecting larger spikes or troughs of Mn concentrations (Fig. 4c). White mica in the metachert has phengitic compositions with Si pfu 3.3–3.5 (Fig. 4b). Amphibole is dominantly glaucophane, with thin rims of magnesioriebeckite (Fig 3a). Glaucophane may appear fractured and pulled apart, in which cases magnesioriebeckite rims entirely surround the two pieces.

4.1.2.2. Metapelite

The metapelitic sample contains two distinct compositional domains: blueschist quartz-rich domains that are associated with sodium amphibole and white mica, and greenschist actinolite- and chlorite-dominated domains. These domains define a blue-green striping at the cm-scale that is visible in outcrop (Fig. 2c, 3a).

Blueschist quartz-rich domains appear to be dominantly quartz, occurring as isoclinally folded layers. These folded quartz layers contain minor inclusions of Na-amphibole, phengite, and biotite, and more “pelitic” zones of Na-amp + Ph + Chl + Ep + Qz + Ttn that include a foliation defined by the shape preferred orientation of amphibole and mica, and isoclinal folds thereof that are co-planar with the isoclinally folded quartz layers. Epidote porphyroblasts include Na-amp + Grt + Qz + Ttn, with Na-amphibole long-axes defining a fabric oblique to that in the surrounding matrix. Sodium amphiboles in the matrix of this domain have dominantly ferro-glaucophane compositions, with little difference between core- and rim-analyses. The Na-amphibole in this domain have $\text{Mg}/(\text{Mg} + \text{Fe}^{2+})$ values that extend to lower values than Na-amphiboles in the metabasalts.

The greenschist actinolite-rich domains include Act + Chl + Ab + Ep + Ph + Qz + Ttn, as

well as trace Na-amphibole. The foliation in this domain is defined primarily by the shape preferred orientation of actinolite and chlorite. Epidote and albite porphyroblasts in this domain contain inclusions of Ph + Qz + Grt + Ttn, and epidote contains inclusions of albite and vice versa. Folded inclusion trails in both porphyroblasts are defined by layers of quartz and titanite. White mica analyses with the lowest Si pfu analyzed in the metapelite come from the cores of micas that are found as inclusions in albite in this layer.

Garnet and phengite show similar zoning patterns and compositions between the matrix of both blueschist and greenschist domains. Garnet inclusions are euhedral and not present elsewhere in the matrix. These have more intermediate and variable compositions than those in the metachert with $X_{\text{Sps}}=0.30\text{--}0.52$, $X_{\text{Alm}}=0.24\text{--}0.32$, $X_{\text{Grs}}=0.17\text{--}0.36$, and $X_{\text{Py}} = 0\text{--}0.01$. The garnets typically exhibit Mn-rich cores, which grade to more Ca- and Fe-rich rims, often with one Mn-rich zone in the mantle (Fig. 4c). Garnets that are included within albite porphyroblasts tend to show stronger zoning than those included in epidote, with albite-included garnets typically reaching higher X_{Alm} and X_{Grs} at their rims than those in epidote. White mica is dominantly phengitic, with core–rim zoning showing more muscovite/paragonite content in the cores and rims with more celadonic compositions (Fig. 4b).

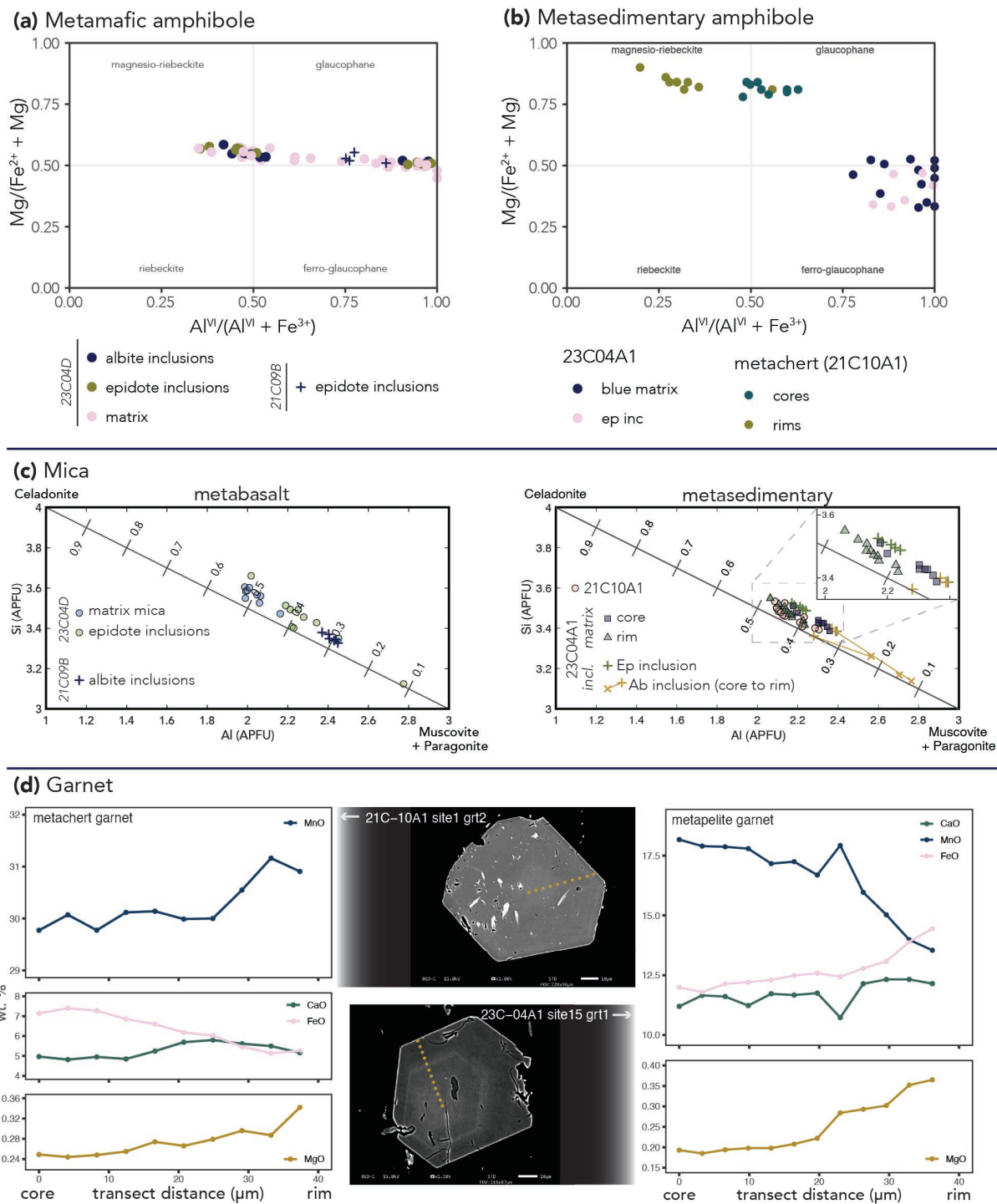


Figure 4. Plots of EPMA mineral analyses. (a, b) Na-amphibole compositions from metabasaltic and metasedimentary samples, respectively. (c) white mica analyses; (d) representative transects from garnets in both the metachert and the metapelite. Garnets in the metachert occur in the matrix of that rock, while garnets appear only as inclusion in epidote and albite in the metapelite sample.

4.2 Phase equilibrium modeling results

A major factor affecting the equilibrium phase assemblage of these rocks is the ratio of $\text{Fe}^{3+}/\text{Fe}^{\text{Total}}$, which we measure directly within the rock for some samples via Fe titration. However, but this ratio represents only the current value in each sample, and will vary depending on oxygen fugacity's evolution over the course of prograde and retrograde metamorphism. Thus, the modern ratio is not necessarily the ratio of the effective bulk composition during metamorphism. Varying $\text{Fe}^{3+}/\text{Fe}^{\text{Total}}$ in both models has a significant impact on the topology of the pseudosection and affects the degree of correspondence between the P - T conditions predicted by the phase equilibrium models.

Rather than including Fe_2O_3 as a component, we vary $\text{Fe}^{3+}/\text{Fe}^{\text{Total}}$ in the models by adding the amount of O_2 necessary to oxidize enough Fe to achieve the target $\text{Fe}^{3+}/\text{Fe}^{\text{Total}}$. This allows us to extract the chemical potential O_2 (μO_2) at a set of relevant P - T conditions and then impose this potential in phase equilibrium models of other bulk compositions. We can therefore approximate adjacent rocks as having experienced the same oxygen fugacity at certain points in their metamorphic evolution. A basic assumption of the modeling presented here is that thermodynamic equilibrium with respect to oxidation is achieved among metabasaltic rocks and metasedimentary rocks, respectively, but not necessarily between the different lithologies.

4.2.1. Metabasalt models

The epidote-blueschist and actinolite-rich metabasalt samples differ in the measured bulk elemental compositions (Table 1). The actinolite-rich metabasalt exhibits higher $\text{MgO}/\text{FeO}^{\text{T}}$ than the glaucophane-rich sample (0.42 vs 0.31), and the latter has more K_2O , FeO^{T} , and Na_2O than the former. Further, the actinolite-rich sample has several wt. % higher CaO concentrations than the glaucophane-rich rock. These variations help account for the differences in the observed mineral assemblages, but the topology of the pseudosections we construct with *Perple_X* is also sensitive to variations in $\text{Fe}^{3+}/\text{Fe}^{\text{T}}$, so we explore metabasalt models with a range of $\text{Fe}^{3+}/\text{Fe}^{\text{T}}$ of 0.2–0.3. This range is higher than the reported average value of 0.16 for MORB (Cottrell and Kelley, 2011), but accounts for likely oxidation of these rocks during seafloor alteration. Although we have a measured bulk rock $\text{Fe}^{3+}/\text{Fe}^{\text{T}}=0.45$ for the blueschist metabasalt, this value is rather high and may be influenced by later weathering of the rock.

The preferred metabasalt models we present here use $\text{Fe}^{3+}/\text{Fe}^{\text{T}} = 0.3$ for the blueschist metabasalt because pseudosections using $\text{Fe}^{3+}/\text{Fe}^{\text{T}} = 0.2$ do not include a mineral assemblage field corresponding to that observed in the rock ($\text{Gln} + \text{Ph} + \text{Ep} + \text{Ab} + \text{Qz} + \text{Chl} + \text{Ttn} + \text{Bt}$). We use this pseudosection to determine the P - T conditions represented by the matrix assemblage, and the conditions near which the inclusions in albite and epidote porphyroblasts might have formed. At the lowest model temperatures of 200–250°C, the inclusion assemblage lawsonite + albite + Na-amphibole is stable between 0.38 and 0.62 GPa (Fig. 5). This is at lower temperatures, but adjacent to the fields in which epidote and white mica are stable. The matrix equilibrium mineral assemblage exists from 325 °C to 475°C and from 0.49 to 0.82 GPa (Fig. 4). We then extract μO_2 from the model at 450 °C and 0.74 GPa, which is a point within this field. This μO_2 of -405340 J/mol is then imposed on the actinolite-rich metabasalt.

The predicted equilibrium mineral assemblage that matches that of the actinolite-rich

rocks extends from 290 °C to 490 °C, and from the lower pressure limit of our model, 0.1 GPa, to 0.82 GPa. This overlaps with the equilibrium mineral assemblage field of the blueschist metabasalt around 420–450°C and 0.6–0.73 GPa.

4.2.2. Metapelite model

We model the greener, actinolite- and chlorite-rich layers of the metapelite sample separately from the bluer, quartz- and glaucophane-rich layer, on the assumption that the effective bulk composition is distinct for these layers (Table 1). Similar to the $\text{Fe}^{3+}/\text{Fe}^{\text{Total}}$ sensitivity of the metabasalt models, however, models of the actinolite-rich layers of the metasedimentary rock fail to stabilize actinolite at all without low $\text{Fe}^{3+}/\text{Fe}^{\text{Total}}$ (<0.15), so the matrix assemblages of these rocks appear to have developed under different $f\text{O}_2$ conditions than the metabasalt. We therefore estimate the $\text{Fe}^{3+}/\text{Fe}^{\text{Total}}$ of the actinolite-rich layer from our mineral assemblage-based bulk rock calculations to be 0.1. Similar to the process applied for modeling the metabasalts we interrogate the set of phase equilibrium models for the actinolite- and glaucophane-dominated layers in which we first model the actinolite-dominated layer and extract μO_2 at the same 450 °C and 0.74 GPa as we did in the metabasalt models. We impose this $\mu\text{O}_2 = -433730 \text{ J/mol}$ on a model of the glaucophane-dominated layer. From these models, we determine the P - T conditions that correspond to the stability of the matrix mineral assemblage, and extract garnet compositional isopleths to compare with our measured garnet compositions.

In both actinolite- and glaucophane-rich layer models, garnet isopleths converge near 470–520°C, and over a wide range of pressures (Fig. 5), but these do not overlap between the models. The blueschist layer garnet compositions in particular do not provide well-constrained pressure estimates and extend up to the maximum pressure limit of our models. Garnet compositions from greenschist layer on the other hand only converge at 0.4–0.7 GPa.

Similar to the greenschist metabasalt, the field corresponding to the matrix assemblage of the actinolite-rich metasedimentary layer extends over a broad 280–485°C and from the lower pressure limit of our models to 0.82 GPa. While imposing the μO_2 estimated from the actinolite-rich layer on the glaucophane-dominated layers produces stable phase assemblages that match those observed in the rocks, this field does not overlap in P - T space with the matrix assemblage of the actinolite-rich layer, but lies at ~ 0.1 GPa higher pressures than the actinolite-rich assemblage (Fig. 5). The blueschist assemblage field also extends up to 1.35 GPa and from 320°C to >500 °C. The lack of overlap between the actinolite-rich and glaucophane-rich layers are due to the positions of the upper-pressure limit of actinolite and the lower-pressure limit of glaucophane, respectively, in each pseudosection.

4.2.3. Metachert model

Because of the spatial proximity of the metachert horizon and the metasedimentary sample location, we also apply the chemical potential of O_2 estimated from the actinolite-rich metasedimentary layer to our phase equilibrium model of the metachert. This model predicts dominantly spessartine garnet over the whole range of P - T conditions over which garnet is stable, but the compositional isopleths of garnet that correspond to those measured in our sample fail to converge.

The phase equilibrium model does, however, contain a field that corresponds to the matrix phase assemblage of the more pelitic layer of the metachert (Fig. 5). Similar to that predicted for the glaucophane-rich layer of the metasedimentary sample, this field extends to high pressures from 0.75 GPa, and over a wide temperature range from 335°C to >500°C. Here again, the lower pressure bound is controlled by the limit of Na-amphibole stability.

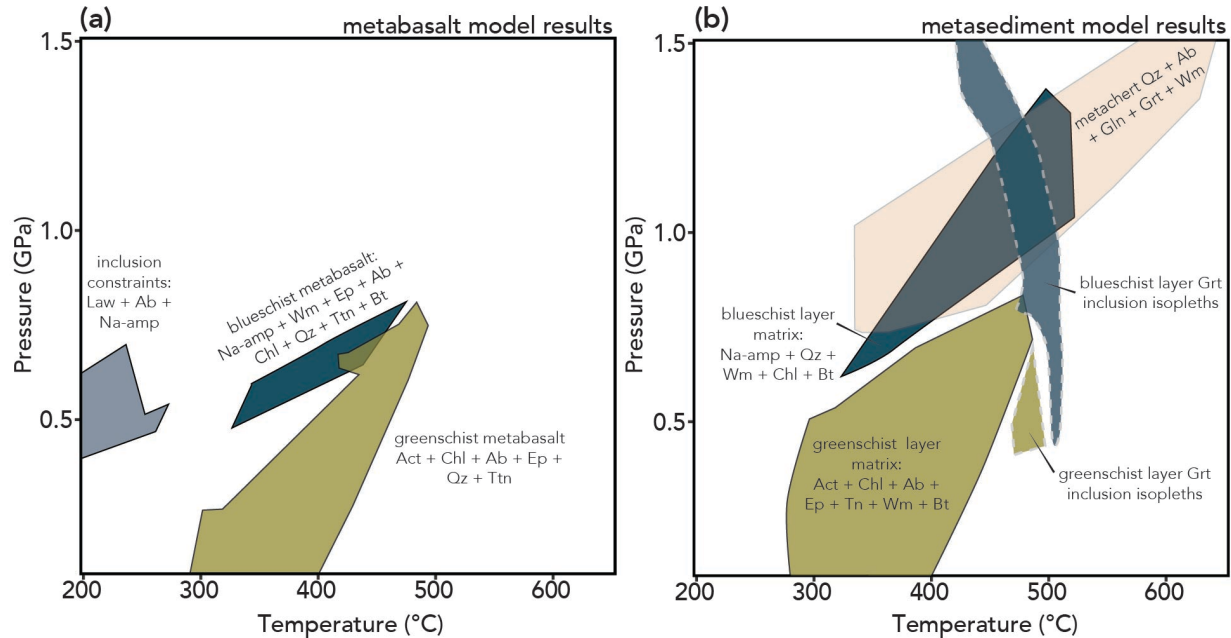


Figure 5. Summary of *P-T* conditions estimated by phase equilibrium for (a) metabasaltic and (b) metasedimentary samples. For metabasalts, inclusion assemblages provide an indication of metamorphic conditions experienced early during subduction. For the metasedimentary rocks, dashed-outlined fields show the regions in which modeled garnet compositional isopleths match the range measured in the respective layers of the metapelite. These are generally at hotter conditions than the conditions corresponding the equilibrium mineral assemblage in the matrix of the metapelite. Greenschist and blueschist layers of the metapelite do not overlap, but are within pressure uncertainties of ~0.1 GPa for phase equilibrium modeling.

4.3 EBSD analyses

We present EBSD maps and data on representative portions of the matrix of both the blueschist and greenschist metabasalts to investigate the deformation mechanisms of amphiboles, as well as the isolated quartz-filled necks between boudinaged albite porphyroblasts. We also present EBSD data on quartz from the metachert sample, which we interpret using the quartz recrystallized grain size paleopiezometer using the approach of Cross et al. (2017).

4.3.1. Amphibole in the metabasalts

EBSD maps of the amphibole-rich matrix from both blueschist and retrogressed greenschist samples show little evidence for intragranular deformation of amphibole. Strong CPOs are characterized by girdles in [001] and [010] axes in the foliation plane, and [100]

alignment perpendicular to foliation (Fig 6c,g). We suspect that the misalignment between the diffuse point maximum in the [001] axes and the edge of the thin section (X direction) is a result of the thin section cut being slightly oblique to the true lineation of the sample, which is faint and difficult to observe in hand sample.

Misorientation-to-mean orientation maps (Mis2Mean) show the misorientation of a point in a grain relative to the mean orientation of that entire grain (Fig. 6). In both glaucophane and actinolite, these show relatively low degrees of misorientation and show little evidence for subgrains, although glaucophane shows slightly higher degrees of misorientation relative to actinolite.

Histograms of random- and neighbor-grain pair misorientations show the misorientation of individual grains with respect to each other and with respect to a theoretical random distribution. Neither histogram follows the random distribution, and neighboring grain pair histograms show higher frequencies of low-angle misorientations (Fig. 6).

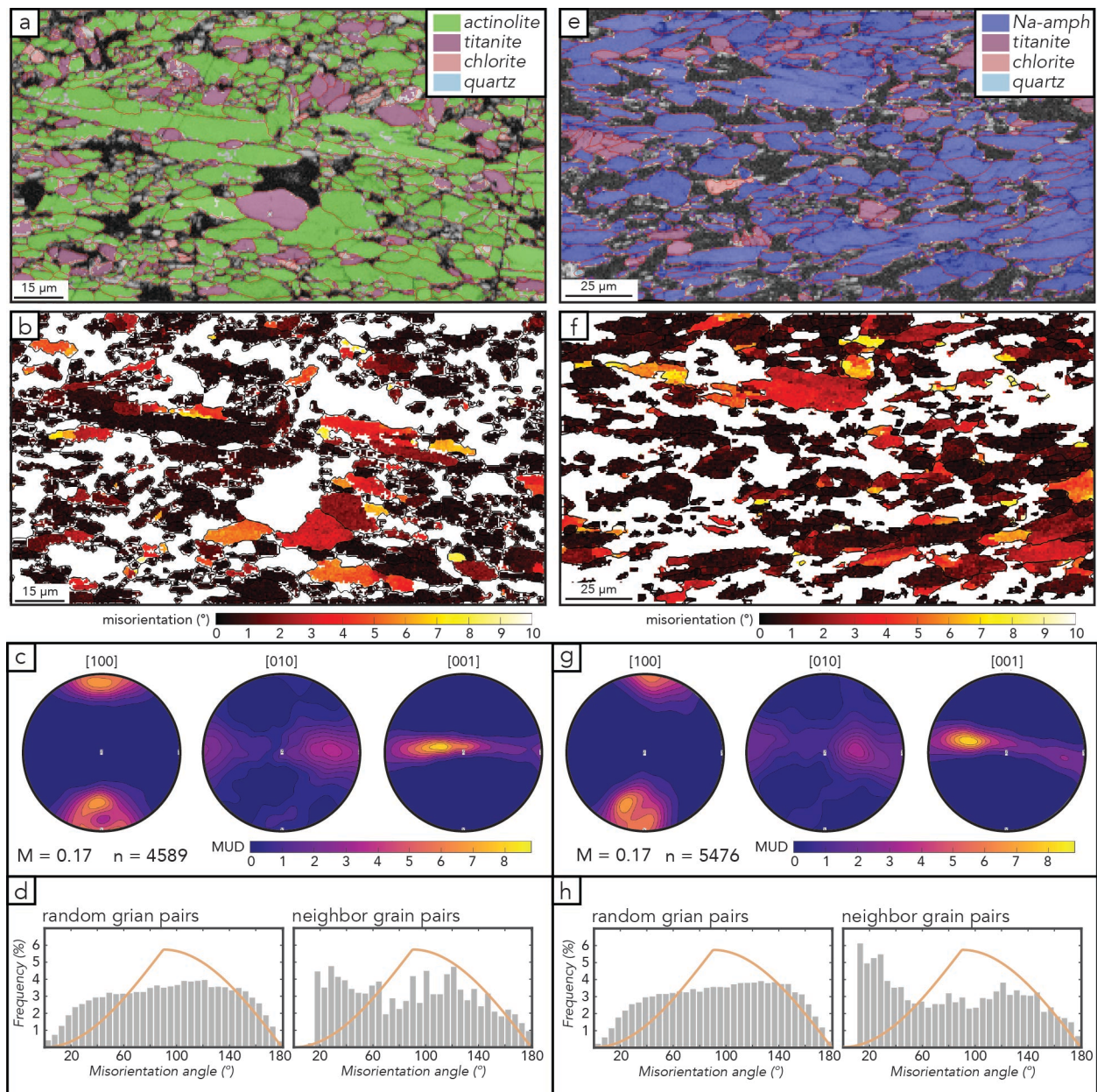


Figure 6. Representative EBSD data from amphibole in the epidote-actinolite greenschist metabasalt (left column, a–d), and the epidote-blueschist metabasalt (right column, e–f). Pole figures are upper-hemisphere equal-angle projections.

