Rheology of Metasedimentary Rocks at the Base of the Subduction Seismogenic Zone

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Abstract
The Arosa Zone, Central Alps, represents a Cretaceous to Paleogene fossil subduction interface from paleodepths of 10 to >35 km. Metasedimentary schists directly below a preserved seismogenic zone exhibit viscous (foliations, folds, and lineations) and brittle (crack-seal quartz ± calcite veins) deformation features. Si-in-phengite barometry and published temperature estimates yield deformation and metamorphism conditions of 0.8–0.9 GPa and 300–350°C for an albite-rich pelitic schist and 0.9 GPa and 420°C for a quartz and calcite-rich calcareous schist. Petrography, microprobe element maps, and electron backscattered diffraction data constrain the modal mineralogies, microtextures, and active deformation mechanisms of these schists. There is evidence of minor crystal plasticity of quartz in both rocks and calcite in the calcareous schist. However, the most prominent deformation microstructures formed during pressure solution creep of quartz and albite. Pressure solution creep was activated due to pinning by ubiquitous fine grained phengite. Rheologic modeling indicates that pressure solution of quartz and albite can accommodate tectonic strain rates at stresses of <30 MPa and is weaker than dislocation creep at these same conditions; results that are consistent with our microstructural observations. However, to accommodate slow slip strain rates by either pressure solution or dislocation creep, stresses would need to be >>100 MPa. Triple oxygen isotope data from veins in these rocks combined with published data suggest a metamorphic mafic source for these vein filling and pressure solution accommodating fluids, indicating that dehydration from subducting oceanic lithosphere likely provided the fluids that facilitated viscous deformation in these rocks.

Plain Language Summary
Subduction zone faults are locked at shallow depths and release built up energy as catastrophic megathrust earthquakes. At deeper depths along this fault, the rocks typically flow rather than breaking, deforming viscously. This study examines the way metamorphosed sedimentary rocks from an ancient fossil subduction fault deformed viscously just below what was once a locked seismogenic zone. Detailed analysis reveals that minerals like quartz and albite deformed viscously by a mechanism called pressure-solution creep, where minerals dissolve and reprecipitate in fluid. Chemical data from veins in these rocks suggests that the fluid that facilitated pressure solution creep came from metamorphic reactions in oceanic rocks that were also subducting during this viscous deformation. Combining our observations with known relationships of strength of rocks deforming by pressure solution creep allows us to bracket the stresses these rocks felt during deformation, and the speeds of deformation (strain rate) they facilitated. Together, these data suggest that slow tectonic creeping strain rates are facilitated by this deformation, but a phenomena called slow earthquakes are likely not the result of pressure solution creep in metamorphosed sedimentary rocks.

1. Introduction
Subduction plate boundary faults are seismogenic and host megathrust earthquakes to depths of 30–50 km, beyond which they slip aseismically (Heuret et al., 2011; Hyndman et al., 1997). The depth of the transition from seismogenic frictional slip to continuous viscous deformation influences societal hazards by controlling how far downdip earthquakes initiate and propagate, and therefore, how far landward ground shaking might occur (e.g., Hyndman, 2013). Because viscous deformation at the base of the seismogenic zone controls the state of stress and loads the seismogenic zone above it, constraining the rheology of this portion of the plate interface is an integral part of understanding deformation cycles at subduction plate boundaries. Modeling work has suggested that
viscous deformation and particularly pressure solution creep may be particularly important at these conditions (Noda, 2016).

There are limited constraints on the mechanics that occur at the transition from frictional to viscous deformation along the subduction plate boundary, including which lithologies host viscous deformation and exactly which mineral phases control the viscous rheology. Metasomatism of subducted sediments and oceanic lithosphere results in the development of new lithologies with rheologic properties that must be better described and understood (e.g., Bebout & Barton, 1993; French & Condit, 2019; Hirauchi et al., 2020). In addition to contributing to where and how the frictional to viscous transition occurs, constraining the rheology of viscous deformation along the plate interface has important implications for subduction geodynamics. For example, recent geodynamic modeling suggests that the subduction of weak rocks, particularly sedimentary lithologies, may result in significant increases in convergence rates (Behr & Becker, 2018).

The transitional region from seismogenic to aseismic slip also hosts enigmatic slip behaviors of episodic tremor and slip (ETS; Beroza & Ide, 2011; Bürgmann, 2018; Rogers & Dragert, 2003). Recent work proposes that viscous deformation may accommodate slow slip at these conditions (e.g., Hayman & Lavier, 2014; Ujiie et al., 2018), while other workers have suggested frictional mechanisms activated by high pore fluid pressure accommodate slow slip (e.g., French & Condit, 2019; Gao & Wang, 2017; Segall et al., 2010). To constrain the strength of the plate interface at the seismogenic to aseismic transition and begin to understand the mechanisms that may accommodate slow slip, we need better constraints on the lithologies that exist at the base of the seismogenic zone, the deformation mechanisms by which they deform, and the resulting rheology (Angiboust et al., 2015; Behr & Bürgmann, 2020; Kirkpatrick et al., 2021; Oncken et al., 2021). Additionally, this portion of the subduction plate interface is a zone of chemical change, transformation, fluid flow, and devolatilization (e.g., Bebout & Penniston-Dorland, 2016; Condit et al., 2020; Epstein et al., 2021), and these chemical transformations likely play an important role in slip mode at these conditions.

We present a case study of an ancient subduction plate boundary exhumed from the base of the seismogenic zone to constrain the lithologies that result from metamorphosed sediments, the mechanisms that accommodate viscous deformation of these lithologies, and the sources of fluids that may contribute to their deformation. We constrain the processes and conditions recorded by deformation of pelitic and calcareous schists from the South Penninic domain of the Central Alps, a well-studied exhumed subduction interface (e.g., Bachmann, Glodny, et al., 2009; Bachmann, Oncken, et al., 2009; French & Condit, 2019; Ioannidi et al., 2020; Jaeckel et al., 2018; Ring et al., 1989, 1990). To determine the sedimentary origins of the schists and the conditions of deformation, we use bulk rock and mineral geochemistry and Si-in phengite barometry, respectively. We use quantitative microstructural analyses to determine the active deformation mechanisms, the strength of these rocks at the subduction interface, and constrain the state of stress. Finally, we use published stable isotope data and newly collected triple oxygen isotope measurements of quartz veins that crosscut these metasedimentary rocks to constrain the source of fluids that assist deformation. Through the integration of these analyses, we establish that viscous deformation of metasedimentary units at the base of seismogenic zone is largely accommodated by diffusive mass transfer of both quartz and albite, which is promoted by pinning of these phases by phengite. While this viscous deformation can readily accommodate tectonic strain rates, it requires unrealistically high shear stresses along the plate interface to facilitate slow slip strain rates. We also find that the source of the fluids that assist deformation is likely the metamorphic dehydration of mafic lithologies (e.g., the subducting slab).

2. Geologic Setting

The plate boundary interface that records subduction of the Penninic lithosphere beneath the Austroalpine plate during the late Cretaceous and early Paleogene is now exhumed and exposed along a suture zone within the western and central Alps (Schmid et al., 2004). The hyperextended Penninic lithosphere and overlying Jurassic to Cretaceous oceanic sediments are preserved as the North, Middle, and South Penninic domains of the central Alps (Figure 1a; Bachmann, Glodny, et al., 2009; Manatschal & Nievertgelt, 1997). The subduction plate boundary shear zone between the Penninic groups and the overriding Austroalpine plate is recorded in the South Penninic (Bachmann, Glodny, et al., 2009; Bachmann, Oncken, et al., 2009), and the preserved shear zone ranges in thickness from tens of meters to 2,500 m wide (Figure 1a; Bachmann, Glodny, et al., 2009; Ring, 1992; Ring et al., 1990). The structures preserved in the South Penninic domain are interpreted to preserve prograde
subduction related deformation, metamorphism, and mass transfer (Bachmann, Oncken, et al., 2009; Jaeckel et al., 2018; Ioannidi et al., 2020; Ring et al., 1989, 1990). For instance, Rb-Sr geochronology on synkinematic micas from this domain and the overriding Austroalpine plate reveal that subduction related deformation and metamorphism occurred up to ∼50 Ma; little to no subsequent overprinting by viscous deformation has occurred (Bachmann, Glodny, et al., 2009).

The South Penninic domain is also often referred to as the Arosa Zone (Bachmann, Glodny, et al., 2009; Bachmann, Oncken, et al., 2009; French & Condit, 2019), and it exposes rocks deformed at conditions that range from diagenetic rocks in the northernmost exposure (<10 km paleodepths, ∼100°C) to epidote-amphibolite facies (∼430°C, >30 km paleodepths) in the southernmost exposure (Figure 1c; Bachmann, Oncken, et al., 2009; Jaeckel et al., 2018). Subduction-aged earthquakes are recorded by pseudotachylite found in the overriding Austroalpine plate within tens of meters the plate interface (Bachmann, Oncken, et al., 2009). The base of the subduction seismogenic zone is interpreted to be at the deepest extent of this pseudotachylite at approximate paleo-depths of 20 km and paleo-temperatures of 300°C. We focus our work at paleo-depths below this last published observation of pseudotachylite, at the base of the interpreted fossil seismogenic zone, where nearby constraints on paleo-temperatures estimate deformation occurred at a peak temperature of >300°C to ∼430°C (Jaeckel et al., 2018). Thermochronology from this region suggests that these rocks have not been above these temperatures since subduction related deformation ended during the mid-Eocene (Price et al., 2018).

Figure 1. (a) Simplified geologic map of the Central Alps after (Bachmann, Oncken, et al., 2009). (b) Geologic map of the study area AR1 and simplified cross-section. (c) Geologic map of study area AR2 and simplified cross-section of the plate interface and Arosa Zone. Map units in panels (b and c) from SwissMap 1:100k scale and our own mapping.
We conduct detailed analyses of the deformation fabrics, textures, and mechanisms of two metasedimentary units from the deepest portions of the South Penninic domain (Figures 1b and 1c). The thickness of the subduction interface ranges from ∼100 to 300 m at site AR1 near Piz Lunghin (Figure 1b), to ∼500 m thick at site AR2 near Piz Corvatsch (Figure 1c). At both localities, the interface is a lithologically diverse and highly deformed zone containing tabular bodies of foliated and vein-rich calcareous and pelitic schists, metabasites, marbles, and serpentinites (Figures 1b and 1c). At site AR1, the plate boundary contact with the Austroalpine domain is preserved as a rotated low angle NE-SW trending fault between schists and gneiss of the overriding Austroalpine plate and South Penninic rocks (Figure 1b). At this site, the latest stage of subduction-related deformation is estimated to have occurred in the range of 300°C–350°C and pressures of >0.7 GPa (Bachmann, Glodny, et al., 2009). At AR2, the plate boundary contact is preserved as a chlorite rich fault that gently dips to the east. The overriding plate is comprised of mylonitized phyllosilicate-rich orthogneiss, while the south Penninic domain contains calcareous and pelitic schists, serpentinites, chlorite schists, and metabasalts (Figure 1c). Work nearby has suggested that some of the subduction material in the Malenco area has been exhumed from deeper depths than the Austroalpine overriding plate (e.g., Ioannidi et al., 2020), however, such constraints are not present at AR2. Jaeckel et al. (2018) estimate the temperature of deformation to be 430°C ± 30°C using Raman spectroscopy on foliation-defining carbonaceous material from one of the same calcareous schist units that we investigate. They interpret this temperature to record conditions of peak metamorphism and by extension foliation development which is defined by the peak mineral assemblage. At each of the locations, we mapped and sampled metasedimentary units to determine the composition, deformation mechanisms, and rheology of metasedimentary rocks at the base of the subduction seismogenic zone. We focus our work on two metasedimentary lithologies: a foliated quartz- and albite-rich pelitic schist crosscut by quartz veins from site AR1 (Figures 1b and 2a–2c), and a foliated calcareous schist with folded and recrystallized calcite ± quartz veins (Figures 1c and 2d–2e) from site AR2.

