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Geophysical and geological perspectives on accretionary wedge
evolution, splay faulting, and megathrust earthquake processes
at the Cascadia Subduction Zone

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A dissertation
submitted in partial fulfillment of the
requirements for the degree of

Doctor of Philosophy

University of Washington

2026

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Abstract

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Accretionary wedges, submarine fold-and-thrust belts that accumulate at subduction zones, are the main locus of brittle deformation during megathrust earthquakes, and this deformation is in turn the main driver of tsunamis. Though accretionary wedges were initially thought aseismic, recent megathrust earthquakes have revealed that coseismic slip on the shallow megathrust and accompanying high-angle splay faults is common and can be devastating. In this Dissertation, I apply geophysical and geological methods to better constrain accretionary wedge structure and deformation as it applies to megathrust earthquake hazards at the Cascadia Subduction Zone.

In Chapter 2, I use nested scales of seismic reflection imaging to better constrain splay fault activity in the landward vergence zone, which reveals that active splay faults concentrate near the deformation front. These splay faults are likely to slip coseismically along with the shallow megathrust during earthquakes. In Chapter 3, I then reexamine the position of the megathrust fault in the landward vergence zone by applying structural geological techniques to newly-collected crustal-scale seismic reflection imaging. I find that the megathrust fault is never co-located with the sediment-basement interface, and instead facilitates on average ~ 500 m of sediment underthrusting in the active outer wedge. These studies have distinct impli-

cations for better understanding the characteristics of future megathrust earthquakes at Cascadia, including informing fault slip scenarios and fault physical properties.

Next, I turn to the portion of Cascadia's accretionary wedge exposed onshore, known as the Olympic Subduction Complex. This area has been identified as an accretionary complex for decades, but there has been little modern geological work in the region. In Chapter 4, I focus on a site known as Lakes of the Gods, where I identify a paleomegathrust interface comprised of multiple anastomosing fault strands contained in a ~ 450 m thick section of deformed *mélange*. In Chapter 5, I design an experiment to test how strain partitions across the various structural domains of the paleomegathrust interface. I find that interseismic, *mélange*-forming processes account for negligible amounts of plate boundary strain, while localized, coseismic, fault zone processes must account for the remainder. This is consistent with the geophysical observations of modern subduction zones, but geologically counterintuitive because *mélange* is vastly volumetrically dominant in the rock record compared to fault rocks.

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ACKNOWLEDGMENTS

First, thank you to Harold Tobin, my advisor, whose unfailing support over the last five years has been instrumental to this work. Over many hours from the field to the office, you have provided endless scientific insight, mentorship, gossip, and frustration. I don't think I would have enjoyed graduate school nearly as much with a different advisor, or without your dog, Lulu.

I am also grateful to Cailey Condit, Nathan Miller, and Janet Watt for providing substantial guidance and mentorship on components of this Dissertation. Thank you to the staff of Olympic National Park who were instrumental in making this research possible. Thank you to Steve Goldstein for showing me my first accretionary wedge and for convincing me becoming a field geologist might be a good idea.

Thank you to my family, whose sacrifices helped make this Ph.D. possible. Thank you to my best friends in ESS, whose company over daily lunches and weekly beers made this Ph.D. fly by. I will miss you. Finally, thank you to Peter Lindquist, my partner and on-call structural geology consultant, who taught me how to use a Brunton and how to cook good tofu, which were both equally important to the creation of this Dissertation.

Chapter 1

INTRODUCTION

1.1 Motivation

Subduction zones are responsible for the world’s largest earthquakes, which in turn produce the most devastating tsunamis. The last century’s most destructive megathrust earthquakes—including the 1964 M9.2 Alaska, the 2004 M9.1 Sumatra, and the 2011 M9.1 Tohoku earthquake—have resulted in hundreds of thousands of casualties. In accretionary wedges, the subduction zone plate boundary fault slip which is the source of these events occurs on a low angle décollement thrust fault. However, subsidiary faults known as splay faults branch from the underlying décollement through the accretionary wedge and may also host slip during large megathrust events. The coseismic activation of the shallow megathrust and/or splay faults—once fault improbable due to the unconsolidated nature of accretionary wedge sediments—has been well-documented for each of the aforementioned megathrust earthquakes. Furthermore, shallow accretionary wedge deformation is the primary driver of tsunamis (Wilson and Ma, 2021). In the U.S. alone, more than 3 million residents live in tsunami flooding areas associated with subduction zones (Wood et al., 2025), so these events pose a large geohazard.

This dissertation focuses on the Cascadia subduction zone, where the Juan de Fuca plate subducts under the North American plate off the coast of western North America. In many ways, Cascadia is a unique subduction zone. It has a thick incoming plate sediment package among subduction zones globally (Heuret et al., 2012), and has, in particular, experienced high recent sedimentation due to glaciation in the region (e.g., Underwood, 2002; Barnard, 1978). The thick sediment cover on the hot,

young Juan de Fuca plate renders it among the hottest subduction zones globally (Salmi et al., 2017). This combination creates a rare scenario in which the seismogenic zone, the region where it is frictionally favorable for earthquakes to nucleate, to extend all the way to the deformation front (e.g., Salmi et al., 2017; Hüpers et al., 2017a). Also, there is a notable absence of seismicity (e.g., Stone et al., 2018; Morton et al., 2023), and the last megathrust earthquake was in 1700 CE (Atwater and Hemphill-Haley, 1997; Satake et al., 1996). Therefore, the understanding of hazards associated with particular fault slip scenarios in future megathrust events is limited.

The accretionary wedge offshore Washington state has atypical morphology; it is particularly flat and wide, and dominated by landward-vergent (seaward-dipping) faults. Such landward-vergent faults are globally rare, only present in large areas elsewhere in the Sumatra subduction zone (Moeremans et al., 2014; Frederik et al., 2015; Dean et al., 2010; McNeill and Henstock, 2014; Henstock et al., 2006) and the Niger Delta (Maloney et al., 2010; Bilotti and Shaw, 2005). Landward vergence has been difficult to explain using the theoretical framework that governs accretionary wedges (critical wedge theory, Dahlen et al., 1984; Davis et al., 1983). The landward vergence zone is also coincident with the most highly-locked region of Cascadia (e.g., Li et al., 2018; Lindsey et al., 2021), and can therefore be considered the most region to host a large future megathrust earthquake.

Coseismic slip on the shallow megathrust fault and the high-angle splay faults which branch from it was thought impossible until seen in recent megathrust earthquakes—such as the 1964 Alaska M9.2, 2004 Sumatra M9.1, and 2011 Japan M9.1 events—and can significantly increase expected tsunami size. It has also been well-established that sediment-rich subduction zones present a unique, heightened hazard; when the fault zone is hosted in sediments, it often promotes higher levels of locking and less frequent, but much larger, earthquakes (Ruff, 1989). These recent earthquakes highlight the under-investigated hazard that shallow accretionary wedge deformation poses and inform the goal of this Dissertation to better understand the

structure of the accretionary wedge and its implications for future megathrust earthquakes.