4.3.2. Quartz

Optical images showing undulose extinction and subgrain textures, and EBSD maps of quartz in both the greenschist metabasalt sample where it fills necks between boudinaged albite, and in the quartz-dominated metachert show evidence for significant intragranular deformation (Fig. 2, 6). We apply the EBSD-based calibration of the recrystallized grain size piezometer developed by Cross et al. (2017) to the quartz in both samples. Following the method of Cross et al. (2017), we process EBSD maps of quartz-rich zones to measure the grain size of recrystallized

quartz crystals using an automatic thresholding method based on the distribution of grain orientation spread (GOS) values for the analyzed grains. Results of these analyses are summarized in Table 4.

4.3.2.1 Quartz in the metabasalt

In the metabasalt sample, quartz occurs in and around clusters of albite and epidote porphyroblasts and in strain shadows thereon, in fractures within those porphyroblasts, and in boudin necks of pulled-apart albite porphyroblasts. In all settings, it exhibits undulose extinction and subgrain textures, and warping of the surrounding actinolite and chlorite foliation into boudin necks (Fig. 7a) suggests deformation of quartz by crystal plastic deformation mechanisms contemporaneous with the foliation-forming deformation. Intragranular deformation is demonstrated by Mis2Mean plots, which show very high degrees of misorientation within some quartz crystals, and the CPO of quartz is moderately strong ($M = 0.1$), especially in the alignment of the c-axis (Fig. 7d). We measure recrystallized quartz grain size in the boudin necks between pulled-apart albite porphyroblasts and find recrystallized grain sizes of 6.0–35.6 μm , which indicate stresses of 47.1–166 MPa via the recrystallized grain size piezometer of Cross et al. (2017).

4.3.2.2. Quartz in the metachert

In the metachert, quartz is ubiquitous and shows similar microtextures throughout the sample. We analyze both a representative section of the quartz-dominated, mica-free matrix and a section of the more mica-rich layer, as well as an area within a C' shear band in the mica-rich layer. Quartz in mica-free domains typically exhibits large grain sizes with elongate ribbons (Fig. 6b), while quartz grain sizes are smaller in the more mica-rich domains, likely as a result of grain boundary pinning by micas. In both domains, undulose extinction and subgrain textures are common. The optical evidence for intragranular deformation is further evidenced by Mis2Mean maps which show large degrees of intragranular misorientations in quartz crystals (Fig. 7). The CPO of quartz in the metachert is moderately strong ($M=0.2$), particularly in c-axis alignment, as in the quartz in the metabasalt (Fig. 7i).

Despite apparent overall grain size differences between the mica-free and mica-rich domains, quartz recrystallized grain sizes are actually quite comparable (~12–32 μm). As expected, however, recrystallized grain sizes within the C' shear band are significantly smaller (~1.5–4.5 μm) than in the surrounding matrix.

Table 4. Summary of quartz paleopiezometry results

sample	lithology	raw indexing rate	no. relict grains	no. rex grains	RMS rex grain size (μm)	stress (Mpa)	M index	J index
21C09B1 site 1	metabasalt	82%	38	441	22.2 ± 13.4	47.1–127	0.101	3.3873
21C09B1 site 2	metabasalt	90%	52	410	16.3 ± 10.3	57.9–166	0.1082	3.316
21C10A1 site 4	metachert	89%	58	351	24.3 ± 13.7	44.9–111	0.1951	3.5174
21C10A1 site 3	metachert	85%	59	263	20.4 ± 10.1	52.5–113	0.106	3.0018

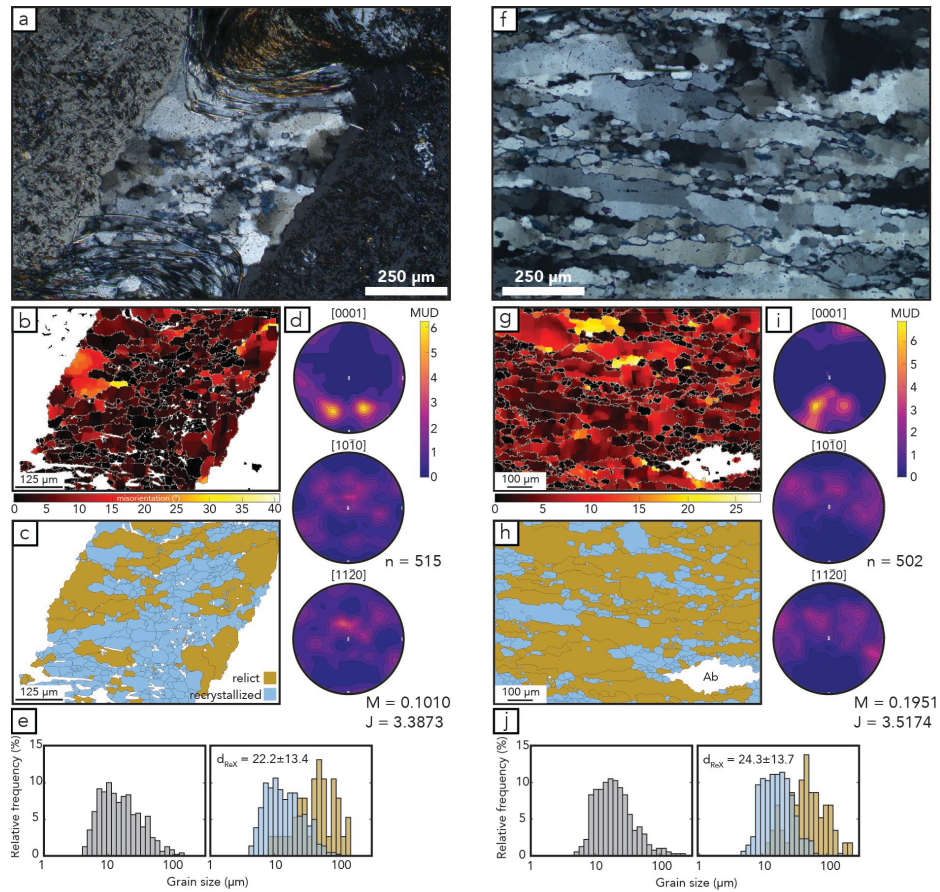


Figure 7. XPL photomicrographs (a, f) and representative EBSD data from quartz in the epidote-

actinolite greenschist metabasalt (left column, a–e), and the metachert (right column, f–j). Pole figures are upper-hemisphere equal-angle projections.

5 Discussion

We combine our phase equilibrium models with textural and chemical observations to reconstruct the P - T path of these rocks. Next, we link structural and textural observations to this metamorphic framework to understand the deformation mechanisms over the prograde to underplating subduction processes and find that the fabrics in the Epidote Blueschist Unit predominantly reflect underplating deformation at higher stresses than were likely present on the plate interface, but also preserve evidence for early subduction-related deformation.

5.1. Metamorphism of the Epidote Blueschist Unit

5.1.2. P - T path of the Epidote Blueschist Unit

Based on the estimated P - T conditions represented by mineral assemblages and garnet compositions in our phase equilibrium models, we present a reconstruction of the metamorphic history of the Epidote Blueschist Unit (Fig 5). This reconstruction is limited by uncertainties in the phase equilibrium modeling approach, which arise from uncertainties in the effective bulk compositions (e.g., Guevara and Caddick, 2016) and oxidation state used in our models, as well as in the thermodynamic data for the end-member phases and activity-composition models that go into the thermodynamic calculations. The latter uncertainties are commonly estimated to be $\pm 50^\circ\text{C}$ and ± 0.1 GPa at the 2σ level (e.g., Powell and Holland, 2008; Hodges and McKenna, 1987). Combined with our petrologic observations and geochemical analyses, however, we can still reconstruct a first order understanding of the metamorphic history of these rocks.

The first constraint on the P - T evolution of the Epidote Blueschist Unit is derived from the minerals that are included within albite and epidote porphyroblasts in the metabasaltic rocks: lawsonite + white mica + Na-amphibole + titanite (Fig 5). The type and composition of these inclusions are consistent with the growth of these phases early in the prograde metamorphic history of these rocks. Lawsonite is stable at low temperatures, but over a wide pressure range, but the co-occurrence of all these phases restricts the range of possible pressures that these rocks passed through. At 200–250°C, the low-temperature limits of our phase equilibrium models, Law + Ab + Na-amp co-exist and about the stability fields of white mica and epidote at pressures of 0.4–0.7 GPa (Fig. 5). We infer that the growth of porphyroblasts and the inclusion of these minerals likely lasted to higher temperatures because of the zoning patterns preserved in some mica and amphibole inclusions. The latter, for example, shows evidence for magnesioriebeckite cores and glaucophane rims, which suggests continued equilibration with the matrix during subduction rather than immediate inclusion and chemical isolation from the rest of the rock.

On the hotter end of the metamorphic history, the regions over which the isopleths of garnet inclusions in the metasedimentary rock converge represent the highest temperature estimates for this unit. At 470–520°C, these temperatures lie at the hotter end of, but overlap within uncertainty of, those measured via Raman spectroscopy of carbonaceous materials for this unit, which yielded up to $470 \pm 26^\circ\text{C}$ (Platt and Schmidt, 2024). The pressures represented by the garnets in the glaucophane-rich layer are relatively unconstrained, but considering that they

overlap in pressure with the actinolite-rich layer, we suggest that the garnets from the actinolite-rich layer constrain the pressures to 0.4–0.7 GPa (Fig. 4). The patterns of zoning in these garnet show decreasing spessartine and increasing grossular and almandine fractions from core to rim. The pattern of garnet isopleths estimated by our phase equilibrium models, combined with the inclusion of these garnets in epidote and albite porphyroblasts, suggest that these garnets preserve some snapshot of prograde or peak garnet growth.

For both metabasaltic and metasedimentary rocks, the regions where our models contain an equilibrium mineral assemblage that matches the observed matrix mineral assemblage exist at similar but not entirely overlapping P - T conditions. In the metasedimentary models, the actinolite- and glaucophane-rich layers do not overlap in pressures, but do lie within estimated uncertainty of 0.1 GPa of one another at $T < 480^{\circ}\text{C}$. Between all five models, the matrix mineral assemblage fields overlap in pressure between 0.6–0.8 GPa. We take the 420–450°C and 0.6–0.73 GPa region where blueschist and greenschist metabasalts overlap in figure four of as the best constraint on P - T conditions represented by the matrix mineral assemblages of these rocks. This overlaps with the stability field of the matrix assemblage for the greenschist metasedimentary layer, and is within uncertainty of those of the metachert and blueschist metasedimentary lithologies as well (Fig. 5).

We argue that the matrix of all the rocks discussed here exhibit some degree of retrograde overprint. This follows from the observations of the lack of garnets in the matrix of the either the glaucophane- or actinolite-rich metapelitic layers, the overprinting of likely prograde compositional zoning in amphibole in the blueschist metabasalt, and the fact that the P - T conditions predicted for the matrix mineral assemblages in both the metabasalt and metasedimentary rocks lie at lower temperatures than those predicted by the garnet isopleths, and maximum temperature estimates from Platt and Schmidt, (2024). Although garnet stability technically overlaps with the predicted stability fields of the matrix phase assemblages in our pseudosections, the predicted modal abundance is so low (<1 vol. %), that we do not consider these boundaries for constraining the matrix stability fields. The core–rim amphibole zoning of magnesioriebeckite to glaucophane compositions that we observe in amphibole inclusions is consistent with prograde compositional changes in amphibole. Where this zoning pattern is preserved in the matrix, it is typically truncated by another rim of magnesioriebeckite that we suggest grew during retrograde metamorphism at cooler temperatures.

We also suggest that the metachert is affected by retrograde metamorphism. Our phase equilibrium models predict that spessartine fraction decreases with increasing temperature, and we expect Mn to be sequestered into garnet early in garnet growth. We therefore suggest that garnets in the metachert, which occur throughout the matrix and have a trend of moderately increasing spessartine fraction towards the rim, may record some retrograde overprint. This is consistent with either equilibration of the garnet during retrogression, or with resorption of garnet and the enrichment of Mn in garnet rims. This retrogression interpretation is supported by the observation of magnesioriebeckite rims on glaucophane throughout the matrix as well. These retrograde trends in garnet composition from core to rim stand in contrast to those trends observed in the more pelitic metasedimentary rocks. In summary, our phase equilibrium modeling and petrographic observations suggest that the Epidote Blueschist Unit experienced: (1) early crystallization of lawsonite-blueschist facies minerals at 200–300°C and 0.4–0.7 GPa, followed by near isobaric heating to; (2) peak metamorphic conditions of 470–520°C and 0.4–

0.7 GPa, followed by cooling to; (3) retrograde metamorphic conditions at 420–450°C and 0.6–0.73 (Fig. 5).

The *P-T* path we reconstruct here is consistent with a shallow subduction angle under relatively thin forearc crust (Fig. 8). The Epidote Blueschist Unit *P-T* path roughly follows that reconstructed for the higher-grade, structurally higher and slightly older Epidote Amphibolite Unit, which showed near isobaric heating from ~300°C to >500°C (Lindquist et al., 2026). Both *P-T* paths have peak pressure conditions below 1 GPa during prograde and early retrograde metamorphism. The path we estimate for the Epidote Blueschist Unit is similarly consistent with shallow or flat-slab subduction of the Farallon plate beneath North America by ca. 100 Ma. As suggested by Lindquist et al. (2026), this slab shallowing can help explain the apparent cooling of the subduction zone that is required to produce the apparent inverted metamorphic field gradient observed in the Catalina Schist. The possible retrogression represented by the matrix mineral assemblages in these rocks is consistent with continued cooling following underplating.

Our estimated temperature conditions overlap with the calculated RSCM estimates of 390–470°C, but our pressure estimates conflict with those previously estimated for the Epidote Blueschist Unit (Platt and Schmidt, 2024). Previous pressure estimates for the Epidote Blueschist Unit rocks of 0.7–1.3 GPa have been based on approximate stability limits for Na-amphibole as a lower bound and the albite = jadeite + quartz reaction line as an upper bound (e.g., Platt and Schmidt, 2024). We argue that our phase equilibrium modeling provides better pressure constraints for the Epidote Blueschist Unit because it uses compositions that more accurately reflect the composition of the rocks. The albite = jadeite + quartz reaction, for example, is based in the three-component Na₂O–Al₂O₃–SiO₂ system, and at the temperatures relevant to the peak metamorphic conditions of these rocks this does not appear to be a relevant reaction. Our *P-T* pseudosections instead suggest that the albite-out reaction in the blueschist metabasalt, for example, instead involves the reaction of amphibole, biotite, and albite to produce more glaucophanic amphibole and phengite. Platt and Schmidt (2024) also commented that their estimated *P-T* conditions for the Epidote Blueschist Unit lie within the experimentally-determined stability field of lawsonite and that the occurrence of epidote-group minerals rather than lawsonite is a result of low water or high oxygen fugacities. We suggest instead that the lack of lawsonite is more fundamentally a result of differences in the bulk composition of these metabasalts and metasedimentary rocks relative to the compositions relevant to the referenced experiments, which were conducted on synthesized samples to explore the simplified CaO–Al₂O₃–SiO₂–H₂O system (Liou, 1971).

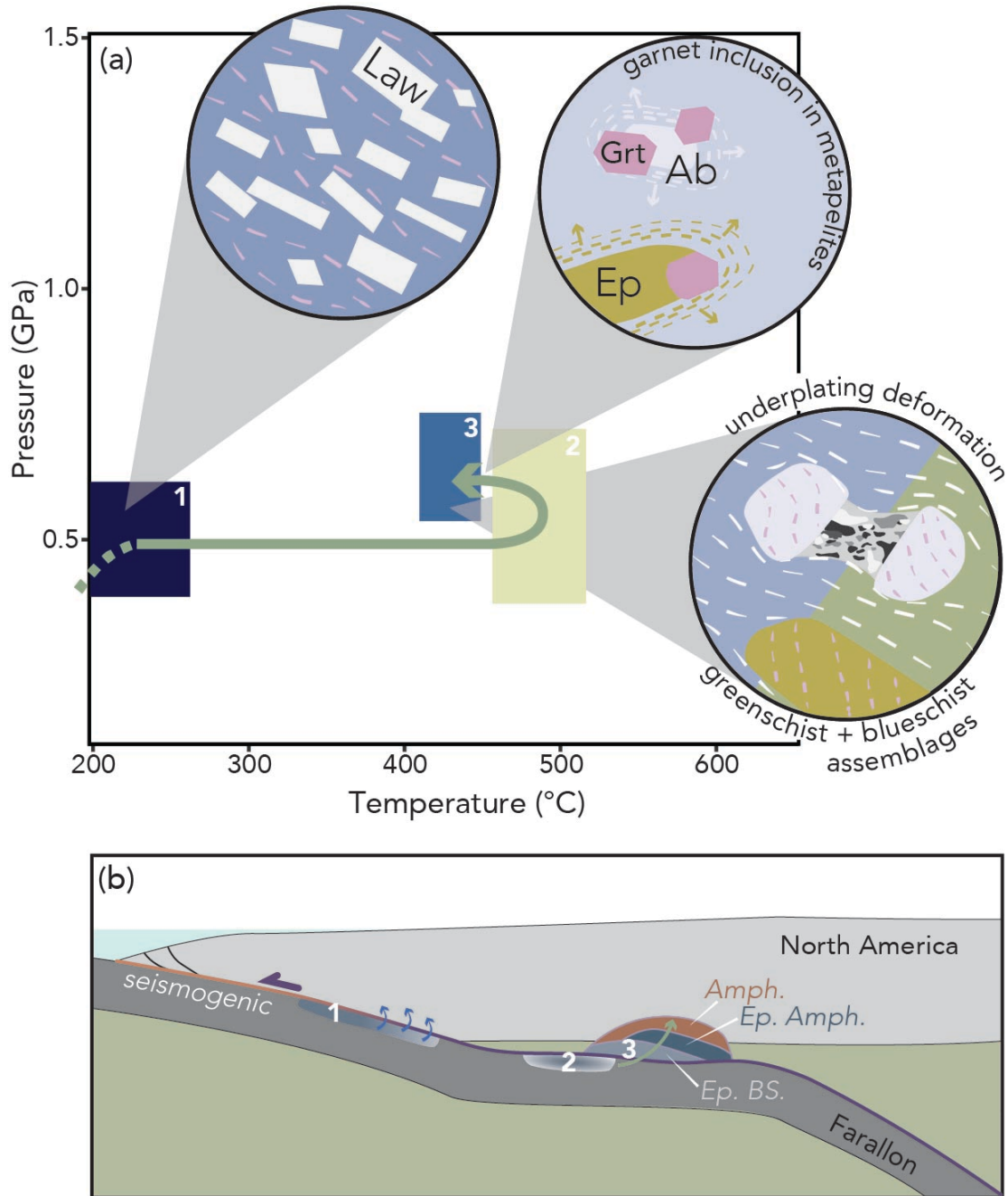


Figure 8. Summary P - T diagram and tectonic cartoon. (a) shows highlights of the metamorphic and structural evolution of the Epidote Blueschist Unit from (1), early fabric development at the lowest T temperature constraints, (2) peak P - T constraints from garnets in the metapelite, which become included in epidote and albite during retrogression, and (3) early retrograde metamorphic conditions associated with underplating-related deformation affecting all lithologies. (b) shows these metamorphic stages of the Epidote Blueschist Unit interpreted in tectonic context.