3. Methods

We employ a suite of macro- and microstructural, textural, petrologic, and geochemical tools to evaluate the (a) conditions of deformation, (b) active deformation mechanisms and resulting rheology during subduction, and (c) sources of aqueous fluids that facilitate deformation. We describe these methods and the role of each in our analyses below.

3.1. Macro and Microstructural Analysis

To constrain the broad structural and field context of the metasedimentary schists, we conducted detailed field work and collected macrostructural measurements of the schists’ foliations, folds and veins networks. We collected a suite of samples from these schists that encompass the full structural and mineralogical expression that we observed in the field. We created petrographic thin sections from these samples to observe microstructures, mineralogies, and mineral textures using a petrographic microscope. These field observations and petrographic data create a framework and provide context for more detailed chemical and textural data detailed below.

3.2. WDS Maps and Mineral Compositions

We measured the compositions of foliation defining minerals to estimate the conditions of deformation and collected maps of mineral compositions to determine phase relationships and qualitative mineral zoning within the metasedimentary units. Silicate compositions and quantitative wavelength dispersive spectroscopy (WDS) maps were collected from the pelitic schist at the Rice University on the JEOL JXA 8530F Hyperprobe and from the calcareous schist at MIT on the JEOL JXA-8200. An accelerating voltage of 15 kV and a current of 20 nA were used for both acquisitions, and mineral compositions were collected with a focused beam for all phases except phengite and feldspars, for which we used defocused beam diameter of 10 μm. Quantitative WDS maps were collected over areas of 3,000 × 2,750 μm using a dwell time of 10 ms, step sizes of 2.5 μm, and a defocused beam of 5 μm. These maps were used to determine the bulk compositions and modal mineralogy of the schists. Modal mineralogy was calculated using ImageJ to stack several WDS mapped elements with false RGB color scheme and then Adobe Photoshop to count pixels for each. These 2D modal abundances were then used in combination with phase compositions to calculate the bulk rock compositions of our schists using the methodology detailed in Condit and Mahan (2018). We also collected micro X-ray fluorescence (XRF) data to document silica
concentrations (intensity) around vein margins. Details of this analytical procedure are in the Text S1 in Supporting Information S1.

3.3. EBSD Analysis

Electron backscattered diffraction (EBSD) data was collected on one representative sample of pelitic schist and one representative sample of calcareous schist to constrain mineral grain sizes, crystallographic preferred orientations, and any shape preferred orientations that might occur, and to aid in interpretation of deformation mechanisms active within these metasedimentary units. Fully automated large-area EBSD maps were analyzed using an Oxford Symmetry EBSD detector fitted on an FEI Apreo LoVac field emission gun scanning electron microscope at the University of California, San Diego. Operating conditions were as follows: working distance between 26 and 28 mm, current of 26–51 nA, and accelerating voltage of 20 kV. A step size of 1 μm was used for data collection. The Oxford AzTec software was used to clean raw EBSD data by removing wild spikes, followed by iterative filling of orientations of non-indexed pixels down to five nearest neighbors. Typically, cleaning was minimal due to high indexing rates of albite, quartz, and calcite (90% or greater). White mica indexed quite poorly. EBSD data were then further processed using the open-source MTEX Matlab toolbox (Bachmann et al., 2010). Grains were defined as areas of the same phase with less than 10° misorientation. When cleaning the data in MTEX, we stripped out grains that comprised less than four pixels (<4 μm² area grains). Pole figures are plotted as orientation distribution functions after Hielscher and Schaeben (2008).

3.4. Triple Oxygen Isotope Geochemistry

We used laser fluorination to measure the triple oxygen isotope composition of six pure quartz vein samples that crosscut the pelitic schist (Hayles et al., 2018; Sharp et al., 2016; Yeung et al., 2018) to characterize and constrain the source of the fluids that precipitated veins and may have facilitated viscous deformation along the exhumed subduction interface. Details of the analytical procedures are provided in Text S2 in Supporting Information S1. Measurements reported in this study were collected over the same time period and using the same standards as reported in Yeung et al. (2018), and are in good agreement with values given by Valley et al. (1995) and other laboratories using VSMOW2-SLAP scaling.

4. Results

4.1. Pelitic Schist

4.1.1. Field Observations, Structural Measurements, and Petrologic Observations

We study a foliated and vein-bearing pelitic schist exposed as a tabular ~40 m wide body that trends NE-SW and outcrops ~60 m structurally below the contact between the Austroalpine plate and the South Penninic domain at site AR1 (Figure 1b). The pelitic schist is structurally above a calcareous schist/marble unit and structurally below a serpentinite unit (Figure 1b). The schist contains a modal mineral assemblage of albite (36.9%), white mica (33.0%), quartz (20.7%), and actinolite (4.9%), with minor stilpnomelane, epidote, titanite, and apatite (Figure 3a, Table 1). This mineralogy is consistent with the pressure and temperature constraints from previous work at site AR1 (e.g., Bachmann, Oncken, et al., 2009). The foliation of the schist is observable in outcrop, hand sample, and thin section and it strikes NE-SW and dips to the NW, sub-parallel to the plate boundary contact (Table S1 in Supporting Information S1, Figures 2b, 2c, and 3d). The foliation is crosscut by dominantly quartz crack seal veins (Figures 2a–2c). Three sets of crack seal quartz veins intersect the foliation at a high angle, and

Table 1
Modal Mineral Abundances and Bulk Rock Composition (in wt. % Oxides) of the Pelitic and Calcareous Schists

|                | Pelitic schist | Calc. schist |
|----------------|---------------|--------------|
| **Modal abundances** |               |              |
| Mineral        |               |              |
| Quartz         | 20.7          | 39.7         |
| Albite         | 36.9          | 2.9          |
| Actinolite     | 4.9           | 0.0          |
| Calcite        | 0.0           | 13.3         |
| Stilpnomelane  | 2.3           | 0.0          |
| Wt. mica       | 33.0          | 42.9         |
| Apatite        | 1.3           | 0.9          |
| Epidote        | 0.6           | 0.0          |
| Titanite       | 0.4           | 0.4          |
| **Bulk rock compositions (wt. % oxide)** | | |
| SiO₂           | 65.86         | 60.81        |
| TiO₂           | 0.58          | 0.23         |
| Al₂O₃          | 16.77         | 12.31        |
| FeO            | 3.63          | 1.24         |
| MgO            | 1.80          | 1.42         |
| MnO            | 0.05          | 0.07         |
| CaO            | 1.67          | 10.20        |
| Na₂O          | 4.19          | 0.38         |
| K₂O            | 3.65          | 4.24         |
| CO₂            | –             | 7.36         |
| H₂O            | 1.79          | 1.74         |
| **Total**      | 100.00        | 100.00       |
they are (a) conjugate arrays of rotated en echelon veinlets, (b) tabular veins with minor apparent shear offset, and (c) tabular veins with no shear offset (Figure S1 and Table S1 in Supporting Information S1). The three vein sets mutually crosscut one another at the macro- (Figures 2b and 2c) and microscales, demonstrating they likely formed contemporaneously. Based on vein orientations and shear sense on the en echelon vein sets, these veins are kinematically consistent with foliation and may have formed synchronously (Figure S1 and Table S1 in Supporting Information S1).

The foliation of the schist is defined by sub-mm (100s μm) spaced cleavage with cleavage domains of white mica ± amphibole and microlithon domains of albite + white mica ± amphibole and quartz + white mica ± actinolite, and occasional microlithons of pure quartz (Figures 2c and 3a). White mica and amphibole grains occur with their long axes sub-parallel to the compositional foliation. The majority of quartzofeldspathic domains are composed of fine-grained albite pinned by mica and amphibole grains, and quartz is a less dominant phase within these layers (Figures 3c and 3d). There are layered domains of fine-grained quartz pinned by white mica and amphibole; however, these domains are less volumetrically abundant than the albitic domains (Figure 3d, Table 2). In domains of albite or quartz with white mica and amphibole, albite and quartz grains are tabular and are well mixed with the white mica and actinolite grains (Figures 3e and 3d).

The domains of pure quartz are lenticular and ∼100 μm thick with grains 20–50 μm in diameter. These quartz grains exhibit dynamic recrystallization microstructures such as subgrain formation and grain boundary migration (Stipp et al., 2002; Figure 3b).

### 4.1.2. EBSD Data

In the pelitic schist, albite, quartz, amphibole (actinolite), apatite, and titanite all indexed at a high rate in the EBSD data (Figure 4a). Unsurprisingly, we were not able to index the white mica, which is common due to the notorious difficulty indexing micas in EBSD (Inoue & Kogure, 2012). Within this rock, the average grain sizes...
of albite and quartz grains are both $\sim$15 μm (Figure 4d). Broadly, both quartz and albite show relatively low intragranular misorientation, with albite showing almost no intracrystalline deformation features (Figure 4c). However, within the layers of nearly pure quartz, quartz grains do exhibit relatively high degrees of intragranular misorientation (Figure 4b).