Finally, while geophysical tools and modeling work have allowed us to probe accretionary wedges and understand subduction zone architecture, there is a mismatch between the low resolution that these methods allow and the small-scale heterogeneity of fault zone properties. Little is known about the geological processes that occur along megathrust faults and within the wall rock of the surrounding accretionary wedge that lead to earthquake nucleation and propagation and allow them to accommodate such large slip (up to 100 meters) over such large areas (over 1000 km along strike). Direct evidence of the conditions that promote coseismic slip can only be found by examining the plate boundary itself, either through oceanic drilling or the study of exhumed subduction zone terranes. While oceanic drilling has resulted in significant advances, it has only allowed drilling of the shallowest plate interfaces, and through small diameter boreholes.

Therefore, the only way to understand geologic processes and conditions at the plate interface deeper in the seismogenic zone is through the study of exhumed subduction terranes, which allow both unprecedented detail as well as large areas of exposed outcrop to understand the heterogeneity of the interface (e.g., Hilley et al., 2022). However, such exhumed terranes are globally rare (e.g., Rowe et al., 2013). In this work, we identify the first known example of a paleomegathrust interface associated with the modern Cascadia subduction zone. We characterize its structure, properties, and lithologies, in particular focusing on the interplay between fast, coseismic deformation localized along faults, and slow, aseismic deformation in the surrounding *mélange*.

1.2 Scientific significance and description of chapters

1.2.1 Chapter 2: Late Quaternary surface displacements on accretionary wedge splay faults in the Cascadia Subduction Zone: Implications for megathrust rupture

For this work, I analyze nested scales of seismic reflection and bathymetric data to identify active splay faults which link to the megathrust fault at depth. While splay faults have been mapped offshore for decades and are present in publications like the Quaternary Faults Database, the use of high-resolution seismic reflection data allows me to identify the most recently active faults. While the research in this chapter is published (Ledeczi et al., 2024), the faults I identify therein will still be published as an open-access USGS-hosted data release (Watt et al., in review) and incorporated into the Cascadia Region Earthquake Science Center’s Community Fault Model v1.0.

Work primarily at Nankai first documented a large, continuous, and long-lived splay fault known as a megasplay (e.g., Moore et al., 2007a; Park et al., 2002; Strasser et al., 2009). This megasplay has been said to localize at the inner-outer wedge transition, also known as the backstop, which is the rheological boundary between younger, less consolidated outer wedge sediments and older, more consolidated inner wedge sediments. It has been largely assumed that outer wedge splay faults form aseismically.

However, at Cascadia, the previously-proposed megasplay fault (Witter et al., 2013) is used for tsunami models and coastal hazards assessments but has been shown to be inactive to absent along most of the margin (Lucas et al., 2025). Instead, I document a previously unknown zone of active splay faults 30 km in width near the deformation front in the outermost wedge (Ledeczi et al., 2024). I suggest that, during megathrust earthquakes, the inner wedge and inactive part of the outer wedge translate seaward without splay activation; meanwhile, slip is transferred up the megathrust and distributed onto one or several splay faults in the outermost, active domain (Ledeczi et al., 2024).

This work has resulted in a significant change in our understanding of splay fault activity at Cascadia, and it is part of a growing body of literature that presents geologic evidence for shallow wedge deformation, either on the shallow megathrust or splay faults, during Cascadia megathrust earthquakes. Because many have argued that shallow splay faults cannot slip coseismically, my work in *Chapter 2* further motivated dynamic rupture modeling of splay faults in the outer wedge and showed that they are readily coseismically activated by slip on the megathrust (Biemiller et al., 2025). Previous work (Cubas et al., 2016) has argued that the existence of landward-vergent splay faults is itself evidence for coseismic slip because their formation requires extremely low coefficients of friction on the megathrust fault (< 0.03) that can only be achieved dynamically during earthquake slip. Watt et al. (in review) present further evidence for coseismic slip of the frontal thrust from high-resolution seismic reflection and bathymetric data. Hill et al. (2026) document mass transport deposits in sediment cores that suggest repeated, coseismic deformation of the outer wedge. This preponderance of evidence suggests slip along the megathrust fault to the toe of the wedge, likely involving slip on multiple splay faults as well, is a common rupture mode at Cascadia.

1.2.2 Chapter 3: Evidence for a sediment-hosted plate boundary fault at the Cascadia Subduction Zone

In this work, I apply quantitative structural geology methods to better understand accretionary wedge structure in the landward vergence zone’s outer wedge. While *Chapter 2* uses mainly high-resolution seismic reflection data, here I focus on the newly-collected, crustal-scale data from the Cascadia Seismic Imaging Experiment in 2021 (CASIE21; Carbotte et al., 2023).

The work in this chapter is inspired by a decades-long discrepancy in the literature about the position of the megathrust fault in the landward vergence zone—specifically, whether the fault localizes at the sediment-basement interface (“top of crust”) or

within the sediment section. Previous work originally established that the megathrust is localized within the sediments but close (<1 km above) to the basement in this region (Adam et al., 2004; Booth-Rea et al., 2008; MacKay, 1995; MacKay et al., 1992; Webb, 2017; Gutscher et al., 2001). While more recent geophysical studies (Peterson and Keranen, 2019), including a new study based on the CASIE21 dataset (Carbotte et al., 2024) has suggested the fault may instead form at the top of crust, this work was not focused on detailed structural geology-informed seismic interpretation.

The application of kinematic fault models (e.g., Suppe, 1983; Suppe and Medwedeff, 1990; Suppe et al., 2004) and area-depth theory (Groshong and Epard, 1994; Groshong et al., 2012) to seismic interpretation ensures that the final interpretation is robust and admissible (e.g., Eichelberger et al., 2015; Shaw et al., 2005). We apply these methods to constrain the depth to which splay faults sole, thereby identifying the through-going décollement. In all cases, we find that splay faults sole into detachment levels within the sediment section, on average ~ 500 m above the top of crust in the landward vergence zone.

We then review previous work on outer accretionary wedge structure at other sediment-rich margins and find that the megathrust fault has been similarly interpreted as localizing at various levels above, but only rarely co-located with, the sediment-basement interface; this includes work from the Hikurangi (Barnes et al., 2010; Ghisetti et al., 2016; Plaza-Faverola et al., 2012; Watson et al., 2019), Barbados (Labaume et al.; Laigle et al., 2013; Shipley et al., 1997), Alaska (Cortés-Rivas et al., 2025; von Huene et al., 2021), Makran (Burg, 2018; Grando and McClay, 2007; Kopp et al., 2000), and Nankai (Bangs et al., 2009; Moore et al., 2007a) subduction zones.