5.1.2. Bulk composition versus oxygen fugacity controls on mineral assemblage

Platt (1976) suggested that the fine-scale layering of blueschist and greenschist mineral assemblages in the Epidote Blueschist Unit (e.g., Fig. 2c) is a result of variations in the Fe^{3+} content of the layers as a result of variable oxygen fugacity. However, our phase equilibrium modeling has demonstrated that simple variations in the bulk composition of the lithologies in this unit at the same $f\text{O}_2$ conditions can account for the observed variations in mineral assemblage. We note that in our phase equilibrium modeling, more “oxidized” compositions tend to produce more Na-amphibole and more “reduced” compositions tend to produce more actinolite, but it is not necessary to invoke cm-scale variation in $f\text{O}_2$ to explain the observed layering. Various other workers have documented similar blueschist–greenschist layering and also attributed the alternation to variability in the composition of the layers (e.g., Dungan et al., 1983; Muñoz-Montecinos et al., 2020). The fine scale of blueschist–greenschist alteration in places makes it unlikely that significant differences in oxidation state persisted during metamorphism.

In the case of the metabasalts, we suggest that the compositional differences result from initial compositional heterogeneity between pillow cores, rims, and interpillow hyaloclastite that have now been deformed and sheared into distinct compositional layers (e.g., Muñoz-Montecinos et al., 2020). For metasedimentary rocks, simple variation in the composition of the deposited sediments may explain the layering. These initial pre-subduction variations may have been further accentuated by metasomatism and the preferential flow of fluids in more permeable layers or networks that resulted in localized metasomatism. We note that our models do suggest differences in the oxidation state of metabasalts and metasedimentary rocks, which is consistent with the distinct origins and pre-subduction histories of the two lithologies. While the basalts formed at a mid-ocean ridge and would have likely experienced hydrothermal alteration associated with that setting, the metasedimentary protoliths would have been deposited much later and nearer the trench and terrigenous sediment and volcanic sources.

One significant though unsurprising point that arises from this is the importance of initial compositional heterogeneity in producing cm- to m-scale variations in mineral assemblage and therefore rheology within nominally the same rock type. In the case of the Epidote Blueschist Unit the rheology of these layers seems to have been comparable (isoviscous) due to the persistence of layering and lack of block-in-matrix structures of blueschist and greenschist lithologies. However, the same mineralogical heterogeneity at different P - T conditions, or different compositional and mineralogical pairings could introduce significant rheological contrasts that lead to strain partitioning.

5.2 Deformation of the Epidote Blueschist Unit

We use the microstructural observations and measurements on amphibole and quartz presented in Section 4.3 to determine the micromechanisms of deformation that were active in the Epidote Blueschist Unit. In the subsequent section (5.3), we interpret these mechanisms in the context of the metamorphic history established in Section 5.1.1.

The matrix fabrics that we observe in the metabasalts suggest that amphibole (both glaucophane/magnesioriebeckite and actinolite) deformed via dissolution-precipitation creep and rigid-body rotation, with possible minor dislocation creep. The CPO and neighbor-grain pair misorientations are suggestive of rigid rotation and dissolution-precipitation creep as the active

deformation mechanisms for amphibole during foliation formation. The low degree of misorientations and apparent lack of subgrains suggests a lack of significant crystal plastic deformation via dislocation creep (Fig. 6). The peak at low angles in the neighbor-grain pair distribution (Fig. 6) of amphibole may suggest the activity of dislocation creep, but this may also be explained by rigid-body rotation. Indeed, some analyzed areas of matrix amphibole deformation in both metabasalts also show peaks in neighboring grain pair misorientations near 180°, which is more suggestive of rigid rotation of grains (Fig. S8). The greater evidence of intragranular misorientations and the larger peak of low-angle neighbor-grain pair misorientations for glaucophane relative to actinolite is consistent with the easier activation of dislocation glide in glaucophane relative to Ca-amphiboles (Reynard et al., 1989).

Nonetheless, the zoning patterns of amphiboles and apparent microboudinage textures, provide support for the contribution of dissolution-precipitation to the development of amphibole CPO and SPO, and subsequent boudinage during periods of elevated chemical potentials. Both Na-amphibole and actinolite show evidence for some microboudinage. Na-amphiboles show the most apparent evidence for microboudinage because of the obvious compositional differences that are apparent in BSE and elemental maps, while actinolite in the greenschist metabasalt only shows incipient fracture patterns in some BSE images (Fig. S7e). Diffusion limited microboudinage creep of amphiboles has been documented in natural and experimentally deformed rocks (e.g., Misch, 1969; Kimura et al., 2010; Tokle et al., 2023).

Quartz in both the metabasalts and the metasedimentary rocks appears to deform via dislocation creep involving subgrain rotation recrystallization. The evidence for undulose extinction, subgrain structures, a strong CPO, and high degrees of intragranular misorientations (Fig. 6) all point to the activity of dislocation creep. The results of our quartz recrystallized grain size piezometry further indicate that this deformation occurred at differential stresses of 50–180 MPa, with similar estimates for both quartz in metabasalt and in metacherts.

5.3 Synthesizing metamorphic and deformation histories

Combining our petrographic and microstructural analyses, we can place the microstructures we observe and their implications for deformation mechanisms within the context of our reconstructed *P-T* path combined with the structures these mechanisms accommodated (Fig. 8). These shed light on the deformation mechanisms and stress conditions in rocks that would have been found near the megathrust plate interface over the lifetime of these rocks.

First, the occurrence of lawsonite-containing inclusion trails defining relict and folded fabrics in porphyroblasts of the metabasalts suggests that these rocks were deforming and forming fabrics early in their subduction history (<300°C, Fig. 8). This indicates deformation of metabasalts at seismogenic zone conditions, although the small size of inclusions and isolation from each other does not allow us to infer what the mechanisms of deformation were, nor what stresses these rocks experienced. Not only are there relict foliations, but the apparent folding of some inclusion trails (Fig. 3c) indicates that progressive deformation occurred in the subducting slab. This all indicates that the metabasaltic rocks of the oceanic crust must have been relatively weak in order to accommodate deformation and develop fabrics likely defined by Na-amphibole and white mica, which were ultimately preserved as inclusion trails in epidote and albite.

The dominant deformation fabrics in metabasalts of the Epidote Blueschist Unit are the foliations defined by Na-amphibole and phengite in the epidote-blueschists and by actinolite and chlorite in the epidote-greenschists. The association of the matrix mineral assemblages in these rocks with likely early retrograde metamorphism therefore suggests that these fabrics record deformation during underplating rather than prograde- or peak- subduction metamorphism. A retrograde overprint is further supported by the truncation of earlier prograde magnesioriebeckite-glaucophane zoning in the matrix of the epidote blueschist and the predominance of secondary magnesioriebeckite rims (Fig. S7f). The deformation of both Na- and Ca amphiboles in these rocks via rigid-body rotation, dissolution-precipitation creep, and potentially minor dislocation creep therefore reflects amphibole deformation at the conditions and stresses associated with underplating of the Epidote Blueschist Unit.

Quartz deformation via dislocation creep recorded in the boudin necks between albite in the epidote-greenschist metabasalt similarly reflects deformation during underplating and retrogression. The warping of the amphibole and chlorite foliation and folding of these minerals into the boudin necks is consistent with boudinage and quartz precipitation contemporaneous with foliation development. We therefore infer that the stresses recorded by the recrystallized quartz (51–180 MPa) are the same as those that formed the surrounding amphibole and chlorite/mica-dominated foliation, and that this deformation occurred during the underplating of these rocks. Quartz-filled fractures and boudin necks also suggest the presence of a fluid to facilitate the diffusion and precipitation of quartz into these fractures, which is consistent with a fluid-rich environment at the plate interface and within underplated slices of the oceanic crust, and a necessary condition for the dissolution-precipitation creep mechanism in amphiboles.

Quartz microstructures in the metachert are similar to those in the quartz-rich domains of the metabasalts and also appear to record subgrain rotation recrystallization. The fact that both metachert and metabasalts record top-to-the-SW kinematics and our stress estimates from recrystallized quartz grain sizes are the similar suggests that the deformation fabrics in these rocks record the same stage of deformation. Furthermore, the pattern of magnesioriebeckite rims on glaucophane in the metachert is consistent with the deformation occurring during early exhumation and retrograde metamorphism. The occurrence of magnesioriebeckite as asymmetric tails on glaucophane fish and as growths in the fractures of microboudinaged glaucophane suggest that the magnesioriebeckite growth was contemporaneous with the matrix deformation.

In the context of our P - T path and the underplating deformation represented by the fabrics in the Epidote Blueschist Unit, our recrystallized quartz grain size paleopiezometry results indicate the stresses experienced by these rocks during underplating. With calculated stresses of >50–100 MPa, this is consistent with deformation not at the subduction plate interface, which is expected to have low differential stresses of <10 MPa (e.g., Wassmann and Stöckhert, 2011). Behr and Platt (2013) similarly showed stresses may be low during subduction, but may increase during exhumation in a subduction channel to over ~100 MPa. Multiple workers have suggested that pressure solution is likely the dominant viscous deformation mechanism in subduction zones and that quartz dislocation creep is unlikely to play a significant role accommodating prograde subduction-related deformation (e.g., Wassmann and Stöckhert, 2011; Condit et al., 2022; Fisher and Hirth, 2024, Condit et al., 2025). This is all consistent with the deformation that is recorded by the Epidote Blueschist Unit fabrics and reflecting stresses >50 MPa reflecting deformation during or after underplating. Schmidt and Platt (2022) have also

documented a switch from pressure-solution dominated deformation to quartz dislocation creep deformation associated with the initiation of underplating and exhumation. Furthermore, the relatively rare occurrence of asymmetric fabrics in the Epidote Blueschist Unit are consistent with dominantly coaxial deformation. Exhumation of the Cycladic Blueschist Unit on Syros similarly records coaxial deformation fabrics during underplating (Kotowski et al., 2022).

While deep tremor and slow slip are typically assumed to occur on the subduction plate interface, and indeed seismological studies often relocate tremor to near the plate interface, recent work from Cascadia has suggested a link between tremor and underplating (Littel et al., 2025). Sammis and Bostock (2025) have also suggested that some slow slip in the same part of Cascadia occurs in a zone above the plate interface. The underplating story that is preserved in rocks like the Epidote Blueschist Unit is therefore not only important for being able to disentangle prograde processes from retrograde overprints, but also for having a more complete picture of subduction processes themselves.

We also note that the well-developed deformation fabrics described in the Epidote Blueschist Unit stand in contrast with the relatively undeformed appearance of the structurally higher Epidote Amphibolite Unit. The rocks from both units include rocks from what would have been the top of the oceanic crust and would have been similarly proximal to the plate interface shear zone. Although cooling to lower metamorphic temperatures and the resulting changes in mineral assemblage has been proposed as a contributor to rheologic weakening in other rocks (Kotowski et al., 2026), these two units likely followed a similar path through P - T space. The difference in their behavior is thus likely compositionally controlled. We speculate that differences in the bulk composition of the basaltic rocks in each unit—perhaps influenced by seafloor hydrothermal alteration—contribute to the observed differences in deformation. The same P - T path with a different bulk composition may have favored weaker assemblages in the Epidote Blueschist Unit that allowed it to deform during subduction. Additionally, if the stresses involved in the deformation of both units are only associated with underplating and the near-peak mineral assemblage, the Epidote Blueschist assemblage itself may be weaker than the Epidote Amphibolite.

6 Conclusions

In this study we present new constraints on the P - T evolution of the Epidote Blueschist Unit of the Catalina Schist, and interpret the deformation textures preserved in these rocks in the context of these constraints. Our application of phase equilibrium modeling has provided updated P - T estimates for the Epidote Blueschist Unit that are lower than previously inferred (e.g., Sorensen, 1986; Platt and Schmidt, 2024), but which are consistent with recent work on the Epidote Amphibolite Unit, which also exhibits evidence for shallow subduction and underplating at < 1 GPa (Lindquist et al., 2026). These results shed new light on the metamorphic history and tectonic setting of the Catalina Schist.

The deformation microstructures that we have analyzed in the Epidote Blueschist Unit also provide snapshots of the deformation of metabasaltic and metasedimentary rocks during subduction and underplating. We have shown that a weak oceanic crust may develop deformation fabrics during subduction through seismogenic conditions as indicated by inclusion trails of lawsonite-blueschist facies minerals. Evidence for dynamic recrystallization in quartz

also shows that stresses in underplated packages of oceanic crust were likely ~50–180 MPa during early exhumation and retrograde metamorphism, and dissolution-precipitation creep was likely a dominant deformation mechanism in both Na-amphiboles and actinolite in metabasalts during underplating. These results provide further documentation of the deformation mechanisms active in amphiboles and add to an existing literature documenting underplating processes and the evolution of deformation mechanisms during subduction metamorphism and underplating.

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Chapter 4 References

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Chapter 5: Conclusions

We have demonstrated several ways in which fluid-related chemical processes are promoted and modify the environment in which slow slip and tremor occur. From providing a fluid source through dehydration reactions in subducting slab to producing rheologically significant metasomatic rocks, chemical interactions between fluids and rocks exert a critical first-order control on subduction processes. Through modeling and field studies (chapter one and two, respectively), we have shown that slab components like metabasalts are an important *local* fluid source at the conditions of slow slip. Furthermore, this work demonstrates that detailed field, petrologic, and geochemical analyses are necessary to place structural and deformation interpretations in a petrologic and tectonic context in order to establish that the rock record represents the subduction processes of interest.

To explore the effects of silica metasomatism on serpentinites at the plate interface in the modern Mexican subduction zone, I performed phase equilibrium modeling of both subducting Cocos plate lithologies and mantle-wedge serpentinites. The former represent a source of Si-rich fluids and the latter a sink in which metasomatism will promote the generation of talc. This talc is rheologically significant because it is weak, easily localizes deformation, and is a candidate for hosting slow slip events. Using these phase equilibrium models, I construct a simple mass balance model to estimate how much talc might be produced along the flat slab segment of the Cocos plate beneath Mexico and suggest that solely advective transport of Si is unlikely to produce significant volumes of talc along the entire plate interface. However, regions near large dehydration pulses occurring in the subducting slab could produce large volumes of talc schists from the addition of Si to serpentinite at the plate interface.

To look for evidence of metamorphic dehydration reactions—such as those that we predict in chapter one—in the geologic record of subduction rocks, I conducted detailed petrologic and geochemical analyses of rocks from the Catalina Schist in chapters two and three. Epidote textures and compositions in the Epidote Amphibolite Unit hint at the metamorphic evolution of metabasalts during subduction, including providing evidence for a major dehydration reaction that likely occurred at the base of the seismogenic zone. This pumpellyite to epidote dehydration reaction also led to the formation of epidote-rich zones, which form mm–cm scale networks that represent limited channel-forming reactive porosity structures that may have helped to drain fluids from the subducting slab. This detailed petrologic work also provided new constraints on the P - T conditions of Epidote Amphibolite Unit metamorphism of $\sim 550^{\circ}\text{C}$ and 0.6–0.9 GPa. We are also able to resolve some of its evolution through P - T space, which is important not only for understanding the dehydration processes potentially applicable to other subduction zones, but provides new insight into the Cretaceous tectonic evolution of the Farallon–North American subduction zone, suggesting an earlier onset of shallow or flat slab subduction than previously thought.

Similar to chapter two, chapter three details the metamorphic evolution of the Epidote Blueschist Unit, which was similar to that of the Epidote Amphibolite unit and followed a shallow subduction P - T path. The Epidote Blueschist Unit, however, reached lower temperature

conditions of 470–520°C at 0.4–0.7 GPa. These rocks also exhibit much more well-developed deformation fabrics for which we present EBSD data on quartz and amphibole. The deformation fabrics and microstructures in these rocks suggest that the foliation we see preserved is likely a result of underplating deformation, although inclusions in porphyroblasts record an earlier subduction-related phase of deformation that would have occurred at seismogenic zone conditions. The underplating deformation was accommodated in part by dissolution precipitation creep of amphibole (Na- and Ca-amphibole), and dislocation creep of quartz. We apply a recrystallized grain size piezometer for quartz to show that stresses in this underplated package of oceanic crust were likely ~50–180 MPa during early exhumation and retrograde metamorphism.

Throughout this work, I have used a combination of thermodynamic modeling, field, and geochemical analysis techniques to provide new insights into subduction zone environments near the plate interface below the seismogenic zone. A highlight of these results is the importance of chemical changes in rocks before, during, and after subduction, and the knock-on effects of those changes such as fluid release. These chemical changes and their impacts are traceable through detailed petrologic and geochemical analysis, which is critical to tackling process-based questions such as what is the fluid budget of the subduction forearc, and how rocks are deforming at various stages of their metamorphic evolution.

Appendix 1

Supporting Information for Chapter 2: Metasomatism and slow slip: talc production along the flat subduction plate interface beneath Mexico (Guerrero)

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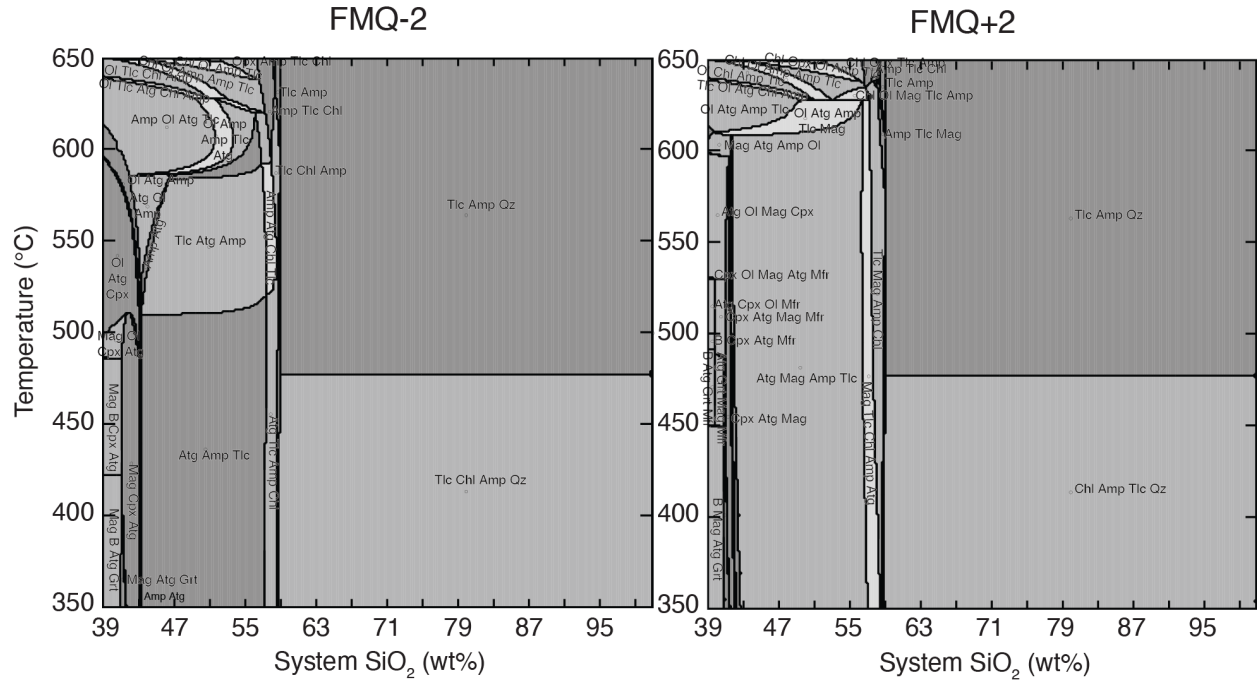


Figure S2. Pseudosections for serpentinite T-X models used for calculations of talc zone thickening rate in section 5.4. These models differ from those in Figure 2 only in that they represent more extreme degrees of silicification so the final composition (C1) is 100 wt. % SiO₂. This is used to determine the maximum amount of talc that may stabilize in each segment of the flat slab and how much silica is required to achieve that. The initial, un-silicified composition is the same as those in Figure 2 (C0 in Table 1).