Quartz and albite both have extremely weak CPOs in the pelitic schist (Figures 4e and 4f). The $J$ and $M$ indices are measures of CPO strength, where higher values indicate stronger fabrics ($J$ ranging from 1 to infinity, and $M$ ranging from 0 to 1; Bunge, 1982; Skemer et al., 2005). Quartz has $J$ and $M$ indices of 1.08 and 0.0082, respectively, whereas albite has $J$ and $M$ indices of 1.187 and 0.0052, respectively. In quartz, there is a weak $c$-axis cross girdle and $a$-axis point maxima within the foliation plane (Figure 4e). The $c$-axis cross-girdle is inclined relative to the foliation and asymmetric with a stronger girdle branch in the direction of shear. These fabrics are consistent with crystal plastic deformation during bulging recrystallization (300°C–380°C) and the low-temperature range of subgrain rotation recrystallization ($\sim$400°C) (Passchier & Trouw, 2005; Stipp & Kunze, 2008; Stipp et al., 2002), and the asymmetric $c$-axis pattern in quartz has been correlated to relatively low shear strain ($\gamma \approx 4$) in deformation experiments (Bellan & Tullis, 1989; Heilbronner & Tullis, 2006). Albite characterized by a diffuse single $c$-axis cross girdle that is asymmetric and sub-parallel to foliation, and a diffuse $b$-axis point maximum at a high angle to foliation (Figure 4f). There is much less information on which to base an interpretation of albite CPO compared to quartz, particularly in low grade metapelites. These albite CPOs are not similar to those formed during magmatic or crystal plastic processes in better studied gabbroic systems (Satsukawa et al., 2013). Together, all these data indicate minor plasticity in both the quartz and albite, and in the quartz the CPO patterns are consistent with relatively small crystal plastic strain at greenschist facies conditions.

Figure 3. Microstructures of pelitic schist from site AR1. (a) Plane-polarized light photomicrograph showing the mineral assemblage and foliation microstructures. (b) Inset area from panel (a) dominated by quartz that exhibits dynamic recrystallization microstructures. (c) Cross-polarized light image of foliation showing the small grain size of albite, phengite, and quartz. (d) False color map of stacked x-ray maps (red colors = Al rich phases, green colors = Mg rich phase, and blue colors = Ca rich phases) showing phase relationships and mineral alignment that defines foliation.
4.1.3. Petrologic Constraints on Deformation Conditions at AR1

A combination of previous work and our own Si-in-phengite barometry indicate that the temperature and pressure of deformation at site AR1 are 300°C–350°C and 0.8–1.0 GPa, respectively. Deformation temperatures were constrained to be 300–350°C near AR1 by Handy et al. (1996) based on deformation microstructures and the broad range of minerals present. Mineral compositions from the microprobe listed in Table 2 indicate that amphiboles are actinolite, and white micas range from muscovite to phengite (Figure 5a). These temperature estimates are consistent with the foliation defining mineral assemblage of albite + quartz + actinolite + phengite that we document, and is also consistent with dominantly bulging microstructures within the coarse-grained pure quartz lenses (Stipp et al., 2002). Using these temperature estimates, we constrained the pressure at which foliation developed using Si-in-phengite barometeretry (Massonne & Schreyer, 1987). To apply this barometer, a specific buffering mineral assemblage is required, including potassium feldspar, which we lack. However, as long as there is a Mg-Fe phase in equilibrium with the phengite in the rock, in our case actinolite, the Si-in Phengite pressure can be considered a minimum (Massonne & Schreyer, 1987).

White mica occurs variably as rounded grains with diameters of ~300 μm, as foliation defining elongate grains ~50–100 μm wide and up to 300 μm long, and as fine grains within well mixed albite domains and quartz domains (Figure 3d). These white micas have different compositions (Table 2, Figure 5a); the large, rounded white

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**Table 2**

**Oxide wt. % of Minerals Within Pelitic and Calcareous Schists**

| Mineral          | White mica (Pelitic schist—17C-04) | Calc. schist—17c-03 | Feldspar | Actinolite | Epidote |
|------------------|------------------------------------|----------------------|----------|------------|---------|
|                  | Musc. core | Phg rim | Matrix Phg | S1 | S1-S2 | S2 | Matrix Ab | Matrix Ab | Vein Kfsp | Matrix Act |
| SiO₂             | 46.98      | 53.13   | 51.37      | 49.95 | 48.96 | 49.05 | 68.51 | 68.18 | 63.81 | 53.10 | 39.16 |
| TiO₂             | 0.07       | 0.13    | 0.07       | 0.15 | 0.14 | 0.16 | 0.00 | 0.00 | 0.00 | 0.03 | 0.06 |
| Al₂O₃            | 33.90      | 26.77   | 26.25      | 27.38 | 27.33 | 27.59 | 20.19 | 20.03 | 19.08 | 1.05 | 25.34 |
| Fe₂O₃            | nm         | nm      | nm         | nm | nm | nm | 0.00 | 0.00 | 0.02 | 0.00 | 11.24 |
| FeO              | 2.08       | 5.63    | 5.43       | 2.88 | 3.55 | 3.32 | 0.00 | 0.00 | 0.00 | 19.62 | 0.00 |
| MgO              | 1.04       | 2.75    | 2.87       | 3.30 | 3.95 | 3.41 | 0.00 | 0.00 | 0.00 | 10.45 | 0.01 |
| MnO              | 0.03       | 0.06    | 0.01       | 0.00 | 0.00 | 0.02 | 0.00 | 0.00 | 0.10 | 0.35 | 0.15 |
| CaO              | 0.00       | 0.01    | 0.00       | 0.02 | 0.01 | 0.00 | 0.15 | 0.12 | 0.00 | 11.70 | 21.9 |
| Na₂O             | 0.35       | 0.06    | 0.05       | 0.15 | 0.20 | 0.16 | 11.40 | 11.47 | 0.19 | 0.33 | 0.0 |
| K₂O              | 11.04      | 7.26    | 10.69      | 9.88 | 9.50 | 9.83 | 0.03 | 0.00 | 16.61 | 0.01 | 0.03 |
| Total            | 95.49      | 95.80   | 96.74      | 93.71 | 93.64 | 93.54 | 100.28 | 100.70 | 99.81 | 96.64 | 97.89 |

Cations p.f.u.*

| Si     | 3.13 | 3.49 | 3.43 | 3.38 | 3.33 | 3.34 | An     | 0.01 | 0.01 | 0.00 | Mg# | 0.48 |
| Ti     | 0.03 | 0.10 | 0.00 | 0.01 | 0.01 | 0.01 | Ab     | 0.99 | 0.99 | 0.02 |
| Al     | 2.66 | 2.07 | 2.06 | 2.19 | 2.19 | 2.21 | Or     | 0.00 | 0.01 | 0.98 |
| Fe     | 0.12 | 0.31 | 0.30 | 0.16 | 0.20 | 0.19 |
| Mg     | 0.10 | 0.27 | 0.29 | 0.33 | 0.40 | 0.40 |
| Mn     | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Ca     | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| Na     | 0.05 | 0.01 | 0.01 | 0.02 | 0.03 | 0.02 |
| K      | 0.94 | 0.61 | 0.91 | 0.85 | 0.83 | 0.85 |
| Mg#    | 0.47 | 0.47 | 0.48 | 0.67 | 0.66 | 0.65 |

Note: nm = not measured.

*Normalized to 12 O.
mica grains are dominantly muscovite, but have 10–20 μm-thick Si-rich white mica rims, while the white mica grains that define the foliation microlithons and those that occur as fine grains are phengite (Figure 3d). We interpret that muscovite grains are inherited and that rims and matrix-defining phengite grew contemporaneously during foliation development. Quantitative electron microprobe measurements of white mica show that the foliation defining phengite grains have a composition of 3.41–3.43 Si p.f.u and a Mg # of 0.46–0.48 (Table 2). In contrast,
the muscovite cores of rounded white mica porphyroblasts have 3.1–3.2 Si p.f.u and the surrounding Si-rich mica rims have 3.3–3.5 Si p.f.u and Mg # 0.46–0.47 (Table 2, Figure 5a). Using the compositions of matrix-defining phengite and the phengite rims that surround muscovite grains in the Si-in-phengite barometer of Massonne and Schreyer (1987), we estimate the pressure of deformation to be >0.8–1.0 GPa (Figure 5b), this is consistent with pressure estimates of >0.7 GPa from (Bachmann, Oncken, et al., 2009).

4.2. Calcareous Schist

4.2.1. Field Observations, Structural Measurements, and Petrologic Observations

We study a heterogenous outcrop of foliated calcareous to graphitic schists from site AR2. This body is ∼40 m thick; it outcrops above a marble unit and its upper contact obscured by quaternary deposits. This unit contains evidence of at least two phases of deformation, commensurate with a dominant foliation (S₁), foliation parallel folded veins (F₁), and subsequent small scale upright crenulation folds (F₂) and transposition in places of S₁ into a new S₂. The dominant foliation (S₂) is defined by mm-scale alternating layers of quartz, calcite, and micas. This foliation strikes SSW and gently dips to the NW. The schists contain folded and transposed calcite ± quartz veins that are subparallel to foliation and in many cases have been ductilely deformed, forming foliation-parallel folded veins (F₁; Figures 2c–2e). Veins are composed of pure calcite or a combination of quartz and calcite, and range in thickness from a few mm to several cm. Carbon and oxygen isotope data from these veins are reported by Jaeckle et al. (2018), and based on their temperature estimates of ∼430°C for interpreted peak foliation development, formed from a fluid with a relatively heavy δ¹⁸O composition of ∼9–11‰. The S₁ foliation is variably and weakly folded by upright crenulation folds (F₂). Where these folds occur, they are spaced at the dm- to cm-scale (Figure 2e) and occur orthogonal to the foliation with hinge lines of 4° → 284° and an axial plain of 279°, 76°. In the hinge areas of these F₂ folds, S₁ is transposed to define a nascent S₂ foliation surface (Figures 2e and 6a).

The sample we focused on for detailed microstructural and compositional analyses contains a modal mineral assemblage of phengite (42.9%), quartz (39.7%), and calcite (13.3%) with minor albite, apatite, and titanite (Table 1). We focused on this sample because mineralogically, it is representative of the calcareous schist, and
it contains both $S_1$ foliation domains and minor $F_2$ folding (Figure 6a). The $S_1$ foliation is defined by mm-scale compositional banding with domains of phengite and quartz, domains of pure phengite, and domains of calcite and quartz (Figure 6). Domains of phengite and quartz are $\sim$300–500 $\mu$m thick, quartz grains within these domains are tabular in shape and $\sim$<30 $\mu$m in diameter, and quartz and phengitic grains are well-mixed. Quartz grains in these domains do not exhibit evidence of dynamic recrystallization. Domains of calcite and quartz are up to 1 mm thick (Figures 6a and 6c). Within these domains, calcite grains are $\sim$100s $\mu$m in diameter, exhibit some twinning, and have irregular to lobate grain boundaries indicative of bulging microstructures (Figures 6b and 6c). Quartz within these domains are coarser grained than in domains dominated by quartz and phengite, form embayments that are $\sim$100–200 $\mu$m across and are bounded on at least three sides by calcite; some of these quartz grains exhibit undulose extinction (Figures 6b and 6c). Layers of dominantly phengite are $\sim$300–500 $\mu$m
thick. Where the compositional foliation is folded into F₂ crenulation folds, phengitic mica is kinked and forms axial planar parallel grains (Figure 6a). As a result, phengite is transposed into a nascent S₂ orientation along the limbs of F₂ folds (e.g., Figure 6a).