While ~ 500 m of sediment underthrusting may not seem significant, the fault zone’s host rock—and therefore its fundamental characteristics like its friction and elastic properties—is the primary factor governing earthquake rupture nucleation, propagation, and arrest. Taken together, this work shows that plate boundary faults preferentially are hosted within the sedimentary section, as opposed to along the

sediment-basement interface or within basaltic crust. This work is consistent with global observations that where thick sediments help cover the rough topography of the oceanic crust, smoothing the plate interface, frictional locking and large megathrust earthquakes are favored (Heuret et al., 2012; Ruff, 1989; Wang and Bilek, 2014). The work presented here favors a sediment-hosted megathrust fault system for the entirety of the landward vergence zone. Moreover, the results of this work are also consistent with the geological record for this margin (Ledeczi et al., 2025), discussed below in *Chapter 4*.

1.2.3 Chapter 4: Structure and properties of the Cascadia plate interface: Evidence from a newly described exhumed paleomegathrust in the Olympic Subduction Complex

This chapter presents a detailed study focused on a newly described exhumed Cascadia paleomegathrust fault exposure (Lake of the Gods, “LOTG”) in the Olympic Mountains, WA. For decades, it has been established that Cascadia’s large, clastic-sediment dominated accretionary wedge continues onshore only on the Olympic Peninsula (e.g., Tabor and Cady, 1978a,b; Brandon et al., 1998; Watt and Brothers, 2020). This unit, known as the Olympic Subduction Complex, was thought to have subducted, underplated, then exhumed (Brandon et al., 1998; Brandon and Calderwood, 1990; Brandon et al., 1992), thereby likely preserving deformation structures associated with plate boundary deformation. However, the rugged terrain of Olympic National Park prevented further investigation, and the most recent geological work conducted west of Mt. Olympus was presented in Tabor and Cady (1978a) and Tabor and Cady (1978b).

There are key differences between geophysical and geological conceptions of megathrust faults. While geophysicists often conceptualized a planar fault with no off fault deformation, the geological record has shown that plate boundaries are comprised of a braided fault network contained in deformed rock known as *mélange* up to kilometers in thickness (e.g., Rowe et al., 2013). Seismological observations are beginning

to show this as well (Chalumeau et al., 2024).

We have conducted four summers of field work in Olympic National Park, with two focused at the LOTG site. At this locality, I describe a braided network of ~ 10 faults cutting across highly disrupted *mélange*, comparable to other models of exhumed megathrust faults globally. Using petrography and microstructural analysis via optical and scanning electron microscopy, I document alternating brittle, coseismic deformation localized in these faults, and plastic, aseismic deformation in surrounding *mélange*, together representing deformation throughout the seismic cycle. I identify only clastic sedimentary rocks at this locality, and no basalts or other components of oceanic plate stratigraphy. Finally, I conduct detailed geothermometry to constrain the maximum paleotemperature these rocks experienced to $260 - 280^\circ\text{C}$, in the heart of the seismogenic zone (Ledeczi et al., 2025). Importantly, this work shows that the megathrust fault was localized within the sedimentary section well above the oceanic crust in the past, consistent with modern observations presented in *Chapter 3*.

*1.2.4 Chapter 5: Off-fault *mélange* formation accommodates only a small fraction of strain in the subduction seismogenic zone*

There is often a difference in perspective between geophysicists studying modern subduction zones where large megathrust earthquakes dominate the deformation between the two plates at seismogenic depths, and geologists studying the rock record where *mélange* surrounding the plate interface is vastly volumetrically dominant compared to thin, localized fault zones that host coseismic slip. Interseismic, off-fault deformation in *mélange* zones is known to accommodate a portion of the relative plate slip, thereby reducing the portion which is released as an elastic response during megathrust earthquakes; but the exact role of these *mélanges*, and how much strain they accommodate, is still debated. This issue is exacerbated because measurements of strain in the exhumed rock record are very difficult to conduct and have been largely unsuccessful.

In this chapter, I address this discrepancy by estimating the amount of total dis-

placement that occurred during subduction and comparing it to deformation recorded in in the LOTG fault zone (Ledeczi et al., 2025). I use a novel combination of geochemistry, rock magnetics, and constitutive relations to quantify the interseismic strain recorded in these rocks, approaching this question from three different angles, and find that the methods all support the same conclusion. I find that the mélange fabrics record negligible amounts of strain—on the order of 0.1% of the plate displacement from the subduction process. Thus, the missing strain signal is most likely accommodated coseismically during megathrust earthquake slip on thin faults. This view is consistent with geophysical observations of megathrust earthquakes. Furthermore, it provides new context for the geological interpretation of subduction fault complexes globally.

Chapter 2

LATE QUATERNARY SURFACE DISPLACEMENTS ON ACCRETIONARY WEDGE SPLAY FAULTS IN THE CASCADIA SUBDUCTION ZONE: IMPLICATIONS FOR MEGATHRUST RUPTURE

Manuscript published in *Seismica* at 10.26443/seismica.v2i4.1158.

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2.1 Abstract

Because splay faults branch at a steep dip angle from the plate-boundary décollement in an accretionary wedge, their coseismic displacement can potentially result in larger tsunamis with distinct characteristics compared to megathrust-only fault ruptures, posing an enhanced hazard to coastal communities. Elsewhere, there is evidence of coseismic slip on splay faults during many of the largest subduction zone earthquakes, but our understanding of potentially active splay faults and their hazards at the Cascadia subduction zone remains limited. To identify the most recently active splay faults at Cascadia, we conduct stratigraphic and structural interpretations of near-surface deformation in the outer accretionary wedge for the ~400 km along-strike length of the landward vergence zone. We analyze recently acquired

high-frequency sparker seismic data and crustal-scale multi-channel seismic data to examine the record of deformation in shallow slope basins and the upper ~ 1 km of the surrounding accreted sediments and to investigate linkages to deeper décollement structure. We present a new fault map for widest, most completely locked portion of Cascadia from 45 to 48°N latitude, which documents the distribution of faults that show clear evidence of recent late Quaternary activity. We find widespread evidence for active splay faulting up to 30 km landward of the deformation front, in what we define as the active domain, and diminished fault activity landward outside of this zone. The abundance of surface-deforming splay faults in the active outer wedge domain suggests Cascadia megathrust events may commonly host distributed shallow rupture on multiple splay faults located within 30 km of the deformation front.

2.2 *Non-technical summary*

Earthquakes and their resulting tsunamis at the Cascadia subduction zone, where the Juan de Fuca plate subducts under the North American plate from northern California to Vancouver Island, pose great hazard to coastal communities. As sediment from the subducting plate is scraped off, piles of sediment known as accretionary wedges can form at these plate boundaries. During large earthquakes, the rupture of the main plate boundary fault may be accompanied by slip on multiple branching faults within the accretionary wedge, which would potentially impact the size and characteristics of tsunami produced by the event. The likelihood of splay fault slip remains largely unknown at Cascadia because the last large earthquake occurred in 1700. We interpret newly collected subsurface images using seismic wave reflections to identify the most recently active offshore faults. We document evidence of widespread recent fault activity within the accretionary wedge in a roughly ~ 30 km wide zone stretching from Washington to northern Oregon. By mapping these active branching faults, we seek to provide inputs to tsunami and earthquake models which may in turn refine predictions and preparations for coastal impacts during future earthquakes at

Cascadia.