Table S1. Solution models for harzburgite models

Solution Phase	Solution Model	Reference
Brucite	B	
Carpholite	Carp	
Chlorite	Chl(W)	White et al. (2014a)
Chloritoid	Ctd(HP)	Holland & Powell (1998)
Epidote	Ep(HP)	Holland & Powell. (1998)
Garnet	Gt(WPH)	White et al. (2007)
Sudoite	Sud	
Clinoamphibole	cAmph(DP)	Diener & Powell (2012)
Pumpellyite	Pu(M)	Massonne & Willner (2008)
Orthopyroxene	Opx(HP)	Powell & Holland (1999)
Clinopyroxene	Omph(GHP)	Green et al. (2007)
Antigorite	Atg(PN)	Padrón-Navarta et al. (2013)
Spinel	Sp(WPC)	White et al. (2002)
Olivine	O(HP)	Holland & Powell (1998)
Talc	T	

Table S2. Solution models for MORB models

Solution Phase	Solution Model	Reference
Biotite	Bi(W)	White et al. (2014a)
Brucite	B	
Carpholite	Carp(M)	Massonne & Willner (2008)
Chlorite	Chl(W)	White et al. (2014a)
Chloritoid	Ctd(W)	White et al. (2014b)
Epidote	Ep(HP11)	White et al. (2014b)
Garnet	Gt(W)	White et al. (2014a)
Sudoite	Sud(M)	Massonne & Willner (2008)
Clinoamphibole	cAmph(G)	Green et al. (2016)
Pumpellyite	Pu	
Orthopyroxene	Opx(W)	White et al. (2014a)
Clinopyroxene	Omph(GHP)	Green et al. (2007)
Ternary feldspar	feldspar	Fuhrman & Lindsley (1988)
Antigorite	Atg(PN)	Padrón-Navarta et al. (2013)
Muscovite	Mica(W)	White et al. (2014b)
Spinel	Sp(WPC)	White et al. (2002)
Olivine	O(HP)	Holland & Powell (1998)
Ilmenite	Ilm(WPH)	White et al. (2000, 2014a)
Stilpnomelane	Stlp	

Table S3. Solution models for metasedimentary rock models

Solution Phase	Solution Model	Reference
Biotite	Bi(W)	White et al. (2014a)
Brucite	B	
Carpholite	Carp(M)	Massonne & Willner (2008)
Chlorite	Chl(W)	White et al. (2014a)
Chloritoid	Ctd(W)	White et al. (2014b)
Epidote	Ep(HP11)	White et al. (2014b)
Garnet	Gt(W)	White et al. (2014a)
Clinoamphibole	cAmph(G)	Green et al. (2016)
Orthopyroxene	Opx(W)	White et al. (2014a)
Clinopyroxene	Omph(GHP)	Green et al. (2007)
Ternary feldspar	feldspar	Fuhrman & Lindsley (1988)
Antigorite	Atg(PN)	Padrón-Navarta et al. (2013)
Muscovite	Mica(W)	White et al. (2014b)
Spinel	Sp(WPC)	White et al. (2002)
Olivine	O(HP)	Holland & Powell (1998)
Ilmenite	Ilm(WPH)	White et al. (2000, 2014a)
Stilpnomelane	Stlp	

Appendix 2

Supporting Information for Chapter 3: Epidote records (de)hydration from the seafloor through subduction and underplating: implications for subduction zone fluid budgets and slow slip

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Introduction

The supplementary information contains elaboration on calculations behind the Sr isotope and W/R ratio results presented in the text (Text S1), additional petrographic observations (Fig. S1,–S2) and additional plots (S3–S4) of *in-situ* epidote REE data using metrics that quantify Ce and Eu anomalies, general normalized REE profile slopes, and the concavity of normalized REE profiles. Also included are the bulk composition and solution models used for Perple_X phase equilibrium models (Table S1 and S2, respectively), and representative mineral compositions for epidote (Table S3), amphibole (Table S4), feldspar (Table S5), and clinopyroxene (Table S6).

Text S1. Initial Strontium Isotope Ratio and W/R Calculations

We calculate initial strontium isotope ratios $\left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_0$ (presented in Figure 8) using the equation

$$\frac{^{87}\text{Sr}}{^{86}\text{Sr}} = \left(\frac{^{87}\text{Sr}}{^{86}\text{Sr}}\right)_0 + \frac{^{87}\text{Rb}}{^{86}\text{Sr}}(e^{\lambda t} - 1)$$

where λ is the decay constant for ^{87}Rb ($1.42 \times 10^{-11} \text{ yr}^{-1}$), and t is the elapsed time since the start of Rb decay. The $\frac{^{87}\text{Rb}}{^{86}\text{Sr}}$ ratio is estimated using the measured Rb/Sr ratio by

$$\frac{^{87}\text{Rb}}{^{86}\text{Sr}} = \left(\frac{\text{Rb}}{\text{Sr}}\right)_{\text{measured}} \cdot \frac{M_{\text{Sr}}}{M_{\text{Rb}}} \cdot \frac{^{87}\text{Rb}/\text{Rb}}{^{86}\text{Sr}/\text{Sr}}$$

where M is molar mass and the numerator and denominator of the final term describes the ratio of the estimated isotopic abundances of ^{87}Rb (0.2783) and ^{86}Sr (0.0966), respectively. Measured Rb/Sr ratios for each sample and raw $^{87}\text{Sr}/^{86}\text{Sr}$ measurements are reported in Table S9.

Initial $^{87}\text{Sr}/^{86}\text{Sr}$ values calculated for the maximum time since closure of the Rb-Sr system (195 Ma) were then used in effective water/rock ratio estimates. These estimates followed the method of Marks et al. (2015) and used the following equation to calculate W/R ratios

$$\frac{W}{R} = \left(\frac{\frac{^{87}\text{Sr}}{^{86}\text{Sr}}_{\text{rock measured}} - \frac{^{87}\text{Sr}}{^{86}\text{Sr}}_{\text{rock initial}}}{\frac{^{87}\text{Sr}}{^{86}\text{Sr}}_{\text{H}_2\text{O initial}} - \frac{^{87}\text{Sr}}{^{86}\text{Sr}}_{\text{rock initial}}} \right) \frac{C_{\text{rock initial}}}{C_{\text{water initial}}}$$

where C represents the initial concentration of Sr in either MORB or seawater. See Section 5.1 for details on assumptions and values used for mass balance equations.

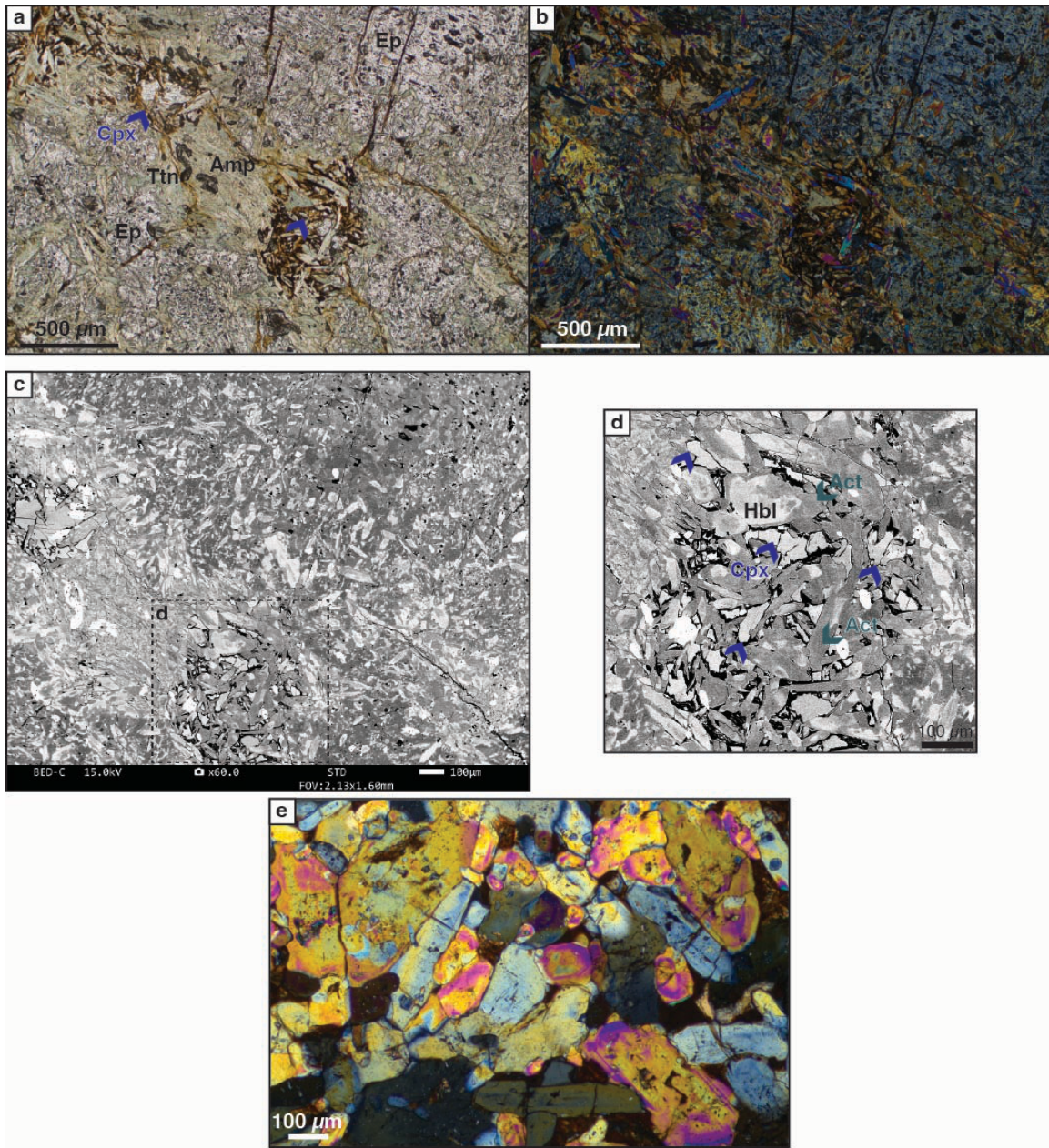


Figure S1. Micrographs from epidosite-containing sample 22C09H2. Optical PPL (a) and XPL (b) images, and (c,d) BSE images of diopside in the matrix from peak P-T conditions being replaced by amphiboles in sample. (e) Close-up view of epidote zoning textures in epidosite.

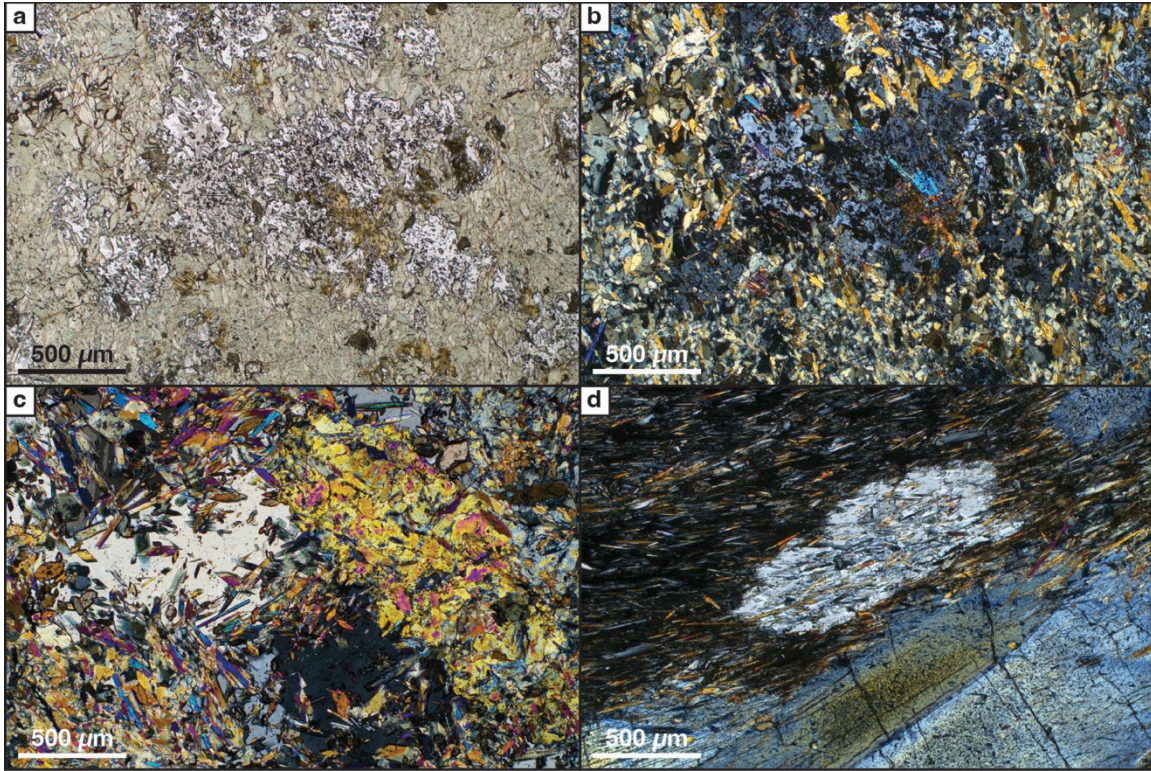


Figure S2. Photomicrographs of plagioclase (albite) occurrence in various samples. (a) PPL and (b) XPL show matrix albite with inclusions of white mica, biotite, and amphibole in 22C10B2. (c) XPL view of albite with amphibole inclusions adjacent to epidote porphyroblast (yellow) in 22C09A2. (d) Albite porphyroblast (center) in deformed sample 22C09F, adjacent to zoned epidote crystals (bottom).

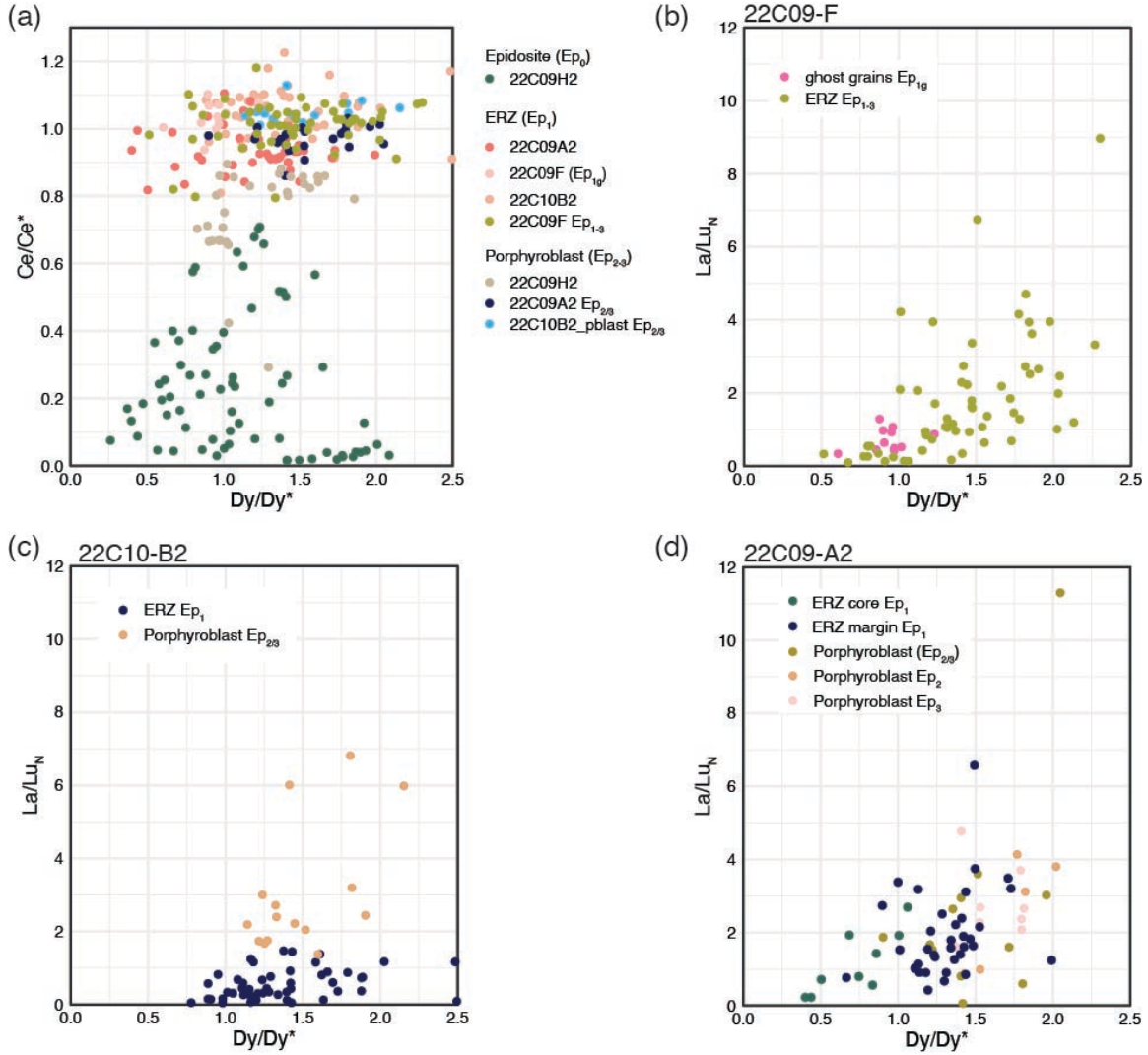


Figure S3. Quantitative analyses of *in-situ* REE data, all plotted against Dy/Dy^* , which is a measure of concavity of REE profiles and is defined as $Dy_N/(La_N^{4/13}Yb_N^{9/13})$ where the subscript N indicates a chondrite-normalized value (Davidson et al., 2013). $Dy/Dy^* > 1$ indicates a concave down profile, while $Dy/Dy^* < 1$ indicates a concave-up profile. (a) Ce/Ce^* is a measure of Cerium anomaly where $Ce/Ce^* = 2Ce_N/(La_N + Pr_N)$, so $Ce/Ce^* < 1$ represents a negative Ce-anomaly, as is evident in all of the analyses from interpillow epidiosites. (b, c, d) La/Lu_N indicates the general slope of the REE profile with $La_N/Lu_N > 1$ indicating negative slopes and $La_N/Lu_N < 1$ indicating positive slopes. In (b), Ep_{1g} analyses cluster at $La/Lu_N \leq 1$ and $Dy/Dy^* \leq 1$, while surrounding epidotes have a greater spread. In (c), Ep_1 values from an ERZ cluster around $La/Lu_N \leq 1$ while porphyroblastic epidote analysis have $La/Lu_N > 1$; all data have a spread in Dy/Dy^* generally > 1 . In (d), there is little distinction between the La/Lu_N values of different textural populations of epidote, and there is much overlap in the concavity of the populations, but ERZ Ep_1 analyses typically show lower Dy/Dy^* than porphyroblastic epidote.

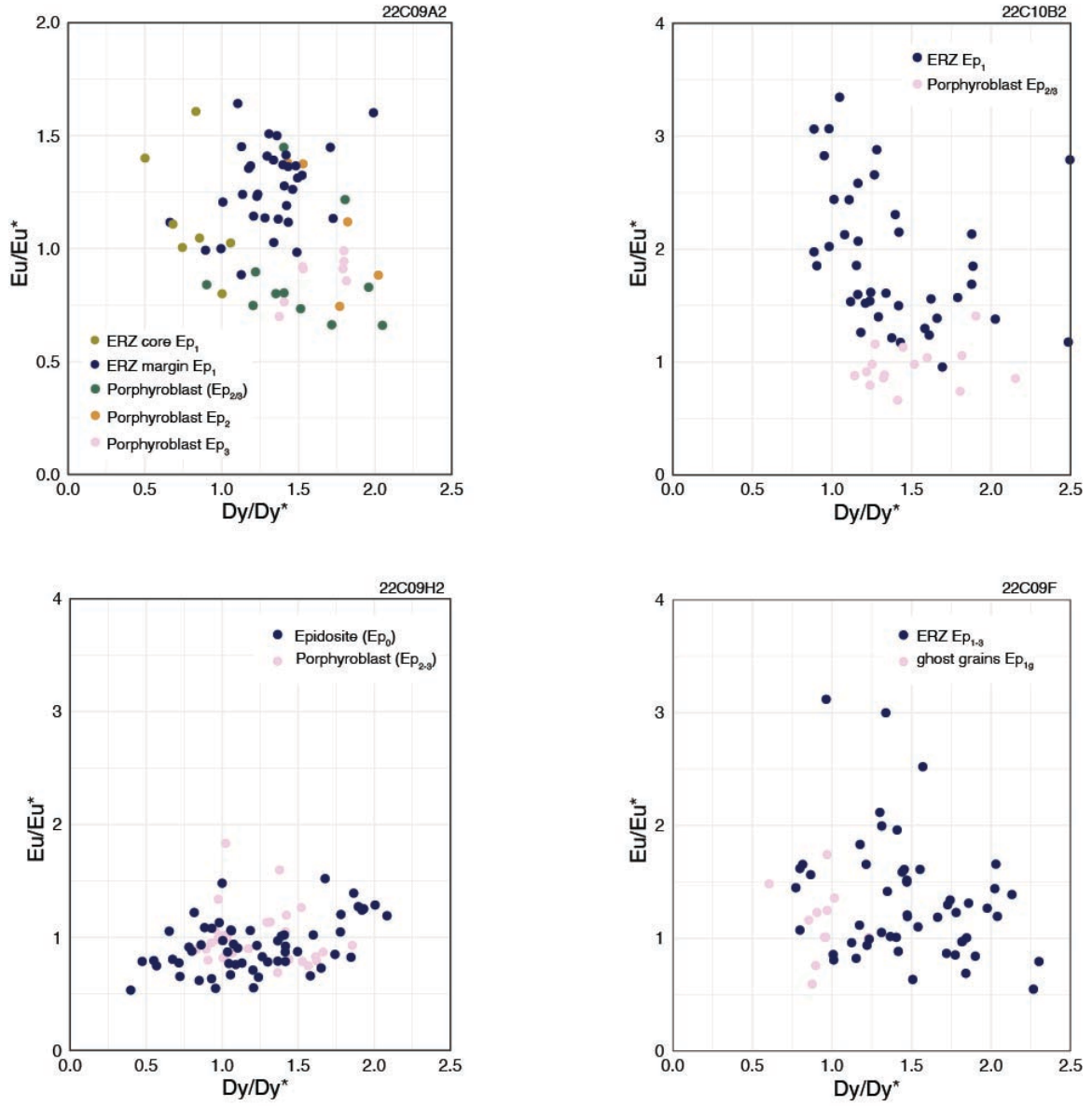


Figure S4. Plots of Eu anomaly, $Eu/Eu^* = 2Eu_N/(Sm_N + Gd_N)$ vs the concavity proxy Dy/Dy^* .