4.2.2. EBSD Data

In the calcareous schist, quartz, calcite, and albite EBSD data all indexed well, but the phengite did not (Figure 7a). From the phase maps, we see qualitative spatial variations in both calcite and quartz grain sizes. Calcite-rich domains contain large calcite grains surrounded by smaller calcite grains that is consistent with dynamic recrystallization microstructures (Figure 7a). Within the calcite-rich domains quartz grains tend to be larger than they are in the quartz + phengite domains. This spatial distribution of grain size is reflected in a secondary peak at ~50 μm in the grain size distribution for calcite (Figure 7). We find that 80% of the calcite by volume has grain diameters greater 50 μm, and the mean grain diameter in this range is 95 μm (Figure 7b). These calcite grains exhibit intragranular misorientations that include evidence of twinning (Figure 7c), consistent with optical observations (Figure 6c). The fine-grained calcite (<50 μm), which represents 20% of the calcite by volume and occurs along the edges of coarser grained calcite, has a mean diameter of 17 μm (Figure 7c). Quartz grains larger than 50 μm comprise 20% of the total quartz by volume, and these have a mean grain size of 68 μm and are found within the calcite domains (Figure 7b). These quartz grains fill embayments along the edges of coarse-grained calcite and have some intragranular misorientation, which consistent with optical observations of undulous extinction (Figure 6c). In contrast with these coarse-grained calcite and quartz domains, 80% of the quartz by volume in the schist occurs as fine grains (<50 μm) within the quartz + phengite layers, which a mean grain size of 15 μm (Figure 7c). These grains exhibit very little intragranular misorientation (Figure 7).

The CPO of calcite is does not vary by grain size, and is relatively weak, with a J-index of 1.08 and an M-index of 0.011. The weak CPO that does occur is consistent with a c-axis point maximum near perpendicular to the S₁ foliation and an a-axis point maximum within the foliation plane and parallel to lineation (Figure 7g). These preferred orientations are consistent with grain boundary bulging, minor subgrain rotation, and new grain formation (e.g., Bestmann et al., 2006). Quartz exhibits an extremely weak CPO with a J index of 1.006 and an M index of 0.0002. Although very weak, the CPO pattern for quartz in the calcareous schist is similar to that of quartz in the pelitic schist and consists of a weak c-axis cross girdle and a-axis point maximum within the foliation plane (Figure 7h). Pole figures that distinguish coarse- and fine-grained quartz and calcite are presented in Supporting Information S1 and we see little difference in texture values or CPOs with grain size (Figure S2 in Supporting Information S1).

4.2.3. Petrologic Constraints on Deformation Conditions at Site AR2

A combination of previous work and our analyses indicate that the temperature and pressure of deformation at site AR2 are 410°C–450°C and 0.8–1.0 GPa, respectively. Jaeckel et al. (2018) determined the precipitation temperature of calcite to be 430°C ± 20°C using Raman spectroscopy on foliation-defining carbonateous materials and interpreted this as synchronous peak metamorphism and foliation development by phases defining that peak amorphous assemblage. The bulging and subgrain rotation microstructures that we observe within coarse-grained quartz occurring in the calcite + quartz domains indicate deformation occurred at temperatures between ~390°C and 420°C (Stipp et al., 2002), consistent with the previous temperature estimates of Jaeckel et al. (Figure 4b).

We therefore assume a temperature of 430°C and use Si-in-phengite barometry on phengite grains that define the primary S₁ foliation and those that form the transposed S₂ fabrics along the limbs of F₂ folds to estimate pressures during deformation. We also note, that for similar rocks in the nearby Dent Blanche region, CO₂ rich fluid infiltration resulted in some lower Si white mica growth (muscovite) even at elevated pressures (Epstein et al., 2021). Thus, our pressure estimates here should be seen as minimums.

All white mica grains in the calcareous schist, including those that define the S₁ and a nascent S₂ foliation, are phengitic in composition. Quantitative silicate compositions from the electron microprobe show that foliation-defining coarse-grained phengite have a core Si p. f.u. composition of 3.24, while rims of these elongate coarse grains have a Si p. f.u. of 3.38 (Table 2). Phengite that is kinked within F₂ folds and transposed into a S₂ orientation have a Si p. f.u. of 3.34 (Table 2). We find that the estimate of S₁ foliation development is >0.9 GPa and the estimate of S₂ foliation developments is >0.7 GPa (Figure 5b) using a temperature of 430°C for both. These temperature and pressure estimates are consistent with near peak conditions along the subduction P–T path estimated for the same region by Bousquet et al. (2008). The pressure estimates for phengite involved in
later minor folding \( (F_2) \) are lower than those determined for the dominant \( S_1 \) foliation. As a result, we propose that the formation of \( S_2 \) may have occurred during an early stage of exhumation, which is consistent with the NW vergence and kinematics of nascent \( F_2 \) folds.

**Figure 7.** Electron backscattered diffraction (EBSD) data from the calcareous schist at site AR2. (a) EBSD phase map showing domains of coarse-grained calcite and quartz, and fine-grained quartz and unindexed phengite. (b) EBSD phase map of the left half of Figure 7a showing only calcite and quartz grains that have a diameter greater than 50 µm. (c) EBSD phase map of the right half of Figure 7a showing only calcite and quartz grains that have a diameter less than 50 µm. (d) Histogram of grain size distribution for calcite and quartz (e) EBSD map of calcite-rich domain indicated in panel (a) showing the degree of intragranular misorientation. Blue arrows show twins (f) EBSD map of quartz-rich domain indicated in panel (a) showing the degree of intragranular misorientation. Scale bar is in degrees. (g) Crystallographic preferred orientation of all calcite grains plotted as one point per grain. (h) Crystallographic preferred orientation of all quartz grains plotted as one point per grain. Pole figures are plotted as equal area, lower hemisphere projections unless otherwise labeled.
4.3. Vein Description and Chemistry

Veins occur in both the pelitic schist and the calcareous schist. We focus on the veins within the pelitic schist, including textural descriptions, mineralogy, and triple oxygen isotopes to complement previously published textural and isotopic data collected on the calcite ± quartz veins from the calcareous schist (e.g., Jaeckel et al., 2018).

All vein sets that crosscut the pelitic schist are dominated by quartz with minor albite and late chlorite, and the en echelon veinlets also contain minor potassium feldspar. Microstructurally, the quartz grains in all vein sets are elongate with their long axes perpendicular to the vein walls (Figures 8a and 8b). The three vein sets exhibit typical crack-seal textures including margin-parallel inclusion bands, margin-perpendicular inclusion trails, and abundant fluid inclusions (Figures 8b and 8c). Vein quartz exhibits evidence of bulging (BLG) and incipient sub-grain rotation (SGR) dynamic recrystallization (Figure 8b), indicating post-precipitation deformation at temperatures of ~390°C–420°C (Stipp et al., 2002). The distance between individual inclusion bands is approximately 20–30 µm in all three types of veins. These observations of cross-cutting relations at the micro (e.g., Figure 8d) and macro scale (e.g., Figures 2a–2c), and crack-seal textures together indicate that these veins developed over many individual fracturing and precipitation events. Micro XRF maps of the veins and surrounding schist show a sharp decrease in Si content at the vein margin and constant values over distances up to several mm away from the vein in the foliation (Figure 8a).

We measured oxygen isotopes from two samples of en echelon veinlets, one shear vein, and three different mode I veins. The triple oxygen isotopic compositions of all six vein samples measured have δ¹⁸O values {ln (¹⁸O_vein / ¹⁶O_vein) / (ln (¹⁸O_NSMOW / ¹⁶O_NSMOW))} between 13.634‰ and 14.793‰ (mean of 14.016 ± 0.405‰; 1σ), and Δ¹⁷O values {δ¹⁷O − 0.5305 × δ¹⁸O} between −0.089‰ and −0.073‰ (mean of −0.073‰ ± 0.006‰; 1σ; Table 3, Figure 9). That the δ¹⁸O and Δ¹⁷O values of all vein sets are nearly identical suggests that they precipitated from the fluids that originated from same reservoir. We determine the isotopic composition of the fluid from which the quartz precipitation assuming a precipitation temperature between 350°C and 430°C, which is based on the quartz microstructures and published constraints on deformation temperatures (e.g., Bachmann, Oncken, et al., 2009). The inferred fluid composition has a δ¹⁸O value between 6.7‰ and 9.9‰ and a Δ¹⁷O value between −0.058‰ and −0.072‰ (Sharp et al., 2016) (Figure 9a, Table 3).

5. Discussion

5.1. Origin of the Metasedimentary Units

The bulk rock geochemistry and mineralogy of the metasedimentary units offers clues for the origins of these lithotypes, and how inherited sedimentary origins have important implications for the rheological behavior at the base of the subduction seismogenic zone. The pelitic schist at AR1 has a bulk composition (Table 1) similar to “typical” pelite (Shaw, 1956) often used in thermodynamic forward modeling papers as an approximation of these rocks (e.g., Baxter & Caddick, 2013; Condit et al., 2020). This bulk composition of the schist controls the occurrence and modal abundance of albite, quartz, and white mica, which has profound influence on the rheology of this rock (see Section 5.3), and the pelitic composition results in a larger modal abundance of albite and white mica at these P-T conditions over quartz as compared to a sandstone (psammite). Within the pelitic schist, there do not appear to be many preserved sedimentary features, however, the muscovite phenocrysts (Figure 3a) may be inherited detrital grains.