2.3 Introduction

Accretionary wedges at subduction zones are fold and thrust belts formed when sediment is scraped off the downgoing plate and accumulates at the leading edge of the overriding plate. The plate boundary décollement forms the base of the accretionary wedge, and numerous higher-angle splay faults, including the frontal thrust, branch from it, bounding the individual thrust sheets that comprise the wedge. As new material is accreted, splay faults activate to accommodate this deformation, adjusting the surface slope of the accretionary wedge through fault activity or erosion to reach a critical taper angle of stability (Davis et al., 1983). It was once thought that accretionary wedges cannot nucleate nor host seismogenic slip near the trench on either the plate boundary décollement or on splay faults because those faults were seen as being hosted in weak, unlithified materials (i.e., frictionally stable) (e.g., Byrne et al., 1988), but this view was gradually supplanted by an understanding that temperature-dependent frictional properties, rather than the absence of sedimentary material, governs seismic locking and slip. A 100 – 150°C isotherm threshold, which corresponds to many transitions in material properties of the host sediments or rock, is now more commonly identified as the updip limit for frictional locking and slip at subduction zones (Hyndman and Wang, 1993; Moore and Saffer, 2001a).

Recent earthquakes have shown that both shallow slip and splay fault activation are possible and even prevalent during large megathrust events. For example, in the M_w 9.1 2011 Tohoku-Oki earthquake, the majority of the rupture occurred above 40 km in depth, and slip peaked at the trench (Ide et al., 2011; Sun et al., 2017). Differential bathymetry, displaced seafloor geodetic monuments, and scarp mapping all indicate that 50 to 120 meters of coseismic slip occurred at the trench axis during this event (Fujiwara et al., 2011; Ito et al., 2011; Kodaira et al., 2012, 2020; Ueda et al., 2023). Similarly, the 2004 M_w 9.1 Sumatra earthquake had largely shallow

coseismic rupture beneath the accretionary wedge that may have extended to the deformation front based on scarps mapped there (Gulick et al., 2011; Henstock et al., 2006; Ishii et al., 2005). Bathymetric surveys following the event document widespread faulting, folding, and mass wasting of the accretionary wedge, attributed to coseismic splay fault rupture (Seeber et al., 2007). Subaerial surface uplift and bathymetric investigation following the 1964 M_w 9.2 Alaska earthquake found coseismic slip of up to 12 m on two splay faults in the wedge (Liberty et al., 2013; Plafker, 1969). Core samples from the offshore frontal thrust and megasplay fault of the Nankai Trough wedge record frictional heating of sediments which indicate that coseismic slip likely propagated to the toe of the accretionary wedge and along the megasplay fault in the past (Sakaguchi et al., 2011; Yamaguchi et al., 2011).

Tsunamis generated by large megathrust events are typically much more destructive than shaking from earthquakes themselves, and shallow slip concentrated in the outer accretionary wedge may play a great role in tsunami generation (Wilson and Ma, 2021). Shallow slip along the megathrust can cause large seafloor uplift in deep water and is thus more efficient at tsunami generation than slip which stops farther downdip. However, because of their relatively steeper dip angle compared to the frontal décollement, splay fault activation can result in even larger tsunamis with multiple flooding episodes compared to megathrust-only fault ruptures, posing a unique hazard to coastal communities (Gao et al., 2018; Wendt et al., 2009; van Zelst et al., 2022).

At Cascadia, dislocation modeling has confirmed that shallow splay fault slip can cause a larger tsunami, as compared to slip only on the megathrust (Gao et al., 2018). The area offshore Washington state has atypical accretionary wedge morphology and is coincident with the zone of greatest geodetic locking at Cascadia (Li et al., 2018; Lindsey et al., 2021; Schmalzle et al., 2014), which suggests this region may experience the most slip during a future megathrust event. However, since the last megathrust event at Cascadia occurred in 1700 (Satake et al., 1996), and there is a notable

absence of recent seismicity in the region (e.g., Stone et al., 2018), our knowledge of what future slip scenarios could look like or what faults could experience slip is limited.

In this study, we systematically evaluate which splay faults offshore Washington and Oregon show evidence of hosting recent near-surface displacement and map their continuity along strike for a > 400 km span from 45° to 48° N latitude. We use high-frequency sparker multi-channel seismic reflection data to characterize sedimentary sequences and structural history of slope basins and the upper 1 km of the accreted wedge, which allows us to identify faults that have been recently active in late Quaternary time throughout the outer wedge in the landward vergence zone (LVZ), the area of Cascadia which displays atypical morphology (Fig. 5.1) and coincides with the largest patch of the megathrust that is apparently fully locked (Lindsey et al., 2021). We analyze slope basins, which are sensitive recorders of activity on accretionary wedge faults as the basins fill with sediments over time from their formation to the present. In particular, we seek to determine whether splay faults are active only at the frontal thrust, at the inner-outer wedge transition, or throughout the outer wedge. We hypothesize that rather than progressing in sequence toward the deformation front, thrusts in the accretionary wedge at Cascadia display a more complex history of thrust activation. Our findings support the hypothesis that many splay faults are potential candidates for hosting coseismic slip, and that Cascadia will likely experience distributed shallow rupture on multiple splay faults located within ~ 30 km of the deformation front in a future megathrust event.

2.4 Tectonic Setting and Accretionary Wedge Mechanics

Offshore Washington, the Juan de Fuca plate is subducting northwestward under the North American plate at ~ 39 mm/yr (Fig. 5.1; DeMets et al., 2010). A 2 to 3 kilometer thick sequence of clastic sediment overlies the oceanic basement of the incoming plate. Most of this sediment is scraped off as the oceanic plate subducts,

forming a wide and rapidly growing accretionary wedge throughout our study area. As splay faults propagate within the wedge, they fold the sedimentary strata of the wedge into anticlinal structures to form trench-parallel ridges. Between the ridges, shallow (~ 100 s of meters deep) wedge-top basins known as piggyback or slope basins are created and fill with continuously depositing sediments. The entire outer wedge in this region, including thrust sheets and slope basin sediments, has been formed over less than 2 million years during the Pleistocene and Holocene (Barnard, 1978).