Table S1. Metabasalt bulk composition used for phase equilibria modelling (sample 22C09A2)

Oxide	XRF/titration measured (wt. %)	Modeled composition (wt. %)
SiO ₂	47.53	47.53
TiO ₂	1.11	1.11
Al ₂ O ₃	13.38	13.38
FeO	8.30	8.30
Fe ₂ O ₃	3.59	3.59
MgO	6.96	6.96
CaO	14.20	14.20
Na ₂ O	1.82	1.82
K ₂ O	0.25	0.00
P ₂ O ₅	0.02	0.00
LOI	1.97	—
H ₂ O	—	sat.*

*assumed water saturated for phase equilibrium modeling

Table S2. Activity-composition models used for phase equilibria modelling

Mineral	Solution Model	Citation
feldspar	feldspar	Fuhrman & Lindsley (1988)
clinoamphibole	cAmph(DP)	Diener & Powell (2012)
chlorite	Chl(W)	White et al. (2014)
pumpellyite	Pu(M)	Massonne & Willner (2008)
clinopyroxene	Omph(GHP)	Green et al. (2007)
orthopyroxene	Opx(HP)	Powell and Holland (1999)
olivine	O(HP)	Holland & Powell (1998)
epidote	Ep(HP)	Holland & Powell (1998)
ilmenite	Ilm(WPH)	White et al. (2000, 2014)
brucite	B	
carpholite	Carp(M)	Massonne & Willner (2008)
sudalite	Sud(M)	Massonne & Willner (2008)
antigorite	Atg(PN)	Padrón-Navarta et al. (2013)
chloritoid	Ctd(HP)	Holland & Powell (1998)
garnet	Gt(WPH)	White et al. (2007)

Table 3. Representative epidote compositions

Epidote generation	Ep ₀	Ep ₀	Ep ₀	Ep _{1g}	Ep _{1g}	Ep _{1g}	Ep ₂	Ep ₂	Ep ₂	Ep ₃	Ep ₃	Ep ₃
analysis	1348	1355	1360	1019	1021	1022	929	934	939	928	936	942
SiO ₂	38.49	38.09	38.28	38.60	38.79	39.07	38.31	38.33	38.19	38.13	38.43	38.77
TiO ₂	0.18	0.58	0.64	0.08	0.21	0.17	0.13	0.12	0.12	0.25	0.24	0.29
Al ₂ O ₃	31.10	29.07	29.75	30.45	29.79	33.14	27.98	28.93	27.97	30.32	30.16	30.39
Cr ₂ O ₃	0.00	0.00	0.00	0.04	0.04	0.04	0.07	0.00	0.04	0.08	0.12	0.07
Mn ₂ O ₃ *	0.23	0.19	0.16	0.17	0.12	0.03	0.34	0.28	0.09	0.15	0.14	0.12
Fe ₂ O ₃ *	5.47	7.51	6.36	6.09	6.67	3.05	8.75	8.20	9.32	5.98	6.31	6.02
MgO	0.06	0.11	0.10	0.04	0.06	0.04	0.02	0.02	0.05	0.06	0.07	0.05
CaO	23.70	23.56	23.84	23.75	23.50	24.21	23.22	23.50	23.04	23.75	23.75	23.70
Na ₂ O	0.00	0.00	0.00	0.00	0.00	0.02	0.01	0.00	0.02	0.00	0.00	0.00
K ₂ O	0.00	0.00	0.00	0.00	0.01	0.01	0.01	0.00	0.00	0.01	0.02	0.01
Total	99.24	99.11	99.12	99.22	99.19	99.78	98.86	99.39	98.83	98.72	99.23	99.42
X _{Ep}	0.29	0.41	0.35	0.32	0.36	0.16	0.48	0.44	0.50	0.32	0.34	0.32

*recalculated from FeO and MnO assuming all Fe and Mn present as Fe³⁺ and Mn³⁺.

Table 4. Representative amphibole compositions

setting	inclusion	matrix	matrix	matrix	inclusion	inclusion
amphibole	actinolite	actinolite	actinolite	magnesio-hornblende	magnesio-hornblende	ferro-pargasite
analysis	956	959	963	964	950	951
SiO ₂	54.00	52.34	53.33	45.19	45.89	43.05
TiO ₂	0.03	0.07	0.04	0.35	0.32	0.53
Al ₂ O ₃	1.76	3.66	2.33	11.06	10.37	13.30
Cr ₂ O ₃	0.04	0.07	0.03	0.04	0.04	0.10
MnO	0.45	0.44	0.42	0.44	0.39	0.42
FeO	13.31	13.58	13.25	16.36	16.01	17.21
MgO	14.52	13.83	14.45	9.74	10.15	8.70
CaO	12.35	12.34	12.44	11.79	12.08	11.55
Na ₂ O	0.29	0.51	0.34	1.38	1.30	1.57
K ₂ O	0.05	0.16	0.11	0.82	0.63	1.22
F	0.00	0.04	0.06	0.03	0.05	0.04
Cl	0.01	0.00	0.00	0.00	0.00	0.00
Total	96.81	97.00	96.79	97.18	97.22	97.65

Table 5. Representative feldspar compositions

Sample	22C-09I2	22C-10B2	22C-10B2
analysis	3	9	15
SiO ₂	67.61	67.68	66.96
Al ₂ O ₃	20.14	21.25	21.15
FeO	0.22	0.03	0.09
MgO	0.12	0.00	0.00
CaO	0.61	1.38	1.50
Na ₂ O	11.11	10.35	10.37
K ₂ O	0.06	0.06	0.07
Total	99.86	100.74	100.15
X _{Ab}	0.97	0.93	0.92
X _{An}	0.03	0.07	0.07

Table 6. Representative clinopyroxene compositions

Sample	22C-09A2	22C-10E	22C-09H2
analysis	993	1175	1318
SiO ₂	52.72	54.57	52.03
TiO ₂	0.02	0.03	0.10
Al ₂ O ₃	0.97	0.38	1.47
Cr ₂ O ₃	0.01	0.08	0.08
MnO	0.53	0.46	0.62
FeO	10.84	6.40	9.31
MgO	10.95	13.95	11.77
CaO	23.42	24.45	23.45
Na ₂ O	0.47	0.31	0.56
K ₂ O	0.02	0.01	0.00
Total	99.98	100.60	99.46

Uploaded Separately to Lindquist et al. (2026) Zenodo repository:

Table S7. Major element concentrations measured via EPMA

Table S8. *In situ* trace element concentrations measured via LA-ICP-MS.

Table S9. Bulk-rock Sr isotope measurements, corrections, and W/R estimates.

Appendix 3

Supporting Information for Chapter 4: Metamorphism and deformation during subduction and underplating: the evolution of the Catalina Schist Epidote Blueschist Unit

P. C. Lindquist¹, C. B. Condit¹, V. E. Guevara²

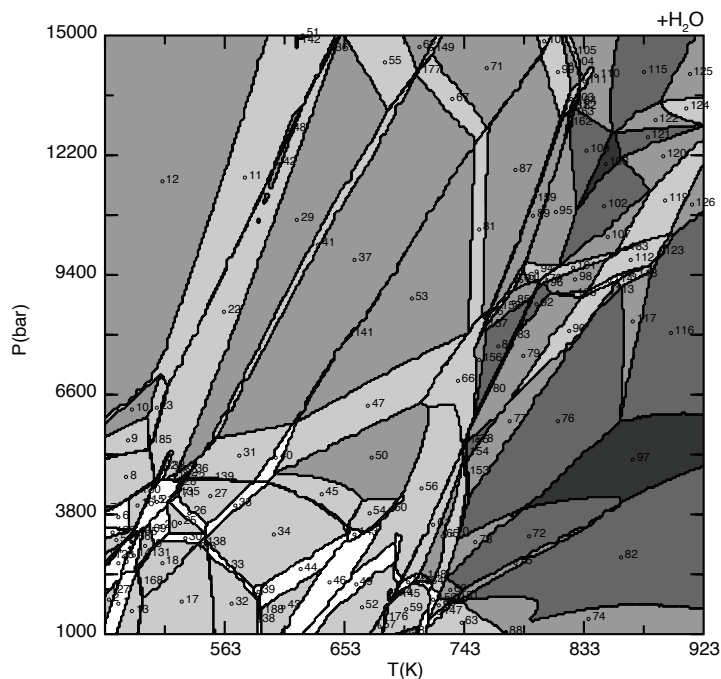
¹ Department of Earth and Space Sciences, University of Washington.

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Contents of this file

Figures S1–S8

Fig S1. Pseudosection for epidote-blueschist metabasalt
23C04D metabasalt Fe³/Fe^T =0.3 using O₂

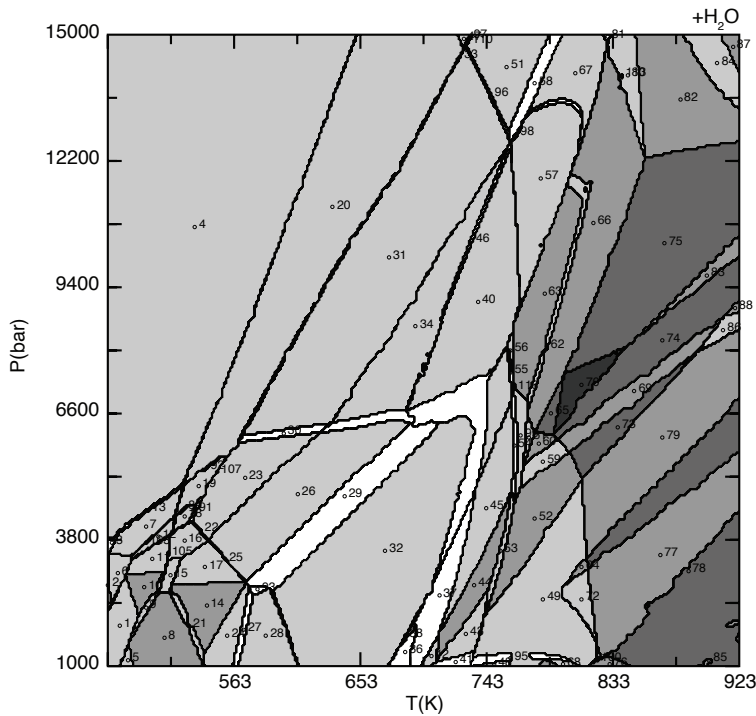


- 1 - Chl Grt Stlp(M) sph rieab ab mic q lmt
- 2 - Chl Grt Stlp(M) sph rieab ab mic lmt
- 3 - Chl Cpx Grt Mica Stlp(M) sph rieab ab lmt
- 4 - Chl Cpx Grt Stlp(M) law sph ab lmt
- 5 - Chl Cpx Cpx Stlp(M) law sph ab lmt
- 6 - Chl Cpx Cpx Stlp(M) law sph rieab ab q
- 7 - Cam Chl Cpx Stlp(M) law sph rieab ab q
- 8 - Cam Chl Cpx Stlp(M) law sph ab q
- 9 - Cam Cpx Mica Stlp(M) law sph ab q
- 10 - Cam Cpx Mica Stlp(M) law sph q
- 11 - Cam Cpx Mica Stlp(M) law sph q ru
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- 13 - Chl Grt sph rieab ab mic q lmt
- 14 - Chl Grt Mica Stlp(M) sph rieab ab q lmt
- 15 - Chl Cpx Cpx Mica Stlp(M) law sph ab q
- 16 - Chl Cpx Cpx Stlp(M) law sph ab q
- 17 - Chl Ep(HP) Grt sph rieab ab mic q
- 18 - Chl Grt Mica sph rieab ab mic q
- 19 - Chl Grt Mica Stlp(M) sph rieab ab q
- 20 - Chl Cpx Grt Mica Stlp(M) sph rieab ab q
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- 27 - Chl Cpx Ep(HP) Mica sph rieab ab q
- 28 - Chl Cpx Mica law sph rieab ab q
- 29 - Cam Chl Cpx Mica law sph q
- 30 - Chl Cpx Grt Mica sph rieab ab mic q
- 31 - Cam Chl Cpx Ep(HP) Mica sph ab q
- 32 - Chl Ep(HP) Grt Bt sph rieab ab q
- 33 - Chl Cpx Ep(HP) Grt Bt sph rieab ab q
- 34 - Chl Cpx Ep(HP) Bt sph rieab ab q
- 35 - Chl Cpx Ep(HP) Bt Mica sph rieab ab q
- 36 - Cam Chl Cpx Mica Stlp(M) law sph q ru
- 37 - Cam Chl Cpx Ep(HP) Mica sph q
- 38 - Chl Ep(HP) Grt Bt sph rieab ab q acti
- 39 - Chl Cpx Ep(HP) Bt sph rieab ab q acti
- 40 - Cam Chl Cpx Ep(HP) Bt Mica sph ab q
- 41 - Cam Chl Cpx Ep(HP) Mica law sph q
- 42 - Cam Chl Mica Stlp(M) law sph q
- 43 - Cam Chl Ep(HP) Bt sph rieab ab q
- 44 - Cam Chl Cpx Ep(HP) Bt sph rieab ab q
- 45 - Cam Chl Cpx Ep(HP) Bt sph ab q
- 46 - Cam Chl Ep(HP) Ilm Bt sph rieab ab q
- 47 - Cam Chl Ep(HP) Bt Mica sph ab q
- 48 - Cam Mica Stlp(M) law sph q ru
- 49 - Cam Chl Ep(HP) Ilm Bt sph ab q mt
- 50 - Cam Chl Ep(HP) Bt sph ab q
- 51 - Cam Cpx Mica Stlp(M) law q ru
- 52 - Cam Chl Ep(HP) Bt sph ab q mt
- 53 - Cam Chl Ep(HP) Bt Mica sph q
- 54 - Cam Cam Chl Ep(HP) Bt sph ab q
- 55 - Cam Chl Cpx Mica law sph q ru
- 56 - Cam Chl Ep(HP) Ilm Bt sph ab q
- 57 - Fsp Cam Chl Ep(HP) Bt sph q mt
- 58 - Fsp Cam Chl Bt sph q mt
- 59 - Fsp Cam Cam Chl Bt sph q mt
- 60 - Cam Cam Chl Ep(HP) Ilm Bt sph ab q
- 61 - Fsp Cam Chl Ilm Bt sph ab q
- 62 - Cam Chl Cpx v law q ru
- 63 - Fsp Cam Cam Chl Ilm Bt q mt

- 64 - Fsp Cam Cam Chl Ilm Bt sph q mt
- 65 - Fsp Cam Chl Ilm Bt sph q
- 66 - Cam Chl Ep(HP) Bt sph ab q ru
- 67 - Cam Chl Cpx Ep(HP) Mica sph q ru
- 68 - Fsp Cam Cam Chl Ilm Ilm Bt q
- 69 - Fsp Cam Chl Ilm Bt q
- 70 - Fsp Cam Chl Ilm Ilm Bt sph q
- 71 - Cam Chl Cpx Ep(HP) Mica q ru
- 72 - Fsp Cam Ilm Ilm Bt q
- 73 - Fsp Cam Ilm Ilm Bt sph q
- 74 - Fsp Cam Cam Ilm Bt q mt
- 75 - Fsp Cam Ilm Ilm Bt q mt
- 76 - Fsp Cam Ilm Bt sph q
- 77 - Fsp Cam Ilm Bt sph q ru
- 78 - Fsp Cam Bt sph q ru
- 79 - Fsp Cam Ep(HP) Bt sph q ru
- 80 - Fsp Cam Ep(HP) Bt sph ab q ru
- 81 - Cam Chl Ep(HP) Bt Mica sph q ru
- 82 - Fsp Cam Ilm Bt q mt
- 83 - Fsp Cam Ep(HP) Bt ab q ru
- 84 - Cam Ep(HP) Bt ab q ru
- 85 - Cam Ep(HP) Bt Mica ab q ru
- 86 - Cam Chl Ep(HP) Bt Mica ab q ru
- 87 - Cam Chl Ep(HP) Bt Mica q ru
- 88 - Fsp Cam Cam Cam Ilm Bt q mt
- 89 - Cam Cam Ep(HP) Bt Mica q ru
- 90 - Fsp Cam Ep(HP) Ilm Bt sph q ru
- 91 - Cam Ep(HP) Bt Mica sph q ru
- 92 - Fsp Cam Ep(HP) Bt q ru
- 93 - Cam Cpx Ep(HP) Bt Mica sph ab q ru
- 94 - Cam Cpx Ep(HP) Bt Mica sph q ru
- 95 - Cam Cpx Ep(HP) Bt Mica q ru
- 96 - Cam Cam Cpx Ep(HP) Bt Mica q ru
- 97 - Fsp Cam Ilm Bt q
- 98 - Fsp Cam Cpx Ep(HP) Bt q ru
- 99 - Cam Chl Ep(HP) Grt Bt Mica q ru
- 100 - Cam Chl Cpx Ep(HP) Grt Mica q ru
- 101 - Fsp Cam Cpx Ep(HP) Bt Mica q ru
- 102 - Cam Cpx Bt Mica q ru
- 103 - Cam Cam Ep(HP) Grt Bt Mica q ru
- 104 - Cam Ep(HP) Grt Bt Mica q ru
- 105 - Cam Cpx Ep(HP) Grt Bt Mica q ru
- 106 - Fsp Cam Cpx Ep(HP) Bt sph q ru
- 107 - Fsp Cam Cpx Bt Mica q ru
- 108 - Cam Cpx Mica q ru
- 109 - Cam Cpx Ep(HP) Mica q ru
- 110 - Cam Cpx Ep(HP) Grt Mica q ru
- 111 - Cam Cam Cpx Ep(HP) Grt Mica q ru
- 112 - Fsp Cam Cpx Ep(HP) Ilm Bt q ru
- 113 - Fsp Cam Ep(HP) Ilm Bt sph q
- 114 - Fsp Cam Cpx Ep(HP) Ilm Bt sph q ru
- 115 - Cam Cpx Grt Mica q ru
- 116 - Fsp Cam Cpx Ilm Bt q
- 117 - Fsp Cam Cpx Ilm Bt sph q
- 118 - Fsp Cam Cpx Ep(HP) Ilm Bt sph q
- 119 - Fsp Cam Cpx Ilm Bt Mica q ru
- 120 - Cam Cpx Ilm Bt Mica q ru
- 121 - Cam Cpx Ilm Mica q ru
- 122 - Cam Cpx Ilm Grt Mica q ru
- 123 - Fsp Cam Cpx Ep(HP) Ilm Bt q
- 124 - Cam Cpx Ilm Grt Bt Mica q ru
- 125 - Cam Cpx Grt Bt Mica q ru