The calcareous schist found at site AR2 likely formed as a carbonaceous shale, however the extent to which the protolith of the rock was deposited with a significant carbonate component or was later modified is unclear. The folded and recrystallized nature of some of these veins is evidence that a significant portion of the calcite within the schist may have originated as calcite-rich veins that precipitated relatively early on the prograde path (Figures 2c and 2d). There are no obvious sedimentary features preserved within this rock, but it does appear that carbonate vein precipitation occurred before or during the major foliation forming S1 event, because these veins are parallel to subparallel to the foliation, and locally folded to define these surfaces (Figures 2c and 2d). We suggest that subsequent F2 upright crenulation folds may have occurred along the exhumation path based on the lower Si-content that indicates a lower pressure of crystallization in the phengite that defines the F2 axial planar (Figure 6).
The occurrence of pelitic and calcareous schists within exhumed subduction interfaces, while perhaps not ubiquitous, is quite common (e.g., Bebout & Penniston-Dorland, 2016). These schists themselves have been grouped within the Schists Lustrés type oceanic sediments that were deposited in the Alpine Tethys realm prior to subduction (Agard et al., 2000; Deville et al., 1992; Epstein et al., 2020; Trümpy, 2003). Similar deformation features have been observed in the calcareous Schists Lustrés in the western Alps (Angiboust et al., 2014, 2015; Epstein et al., 2021), and similar suites of metasedimentary pelites that have been found

Figure 8. Vein microstructures and Si-distribution in the pelitic schist at AR1. (a) Full thin section photomicrograph in XPL showing en echelon veins on the left and shear vein on the right cross cutting the foliation at a high angle. Plots show XRF derived Si normalized content in transects toward the veins. (b) XPL photomicrograph of vein margin and internal portion showing quartz bulging microstructures (orange arrows), albite, and inclusion bands (black arrows). (c) PPL photomicrograph showing inclusion bands (black arrows) and inclusion trains (white arrows) within quartz near a vein margin. (d) En echelon veinlet crosscutting a mode 1 vein. Note. The block textures within each vein.
in other exhumed subduction interfaces (Behr & Platt, 2012; Schmidt & Platt, 2022; Tewksbury-Christle et al., 2021), clearly indicating that the mineralogy, deformation mechanisms, and rheology of these rocks are important for the overall strength of the subduction interface (Behr & Becker, 2018) because of their common occurrences.

| SAMP. | VEIN TYPE | δ¹⁸O (%e) | δ¹⁷O (%e) | 300°C | 350°C | 400°C | 430°C |
|-------|-----------|-----------|-----------|-------|-------|-------|-------|
| MA-1  | Mode 1    | 13.634    | −0.087    | 6.344 | −0.062| 7.851 | −0.069|
|       | Shear     | 13.831    | −0.085    | 6.533 | −0.061| 8.040 | −0.067|
| MB-2  | En echelon| 13.825    | −0.089    | 6.533 | −0.061| 8.040 | −0.067|
| CB-S1 | Mode 1    | 14.793    | −0.082    | 7.456 | −0.060| 8.963 | −0.066|
|       | Shear     | 13.985    | −0.079    | 6.680 | −0.055| 8.188 | −0.061|
| CB-S1 | Mode 1    | 14.026    | −0.073    | 6.720 | −0.049| 8.227 | −0.056|
| Mean  |           | 14.016    | −0.082    | 6.711 | −0.058| 8.216 | −0.064|
| SD    |           | 0.405     | 0.006     | 0.388 | 0.005 | 0.388 | 0.005 |

Table 3: Oxygen Isotope Compositions and Calculated Fluid Compositions for a Range of Geologically Constrained Temperatures

Figure 9. Stable oxygen isotope data. (a) The triple oxygen isotope composition of the quartz veins (red circles). The oxygen isotope compositions for MORB glass, low temperature altered oceanic crust, and mantle minerals are shown for reference (Pack & Herwartz, 2014). The isotopic composition of the vein-forming fluid depends on the precipitation temperature and those values were calculated using the calibration of Sharp et al. (2016) and are shown as a light blue array. (b) Our vein δ¹⁸O measurements are in red and the calculated fluid composition based on temperature constrains is shown in blue. AR2 subduction fluids are from Jaeckle et al. (2018). Accretionary prism fluids are from Voelij et al. (1991) and bulk rock measurements are from Sharp (2007) and Cooperdock et al. (2018).
5.2. Active Deformation Mechanisms

We evaluate the deformation mechanisms that were active in the metasedimentary rocks and their roles in controlling rheology at the base of the subduction seismogenic zone. First, we use our microstructural observations and measurements to determine the deformation mechanisms that were active in the minerals that comprise the pelitic and calcareous schists. Then, we summarize what is known about the rheology of these deformation mechanisms from the rock deformation literature. Finally, we use the both the microstructures and known rheologic constraints to constrain the rate-controlling deformation mechanisms and overall viscous rheology of the schists.

5.2.1. Pelitic Schist

The microstructures that are preserved in the pelitic schist and define the foliation indicate that deformation was largely accommodated by diffusive mass transfer of both quartz and albite, with minor crystal plastic deformation accommodated by both phases. The spaced cleavage foliation is indicative of pressure solution creep, and develops during mass transfer accommodated by a fluid (Passchier & Trouw, 2005; Rutter, 1983; Stockhert, 2002). Because pressure solution creep is a grain size sensitive deformation mechanism, it will be favored in rocks with small grain sizes (Wheeler, 1992). Thus, the occurrence of very fine-grained albite and quartz (∼15 μm) pinned by phengite and actinolite within the microlithons is favorable for this mechanism (Figures 3a, 3d and 3c). Additionally, the very weak CPOs in both quartz and albite (Figures 4e and 4f) and the lack of intragranular misorientations within the albite (Figure 4c) indicate that crystal plastic deformation was minor in these phases. However, quartz seems to have experienced more crystal plasticity than albite, even though it was still not the primary strain accommodating mechanism. For example, where quartz occurs as monomineralic lenses there are clear bulging recrystallization microstructures (e.g., Figure 3b), and quartz grains also have higher degrees of intergranular misorientation than albite (Figures 4b and 4c). This suggests that albite did not deform by crystal plastic mechanisms, while quartz did. Furthermore, the weak CPO pattern is consistent with low temperature and low-strain crystal plastic deformation (Figure 4e; Passchier & Trouw, 2005).

5.2.2. Calcareous Schist

The microstructures preserved in the calcareous schist indicate that pressure solution creep was active in the fine-grained quartz and phengite domains, while minor crystal plastic deformation was active within the coarser grained calcite and quartz layers. Within the phengite and quartz domains, the occurrence of phengite sub-domains separated by fine grained (∼15 μm) tabular quartz grains that exhibit little evidence of crystal plasticity (Figures 7f and 7h), indicate that pressure solution creep was active and accommodating the formation and deformation of the S1 foliation (Stockhert, 2002; Wassmann & Stöckhert, 2013; Wheeler, 1992). Although quartz grains that occur as embayments into the coarse-grained calcite domains have undulous extinction and bulging and subgrain recrystallization microstructures (Figures 6b and 6c), the CPO of this quartz, which only accounts for 20% of the quartz by volume, is extremely weak (Figure 7h, Figure S2 Supporting Information S1). Together, these microstructural observations indicate that minor crystal plastic deformation was active in the coarse-grained quartz, but that quartz deformed primarily by pressure solution creep within the phengite and quartz domains. Calcite, which predominantly occurs as coarse grains (>50 μm; Figures 4a–4d), exhibits bulging microstructures, twinning, and higher degrees of intragranular misorientation indicative of crystal plastic deformation (Figure 7e). Nevertheless, calcite lacks a strong CPO and does not appear to have accommodated significant deformation in these zones. Together, the microstructural data indicate that the S1 foliation developed predominantly by pressure solution creep of quartz in phengite domains with minor dislocation creep and low-temperature plasticity accommodating deformation in the coarser grained calcite and quartz layers.

5.3. Rheology of Metasedimentary Units

Guided by the microstructural record of active deformation mechanisms, we now evaluate the constitutive relations for these minerals and rocks at the pressure and temperatures of in situ deformation. We start by synthesizing the laboratory-constrained flow laws for the deformation mechanisms active in the minerals of the pelitic and calcareous schists, and then evaluate how these may inform the more complex rheology of the polymineralic schists. We evaluate deformation of the pelitic schist at 900 MPa pressure, lithostatic pore fluid pressure of 900 MPa, and 325°C and deformation of the calcareous schist at 900 MPa pressure, lithostatic pore fluid pressure of 900 MPa, and 430°C.
Experimentally constrained flow laws are typically determined using an axially symmetric experimental geometry and present axial strain rate, \( \dot{\varepsilon} \), as a function of differential stress, \( \sigma_d \). Deformation along the subduction plate boundary is more closely approximated as simple shear and the parameters of interest are shear strain rate, \( \dot{\gamma} \), as a function of shear stress, \( \tau \). Accordingly, in our analysis we convert differential stress to an equivalent shear stress (\( \tau_e \)) using the relation \( \tau_e = \sigma_d / \sqrt{3} \) and convert axial strain rate to an equivalent shear strain rate using \( \dot{\gamma}_e = \sqrt{3} \dot{\varepsilon} \) (Paterson & Olgaard, 2000). We present experimentally determined flow laws where \( R \) is the gas constant, \( Q \) is activation energy, temperature, \( T \), is units of K, and grain diameter, \( d \), is in units of \( \mu \)m. Material parameters for the flow laws are provided in Table 4.

### 5.3.1. Single Mineral Rheology

When evaluating how individual deformation mechanisms contribute to the total deformation of a mineral, we assume that mechanisms operate concurrently unless experimental evidence indicates otherwise. Concurrent deformation occurs because mechanisms operate independent of one another such that the total strain rate is the sum of the strain rates of individual mechanisms, and the constitutive behavior is dominated by the fastest mechanism.

Quartz is abundant in both the pelitic and calcareous schists, and because we see evidence for dislocation creep and pressure solution creep in both lithologies, we consider these mechanisms for quartz. For dislocation creep, we use the constitutive law of (Hirth et al., 2001) given by

\[
\dot{\varepsilon}_{dd} = A_{dd} f_{dd} \sigma_d^4 \exp(-Q_{dd}/(RT))
\]

where \( f_{dd} \) is the water fugacity (Table 4). Tokle et al. (2019) demonstrate that the form of the dislocation creep flow law for quartz depends on stress magnitude, but it is unclear how these new laboratory results extrapolate to nature. The flow law of Hirth et al. (2001) that we use represents an average. For pressure solution, we use the thin film model (Rutter, 1976; Weyl, 1959) given by

\[
\dot{\varepsilon}_{df} = A_k V_w \rho_f c D w \sigma_d^4/(\rho_s RT d^3)
\]

\( A_k \) is a geometric constant determined by grain shape, \( w \) is the grain boundary width, and other material parameters are provided in Table 4. We use the geometric constant for tabular grains \( A_k = 32 \). Grain boundary widths have been reported between 1 and 100 nm in silicates with most values between 1 and 10 nm for monomineralic aggregates (Brady, 1983; Joesten, 1983; Ricoult & Kohlstedt, 1983; White & White, 1981). We use the thin-film model over other pressure solution models as it has been successful in modeling pressure solution of quartz along other paleo-subduction plate boundaries (e.g., Behr & Platt, 2013). Furthermore, phyllosilicates along grain boundaries are known to enhance diffusion relative to a single-phase aggregate by increasing the effective grain boundary width by approximately a factor of 30 (Hickman & Evans, 1995), and the thin film model can best describe the enhanced diffusion. Accordingly, we scale the effective width of a monomineralic aggregate (∼5 nm) to 150 nm for the phyllosilicate-bearing polyphase aggregate.