The accretionary wedge is comprised of several distinct morphotectonic regions with variable characteristics along strike (Watt and Brothers, 2020). From 45° to 48° N latitude, the LVZ contains almost exclusively landward-vergent (seaward-dipping) splay faults (MacKay, 1995; Seely, 1977), while thrusting in the wedge elsewhere is mostly seaward-vergent, which is more typical in subduction zones worldwide (Fig. 5.1). Seismic reflection data show that, in the LVZ, landward-vergent thrusts have kilometers of displacement each whereas seaward-vergent thrusts are nearly nonexistent (Adam et al., 2004). Many explanations have been proposed for the LVZ, such as an internally strong wedge with a weak base, perhaps due to high pore pressure (MacKay, 1995; Tobin et al., 1993); a high sedimentation rate (Adam et al., 2004); and/or a compositional difference in wedge sediments (Gutscher et al., 2001; Underwood, 2002). The widespread nature of the landward vergence at Cascadia is unusual among fold-and-thrust belts worldwide, with only two comparable examples. The northern Sumatra region of the Sunda subduction zone also displays dominantly landward vergence at the toe of the accretionary wedge, with seaward-vergent faults located farther landward (Frederik et al., 2015; McNeill and Henstock, 2014). Isolated areas of landward vergence have been observed at the Niger Delta, a submarine fold and thrust belt driven by gravitational collapse rather than tectonics (Bilotti and Shaw, 2005; Maloney et al., 2010).

The oblique component of the subduction angle is thought to be accommodated by numerous left-lateral strike-slip faults striking northwest-southeast across the margin.

These structures were first identified in the 1990s by sidescan sonar, seismic reflection, and bathymetric data, and were found to have Holocene activity by observing offset geomorphic markers (Goldfinger et al., 1992, 1997, 1996). Some of these faults are thought to offset the downgoing plate (Han et al., 2018) whereas some may originate in and be confined to the accretionary wedge sediments of the upper plate. These faults can accommodate arc-parallel strain, facilitating bookshelf faulting of the forearc (Goldfinger et al., 1996). A zone of listric normal faulting on the continental shelf offshore Grays Harbor suggests this region is undergoing extension and is unaffected by the compressive forces dominating the outer wedge (McNeill et al., 1997).

The strength contrast at the inner-outer wedge transition or the dynamic backstop boundary (Frederik et al., 2015; Kopp and Kukowski, 2003; Wang and Hu, 2006) has been hypothesized to localize splay fault activity at other subduction zones, such as Nankai (Gulick et al., 2010; Moore et al., 2007b; Tobin et al., 2019) and Alaska (Liberty et al., 2013; Ramos et al., 2022; von Huene et al., 2021). We prefer the term inner-outer wedge transition over dynamic backstop boundary and use it throughout this paper. At Cascadia, the inner-outer wedge transition marks the boundary between the less consolidated, Pleistocene-dominated outer accretionary wedge and the more consolidated, Miocene inner accretionary wedge sediments (Frederik et al., 2015; Kopp and Kukowski, 2003; Tobin et al., 2019; Wang and Hu, 2006). At Nankai, the continuity of a major active fault along strike at the boundary led to the use of the term “megasplay” to distinguish this major domain-bounding fault system from other splays in the outer wedge (Tobin and Kinoshita, 2006a). In both Nankai and Alaska, there is evidence that splay faults have localized slip in past megathrust events and serve as a potential tsunami source (Moore et al., 2007b; Ramos et al., 2022; von Huene et al., 2021). Similar megasplay faults at the inner-outer wedge transition have been proposed and used for tsunami modeling at Cascadia (e.g., Gao et al., 2018; Witter et al., 2013).

2.5 Methods

2.5.1 Data Acquisition and Processing

The 11 lines of high-resolution multichannel seismic reflection data used in this study were collected aboard the *R/V Rachel Carson* in 2019 (Balster-Gee et al., 2023). The seismic sound source was an Applied Acoustics Delta sparker operated at 1.4 to 3.6 kJ and towed at a nominal depth of 0.75 m. The data were recorded using a 72-channel, $\sim$ 450-m-long Geometrics GeoEel hydrophone streamer sampled at 0.25 ms. Streamer tow depth ranged from $\sim$ 1 to 9 m, depending on sea state. Two calibrated hydrophones were towed below the sparker source to record near-field source signatures at each source point, enabling deconvolution of the mixed-phase and highly variable source (Kluesner et al., 2018). All positioning was determined relative to the ship’s GPS, with streamer and source positions assumed to follow the ship track.

The sparker data were processed through a workflow including: crooked-line common-midpoint (CMP) binning at 3.125 m spacing; resampling from 0.25 to 0.5 ms, 70-400 Hz band-pass filtering; source designation and deghosting using recorded source signatures; trace editing; despiking; F-K filtering; semblance analysis for stacking velocities; normal-moveout correction; receiver static correction and three-trace median filtering in the channel domain; stacking; and post-stack time migration. Versions of the migrated images with a 200 ms automatic gain control (AGC) and spiking deconvolution applied were used for interpretation. The sparker source provides usable reflections up to 1 km below the seafloor but penetration depends heavily on the subsurface geology. Horizontal resolution of the data is set by the 3.125 m CMP spacing. Vertical resolution based on the peak frequencies in the migrated images is between 0.42 and 3.75 m.

We conduct joint interpretation of sparker data with lower-resolution but deeper-penetrating 2-D multi-channel seismic reflection data from the Cascadia Seismic Imaging Experiment (CASIE21) collected on the *R/V Marcus G. Langseth* in 2021 (Car-

botte et al., 2023). The data were collected using a 12 to 15 km streamer with a 6600 cu in 36-airgun array source. Processing of the data to pre-stack depth migration was carried out by ION Geophysical. This study is based upon the analysis of the 11 northernmost sparker dip lines and 8 nearby CASIE21 lines which cover the LVZ (Fig. 5.1a). These lines are spaced from 30 km to a maximum of 60 km apart. The lines traverse the accretionary wedge from the incoming plate to the upper continental shelf, with lengths ranging from 100 to 125 km. The usable sparker penetration reaches about 1 second two-way travel time (TWTT) below seafloor, or slightly less than 1 km, while the CASIE21 lines image to >10 km below seafloor. All seismic reflection profiles shown here face north, with seaward (W) to the left and landward (E) to the right. To aid in the identification of active faulting, we used published GMRT 50-m gridded bathymetry (Ryan et al., 2009a) to conduct geomorphic analysis of bathymetric features such as scarps and anticlines.

2.5.2 Seismic and Bathymetric Interpretation

The time-domain sparker seismic sections were used to conduct a detailed stratigraphic and structural analysis of near-surface deformation in the wedge and in shallow lower slope basins that lie atop the accreted material. Principles of seismic stratigraphy were employed to identify packages within each basin displaying similar seismic reflection characteristics, known as seismic facies (Mitchum et al., 1977; Posamentier et al., 2022). Significant reflectors separating these packages were picked as horizons using the HIS Kingdom Suite seismic interpretation software package. We focus on slope basins because they represent the modern depocenters, where sediments are more likely to accumulate continually than on steeper anticlinal structures, allowing characterization of late-Pleistocene to Holocene deformation.