- 126 - Fsp Cam Cpx Ilm Bt Mica q
- 127 - Chl Cpx Grt Stlp(M) sph rieab ab mic lmt
- 128 - Chl Cpx Grt Mica Stlp(M) law sph ab lmt
- 129 - Chl Cpx Cpx Stlp(M) law sph rieab ab lmt
- 130 - Cam Chl Cpx Cpx Stlp(M) law sph ab q
- 131 - Chl Grt Mica Stlp(M) sph rieab ab mic q
- 132 - Cam Chl Cpx Mica Stlp(M) law sph ab q
- 133 - Cam Chl Cpx Cpx Mica Stlp(M) law sph q
- 134 - Cam Chl Cpx Mica Mica Stlp(M) law sph q
- 135 - Chl Pu(M) Cpx Ep(HP) Mica sph rieab ab q
- 136 - Cam Chl Cpx Mica law sph ab q
- 137 - Cam Chl Cpx Ep(HP) Mica law sph ab q
- 138 - Chl Cpx Ep(HP) Grt sph rieab ab mic q
- 139 - Cam Chl Cpx Ep(HP) Mica sph rieab ab q
- 140 - Cam Chl Mica Stlp(M) law sph q ru
- 141 - Cam Chl Cpx Ep(HP) Bt Mica sph q
- 142 - Cam Cpx Cpx Mica Stlp(M) law sph q ru
- 143 - Cam Cam Chl Ep(HP) Bt sph rieab ab q
- 144 - Cam Cam Chl Ep(HP) Bt sph ab q mt
- 145 - Fsp Cam Chl Bt sph ab q mt
- 146 - Fsp Cam Chl Ilm Bt sph ab q mt
- 147 - Fsp Cam Cam Chl Ilm Ilm Bt q mt
- 148 - Fsp Cam Chl Ilm Bt sph q mt
- 149 - Cam Chl Cpx Ep(HP) Mica law q ru
- 150 - Fsp Cam Cam Chl Ilm Ilm Bt sph q
- 151 - Fsp Cam Chl Ilm Ilm Bt q mt
- 152 - Cam Chl Ep(HP) Bt Mica sph ab q ru
- 153 - Fsp Cam Chl Ilm Bt sph q ru
- 154 - Fsp Cam Chl Bt sph q ru
- 155 - Fsp Cam Chl Ep(HP) Bt sph q ru
- 156 - Cam Chl Ep(HP) Bt ab q ru
- 157 - Cam Cam Chl Ep(HP) Bt ab q ru
- 158 - Cam Cam Ep(HP) Bt Mica ab q ru
- 159 - Cam Cam Chl Ep(HP) Bt Mica q ru
- 160 - Cam Chl Cpx Ep(HP) Bt Mica q ru
- 161 - Cam Cam Ep(HP) Bt Mica sph q ru
- 162 - Cam Ep(HP) Bt Mica q ru
- 163 - Cam Ep(HP) Mica q ru
- 164 - Cam Cam Ep(HP) Grt Mica q ru
- 165 - Chl Cpx Mica Stlp(M) law sph rieab ab lmt
- 166 - Chl Cpx Cpx Mica Stlp(M) law sph ab lmt
- 167 - Chl Ep(HP) Grt sph rieab ab mic q lmt
- 168 - Chl Grt Mica sph rieab ab mic q lmt
- 169 - Chl Grt Mica Stlp(M) law sph rieab ab q
- 170 - Cam Cpx Mica Mica Stlp(M) law sph q ru
- 171 - Chl Pu(M) Cpx Mica law sph rieab ab q
- 172 - Cam Chl Cpx Mica law sph rieab ab q
- 173 - Chl Cpx Ep(HP) Mica law sph rieab ab q
- 174 - Cam Cam Chl Cpx Ep(HP) Bt sph ab q
- 175 - Fsp Cam Chl Ep(HP) Bt sph ab q mt
- 176 - Fsp Cam Cam Chl Ep(HP) Bt sph q mt
- 177 - Cam Chl Cpx Ep(HP) Mica law sph q ru
- 178 - Fsp Cam Ep(HP) Bt Mica ab q ru
- 179 - Cam Cpx Ep(HP) Bt Mica ab q ru
- 180 - Cam Cpx Bt Mica sph q ru
- 181 - Cam Cam Chl Ep(HP) Grt Bt Mica q ru
- 182 - Cam Ep(HP) Grt Mica q ru
- 183 - Fsp Cam Cpx Ep(HP) Ilm Bt Mica q ru
- 184 - Chl Cpx Grt Mica Stlp(M) sph ab lmt
- 185 - Cam Cpx Mica Mica Stlp(M) law sph ab q
- 186 - Chl Ep(HP) Grt Mica sph rieab ab mic q
- 187 - Chl Ep(HP) Grt Bt sph rieab ab mic q
- 188 - Cam Chl Ep(HP) Grt Bt sph rieab ab q
- 189 - Fsp Cam Chl Ep(HP) Ilm Bt sph ab q
- 190 - Fsp Cam Cam Chl Ilm Bt sph q
- 191 - Fsp Cam Chl Ilm Ilm Bt sph q ru
- 192 - Cam Chl Ep(HP) Ilm Bt sph ab q ru
- 193 - Fsp Cam Chl Ep(HP) Bt sph ab q ru
- 194 - Cam Ep(HP) Bt sph ab q ru
- 195 - Cam Ep(HP) Bt Mica sph ab q ru
- 196 - Fsp Cam Ep(HP) Bt Mica q ru
- 197 - Cam Chl Cpx Ep(HP) Grt Bt Mica q ru
- 198 - Fsp Cam Cpx Bt Mica sph q ru

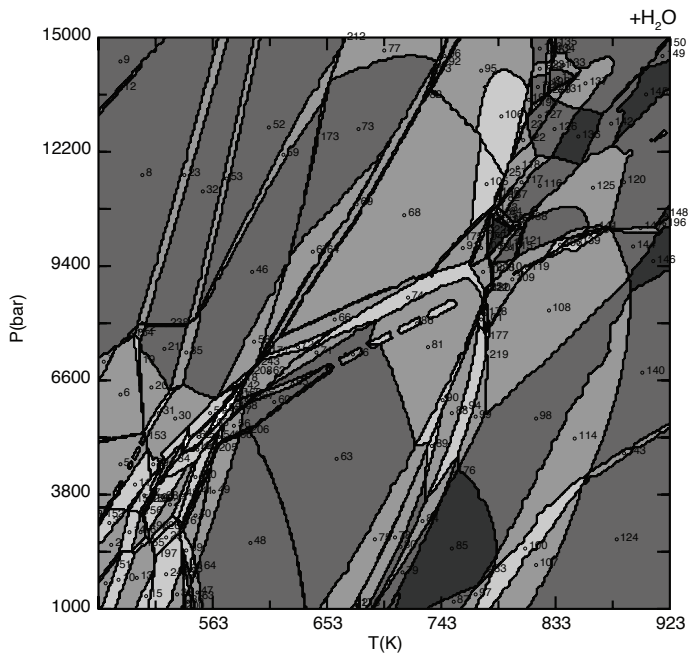
Fig S2. Pseudosection for epidote-actinolite greenschist metabasalt
21C09B metabasalt using fixed activity of O₂ -405340 from 23C04D



- 1 - Chl(HP) Gt(WPPH) Stlp(M) sph ab q lmt
- 2 - Chl(HP) Omph(GHP) Stlp(M) sph rie b ab q lmt
- 3 - Chl(HP) Omph(GHP) Stlp(M) law sph rie b ab q
- 4 - cAmph(DP) Chl(HP) Omph(GHP) Stlp(M) law sph q
- 5 - Chl(HP) Gt(WPPH) sph ab mic q lmt
- 6 - Chl(HP) Omph(GHP) Stlp(M) sph ab q lmt
- 7 - Chl(HP) Omph(GHP) Stlp(M) law sph ab q
- 8 - Chl(HP) Gt(WPPH) sph ab mic q
- 9 - Chl(HP) Gt(WPPH) Stlp(M) sph ab mic q
- 10 - Chl(HP) Gt(WPPH) Stlp(M) sph ab q
- 11 - Chl(HP) Omph(GHP) Gt(WPPH) Stlp(M) sph ab q
- 12 - Chl(HP) Pu(M) Omph(GHP) Stlp(M) sph ab q
- 13 - cAmph(DP) Chl(HP) Omph(GHP) Stlp(M) law sph ab q
- 14 - Chl(HP) Gt(WPPH) Bio(WPH) sph ab q
- 15 - Chl(HP) Omph(GHP) Gt(WPPH) sph ab mic q
- 16 - Chl(HP) Omph(GHP) Gt(WPPH) Mica(CHA) sph ab q
- 17 - Chl(HP) Omph(GHP) Gt(WPPH) Bio(WPH) sph ab q
- 18 - Chl(HP) Pu(M) Omph(GHP) Mica(CHA) sph ab q
- 19 - Chl(HP) Omph(GHP) Mica(CHA) law sph ab q
- 20 - cAmph(DP) Chl(HP) Omph(GHP) Mica(CHA) law sph q
- 21 - Chl(HP) Gt(WPPH) Bio(WPH) sph ab mic q
- 22 - Chl(HP) Omph(GHP) Ep(HP) Gt(WPPH) Mica(CHA) sph ab q
- 23 - Chl(HP) Omph(GHP) Ep(HP) Mica(CHA) sph ab q
- 24 - Chl(HP) Ep(HP) Gt(WPPH) Bio(WPH) sph ab q
- 25 - Chl(HP) Omph(GHP) Ep(HP) Gt(WPPH) Bio(WPH) sph ab q
- 26 - Chl(HP) Omph(GHP) Ep(HP) Bio(WPH) sph ab q
- 27 - Chl(HP) Ep(HP) Gt(WPPH) Bio(WPH) sph ab q acti
- 28 - Chl(HP) Ep(HP) Bio(WPH) sph ab q acti
- 29 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Bio(WPH) sph ab q
- 30 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Mica(CHA) sph ab q
- 31 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Mica(CHA) sph q
- 32 - cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) sph ab q
- 33 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Mica(CHA) law sph q
- 34 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Bio(WPH) sph q
- 35 - cAmph(DP) Chl(HP) Omph(GHP) Mica(CHA) Stlp(M) law sph q
- 36 - feldspar cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) Sp(WPC) sph q
- 37 - feldspar cAmph(DP) cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) sph q
- 38 - feldspar cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) sph ab q
- 39 - cAmph(DP) cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) sph ab q
- 40 - cAmph(DP) cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) sph q
- 41 - feldspar cAmph(DP) cAmph(DP) Chl(HP) Ilm(WPH) Bio(WPH) sph q
- 42 - feldspar cAmph(DP) cAmph(DP) Chl(HP) Bio(WPH) sph q
- 43 - feldspar cAmph(DP) Chl(HP) Ilm(WPH) Bio(WPH) sph q
- 44 - feldspar cAmph(DP) Chl(HP) Bio(WPH) sph q
- 45 - feldspar cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) sph q
- 46 - cAmph(DP) cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Bio(WPH) sph q

- 47 - Chl(HP) Omph(GHP) Mica(CHA) law sph q ru
- 48 - feldspar feldspar cAmph(DP) Chl(HP) Ilm(WPH) Bio(WPH) q ru
- 49 - feldspar cAmph(DP) Chl(HP) Ilm(WPH) Bio(WPH) q ru
- 50 - feldspar cAmph(DP) Chl(HP) Ilm(WPH) Bio(WPH) sph q ru
- 51 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Mica(CHA) q ru
- 52 - feldspar cAmph(DP) Chl(HP) Bio(WPH) q ru
- 53 - feldspar cAmph(DP) Chl(HP) Bio(WPH) sph q ru
- 54 - feldspar cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) sph q ru
- 55 - cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) sph ab q ru
- 56 - cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) sph q ru
- 57 - cAmph(DP) cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) q ru
- 58 - cAmph(DP) cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Mica(CHA) q ru
- 59 - feldspar cAmph(DP) Chl(HP) Bio(WPH) Mica(CHA) q ru
- 60 - feldspar cAmph(DP) Chl(HP) Mica(CHA) q ru
- 61 - feldspar cAmph(DP) Chl(HP) Ep(HP) Mica(CHA) q ru
- 62 - cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) Mica(CHA) q ru
- 63 - cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) q ru
- 64 - feldspar cAmph(DP) Chl(HP) Ilm(WPH) Bio(WPH) q
- 65 - cAmph(DP) Chl(HP) Mica(CHA) q ru
- 66 - cAmph(DP) Chl(HP) Ep(HP) Mica(CHA) q ru
- 67 - cAmph(DP) cAmph(DP) Chl(HP) Ep(HP) Mica(CHA) q ru
- 68 - feldspar cAmph(DP) cAmph(DP) Chl(HP) Ilm(WPH) Bio(WPH) q
- 69 - feldspar cAmph(DP) Bio(WPH) Mica(CHA) q ru
- 70 - cAmph(DP) Mica(CHA) q ru
- 71 - cAmph(DP) cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) Mica(CHA) q ru
- 72 - feldspar cAmph(DP) Chl(HP) Ilm(WPH) Ilm(WPH) Bio(WPH) q
- 73 - feldspar cAmph(DP) Bio(WPH) q ru
- 74 - feldspar cAmph(DP) Mica(CHA) q ru
- 75 - cAmph(DP) Ep(HP) Mica(CHA) q ru
- 76 - feldspar cAmph(DP) Ilm(WPH) Bio(WPH) ta q
- 77 - feldspar cAmph(DP) Ilm(WPH) Ilm(WPH) Bio(WPH) q
- 78 - feldspar cAmph(DP) Ilm(WPH) Bio(WPH) q
- 79 - feldspar cAmph(DP) Ilm(WPH) Bio(WPH) q ru
- 80 - cAmph(DP) Chl(HP) Ep(HP) Gt(WPPH) Mica(CHA) q ru
- 81 - cAmph(DP) cAmph(DP) Chl(HP) Ep(HP) Gt(WPPH) Mica(CHA) q ru
- 82 - cAmph(DP) Ep(HP) Gt(WPPH) Mica(CHA) q ru
- 83 - feldspar cAmph(DP) Ep(HP) Mica(CHA) q ru
- 84 - cAmph(DP) Omph(GHP) Ep(HP) Gt(WPPH) Mica(CHA) q ru
- 85 - feldspar cAmph(DP) cAmph(DP) Ilm(WPH) Bio(WPH) q
- 86 - feldspar cAmph(DP) Ilm(WPH) Bio(WPH) Mica(CHA) q ru
- 87 - cAmph(DP) Omph(GHP) Gt(WPPH) Mica(CHA) q ru
- 88 - feldspar cAmph(DP) Ilm(WPH) Mica(CHA) q ru
- 89 - Chl(HP) Pu(M) Omph(GHP) Mica(CHA) law sph ab q
- 90 - Chl(HP) Omph(GHP) Mica(CHA) Stlp(M) law sph ab q
- 91 - Chl(HP) Pu(M) Omph(GHP) Ep(HP) Mica(CHA) sph ab q
- 92 - cAmph(DP) Chl(HP) Omph(GHP) Mica(CHA) law sph ab q
- 93 - Chl(HP) Omph(GHP) Ep(HP) Bio(WPH) sph ab q acti
- 94 - cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) sph ab q acti
- 95 - feldspar cAmph(DP) cAmph(DP) Chl(HP) Ilm(WPH) Bio(WPH) q ru
- 96 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Mica(CHA) sph q ru
- 97 - Chl(HP) Omph(GHP) Ep(HP) Mica(CHA) law q ru
- 98 - cAmph(DP) cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Bio(WPH) q ru
- 99 - feldspar cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) q ru
- 100 - feldspar cAmph(DP) Ilm(WPH) Ilm(WPH) ta q ru
- 101 - Chl(HP) Omph(GHP) Gt(WPPH) Stlp(M) sph ab q lmt
- 102 - Chl(HP) Pu(M) Omph(GHP) Gt(WPPH) Stlp(M) sph ab q
- 103 - Chl(HP) Pu(M) Omph(GHP) Stlp(M) law sph ab q
- 104 - Chl(HP) Omph(GHP) Gt(WPPH) Stlp(M) sph ab mic q
- 105 - Chl(HP) Omph(GHP) Gt(WPPH) Mica(CHA) Stlp(M) sph ab q
- 106 - Chl(HP) Pu(M) Omph(GHP) Gt(WPPH) Mica(CHA) sph ab q
- 107 - Chl(HP) Omph(GHP) Ep(HP) Mica(CHA) law sph ab q
- 108 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Bio(WPH) Mica(CHA) sph q
- 109 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Mica(CHA) law q ru
- 110 - Chl(HP) Omph(GHP) Ep(HP) Mica(CHA) law sph q ru
- 111 - cAmph(DP) cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) sph q ru
- 112 - cAmph(DP) cAmph(DP) cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) q ru
- 113 - cAmph(DP) Chl(HP) Ep(HP) Gt(WPPH) Mica(CHA) Mica(CHA) q ru
- 114 - Chl(HP) Gt(WPPH) Stlp(M) sph ab mic q lmt
- 115 - Chl(HP) Pu(M) Omph(GHP) Stlp(M) sph ab q lmt
- 116 - Chl(HP) Pu(M) Omph(GHP) Mica(CHA) Stlp(M) sph ab q
- 117 - cAmph(DP) Chl(HP) Omph(GHP) Mica(CHA) law sph q ru
- 118 - cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) ab q ru
- 119 - feldspar cAmph(DP) Chl(HP) Ilm(WPH) Ilm(WPH) Bio(WPH) q ru
- 120 - cAmph(DP) Chl(HP) Ep(HP) Mica(CHA) Mica(CHA) q ru

Fig S3. Pseudosection for greenschist layer of metapelite
23C04A1 green layer metased Fe₃/FeT =0.09 using O₂



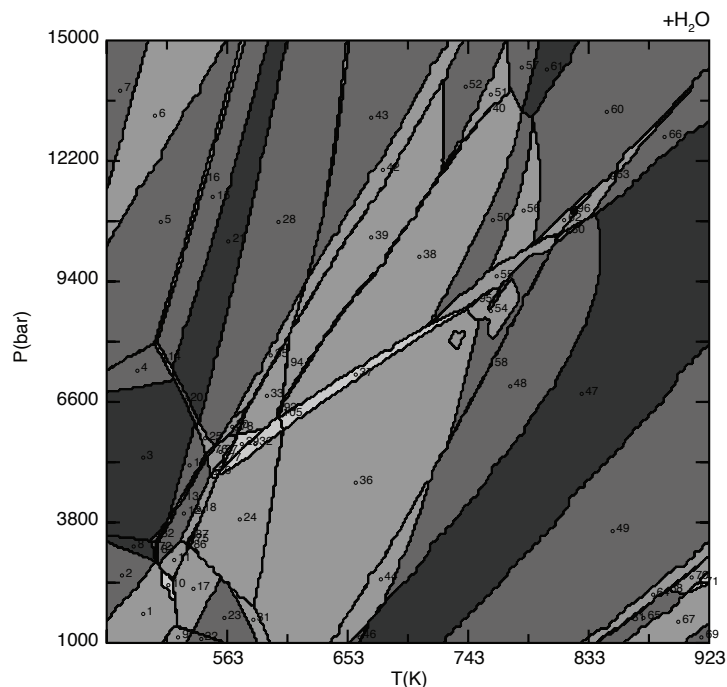
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2 - Chl Mica Stp sph rieab ab Qz lmt
3 - Chl Cpx Mica Stp sph rieab ab Qz lmt
4 - Chl Cpx Mica Stp sph ab Qz lmt
5 - Chl Cpx Mica Stp Law sph ab Qz
6 - Cam Cpx Mica Stp Law sph ab Qz
7 - Cam Cpx Mica Stp Law sph ab Qz Rt
8 - Cam Cpx Mica Stp Law Qz Rt
9 - Cam Cpx Carp(M) Stp Law Qz Rt
10 - Chl Grt Stp sph rieab ab mic Qz lmt
11 - Chl Cpx Mica Stp Law sph rieab ab Qz
12 - Cam Cpx Carp(M) Mica Stp Law Qz Rt
13 - Chl Grt Stp sph ab mic Qz lmt
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15 - Chl Grt Stp sph pre ab mic Qz lmt
16 - Chl Grt Mica Stp sph ab Qz lmt
17 - Chl Grt Mica Stp Law sph rieab ab Qz
18 - Cam Chl Cpx Mica Stp Law sph ab Qz
19 - Cam Cpx Mica Mica Stp Law sph ab Qz
20 - Cam Mica Mica Stp Law sph ab Qz
21 - Cam Mica Mica Stp Law sph Qz
22 - Cam Cpx Mica Mica Stp Law sph Qz
23 - Cam Cpx Mica Mica Stp Law Qz Rt
24 - Chl Ep Grt Stp sph pre ab mic Qz
25 - Chl Pmp Ep Grt Stp sph ab mic Qz
26 - Chl Pmp Grt Mica Stp sph ab mic Qz
27 - Chl Pmp Grt Mica Stp sph ab Qz
28 - Chl Pmp Ep Grt Mica Stp sph ab Qz
29 - Cam Chl Mica Stp Law sph rieab ab Qz
30 - Cam Chl Mica Stp Law sph ab Qz
31 - Cam Chl Mica Mica Stp Law sph ab Qz
32 - Cam Mica Mica Stp Law Qz Rt
33 - Chl Ep Grt sph pre ab mic Qz
34 - Chl Pmp Cpx Mica Stp Law sph ab Qz
35 - Cam Chl Mica Mica Stp Law sph Qz
36 - Chl Ep Bt sph pre ab mic Qz
37 - Chl Ep Bt Stp sph pre ab mic Qz
38 - Chl Ep Bt Stp sph pre ab Qz
39 - Chl Ep Grt Bt Stp sph pre ab Qz
40 - Chl Ep Grt Bt Stp sph ab Qz
41 - Chl Pmp Ep Grt Bt Stp sph ab Qz
42 - Chl Pmp Ep Bt Mica Stp sph ab Qz
43 - Chl Pmp Grt Bt Mica Stp sph ab Qz
44 - Chl Pmp Cpx Ep Mica Stp sph ab Qz
45 - Cam Chl Pmp Cpx Mica Stp sph ab Qz
46 - Cam Chl Mica Stp Law sph Qz
47 - Chl Ep Bt sph pre ab Qz acti
48 - Chl Ep Bt sph ab Qz acti