We evaluate dislocation creep and diffusion creep of albite at the deformation conditions of the pelitic schist only, as it is a minor component of the calcareous schist. We use the constitutive law of Offerhaus et al. (2001) as reported in Rybacki and Dresen (2004) to model dislocation creep, which is given by

\[
\dot{\varepsilon}_{ad} = A_{ad} \sigma_d^3 \exp\left(-\frac{Q_{ad}}{RT}\right)
\]

Water enhances grain boundary diffusion creep of albite (Rybacki & Dresen, 2000), and microstructural evidence indicates that as little as 1 wt% water can result in pressure solution creep of albite (Tullis & Yund, 1991). However, the only laboratory flow law for diffusion creep of wet albite was made with 0.2 wt% added water (Offerhaus et al., 2001), which may be insufficient to allow for dissolution and diffusion through grain boundary fluid films, as occurs during pressure solution. In addition, there are insufficient constraints on the grain boundary diffusivities, \( D \), of the diffusing chemical species in albite to employ the thin film pressure solution creep flow law of Weyl (1959) (Farver & Yund, 1995; M. Liu et al., 1997). Accordingly, we use the empirical flow law of Offerhaus et al. (2001) for diffusion creep of wet albite with 0.2 wt% water given by

\[
\dot{\varepsilon}_{apw} = A_{apw} d^{-3} \sigma_d \exp\left(-\frac{Q_{apw}}{RT}\right)
\]
### Table 4

**Material Constrains in Single Mineral Constitutive Relations**

| Rock          | Pelitic schist | Calcareous schist | Units | References                  |
|---------------|----------------|-------------------|-------|-----------------------------|
| Temperature   | 598            | 703               | K     |                             |
| **Quartz**    |                |                   |       |                             |
| $d$           | 15             | 15, 68            | μm    |                             |
| **Dislocation creep** |          |                   |       |                             |
| $A_{ql}$      | $10^{11.2}$    | $10^{11.2}$       | MPa$^{-5}$s$^{-1}$ | Hirth et al. (2001)         |
| $Q_{ql}$      | 135            | 135               | kJ/mol |                             |
| $\varphi_{H_2O}$ | 280          | 526               | MPa    | Pitzer and Sterner (1994)   |
| **Pressure solution creep** |        |                   |       |                             |
| $A_R$         | 32             |                   |       | Weyl (1959)                 |
| $V_m$         | $2.26 \times 10^{-5}$ | $2.26 \times 10^{-5}$ | m$^3$/mol | Frondel and Hurlbut (1955) |
| $\rho_f$      | 1.069          | 1.017             | kg/m$^3$ | Burnham et al. (1969)      |
| $\rho_s$      | 2.650          | 2.650             | kg/m$^3$ |                             |
| $c$           | $4.88 \times 10^{-4}$ | $1.10 \times 10^{-3}$ |       | Fournier and Potter (1982)  |
| $D$           | $2 \times 10^{-19}$ | $1.10 \times 10^{-17}$ | m/s    | Farver and Yund (2000)      |
| $w$           | 150            | 150               | nm     |                             |
| **Albite**    |                |                   |       |                             |
| $d$           | 15             |                   | μm    |                             |
| **Dislocation creep** |          |                   |       |                             |
| $A_{ad}$      | $10^{7.4}$     |                   | MPa$^{-5}$s$^{-1}$ | Offerhaus et al. (2001)     |
| $Q_{ad}$      | 332            |                   | kJ/mol | Offerhaus et al. (2001)     |
| **Diffusion creep** |        |                   |       |                             |
| $A_{aps}$     | $10^{1.9}$     |                   | MPa$^{-1}$μm$^3$s$^{-1}$ | Offerhaus et al. (2001)     |
| $Q_{aps}$     | 193            |                   | kJ/mol | Offerhaus et al. (2001)     |
| **Calcite**   |                |                   |       |                             |
| **High stress dislocation creep** |        |                   |       | Rutter (1974)               |
| $A_{dch}$     | $10^6$         |                   | s$^{-1}$ |                             |
| $Q_{dch}$     | 197            |                   | kJ/mol |                             |
| $C_{dch}$     | 0.035          |                   | MPa$^{-1}$ |                             |
| **Low stress dislocation creep** |        |                   |       | Renner et al. (2002)        |
| $A_{dcl}$     | $10^{0.5}$     |                   | MPa$^{-5}$s$^{-1}$ |                             |
| $Q_{dcl}$     | 200            |                   | kJ/mol |                             |
| $s_p$         | $S + K d^{-0.5}(T_m - T)$ |                   | MPa |                             |
| $S$           | 7.8            |                   | MPa kK$^{-1}$ |                             |
| $K$           | 115            |                   | MPa kK$^{-1}$ μm$^{0.5}$ |                             |
| $T_m$         | 1,600          |                   | K     |                             |
| **Grain boundary sliding** |        |                   |       | Walker et al. (1990)        |
| $A_{gbs}$     | $10^{10.9}$    |                   | MPa$^{-1}$s$^{-1}$μm$^{1.9}$ |                             |
| $Q_{gbs}$     | 190            |                   | kJ/mol |                             |
| **Diffusion creep** |        |                   |       | Herwegh et al. (2003)       |
| $A_{dif}$     | $10^{7.6}$     |                   | MPa$^{-1}$s$^{-1}$μm$^{1.36}$ |                             |
| $Q_{dif}$     | 200            |                   | kJ/mol |                             |
and acknowledge that this flow law represents a lower bound on the strain rates that may be accommodated at higher water contents during pressure solution creep. We evaluate the flow law as published and, in the case that deformation represents pressure solution creep, we also scale results by a factor of 30 to account for faster diffusion along phyllosilicates in the pelitic schist compared to grain boundaries in the experimental monomineralic rock (e.g., Hickman & Evans, 1995).

We evaluate the contributions of dislocation creep, grain boundary sliding, and diffusion creep to the deformation of calcite, which is abundant in the calcareous schist. At these conditions, pressure solution creep is extremely slow and negligible compared to the other deformation mechanisms (Gratier, 2011; Rutter, 1974), because of the very low solubility of calcite at these temperatures (Fein & Walther, 1987). Based on deformation experiments that show dislocation creep of calcite cannot be described by a single flow law at all conditions, we represent dislocation creep of calcite with the constitutive relation of Renner et al. (2002) at low differential stresses and that of Rutter (1974) at higher differential stresses. The high stress dislocation creep constitutive relation from Rutter (1974) has the form

$$\dot{\varepsilon}_{\text{dch}} = A_{\text{dch}} \exp\left(-\frac{Q_{\text{dch}}}{RT}\right) \exp\left(\alpha \sigma_d^2 \right)$$

(5)

and the low stress dislocation creep relation is

$$\dot{\varepsilon}_{\text{dcl}} = A_{\text{dcl}} \sigma_d^{-1.7} \exp\left(-\frac{Q_{\text{dcl}}}{RT}\right)$$

(6)

where the Peierls stress, $\sigma_p$, is given by

$$\sigma_p = S + K \cdot d^{-0.5} \cdot (T_m - T)$$

(7)

from Renner et al. (2002) (Table 4). For grain boundary sliding, we use the result of Walker et al. (1990):

$$\dot{\varepsilon}_{\text{gbs}} = A_{\text{gbs}} \sigma_d^{1.7} \sigma_d^{-1.9} \exp(-Q_{\text{gbs}}/RT)$$

(8)

Finally, for diffusion creep, we use the flow law of Herwegh et al. (2003) given by

$$\dot{\varepsilon}_{\text{dif}} = A_{\text{dif}} \sigma_d^{1.1} \sigma_d^{-3.26} \exp(-Q_{\text{dif}}/RT)$$

(9)

Phengite is the primary phyllosilicate phase in the pelitic and calcareous schists; however, there are no published flow laws for the strength of phengite. As a result, we approximate the rheology of phengite with that of muscovite, another white mica, because its deformation has been measured in the laboratory. Mica minerals experience dislocation glide within their basal planes at low shear stresses and temperatures (Christoffersen & Kronenberg, 1993; Kronenberg et al., 1990; Shea & Kronenberg, 1992), but constitutive relations for homogenous deformation by crystal plastic and/or diffusion processes have not been constrained for the micas. To approximate deformation of phengite, we use the dislocation glide flow law for muscovite from Mares and Kronenberg (1993), given by

$$\dot{\varepsilon}_{\text{mg}} = A_{\text{mg}} \exp(-Q_{\text{mg}}/RT) \exp(\alpha \sigma_d)$$

(10)

### 5.3.2. Pelitic Schist Rheology

At the deformation conditions of the pelitic schist, quartz deforms by pressure solution creep at shear stresses less than approximately 30 MPa and shear strain rates less than $3 \times 10^{-14}$ s$^{-1}$ (Figure 10a). At higher stresses and strain rates,
Figure 10. Equivalent shear stress as a function of equivalent strain rate for minerals in the pelitic schist. The constitutive relations for individual mechanisms are shown along with the overall constitutive behavior of mineral for temperature of 325°C and lithostatic pore fluid pressure of 900 MPa. References for constitutive relations are provided in Table 4. The grain sizes used in the calculations are shown for each panel. (a) Quartz, (b) albite, including both the calculation for diffusion creep from the Offerhaus et al. (2001) constitutive law and from our calculation assuming enhanced diffusion along phyllosilicate-lined grain boundaries, and (c) mica, representing phengite.
quartz deforms by dislocation creep (Figure 10a). The distribution of phengite throughout the quartz-rich regions will promote pressure solution creep over dislocation creep for two reasons. One is that phyllosilicates enhance grain boundary diffusion, which we took into consideration and is represented in Figure 10a. The other is that the phengite pins the quartz (e.g., Figures 3a and 3d) and inhibits dislocation creep, which is an effect that we did not directly account for. As a result, pressure solution creep may persist to higher stresses and strain rates than we calculated.

Deformation of albite is controlled by grain boundary diffusion creep over the entire range of shear stresses and strain rates that we evaluate (Figure 10b). Grain boundary diffusion creep of albite is predicted to occur about an order of magnitude slower than pressure solution creep of quartz when approximated by the Offerhaus et al. (2001) flow law, which we predict is a lower estimate. If, we consider faster diffusion pathways along phengite-lined grain boundaries, we predict strain rates that are similar to those in quartz. In this case, we would predict albite to accommodate comparable strain rates to quartz at shear stresses below approximately 30 MPa (Figures 10a and 10b). Dislocation creep of albite is extremely slow at these conditions, which indicates that the mechanism contributes negligibly to deformation and is consistent with the lack of a CPO and intragranular misorientation in albite (Figures 4c and 4f).

In the pelitic schist, dislocation glide of mica occurs at significantly lower stresses and higher strain rates than all deformation mechanisms in quartz and albite (Figure 10c). Extrapolation of the exponential flow law for glide of mica results in apparent dramatic weakening at strain rates less than $10^{-12}$ s$^{-1}$. Although strength is likely underestimated at low rates, basal glide of phengite is still expected to be weak compared to the strength of other phases. However, basal glide cannot satisfy von Mises criterion, which requires five independent slip systems for homogeneous deformation. Therefore, we predict that basal dislocation glide of phengite alone is unlikely to accommodate significant shear strain. However, glide within the basal plane of mica does likely occur with other deformation mechanisms in mica and during deformation of the other minerals present in the pelitic schist.