Faults we map here are either surface-breaking or buried (“blind”). For surface-breaking faults, we determine if they are active by whether they deform the surface with a discrete fault scarp, which definitively indicates Holocene activity (e.g., Mc-

Calpin and Nelson, 2009). Activity on buried faults is assessed by documenting the geometry of strata deposited in overlying or adjacent basins. Some active faults have seafloor fold scarps but we do not discriminate between small and large active folds because it is unknown how much of this difference can be explained by potentially varying sedimentation rates between basins or by differences in rate or timing of folding (e.g., Fig. 5.2b-c).

Types of deformation and stratigraphic characteristics considered in our analysis are shown in Fig. 5.2. The most common reflection terminations in our seismic data are onlap, a base-discordant relationship where horizontal or inclined strata terminate against a surface with greater inclination (Mitchum et al., 1977). This is commonly seen at basin boundaries (Fig. 5.2e) or at seismic sequence boundaries within basin sediments (Fig. 5.2d). We also note intrabasinal unconformities similar to others which have been noted in the past (e.g., McNeill et al., 2000). Thickening or thinning of stratal units onto an anticline or a fault, known as growth strata, are deposited during deformation, and are used as a proxy for discerning activity on buried faults underlying anticlines next to each basin (Suppe et al., 1992). Previous work has shown that sediment deposition occurs largely on slopes below 5° , and that sediment bypass is most likely on steeper slopes (e.g., Hill et al., 2017; Ross et al., 1994). Therefore, we consider tilting of basinal units above 5° a result of tectonic activity rather than sediment deposition (Fig. 5.2e) and we mark this angle on each sparker profile accounting for vertical exaggeration. We measure the average angle of each basinal unit adjacent to the flanking anticlines using an assumed 1700 m/s sediment velocity to determine if it exceeds this threshold. In areas where the seafloor is steeper than 5° and nearby basins are not able to record growth, this method cannot be applied.

It is well-established that the age of the outer accretionary wedge in this region is Pleistocene (e.g., Silver, 1972)—with fossil evidence indicating that the outermost ridges are less than 2 Ma (Barnard, 1978)—and that there is a major structural

boundary and age difference between the outer wedge and inner wedge domains (e.g., Fisher et al., 1999; Flueh et al., 1998; McNeill et al., 1997). We determine the recency of deformation based on relative ages of stratigraphic units in slope basins atop the wedge; since basinal sediment deposition continues to the present-day, the characteristics of the uppermost sediments allow us to determine the recency of deformation, which is likely much younger than the age of the wedge itself. Sediment accumulation rates within outer wedge slope basins offshore Washington varied between 2 and 40 cm/ 10^3 years during the late postglacial 0-6750 B.P. and 6 and 29 cm/ 10^3 years during the early postglacial 6750-10,000 B.P. (Barnard, 1978; Caulet, 1995). Assuming a sediment seismic velocity of 1600 m/s and using the maximum sediment accumulation rate, the Holocene represents 0.006 seconds TWTT, which is at most the top few reflectors of the sparker data used in this study. This vertical resolution constraint prevents us from determining whether observed deformation is explicitly Holocene unless discrete fault offset reaches the seafloor, producing a seafloor scarp. Instead, we categorize active faults as late Quaternary, which encompasses the Holocene and the latest Pleistocene, corresponding conservatively to the last 100,000 years, but in many cases much less. Older faults are classified as broadly older Quaternary, indicating activity before 100,000 years ago but since the known outer wedge formation at 2.58 Ma. Our use of high-resolution data allows us to discriminate the faults that exhibit evidence of activity in the very late Quaternary, and in some cases Holocene, from those which do not show evidence of true surface deformation, which we consider inactive. Essentially all of the inactive faults identified herein are listed as active in published datasets like the Quaternary Faults Database (Schmitt, 2017) and the DOGAMI neotectonic map (Goldfinger et al., 2023); hence our work constitutes a significant advance in identifying the most recently active structures.

We employ the GMRT 50m gridded bathymetry (Ryan et al., 2009a) to map bathymetric signatures (fault-propagation folds or fault scarps) of active faults along strike between seismic lines (Fig. 5.3a). We quantify fault strike and length for all lo-

cated faults as well as the height of seafloor signatures identifiable on the bathymetry (Fig. 5.3b). The bathymetry is also used to confirm the vergence of each fault by noting the steeper side of each anticline. For fault-propagation fold anticlines where the fault itself is not imaged in the sparker data, but is visible in the CASE21 MCS data, we map their location at the place where the anticline meets the seafloor, accounting for the fault vergence. These anticlines can be determined to be actively deforming or not based on the onlapping relationship of basin-fill strata. Inferred strike-slip faults are mapped based on the offset of anticlines visible in the bathymetry.

Once faults are identified in individual high-resolution seismic sections in this manner, we compare them to imaged faults in nearby or coincident lower-resolution CASIE21 seismic lines to ensure that identified splay faults are linked to the megathrust at depth (Fig. 5.3d). This analysis helps determine the vergence for faults that are near-vertical as they approach the surface or for those underlying fault-propagation fold anticlines. Combining these observations from the sparker data, CASIE21 data, and the bathymetry allows us to recognize faults which are continuous across multiple lines and link these in our maps. Minor faults which have little offset, are not spatially extensive, and/or are not imaged at depth in the CASIE21 data are shown in seismic profiles but are not included in our final maps.

2.6 Results

2.6.1 Defining the active and inactive domains

We located faults throughout the outer wedge of the LVZ using seismic reflection data (Fig. 5.4). A comparison of slope basins within 30 km of the deformation front to those located in the innermost outer wedge closer to the inner-outer wedge boundary shows that they record a profound difference in deformational history and tectonic stratigraphy. This leads us to establish two main zones of the outer wedge, the *active domain* and the *inactive domain*. The boundary between the active and inactive

domain follows a similar map pattern to the inner-outer wedge transition along the margin. We display transects across the active-inactive domain boundary to show how basinal characteristics differ on either side (Fig. 5.5).

The active domain of the outer wedge varies in width from ~ 15 to 30 km, defined by the prevalence of late Quaternary faulting at the deformation front and landward. It is characterized by prominent seafloor fault signatures and seismically imaged faulting or basinal growth strata to the seafloor. Most basins in the active domain are filled with syndeformational packages which indicates continuous fault activity during sediment deposition throughout the lifetime of the basin (Fig. 5.5a, f). Some basins within the active domain contain prominent faults which offset the seafloor (Fig. 5.5a, f). Others do not contain identified surface-breaking faults but show tilting of topmost packages at greater than 5° indicating activity of faults underlying adjacent anticlines.

We define the inactive domain as the landward portion of the outer wedge which lies between the active domain and the inner wedge and contains mostly older Quaternary faults. It varies in width from ~ 20 to 40 km and is characterized by flat, wide, and thickly sedimented basins with deeply buried faults. Only one seafloor fault signature is evident in the multibeam bathymetry in this region. These basins record syndeformational deposition of sediments in their bottommost units with widespread imaged faults (Fig. 5.5b), thinning of units, or small-scale folding (Fig. 5.5g). However, in contrast to the active domain, basins in the inactive domain have postdeformational topmost packages which show no stratal thinning or thickening (Fig. 5.5b, g). Usually, a distinct unconformity separates these two types of tectonic strata (e.g., Fig. 5.5g).