49 - Chl Ep Bt Stp sph ab Qz acti
50 - Chl Pmp Ep Bt Stp sph ab Qz
51 - Cam Chl Pmp Mica Stp Law sph ab Qz
52 - Cam Chl Mica Stp Law Qz Rt
53 - Cam Chl Mica Mica Stp Law Qz Rt
54 - Chl Pmp Cpx Ep Bt Stp sph ab Qz
55 - Cam Chl Pmp Cpx Ep Mica Stp sph ab Qz
56 - Chl Cpx Ep Bt Stp sph ab Qz
57 - Chl Cpx Ep Bt Mica Stp sph ab Qz
58 - Cam Chl Cpx Mica Stp Law sph Qz
59 - Cam Chl Mica Stp Law sph Qz Rt
60 - Chl Cpx Ep Bt sph ab Qz
61 - Chl Cpx Ep Mica sph ab Qz
62 - Cam Chl Cpx Ep Mica sph ab Qz
63 - Cam Chl Ep Bt sph ab Qz
64 - Cam Chl Cpx Ep Grt Mica Law sph Qz
65 - Cam Chl Cpx Ep Bt sph ab Qz
66 - Cam Chl Ep Grt Mica sph ab Qz
67 - Cam Chl Cpx Grt Mica Law sph Qz
68 - Cam Chl Ep Grt Mica Mica sph Qz
69 - Cam Chl Ep Grt Mica Law sph Qz
70 - Cam Chl Ep Bt sph ab Qz acti
71 - Cam Chl Ep Bt Mica sph ab Qz
72 - Cam Chl Cpx Ep Grt Mica sph ab Qz
73 - Cam Chl Grt Mica Law sph Qz
74 - Cam Chl Ep Grt Bt Mica sph ab Qz
75 - Fsp Cam Chl Ep Bt sph ab Qz
76 - Cam Cam Chl Ep Bt sph ab Qz
77 - Cam Chl Grt Mica Law sph Qz Rt
78 - Fsp Cam Cam Chl Ep Bt sph Qz
79 - Fsp Cam Chl Ilm Bt sph Qz
80 - Fsp Cam Chl Ep Bt sph Qz
81 - Cam Chl Ep Grt Bt sph ab Qz
82 - Cam Chl Ep Grt Mica Mica Law sph Qz
83 - Fsp Cam Cam Chl Ilm Bt Qz
84 - Fsp Cam Chl Ep Ilm Bt sph Qz
85 - Fsp Cam Chl Ilm Bt Qz
86 - Cam Cam Chl Ep Grt Bt sph ab Qz
87 - Fsp Cam Chl Ilm Bt Sp Qz
88 - Fsp Cam Chl Ep Grt Bt sph Qz
89 - Fsp Cam Cam Chl Ep Grt Bt sph Qz
90 - Fsp Cam Chl Ep Grt Bt sph ab Qz
91 - Cam Chl Ep Grt Bt Mica sph Qz
92 - Cam Chl Ep Grt Mica Mica Law Qz Rt
93 - Cam Chl Ep Grt Mica Law sph Qz Rt
94 - Fsp Cam Chl Ep Ilm Grt Bt sph Qz
95 - Cam Chl Ep Grt Mica Mica Qz Rt
96 - Cam Chl Ep Grt Mica Law Qz Rt

97 - Fsp Cam Cam Chl Ilm Bt Sp Qz
98 - Fsp Cam Chl Ilm Grt Bt Qz
99 - Fsp Cam Chl Ep Ilm Grt Bt Qz
100 - Fsp Cam Cam Chl Ilm Grt Bt Sp Qz
101 - Fsp Cam Chl Ep Grt Bt sph Qz Rt
102 - Cam Chl Grt Bt Mica sph ab Qz
103 - Cam Cam Chl Grt Bt Mica sph Qz
104 - Cam Cam Chl Ep Grt Bt Mica sph Qz
105 - Cam Chl Ep Grt Mica Mica zo sph Qz
106 - Cam Chl Ep Grt Mica Mica zo Qz Rt
107 - Fsp Cam Chl Ilm Grt Bt Sp Qz
108 - Fsp Cam Chl Ilm Grt Bt Mica Qz
109 - Fsp Cam Chl Grt Bt Mica Qz Rt
110 - Cam Chl Grt Bt Mica ab Qz Rt
111 - Cam Grt Bt Mica sph ab Qz Rt
112 - Cam Grt Bt Mica sph Qz
113 - Cam Cam Chl Grt Bt Mica Mica sph Qz
114 - Fsp Cam Cam Chl Ilm Grt Bt Qz
115 - Cam Grt Bt Mica ab Qz Rt
116 - Cam Grt Bt Mica Mica Qz Rt
117 - Cam Cam Grt Bt Mica Mica Qz Rt
118 - Cam Cam Chl Grt Mica Mica Qz Rt
119 - Fsp Cam Chl Grt Bt Mica Mica Qz Rt
120 - Fsp Cam Grt Bt Mica Qz Rt
121 - Fsp Cam Grt Bt Mica ab Qz Rt
122 - Cam Cam Chl Grt Mica Mica Mica Qz
123 - Cam Chl Ep Grt Mica Mica Mica zo Qz
124 - Fsp Cam Ilm Grt Bt Sp Qz
125 - Fsp Cam Grt Bt Mica Mica Qz Rt
126 - Cam Cam Grt Bt Mica Mica Qz Rt
127 - Cam Cam Grt Mica Mica Mica Qz
128 - Cam Ep Grt Mica Mica Mica Qz
129 - Cam Ep Grt Mica Mica Mica Qz Rt
130 - Cam Ep Grt Mica Mica Qz Rt
131 - Cam Cam Cpx Grt Mica Mica Mica Qz
132 - Cam Cpx Grt Mica Mica Mica Qz
133 - Cam Cpx Grt Mica Mica Mica Qz Rt
134 - Cam Cpx Ep Grt Mica Mica zo Qz Rt
135 - Cam Cpx Ep Grt Mica Mica Qz Rt
136 - Cam Grt Mica Mica Qz Rt
137 - Cam Cam Cpx Grt Mica Mica Qz Rt
138 - Cam Cpx Grt Mica Mica Qz Rt
139 - Fsp Cam Chl Ilm Grt Bt Mica Mica Qz
140 - Fsp Cam Cam Ilm Grt Bt Qz
141 - Fsp Cam Ilm Grt Bt Mica Mica Qz Rt
142 - Fsp Cam Grt Mica Mica Qz Rt
143 - Fsp Cam Cam Ilm Grt Bt Sp Qz
144 - Fsp Cam Ilm Grt Bt Mica Qz
145 - Fsp Cam Grt Mica Qz Rt
146 - Fsp Cam Ilm Grt Bt Qz
147 - Fsp Cam Ilm Grt Bt Mica Qz Rt
148 - Fsp Cam Grt Bt Qz Rt
149 - Fsp Cam Cpx Grt Mica Qz Rt
150 - Fsp Cam Cpx Grt Mica Mica Qz Rt
151 - Chl Mica Stp sph rieab ab mic Qz lmt
152 - Chl Cpx Mica Mica Stp sph ab Qz lmt
153 - Chl Cpx Mica Mica Stp Law sph ab Qz
154 - Cam Cpx Mica Stp Law sph Qz
155 - Chl Grt Mica Stp sph ab mic Qz lmt
156 - Chl Ep Grt Mica Stp sph ab Qz lmt
157 - Chl Mica Stp Law sph rieab ab Qz
158 - Cam Cpx Mica Mica Stp Law sph Qz Rt
159 - Chl Ep Grt Mica Stp Law sph ab Qz
160 - Cam Chl Cpx Mica Stp Law sph rieab ab Qz
161 - Chl Pmp Ep Bt Stp sph ab mic Qz
162 - Cam Chl Pmp Cpx Mica Stp Law sph ab Qz
163 - Chl Ep Bt sph pre ab Qz
164 - Chl Ep Bt Stp sph pre ab Qz acti
165 - Chl Pmp Cpx Mica Stp sph ab Qz
166 - Chl Cpx Ep Bt Stp sph ab Qz acti
167 - Cam Chl Pmp Ep Mica Stp sph ab Qz
168 - Chl Cpx Ep Mica Stp sph ab Qz
169 - Cam Chl Cpx Ep Mica sph ab Qz
170 - Cam Chl Pmp Cpx Ep Mica sph ab Qz
171 - Cam Chl Ep Grt Mica Law sph ab Qz
172 - Cam Chl Ep Grt Mica Mica sph ab Qz
173 - Cam Chl Grt Mica Stp Law sph Qz
174 - Fsp Cam Chl Ep Bt Sp sph Qz
175 - Cam Chl Ep Grt Bt Mica Mica sph Qz
176 - Fsp Cam Chl Ep Ilm Bt Qz
177 - Fsp Cam Chl Ilm Grt Bt Mica sph Qz

178 - Fsp Cam Chl Grt Bt Mica sph Qz
179 - Fsp Cam Chl Grt Bt Mica sph Qz Rt
180 - Fsp Cam Chl Grt Bt Mica ab Qz
181 - Cam Grt Bt Mica sph ab Qz
182 - Cam Cam Chl Grt Mica Mica sph Qz
183 - Cam Grt Bt Mica Mica sph Qz
184 - Cam Cam Grt Bt Mica Mica sph Qz Rt
185 - Cam Grt Bt Mica Mica sph Qz Rt
186 - Cam Cam Chl Ep Grt Mica Mica sph Qz
187 - Cam Cam Chl Grt Bt Mica Mica Qz Rt
188 - Cam Grt Bt Mica Mica ab Qz Rt
189 - Cam Cam Chl Ep Grt Mica Mica Mica Qz
190 - Cam Cam Ep Grt Mica Mica Mica Qz
191 - Cam Chl Ep Grt Mica Mica Mica Qz
192 - Fsp Cam Chl Ilm Grt Bt Mica Mica Qz Rt
193 - Fsp Cam Chl Ilm Grt Bt Mica Qz Rt
194 - Cam Cam Ep Grt Mica Mica Mica zo Qz
195 - Cam Ep Grt Mica Mica Mica zo Qz
196 - Fsp Cam Ilm Grt Bt Qz Rt
197 - Chl Ep Grt Stp sph ab mic Qz lmt
198 - Chl Pmp Grt Mica Stp sph ab Qz lmt
199 - Chl Pmp Ep Grt Mica Stp sph ab Qz
200 - Chl Ep Grt sph pre ab mic Qz acti
201 - Chl Ep Grt Bt sph pre ab mic Qz
202 - Chl Ep Grt Bt Stp sph ab Qz acti
203 - Cam Chl Pmp Mica Stp sph ab Qz
204 - Cam Chl Mica Mica Stp Law sph Qz Rt
205 - Chl Pmp Ep Bt Stp sph ab Qz acti
206 - Chl Cpx Ep Bt sph ab Qz acti
207 - Chl Cpx Ep Bt Mica sph ab Qz
208 - Cam Chl Cpx Ep Mica Stp Law sph ab Qz
209 - Cam Chl Cpx Ep Mica Law sph ab Qz
210 - Cam Chl Cpx Ep Bt Mica sph ab Qz
211 - Cam Chl Cpx Ep Grt Bt Mica sph ab Qz
212 - Cam Chl Grt Mica Stp Law Qz Rt
213 - Fsp Cam Cam Chl Ep Bt Sp sph Qz
214 - Cam Chl Ep Grt Bt Mica Mica sph ab Qz
215 - Cam Chl Ep Grt Mica Mica sph Qz Rt
216 - Fsp Cam Cam Chl Ep Grt Bt sph ab Qz
217 - Cam Chl Grt Mica Law Qz Rt
218 - Fsp Cam Chl Ep Ilm Grt Bt sph Qz Rt
219 - Fsp Cam Chl Ep Ilm Grt Bt Mica Qz
220 - Fsp Cam Chl Ep Grt Bt Mica sph Qz
221 - Fsp Cam Chl Grt Bt Mica sph ab Qz
222 - Cam Chl Grt Bt Mica Mica ab Qz
223 - Cam Chl Grt Bt Mica sph ab Qz Rt
224 - Cam Cam Grt Bt Mica sph Qz
225 - Cam Cam Grt Bt Mica Mica sph Qz
226 - Cam Grt Bt Mica Mica sph ab Qz Rt
227 - Cam Cam Chl Grt Bt Mica Mica Qz Rt
228 - Cam Cam Grt Mica Mica Mica Qz Rt
229 - Cam Grt Mica Mica Mica zo Qz
230 - Cam Cam Grt Mica Mica Mica zo Qz
231 - Cam Ep Grt Mica Mica Mica zo Qz Rt
232 - Cam Ep Grt Mica Mica zo Qz Rt
233 - Cam Cam Cpx Grt Mica Mica Qz Rt
234 - Fsp Cam Ilm Grt Bt Mica Mica Qz
235 - Chl Grt Mica Stp sph rieab ab mic Qz lmt
236 - Chl Pmp Ep Mica Stp Law sph ab Qz
237 - Chl Pmp Ep Grt Bt Mica sph ab Qz
238 - Cam Mica Mica Stp Law sph Qz Rt
239 - Chl Ep Bt sph pre ab mic Qz acti
240 - Chl Pmp Ep Grt Bt Mica Stp sph ab Qz
241 - Chl Pmp Cpx Ep Bt Mica Stp sph ab Qz
242 - Cam Chl Pmp Ep Mica Stp Law sph ab Qz
243 - Cam Chl Cpx Ep Grt Mica Law sph ab Qz
244 - Cam Chl Cpx Grt Mica Stp Law sph Qz
245 - Cam Cam Chl Cpx Ep Bt sph ab Qz
246 - Cam Chl Grt Mica Stp Law sph Qz Rt
247 - Fsp Cam Cam Chl Ep Bt sph ab Qz
248 - Cam Cam Chl Grt Bt Mica sph ab Qz
249 - Fsp Cam Chl Ep Ilm Grt Bt Mica sph Qz
250 - Fsp Cam Chl Grt Bt Mica Mica ab Qz
251 - Cam Chl Grt Bt Mica Mica sph ab Qz
252 - Fsp Cam Chl Grt Bt Mica ab Qz Rt
253 - Cam Chl Grt Bt Mica Mica sph ab Qz Rt
254 - Cam Chl Grt Bt Mica sph Qz
255 - Cam Cam Chl Ep Grt Bt Mica Mica sph Qz
256 - Cam Cam Ep Grt Bt Mica sph Qz
257 - Cam Cam Chl Ep Grt Mica Mica Qz Rt
258 - Cam Chl Ep Grt Mica Mica Mica Qz Rt
259 - Cam Cpx Ep Grt Mica Mica Mica Qz Rt

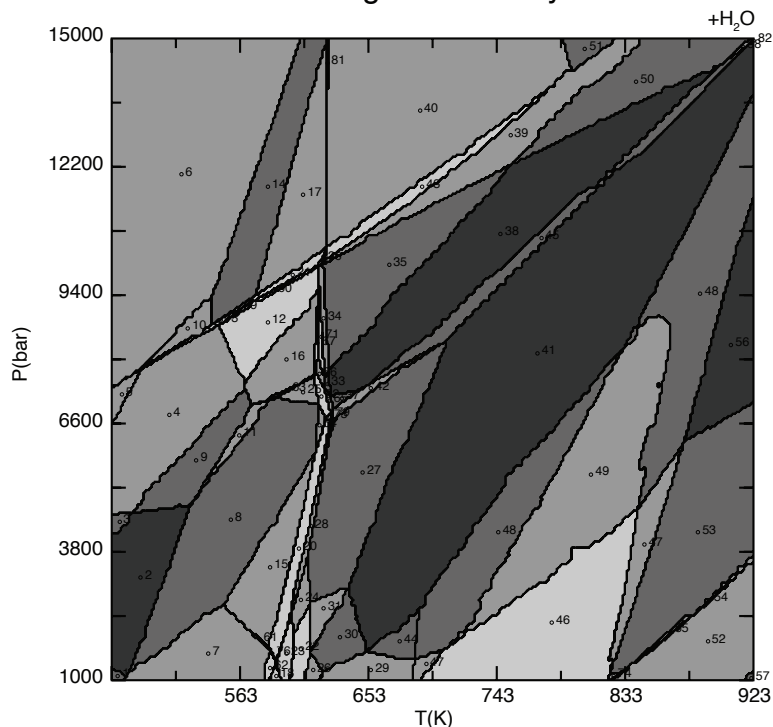
Fig S4. Pseudosection for blueschist layer of metapelite
23C04A1 blue layer metased using fixed activity of O₂ -424310 imposed from green layer



- 1 - Chl(HP) Mica(CHA) Stlp(M) ab mic q lmt
- 2 - Mica(CHA) Stlp(M) ab mic q lmt
- 3 - Mica(CHA) Stlp(M) law ab q
- 4 - Mica(CHA) Mica(CHA) Stlp(M) law ab q
- 5 - Omph(GHP) Mica(CHA) Mica(CHA) Stlp(M) law q
- 6 - Omph(GHP) Mica(CHA) Mica(CHA) Stlp(M) law q ru
- 7 - Omph(GHP) Mica(CHA) Stlp(M) law q ru
- 8 - Mica(CHA) Stlp(M) ab q lmt
- 9 - Chl(HP) Mica(CHA) Stlp(M) ab mic q wrk
- 10 - Chl(HP) Mica(CHA) Stlp(M) sph ab mic q lmt
- 11 - Chl(HP) Pu(M) Mica(CHA) Stlp(M) ab mic q
- 12 - Chl(HP) Pu(M) Mica(CHA) Stlp(M) ab q
- 13 - Chl(HP) Pu(M) Mica(CHA) Stlp(M) law ab q
- 14 - cAmph(DP) Mica(CHA) Mica(CHA) Stlp(M) law ab q
- 15 - cAmph(DP) Mica(CHA) Mica(CHA) Stlp(M) law q
- 16 - cAmph(DP) Omph(GHP) Mica(CHA) Mica(CHA) Stlp(M) law q
- 17 - Chl(HP) Ep(HP) Mica(CHA) Stlp(M) ab mic q
- 18 - Chl(HP) Pu(M) Bio(WPH) Mica(CHA) Stlp(M) ab q
- 19 - Chl(HP) Mica(CHA) Stlp(M) law ab q
- 20 - cAmph(DP) Mica(CHA) Stlp(M) law ab q
- 21 - cAmph(DP) Mica(CHA) Stlp(M) law q
- 22 - Chl(HP) Mica(CHA) ab mic q wrk
- 23 - Chl(HP) Ep(HP) Mica(CHA) ab mic q
- 24 - Chl(HP) Ep(HP) Bio(WPH) Mica(CHA) Stlp(M) ab q
- 25 - cAmph(DP) Chl(HP) Mica(CHA) Stlp(M) law ab q
- 26 - cAmph(DP) Chl(HP) Pu(M) Bio(WPH) Mica(CHA) Stlp(M) ab q
- 27 - cAmph(DP) Chl(HP) Pu(M) Mica(CHA) Stlp(M) ab q
- 28 - cAmph(DP) Chl(HP) Mica(CHA) Stlp(M) law q
- 29 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Mica(CHA) Stlp(M) ab q
- 30 - cAmph(DP) Chl(HP) Pu(M) Mica(CHA) Stlp(M) law q
- 31 - Chl(HP) Ep(HP) Bio(WPH) Mica(CHA) ab mic q
- 32 - cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) Mica(CHA) Stlp(M) ab q
- 33 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Mica(CHA) Stlp(M) q
- 34 - cAmph(DP) Chl(HP) Pu(M) Ep(HP) Mica(CHA) Stlp(M) law q
- 35 - cAmph(DP) Chl(HP) Omph(GHP) Mica(CHA) Stlp(M) law q
- 36 - Chl(HP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) ab q
- 37 - cAmph(DP) Chl(HP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) ab q
- 38 - cAmph(DP) Chl(HP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) q
- 39 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Gt(WPPH) Mica(CHA) q
- 40 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) q
- 41 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Gt(WPPH) Mica(CHA) law q