We use our constraints on the deformation mechanisms and constitutive relations of individual minerals to evaluate their contributions to the overall deformation of the pelitic schist. We infer that the quartz and albite-rich domains deform concurrently and independent of one another based on our structural observations which show that these domains occur subparallel to one another and to the foliation they define (e.g., Figures 3a, 3d and 6a). Within these domains, quartz and albite are mixed with much weaker phengite grains, which do not form interconnected layers beyond a few grain diameters. Given these interpretations, we expect that deformation of the relatively weak phengite is limited by deformation of the surrounding quartz and phengite.

To estimate the effects of distributed phengite grains on the rheology of the quartz and albite-rich domains, we use the approximation of Handy (1994) for a relatively weak mineral distributed within a load-bearing framework. In this case, the strain rate in the weak and strong phases are equal and controlled by the rate-limiting load-bearing mineral. The total stress is the volume averaged sum of the stresses supported by each mineral at that strain rate. In our calculations, we assume that phengite is evenly distributed between quartz-rich and albite-rich domains. The resulting rheology demonstrates that the composite is weakened relative to pure quartz and albite phases, but closely reflects the mechanisms operating in those phases (Figure 12a). The quartz and albite-rich domains are predicted to deform at rates within an order of magnitude of each other below shear stresses of approximately 10–20 MPa, given the uncertainty caused by our limited understanding of diffusive mass transfer in albite (Figures 10a, 10b and 12a). At higher stresses, dislocation creep in quartz will result in higher strain rates in the quartz-rich layers relative to the albite rich layers (Figures 10 and 12a).

Because of the subparallel arrangement of the albite and quartz-rich domains, we infer that they likely deform synchronously and independent of one another such that the shear stresses acting across them are similar, but the strain rates within them sum to the total strain rate. Thus, the overall deformation of the schist is controlled by whichever is the fastest deforming domain, which is uncertain without more constraints on albite deformation. If deformation of albite is enhanced by fluid transport along phengite grain boundaries, albite may be the weakest phase within these rocks (Figure 12a) and control the strength of this rock at these P-T-fluid conditions. Given that the microstructures predominate indicate diffusive mass transfer processes, we infer that deformation of the quartz and albite rich layers both occur at rates within an order of magnitude of one another at stresses less than approximately 10 MPa. At total shear stresses of 10–20 MPa dislocation creep becomes dominant in quartz and given the microstructural evidence for minor but not dominant dislocation creep in quartz, deformation may have occurred within this range of stresses (Figure 4). This range of stresses corresponds to strain rates of
approximately $10^{-13}$ s$^{-1}$, and is consistent with other stress estimates at these conditions along the place interface (Behr & Platt, 2013; Wassmann & Stöckhert, 2013).

### 5.3.3. Calcareous Schist Rheology

In the calcareous schist, we evaluate the grain size sensitive mechanisms for the large and small grain size fractions of quartz and calcite separately, as the size fractions are spatially segregated. The smallest grain size fraction of quartz, which is volumetrically dominant, deforms by pressure solution creep at shear stresses less than 30 MPa and strain rates less than approximately $10^{-12}$ s$^{-1}$ and by dislocation creep at higher stresses and strain rates (Figure 11a). The rate of pressure solution creep is lower for the larger grain size fraction, and as a result, dislocation creep dominates the rheology at shear stresses greater than 7 MPa and strain rates greater than 10$^{-15}$ s$^{-1}$ (Figure 11b).

For the larger size fraction of calcite, which is volumetrically dominant, grain boundary sliding is the dominant deformation mechanism at shear stresses below 10 MPa and strain rates less than $10^{-13}$ s$^{-1}$, above which dislocation creep is dominant (Figure 11d). In contrast, for the small size fraction, grain boundary sliding and diffusion creep occur at comparable rates and stresses and are the dominant deformation mechanisms at shear stresses less than 20 MPa and strain rates less than $10^{-8}$ s$^{-1}$, with dislocation creep occurring at higher stresses and strain rates (Figure 11e).

At the deformation conditions of the calcareous schist, dislocation glide of mica occurs at lower stresses and higher strain rates than most of the deformation mechanisms active in the calcite and quartz, like its behavior at the deformation conditions of the pelitic schist. The one exception is that the small size fraction of calcite grains deforms by grain boundary sliding at lower stresses and higher strain rates than glide of mica (Figure 11c).

We focus on the rheology recorded by the microstructures of the $S_1$ foliation, which we interpret records syn-subduction deformation. We infer that the quartz and phengite, calcite and quartz, and phengite domains will deform synchronously given their subparallel arrangement. In the domains dominated by the small size fraction of quartz and phengite, we use the same approach of Handy (1994) that we used for the pelitic schist to approximate the overall rheology of the domain. Deformation of weak phengite distributed within a fine-grained quartz matrix is predicted to be dominated by the quartz rheology, although shifted to lower stresses and higher strain rates due to deformation accommodated by the weaker phengite (Figures 11a and 12b).

In the domains dominated by calcite and quartz, the phases are mixed and have interlocking textures. Estimation of the rheology of these domains is challenging because it is not clear that they can be approximated as either having a load bearing framework or interconnected layers of a weak phase (e.g., Handy, 1994). As a very general approximation, we look at the end member cases that the coarse-grained calcite is distributed within a matrix of quartz (load bearing framework) and that the coarse-grained calcite forms interconnected layers (interconnected layers of a weak phase). Our microstructural observations show regions of both distributions (Figures 6a and 7). For simplicity, we ignore the fine-grained calcite which is a volumetrically small proportion of the domains (~13%) and does not form interconnected layers, so is not expected to cause dramatic weakening of the coarse-grained calcite and quartz mixtures. As expected, the composite rheology of the calcite and quartz layers is dominated by the quartz rheology for the end member case that quartz forms a load bearing matrix and by calcite rheology in the end member that calcite forms an interconnected weak phase (Figure 12b). In reality, the composite rheology of these layers is most likely intermediate to these end members and likely also varies spatially depending on the local geometric arrangement of calcite and quartz.

The partitioning of deformation amongst the different compositional domains of the calcareous schist is probably more complex than it is in the pelitic schist. Because the quartz and phengite domains, calcite and quartz domains, and pure phengite domains are subparallel to one another, we assume that these domains operate independent of one another, and that the fastest deforming domains will accommodate most of the deformation. The domains of pure phengite are predicted to be the weakest (Figure 12b). However, the rheologic constraints are based on flow laws for basal glide, which, as discussed above, is not expected to be capable of accommodating homogeneous and steady-state deformation. Nevertheless, these domains probably do contribute considerably to deformation, but much more work is needed to constrain the rheology of micas, including phengite, and to enable better indexing of these phases in EBSD so that active deformation mechanisms can be evaluated with more certainty. Quartz is predicted to be rate-limiting both in the quartz and phengite domains and in the calcite and quartz domains where calcite is poorly interconnected (Figure 12b). Between these domains, the quartz and
phengite domains are weaker at stresses less than \( \sim 10 \) MPa owing to the small grain size, and then comparable at higher stresses where both domains are predicted to deform by dislocation creep. Domains in which calcite is interconnected should accommodate higher deformation rates than those rate-controlled by quartz except for at the extremely low stresses less than 1 MPa. However, our structural observations and measurements show very little evidence of significant strain accommodated by calcite, including very weak CPOs and a lack of textures that indicate diffusion creep and grain boundary sliding (Figures 6 and 7). As a result, we posit that the domains in which calcite can be approximated as forming an interconnected network are relatively uncommon and that calcite-rich domains probably accommodate relatively little of the overall deformation compared to the quartz-rich
domains. This conclusion is somewhat counterintuitive, given calcite is predicted to be weaker than quartz in this rock, and suggests that calcite strength may not control the rheology of all calcite-bearing metasedimentary rocks. These inferences, based on structural observations and our understanding of the rheology of the constituent minerals, provide some broad constraints on the deformation conditions.

The observation that deformation of the quartz and phengite domains predominantly occurs by pressure solution creep with evidence for minor dislocation creep indicates that shear stresses were in the ballpark of 1–10 MPa, which is also consistent with the weak dislocation creep microstructures documented in the calcite (Figures 11d and 11e). If the quartz and phengite domains are rate-limiting, then the corresponding strain rates in those domains are on the order of $10^{-15}$ to $10^{-13}$ s$^{-1}$, whereas if calcite does form a network of interconnected grains, then shear strain rates in those domains could be as high as $10^{-12}$ to $10^{-11}$ s$^{-1}$ (Figure 12b). Together, this rheological modeling, guided by our microstructural observations suggests that stresses were relatively low, and quartz pinned by phengite and deforming by pressure solution creep is the most important phase controlling the rheology of the calcareous schist.

5.4. Fluid Sources

The diffusive mass transfer mechanisms active in the pelitic and calcareous schists indicate that fluids play an important role in viscous deformation, and likely play an important role in brittle deformation at these conditions through fracture and vein formation or frictional activation of slip at elevated pore fluid pressure (Fagereng et al., 2018; Fisher & Brantley, 2014; French & Condit, 2019). We can use the chemistry of vein-filling quartz to constrain the sources of fluids at approximate in situ deformation conditions within both of the metasedimentary units studied here. Previously published data from the calcite and quartz veins in the calcareous schist at AR1 were interpreted to have a vein filling fluids with a $^{18}$O value between 9‰ and 11‰ sourced from dehydrating mafic and ultramafic oceanic lithosphere (Jaeckel et al., 2018). This value is similar to the calculated $^{18}$O values for the vein-filling fluids in the pelitic schist that we report (Table 3, Figure 9).

Dynamic recrystallization microstructures in the quartz crack seal veins within the pelitic schist, which include bulging and incipient subgrain rotation microstructures, form at temperatures of $\sim$380°C–420°C (Stipp et al., 2002). This is above the peak subduction temperature estimates of 300°C–350°C (Bachmann, Gladny,
et al., 2009) for this portion of the Arosa Zone. Low temperature thermochronology from this region indicates that these rocks were last at temperatures greater than 300°C during the early Eocene, constraining the formation of the veins in the pelitic schist to be during or directly after subduction and before the subsequent Alpine Orogeny (Price et al., 2018). These data taken together suggest that the veins in the pelitic schist formed either at peak pressure and temperature subduction conditions and recrystallized at higher temperatures along the exhumation path or actually formed early along the exhumation path at these slightly elevated temperatures (Figure 5). This is consistent with the microstructural observations of the veins, and with the kinematic compatibility of the vein networks with the viscous foliation development. Thus, these veins appear to record subduction or early exhumation-related fluid compositions. Within a temperature range of 300°C−430°C for precipitation, δ¹⁸O values for the fluids that precipitated these veins are 6.7‰−10‰, remarkably consistent with the vein filling fluids reported for the calcareous schist’s calcite and quartz veins.