2.6.2 *The active domain*

Concentrated late Quaternary faulting defines the active domain, where 35 of 38 active faults in the map area are located (Fig. 5.4). Landward-vergent thrust faults

dominate the active domain, with only four seaward-vergent active faults imaged. Fault strike of thrusts in the active domain varies between 322° at the north end to 17° at the south end of the map area, following a change in the obliquity of subduction along the margin. Fault length varies from 7.5 to 70.2 km, with an average length of 29.7 km.

The active domain includes the deformation front which rather than being a single continuous fault is segmented along strike by mostly WNW-striking strike-slip faults into at least seven parts (Fig. 5.4). The deformation front region is characterized by thin basinal sediments with some faults which produce measurable seafloor scarps, indicating Holocene activity. The frontal thrust in the map area is primarily landward-vergent, but some short seaward-vergent frontal thrusts are present south of Willapa Canyon. Incipient thrust formation in multiple locations indicates the continued advance of the deformation front seaward. Eastward of the deformation front, the geomorphology of the active domain is typical of that of the LVZ, with widely spaced thrust sheets overlain by thick basinal sediments. Instead of seafloor scarps that characterize frontal thrust, evidence of faulting such as folding or warping of topmost sediment packages are more prevalent in the bathymetry.

At the northern end of the LVZ, the active domain is narrow, about 15 km in width, and contains landward-vergent faults which underlie closely spaced anticlines separated by small basins. The frontal thrust is cut by strike-slip Fault A at a large landslide scarp on the frontal anticline around 47.6°N , causing a 2 km left-lateral offset of the frontal thrust (Fig. 5.4). Here, the active domain widens to 28 km but the fault spacing remains similar to north of Fault A at 8 km or less between each landward-vergent splay fault. A new landward-vergent thrust is the dominant frontal thrust south of Fault A, which is again cut by strike-slip Fault B, showing less (~ 1 km) apparent left-lateral offset. The first fault east of the deformation front is an active landward-vergent thrust with a prominent 13 m high seafloor scarp 24 km in length, located in the middle of an 8 km wide basin (Fig. 5.3). This fault is well-

imaged in two sparker lines as well as a CASIE21 MCS line, allowing us to trace it from the décollement to the near-surface (Fig. 5.3c-d). Though the fault tip does not break the surface at the available resolution of imaging, the uppermost sediment package is warped, and there are abundant growth strata in the basin, indicating late Quaternary activity of this fault (Fig. 5.3c).

Two active landward-vergent faults appear south of Fault B in an 18 km wide basin that crosses the active-inactive domain boundary (Fig. 5.3d). The westward fault in the fault pair has a seafloor fold 48 m high (Fig. 5.5a). These faults branch in the near-surface but connect at depth as well as merging into one fault along strike at the South Nitinat Fault (Fig. 5.5d). Between the North and South Nitinat strike-slip faults, a new 16 km long incipient frontal thrust forms seaward of the previously dominant frontal thrust. In this area, the deformation front is composed of two imbricated fault-propagation folds underlain by landward-vergent faults approximately 1 km apart.

Just south of the South Nitinat Fault, the two frontal thrusts are spread farther apart to 4 km, and the westward thrust is the dominant frontal thrust for 61 km along the margin, while the second thrust is continuous for 52 km (Fig. 4.6a). In this region, the active domain is 20 km wide and is characterized by fewer anticlines separated by wider basins with a thicker and smoother sediment cover. The deformation front is defined by an imbricated fault-propagation fold where two landward-vergent faults break to the seafloor; the eastward one has a scarp of 70 m visible on bathymetry (Fig. 4.6b). At 46.7°N, the frontal thrust scarp is 44 m in height where it is visible in a large coulee (Fig. 4.6a; described in Beeson et al., 2017). This frontal thrust is segmented at 46.4°N where the right-lateral offset on Fault E displaces the anticline seaward by 4 km. At the active-inactive domain boundary, two short, 9 km long inactive seaward-vergent faults are imaged.

We identify two, short (~ 14 km) incipient frontal thrusts at 46.3°N forming 5 km seaward of the well-developed frontal thrust offset by Fault E (Fig. 4.6a). The seaward-vergent incipient thrust is cut by the more active landward-vergent incipient

thrust, which produces a seafloor signature (Fig. 4.6c). Just south of 46.2°N , both incipient frontal thrusts as well as the previously dominant frontal thrust end as the deformation front steps landward by 15 km to a seaward-vergent thrust for a 32 km span, which decreases the width of the active domain from 22 km to 15 km. Both incipient frontal thrusts, the well-developed landward-vergent frontal thrust, and the seaward-vergent frontal thrust are imaged in Fig. 4.6c.

Between the Willapa Canyon Fault and Fault F, mixed vergence dominates both the active and inactive domains (Fig. 5.4). The width of the active domain is at its minimum value, around 15 km, when crossing Astoria Canyon, and begins to increase again to the south. Regularly spaced, 15 km wide thrust sheets with a smooth sedimentary cover are typical in this area. Faults in basins which create seafloor folds or fault propagation fold anticlines with basinal growth strata to the surface remain common in this region (Fig. 4.6d).

South of Astoria Canyon, the deformation front is defined by a continuous landward-vergent frontal thrust 65 km in length (Fig. 4.6a). While this fault itself is deep-seated and not well-imaged in sparker data, sediment packages onlapping onto the frontal anticline indicate recent activity (Fig. 4.6d). This landward-vergent frontal thrust cuts a less active, shorter, 13 km long seaward-vergent thrust at the deformation front (Fig. 4.6f). Farther south at 45.3°N , the offset on this fault is smaller, so we classify it as a protothrust and do not extend it on the map. The second active fault at 45.5°N comes close to the seafloor, producing a large fold with a seafloor signature (Fig. 4.6d-f). This fault is the dominant frontal thrust at 45.3°N and produces a prominent 20 m seafloor scarp in this region (Fig. 4.6f).

From 45.5°N to 45.1°N , we observe active faulting to the inner-outer wedge boundary, so we extend the active domain to 40 km in width to encompass the entire outer wedge (Fig. 4.7a). Fault spacing decreases to 10 km and anticlines become more closely spaced and less distinct. Thinning and warping of the uppermost sediment package is observed above three active landward-vergent thrusts at 45.5°N (Fig. 4.7b).

The westernmost active fault imaged in Fig. 4.7b has an 80 m high fold at the seafloor. Farther south, by 45.3°N , the typical geomorphic character of the LVZ is replaced by an uplifted plateau-like region without distinct anticlines. In this area, faulting to the surface is widespread (Fig. 4.7c). Onlapping sediment packages with no stratal thinning appear to the east of this active plateau, but they do not drape the uplifted zone, so it is uncertain whether the surface-reaching faults we image are active in this area. Regardless, we classify this area as active, though its features are much different from the active domain we observe in most of the LVZ.