- 42 - cAmph(DP) Chl(HP) Omph(GHP) Gt(WPPH) Mica(CHA) law q
- 43 - cAmph(DP) Chl(HP) Gt(WPPH) Mica(CHA) law q
- 44 - feldspar Chl(HP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) q
- 45 - cAmph(DP) Chl(HP) Gt(WPPH) Mica(CHA) Stlp(M) law q
- 46 - feldspar Chl(HP) Gt(WPPH) Bio(WPH) Mica(CHA) q
- 47 - feldspar Gt(WPPH) Bio(WPH) Mica(CHA) q
- 48 - feldspar Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) q
- 49 - feldspar feldspar Gt(WPPH) Bio(WPH) Mica(CHA) q
- 50 - cAmph(DP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) q
- 51 - cAmph(DP) Chl(HP) Omph(GHP) Gt(WPPH) Bio(WPH) Mica(CHA) q
- 52 - cAmph(DP) Chl(HP) Omph(GHP) Gt(WPPH) Mica(CHA) q
- 53 - Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) ab q
- 54 - Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) Mica(CHA) ab q
- 55 - cAmph(DP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) ab q
- 56 - cAmph(DP) Omph(GHP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) q
- 57 - cAmph(DP) Chl(HP) Gt(WPPH) Bio(WPH) Mica(CHA) q
- 58 - feldspar Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) ab q
- 59 - cAmph(DP) Omph(GHP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) ab q
- 60 - cAmph(DP) Omph(GHP) Gt(WPPH) Bio(WPH) Mica(CHA) q
- 61 - cAmph(DP) Gt(WPPH) Bio(WPH) Mica(CHA) q
- 62 - cAmph(DP) Omph(GHP) Gt(WPPH) Bio(WPH) Mica(CHA) ab q
- 63 - feldspar cAmph(DP) Omph(GHP) Gt(WPPH) Bio(WPH) Mica(CHA) q
- 64 - feldspar feldspar Gt(WPPH) Bio(WPH) Mica(CHA) hCrd q
- 65 - feldspar feldspar Ilm(WPH) Bio(WPH) hCrd q
- 66 - feldspar Omph(GHP) Gt(WPPH) Bio(WPH) Mica(CHA) q
- 67 - feldspar feldspar Ilm(WPH) Bio(WPH) hCrd Sp(WPC) q
- 68 - feldspar feldspar Ilm(WPH) Gt(WPPH) Bio(WPH) Mica(CHA) hCrd q
- 69 - feldspar feldspar Bio(WPH) hCrd Sp(WPC) q
- 70 - feldspar feldspar Ilm(WPH) Gt(WPPH) Bio(WPH) hCrd q
- 71 - feldspar feldspar Ilm(WPH) Gt(WPPH) Bio(WPH) hCrd Sp(WPC) q
- 72 - Pu(M) Mica(CHA) Stlp(M) ab q lmt
- 73 - Chl(HP) Ep(HP) Mica(CHA) Stlp(M) ab mic q lmt
- 74 - Chl(HP) Pu(M) Ep(HP) Mica(CHA) Stlp(M) ab mic q
- 75 - Chl(HP) Gt(WPPH) Bio(WPH) Mica(CHA) Stlp(M) ab q
- 76 - cAmph(DP) Chl(HP) Pu(M) Mica(CHA) Stlp(M) law ab q
- 77 - cAmph(DP) Chl(HP) Pu(M) Omph(GHP) Mica(CHA) Stlp(M) ab q
- 78 - cAmph(DP) Chl(HP) Pu(M) Mica(CHA) Stlp(M) q
- 79 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Mica(CHA) Stlp(M) law q
- 80 - feldspar cAmph(DP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) q
- 81 - feldspar feldspar Ilm(WPH) Bio(WPH) Mica(CHA) hCrd q
- 82 - Pu(M) Mica(CHA) Stlp(M) law ab q
- 83 - Chl(HP) Pu(M) Mica(CHA) Stlp(M) ab q lmt
- 84 - Chl(HP) Ep(HP) Mica(CHA) Stlp(M) ab mic q wrk
- 85 - Chl(HP) Ep(HP) Bio(WPH) Mica(CHA) Stlp(M) ab mic q
- 86 - Chl(HP) Gt(WPPH) Mica(CHA) Stlp(M) ab mic q
- 87 - Chl(HP) Pu(M) Gt(WPPH) Bio(WPH) Mica(CHA) Stlp(M) ab q
- 88 - Chl(HP) Ep(HP) Mica(CHA) ab mic q wrk
- 89 - Chl(HP) Pu(M) Ep(HP) Bio(WPH) Mica(CHA) Stlp(M) ab q
- 90 - cAmph(DP) Chl(HP) Pu(M) Ep(HP) Mica(CHA) Stlp(M) q
- 91 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Bio(WPH) Mica(CHA) Stlp(M) ab q
- 92 - cAmph(DP) Chl(HP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) Stlp(M) q
- 93 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Bio(WPH) Mica(CHA) Stlp(M) q
- 94 - cAmph(DP) Chl(HP) Omph(GHP) Ep(HP) Gt(WPPH) Mica(CHA) Stlp(M) q
- 95 - cAmph(DP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) Mica(CHA) ab q
- 96 - cAmph(DP) cAmph(DP) Omph(GHP) Gt(WPPH) Bio(WPH) Mica(CHA) ab q
- 97 - Omph(GHP) Mica(CHA) Mica(CHA) Stlp(M) law ab q
- 98 - Chl(HP) Pu(M) Mica(CHA) Stlp(M) sph ab mic q lmt
- 99 - Chl(HP) Pu(M) Gt(WPPH) Mica(CHA) Stlp(M) ab mic q
- 100 - Chl(HP) Gt(WPPH) Bio(WPH) Mica(CHA) Stlp(M) ab mic q
- 101 - Chl(HP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) Stlp(M) ab q
- 102 - cAmph(DP) Chl(HP) Pu(M) Ep(HP) Mica(CHA) Stlp(M) ab q
- 103 - cAmph(DP) Chl(HP) Pu(M) Omph(GHP) Ep(HP) Mica(CHA) Stlp(M) q
- 104 - cAmph(DP) Chl(HP) Pu(M) Omph(GHP) Mica(CHA) Stlp(M) law q
- 105 - cAmph(DP) Chl(HP) Ep(HP) Bio(WPH) Mica(CHA) Stlp(M) q
- 106 - feldspar Chl(HP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) ab q
- 107 - feldspar cAmph(DP) Omph(GHP) Ep(HP) Gt(WPPH) Bio(WPH) Mica(CHA) q

Fig S5. Pseudosection for metachert
21C10A1 metachert using fixed activity of O₂ -424310 imposed by green layer



- 1 - Mica Stp sph ab mic Qz
- 2 - Mica Stp sph ab Qz
- 3 - Mica Stp sph ab Qz Rt
- 4 - Mica Mica Stp sph ab Qz Rt
- 5 - Mica Mica Stp Law ab Qz Rt
- 6 - Cpx Mica Mica Stp Law Qz Rt
- 7 - Chl Mica Stp sph ab mic Qz
- 8 - Chl Mica Stp sph ab Qz
- 9 - Mica Mica Stp sph ab Qz
- 10 - Cpx Mica Mica Stp sph Qz Rt
- 11 - Chl Mica Mica Stp sph ab Qz
- 12 - Cam Mica Mica Stp sph ab Qz Rt
- 13 - Cpx Mica Mica Stp sph ab Qz Rt
- 14 - Cpx Mica Mica Stp Qz Rt
- 15 - Chl Bt Mica Stp sph ab Qz
- 16 - Cam Mica Mica Stp sph ab Qz
- 17 - Cam Cpx Mica Mica Stp Qz Rt
- 18 - Chl Grt Mica sph pxmn ab mic Qz
- 19 - Chl Grt Bt Mica sph pxmn ab mic Qz
- 20 - Chl Grt Bt Mica Stp sph ab Qz
- 21 - Cam Cpx Mica Mica Stp ab Qz Rt
- 22 - Chl Ilm Grt Bt Mica pxmn ab Qz
- 23 - Chl Grt Bt Mica sph pxmn ab Qz
- 24 - Chl Grt Bt Mica pxmn ab Qz
- 25 - Cam Chl Mica Stp sph ab Qz
- 26 - Ilm Grt Bt Mica pxmn ab Qz
- 27 - Chl Grt Bt Mica ab Qz
- 28 - Chl Grt Bt Mica Stp ab Qz
- 29 - Ilm Grt Bt Mica ab mic Qz
- 30 - Ilm Grt Bt Mica ab Qz
- 31 - Chl Ilm Grt Bt Mica ab Qz
- 32 - Cam Chl Grt Mica Stp sph ab Qz
- 33 - Cam Grt Mica Stp ab Qz

- 34 - Cam Grt Mica Stp ab Qz Rt
- 35 - Cam Grt Mica ab Qz Rt
- 36 - Cam Cpx Grt Mica Mica Stp ab Qz Rt
- 37 - Cam Chl Grt Mica ab Qz
- 38 - Cam Grt Mica ab Qz
- 39 - Cam Cpx Grt Mica ab Qz Rt
- 40 - Cam Cpx Grt Mica Mica Qz Rt
- 41 - Grt Bt Mica ab Qz
- 42 - Cam Chl Grt Bt Mica ab Qz
- 43 - Cam Cpx Grt Mica Mica ab Qz Rt
- 44 - Grt Bt Mica ab mic Qz
- 45 - Cam Grt Bt Mica ab Qz
- 46 - Fsp Fsp Ilm Grt Bt Mica ab Qz
- 47 - Fsp Ilm Grt Bt Mica ab Qz
- 48 - Fsp Grt Bt Mica ab Qz
- 49 - Fsp Fsp Grt Bt Mica ab Qz
- 50 - Cam Cpx Grt Mica ab Qz
- 51 - Cam Cpx Grt Mica Qz Rt
- 52 - Fsp Fsp Ilm Grt Bt hCrd Qz
- 53 - Fsp Ilm Grt Bt Mica Qz
- 54 - Fsp Fsp Ilm Grt Bt Mica hCrd Qz
- 55 - Fsp Ilm Grt Bt Mica hCrd Qz
- 56 - Fsp Grt Bt Mica Qz
- 57 - Fsp Fsp Ilm Grt hCrd Qz
- 58 - Cam Cpx Grt Bt Mica ab Qz
- 59 - Cpx Mica Mica Stp ab Qz Rt
- 60 - Cam Cpx Mica Mica Stp sph ab Qz Rt
- 61 - Chl Bt Mica Stp sph ab mic Qz
- 62 - Chl Grt Mica Stp sph ab mic Qz
- 63 - Cam Chl Mica Mica Stp sph ab Qz
- 64 - Chl Grt Mica Stp sph ab Qz
- 65 - Cam Chl Grt Mica Stp ab Qz
- 66 - Cam Grt Mica Stp sph ab Qz
- 67 - Cam Grt Mica Mica Stp sph ab Qz
- 68 - Cpx Mica Mica Stp Law ab Qz Rt
- 69 - Chl Grt Bt Mica Stp sph pxmn ab Qz
- 70 - Ilm Grt Bt Mica pxmn ab mic Qz
- 71 - Cam Grt Mica Stp sph ab Qz Rt
- 72 - Cam Cpx Grt Mica Stp ab Qz Rt
- 73 - Cam Chl Grt Bt Mica Stp sph ab Qz
- 74 - Fsp Fsp Ilm Grt Bt Mica Qz
- 75 - Chl Grt Mica Stp sph pxmn ab mic Qz
- 76 - Chl Grt Bt Mica Stp sph ab mic Qz
- 77 - Chl Ilm Grt Bt Mica sph pxmn ab Qz
- 78 - Chl Grt Bt Mica Stp pxmn ab Qz
- 79 - Cam Chl Grt Bt Mica Stp ab Qz
- 80 - Chl Ilm Grt Bt Mica Stp ab Qz
- 81 - Cam Cpx Grt Mica Mica Stp Qz Rt
- 82 - Cpx Grt Bt Mica ab Qz

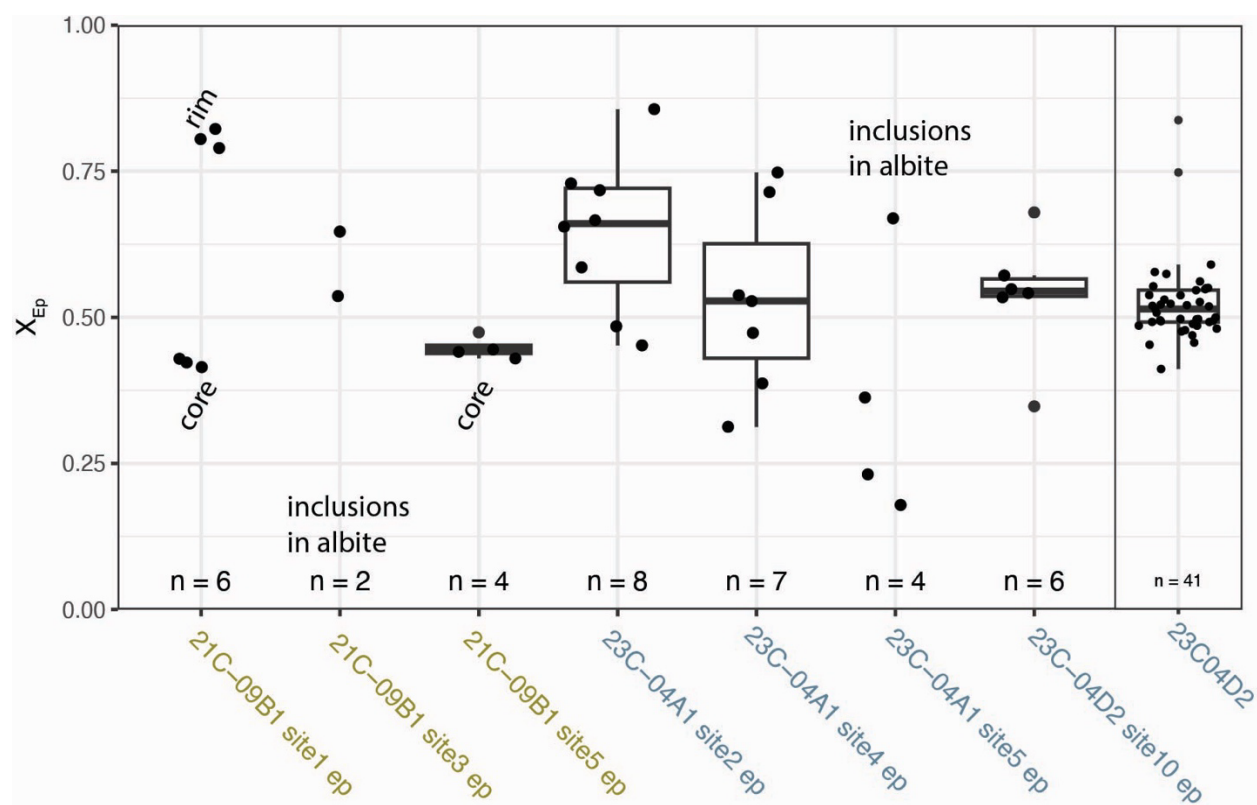


Figure S6. Plot of epidote compositions showing the spread of compositions from clinozoisite ($X_{Ep} = 0$) to epidote ($X_{Ep} = 1$).

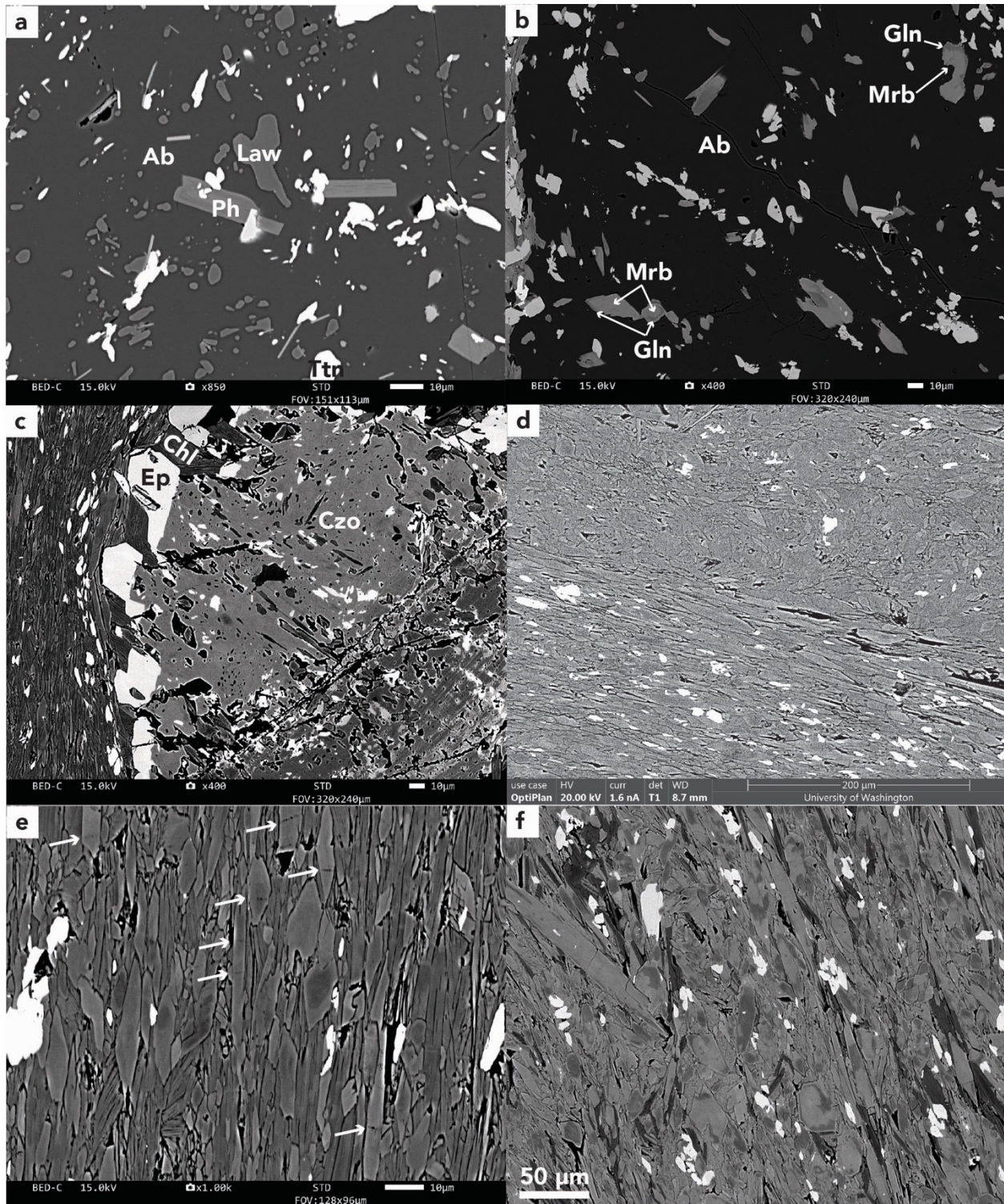
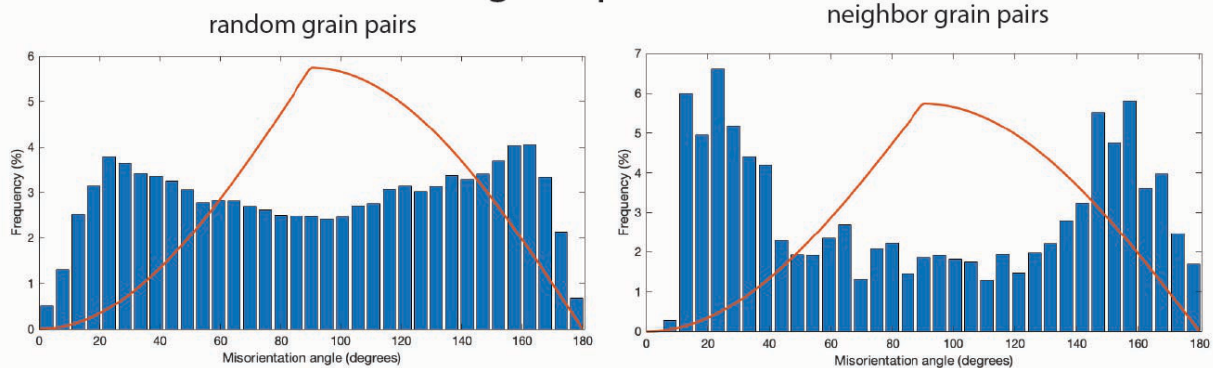


Figure S7. BSE images of important textures. (a) Lawsonite inclusion in albite porphyroblast in the greenschist metabasalt. (b) Inclusions in an albite porphyroblast in the blueschist metabasalt. Amphibole zoning here contains mostly, but not entirely, magnesioriebeckite cores (light) with glauconite rims (dark). (c) A clinozoisite porphyroblast with Fe-rich epidote rims occurring as euhedral growth interfingering with chlorite in the greenschist metabasalt. (d) An example of

textural contrast between foliated matrix actinolite and chlorite and randomly oriented actinolite and chlorite in the strain shadow of the epidote porphyroblast in (c). (e) Examples of incipient fracture textures in actinolite crystals in the greenschist metabasalt suggesting the activity of microboudinage. (f) A region of the blueschist metabasalt matrix showing zoning patterns in Na-amphiboles, with truncated zoning patterns providing evidence for the operation of dissolution precipitation creep.

21C09B1 Site 6 actinolite grain pairs



23C04D2 Site 7 glaucophane grain pairs

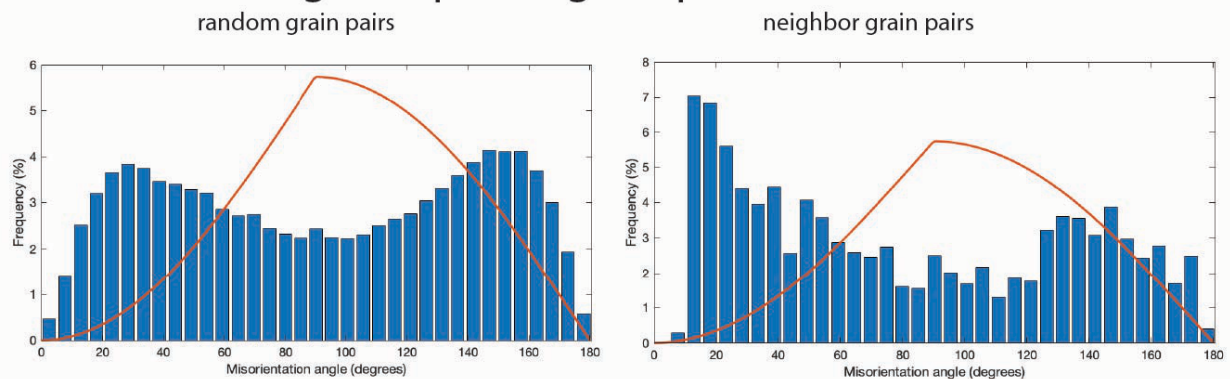


Fig S8. Histograms of amphibole grain pairs from additional analyzed regions in the metabasalt samples. The kick-ups in frequency near 0° and 180° are suggestive of rigid body rotation.