We can use the isotopic composition of these vein-forming fluids to help constrain the source of the fluids that formed the veins and may have facilitated pressure solution creep. For example, fluids that previously interacted with unaltered mafic to ultramafic lithologies at high temperatures should have δ¹⁸O values slightly lower than the partner rock if the interactions occurred at low water-to-rock ratios (Cole et al., 1987). The range of δ¹⁸O values we determined for the vein-precipitating fluid (6.7–10‰) is higher than that of pristine oceanic crust (δ¹⁸O = 5.6‰ (Pack & Herwartz, 2014)), ruling out unaltered oceanic crust as the rock with which the water interacted immediately prior to mineral precipitation. In contrast, altered or hydrated oceanic lithosphere has a wide range of δ¹⁸O values from >10‰ to as low as 4‰ (Barnicoat & Cartwright, 1995; Cooperdock et al., 2018; Epstein et al., 2020; Gregory & Taylor, 1981), so it is plausible that fluids interacted with or were sourced from altered oceanic lithosphere, as was interpreted for the calcite and quartz veins in the calcareous schist by Jaekel et al. (2018). Metasedimentary rocks, similar to those in which the veins are hosted, typically have higher δ¹⁸O values (11 ± 2‰ (Simon & Lécuyer, 2005)), meaning it, and other metasedimentary rocks, cannot be ruled out as a fluid source. Although the absence of silica depletion rinds around the vein margins (Figure 8a) indicates that the silica did not diffuse from the immediately adjacent pelitic schist (e.g., Fisher & Brantley, 2014), potential pervasive water-rock equilibration along fluid flow paths precludes definitive isotopic fingerprinting of the original fluid source.

While our isotope data alone cannot definitively differentiate between different metamorphic dehydration fluid sources, they, along with other regional data (Jaekel et al., 2018), suggest that the vein forming fluids are sourced from metamorphic dehydration reactions from oceanic crust along the plate interface. Devolatilization metamorphic reactions are strongly temperature dependent (e.g., Schmidt & Poli, 1998), and thus the thermal structure of the subduction interface is particularly important for the depths that substantial dehydration occurs. Recent thermodynamic modeling work indicates that at similar P-T conditions to those recorded here in the Arosa Zone, fluids will be dominantly be sourced from mafic lithologies (Condit et al., 2020), and substantial in situ fluid production can occur. Some recent isotopic work from the rocks of the subduction related Namibian Darma belt has suggested vein fluid sources to be local and from a mafic source (Fagereng et al., 2018). Additionally, workers have shown local fluid production and vein formation followed by external fluid infiltration and vein impregnation in other subduction interfaces (Muñoz-Montecinos et al., 2020). In the case of the Arosa Zone, it is possible that these fluids were produced at deeper depths and migrated along the plate interface, and the isotopic compositions recorded in these veins are in part the result of alteration during transport from source to vein (Bebout, 1991).

5.5. Stress State, Fluids, and Deformation Mechanisms at the Base of the Subduction Seismogenic Zone

Based on our microstructural, rheological, and isotopic data from these metasedimentary units, the base of the subduction seismogenic zone exposed in the Arosa zone is an area of very low shear stresses, in the range of <10 MPa to <30 MPa, where fluids play a vital role in facilitating viscous deformation through diffusive mass transfer. This range of shear stresses is similar to estimates from other subduction interface rocks exhumed from similar P-T conditions (Angiboust et al., 2015; Behr & Platt, 2013; Platt et al., 2018; Schmidt & Platt, 2022; Xia & Platt, 2017), and indicates that these low stresses are common along this portion of the plate interface. The fluids at these conditions appear to be sourced from metamorphic dehydration of subducting oceanic lithosphere, and likely also facilitated pressure solution creep and fill crack-seal quartz and calcite veins with δ¹⁸O composition of ~7.8‰−11‰ (Figure 7b; Jaekel et al., 2018). This is similar to other fluid sources at similar conditions (e.g., Fagereng et al., 2018).
Previous work (Behr & Platt, 2013; Platt et al., 2018; Wassmann & Stöckhert, 2013) combined with our microstructural observations in both metasedimentary units investigated in the Arosa Zone suggest that pressure solution creep, facilitated by a fluid-rich environment and small grain sizes, appears to be a key viscous deformation mechanism at the base of the subduction seismogenic zone. These previous studies have identified pressure solution creep in quartz as facilitating much of the ductile deformation within metasedimentary rocks. However, this study demonstrates that quartz is not the only major phase in metasedimentary rocks to deform by diffusive mass transfer at these depths along the subduction interface, particularly in pelitic lithologies. Albite may be as weak or weaker than quartz at these P-T-fluid conditions (e.g., Figure 12a). Pressure solution creep has been recognized as important for albite rocks deforming at greenschist facies conditions previously (Stünitz & Gerald, 1993), but to our knowledge has not been studied in detail in subduction zone rocks. As discussed above, there are relatively few laboratory constraints on pressure solution creep in albite, yet the microstructures that developed in the pelitic schist indicate that albite pressure solution creep is an important mechanism at the base of the subduction seismogenic zone and the diffusive mass transfer flow law for wet albite does either represent or approximate pressure solution (Rybacki & Dresen, 2000). In contrast, while we do observe minor evidence of dislocation creep in un-pinned quartz and calcite, the microstructures and rheological analysis suggests that at these fluid rich conditions and with these low stresses, crystal plastic deformation is not able to accommodate much deformation. Therefore, we posit that dislocation creep is a less important deformation mechanism than fluid-mediated diffusion at these conditions within most metasedimentary rocks, which typically have appreciable phyllosilicates to act as pinning phases.

5.6. Implications

Recent modeling shows that subducting thick packages of sediments can weaken the subduction interface considerably (e.g., Behr & Becker, 2018). Given the microstructural record of low shear stresses along the plate interface (∼<10 MPa to <30 MPa), we estimate the thicknesses of a sedimentary package that could accommodate tectonic strain rates through pressure solution and or dislocation creep. Using the pelitic schist as an example, with a typical subduction velocity of ∼50 mm/yr, and a stress range of 10–30 MPa, accommodating strain rates of ∼10⁻¹² requires a ∼1,500m thick package of metasedimentary rocks along the plate interface. This range of thicknesses is not unreasonable; the preserved thickness of the Arosa Zone at these locations is ∼520 m maximum (French & Condit, 2019), although it is not occupied by metasedimentary rocks exclusively, and other exhumed plate interfaces contain thick metasedimentary packages that may be able to fully accommodate plate rates during subduction (e.g., Agard et al., 2001; Tewksbury-Christle et al., 2021). This suggests that if there are ample sediments subducted, pressure solution creep of quartz and albite could accommodate all viscous deformation during tectonic creeping at the base of the subduction seismogenic zone, regardless if these metasedimentary rocks are calc-schists or metapelites. As noted above, the Arosa Zone is not solely occupied by metasedimentary rocks. French and Condit (2019) investigated the role of variable rheology of these lithologies at a range of pore fluid pressures and found that deformation partitioning between metasedimentary lithologies and metamafic and ultramafic rocks will occur based on effective stresses guided by pore fluid pressures.

Within subduction zones, the cause of the seismogenic to aseismic transition is not known, and the rock types that accommodate aseismic creep, slow slip, and tremor are debated (e.g., Bürgmann, 2018). However, our rheological analyses indicate that by ∼30–35 km depth and 300°C–430°C, viscous deformation of metasedimentary rocks is sufficient to account for all aseismic deformation at a shear stress of <30 MPa (Figure 12). The magnitude of this stress is very similar to previous studies of exhumed subduction interfaces at similar depths (Behr & Platt, 2013; Platt et al., 2018). This is at the low end of the range of stresses inferred for the base of the seismogenic zone using heat flow constraints (England, 2018).

This is also the region of the plate interface where deep slow slip events (SSEs) occur, which have been suggested by some workers to be facilitated by viscous deformation (Behr et al., 2018; Hayman & Lavier, 2014; Kirkpatrick et al., 2021; Ujiie et al., 2018). Given the same range of interface thicknesses we found above (300 m–3 km), and a slow slip velocity range of 300–1,500 mm/yr (Wallace et al., 2012), for pressure solution creep to accommodate slow slip we would need strain rates of ∼10⁻¹¹ s⁻¹ to ∼10⁻¹⁰ s⁻¹, which corresponds to shear stresses in the range of 100s–1,000s of MPa (Figure 12a). Such high shear stresses are at odds with studies of SSEs in active subduction zones that indicate that shear stresses are extremely low (<1 MPa) during these events (Hawthorne & Rubin, 2010). Therefore, deformation by other processes or of other lithologies must be important. Recent
modeling work by French and Condit (2019) shows that these low shear stresses are best explained by frictional deformation of talc and chlorite schists at elevated pore fluid pressures. Similar results have been found by Y. Liu and Rice (2007). Thus, we infer that the microstructures that we observe here in these metasedimentary rocks record tectonic strain rates, rather than SSEs, and during periods of elevated pore fluid pressure deformation will partition into the lithologies that host slow slip.

The veins, particularly those in the pelitic schist, may record cyclic variations in pore fluid pressure. Workers in similar subduction settings have interpreted similar foliation and vein structures to form coeval (e.g., Ujiie et al., 2018). The crack-seal microstructures in each vein set suggest multiple phases of pore fluid overpressure and resulting cracking, precipitation, and sealing. Fluctuations in pore fluid pressure could be due to (a) punctuated fluid delivery or (b) evolving porosity of the subduction interface due to fracturing and precipitation. Lithium isotope diffusion data from Taetz et al. (2018) record transient 1–4 months long periods of fluid release during dehydration and reaction vein formation at eclogite facies (depths ~60–70 km), offering a potential fluid source-dependent mechanism that could regulate the observed cycles in pore fluid pressure updp, particularly if these pore fluids are deeply sourced and transported quickly updp. Alternatively, other workers (e.g., Fisher & Brantley, 2014; Ujiie et al., 2018) showed that the timescale of quartz precipitation within fractures, and the resultant porosity loss and fluid pressurization, is similar to ETS inter-event time scales. This suggests that the cyclic nature of both pore fluid pressure and ETS may be controlled by timescales of silica precipitation rather than fluid fluxes or delivery.

6. Conclusions

We studied metasedimentary rocks exhumed from the base of the subduction seismogenic zone to understand the active viscous mechanisms and corresponding strength at this transition. Viscous deformation was accommodated largely by pressure solution creep of quartz and albite when these phases are pinned by phengite. Stable isotope data from crack-seal veins suggests that the fluids at these conditions were sourced from dehydration reactions within altered oceanic lithosphere. Rheological analysis of these deformation structures suggests that pressure solution creep of quartz and albite can readily accommodate tectonic creeping strain rates in reasonably thick subducted sedimentary packages at stresses of <30 MPa, but that these mechanisms cannot accommodate slow slip strain rates at these nominally low stresses. Given that, these lithologies are common within exhumed subduction terranes, these results are probably broadly applicable to many subduction zones where pelitic and or calcareous sediments are subducted.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Data Availability Statement

Textural and geochemical data from this contribution can be found at https://doi.org/10.5281/zenodo.5800266.

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