2.6.3 *The inactive domain*

In the inactive domain, we identify 26 older Quaternary faults and three late Quaternary active faults, in addition to four inferred but not imaged thrusts. These faults exhibit lengths ranging from 5.4 to 64.2 km, with an average length of 25.1 km. The inactive domain fault strikes vary between 316° in the north and 345° in the south end of the LVZ. Landward vergence is observed for 21 of 29 imaged faults.

The inactive domain at the northern boundary of the LVZ at 48°N is approximately 20 km wide (Fig. 5.4). We map only two faults in this area which are traces of two seafloor anticlines near the active-inactive domain boundary. These faults are short, 19 and 26 km in length, and are not crossed by any seismic lines, so are mapped as inferred thrusts. The Juan de Fuca Canyon turns toward the north at 47.55°N , covering the inactive domain with channel deposits in this region and obscuring possible faults.

South of the Quillayute Channel, distinct anticlines reappear, and minimal channel deposits allow the mapping of many inactive faults in this region. Between 47.4°N and 46.7°N , there is a region of prominent older Quaternary landward-vergent thrusting. The faults here are closely spaced at approximately 4 km apart. We map 11 of these faults in this area, ranging from 11 to 55 km in length. Basins in this region show flat strata with no deformation in the uppermost several sediment packages, indicating

no recent activity, with some examples of buried faults (Fig. 5.5b) and unconformities (Fig. 5.5g). Just south of Quinault Canyon is the widest point of the inactive domain at 41 km across.

South of Grays Canyon, the inactive domain decreases in width again to about 25 km wide (Fig. 4.8b). Channel deposits and wide line spacing limit imaging of deformation in this region, but we do identify two inferred thrust faults 29 and 15 km in length based on seafloor geomorphology (dashed black lines, Fig. 4.8b). Between Willapa Canyon and the southern boundary of the inactive domain, there is a zone of mixed vergence. Anticlines increase in number and have a consistent spacing of approximately 8 km separated by wide, flat basins. The strike-slip Willapa Canyon Fault marks another wide point of the inactive domain where it measures 41 km across. The sparker imaging does not capture the area between Willapa and Astoria Canyons, but we use the CASIE21 data to identify four older Quaternary faults here, two of which have seaward vergence. These faults appear discontinuous and do not cross to the south of Astoria Canyon.

In the southernmost section of the inactive domain, around 45.9°N, vergence of faults alternates and fault spacing is 7 km on average (Fig. 4.8b). There are four inactive landward-vergent and two inactive seaward-vergent thrusts imaged, ranging in length from 11 km to 64 km. The basin landward of the seaward-vergent thrust just south of Astoria Canyon shows growth strata at depth and flat-lying strata in the near-surface separated by at least one prominent unconformity (Fig. 4.8c), confirming that both faults underlying the anticlines adjacent to this basin and imaged in the CASIE data are inactive (Fig. 4.8f). Another nearby basin shows a similar decrease in the presence of growth strata from depth to surface, with little activity at present (Fig. 4.8e). These basins are typical of those imaged throughout the inactive domain.

Faults near the southern termination of the inactive domain also show larger differences in evidence of activity along strike (Fig. 4.8b). For instance, the anticline nearest to the inner-outer wedge boundary has two inactive landward-vergent branches while

the main branch shows late Quaternary activity (Fig. 4.8b). An active splay from the middle of the main thrust has a seafloor anticline height of 160 m above the adjacent basin. Similarly, just south of strike-slip Fault F and near the active-inactive domain boundary, a structural high emerges where a landward-vergent thrust changes activity along strike from inactive to active at 45.65°N (Fig. 4.7a, 4.8b). We terminate the inactive domain at this latitude as active faults become common throughout the entire outer wedge to the inner-outer wedge transition.

2.6.4 The inner-outer wedge transition

Four seaward-vergent inactive faults are adjacent to or co-located with the inner-outer wedge boundary (Fig. 5.4). In the LVZ, we observe the presence of these faults in the CASIE21 data, but none show activity in the near-surface sparker data, indicating they are currently inactive. The northernmost fault associated with this boundary is at 47.5°N where an inactive, 21-km-long fault-propagation fold anticline crosses the boundary at a 19° strike. This strike direction is unique compared to the other faults we image located near the inner-outer wedge transition. In the basin to the west of this anticline, a short, 5 km inactive seaward-vergent thrust and a longer, 14 km active landward-vergent thrust are present.

A second seaward-vergent inactive thrust 13 km in length is located just south of Quinault Canyon (Fig. 4.8b). Like the remainder of the faults we discuss here, this thrust occurs at a slope angle that marks the inner-outer wedge transition. Similarly, a third imaged fault is located at the base of the slope between Grays and Willapa Canyons. This seaward-vergent thrust at the inner-outer wedge transition is longer at 43 km in length. Though this fault is clearly imaged in the CASIE21 data in this region, the sparker data do not image the fault itself (Fig. 4.8a). Rather, they capture the overlying sediments, which form a steep, non-depositional slope. The sedimentary package west of this slope documents the bending of strata at depth unconformably overlain by flat-lying sediment, indicating this region is no longer

uplifting and the fault responsible for the older phase of basin deformation is no longer active. However, the fault imaged in the CASIE21 data appears to intersect the seafloor on a sloping surface that has no recent sediment drape discernible in the sparker imagery. Therefore, we cannot rule out the possibility of recent fault activity along the CASIE21 structure.

The southernmost imaged fault, located south of Astoria Canyon, is the longest one associated with the inner-outer wedge transition in the LVZ, measuring 58 km in length. The basin to the west of the slope break shows growth and tilting at depth but decreasing activity to present (Fig. 4.8d). Similar to the previous examples, we consider this fault inactive.

2.6.5 *Strike-slip faults*

Strike-slip faults at a high oblique angle to the convergence direction cross-cut these zones of activity and may offset both active and inactive thrust fault structures. Strike-slip faults are inferred using bathymetry rather than in the seismic profiles, where oblique structures can be difficult to identify. In general, we observe strike-slip faults offsetting anticlinal ridges in the active and inactive domains resulting in the segmentation of thrust faults into shorter sections along strike. We identify 9 left-lateral and 1 right-lateral strike-slip faults, 7 of which strike west-northwest at 93° to 112° (Fig. 5.4). The remaining 2 faults strike northeast at nearly 90° to the predominant strike-slip fault orientation. Most of the strike-slip faults we map are located in the northern portion of the LVZ between the Juan de Fuca and Willapa Canyons, with two faults south of the Willapa Canyon Fault.

Fault A is 26 km in length and newly identified in this study based on the 2 km offset of the frontal thrust at 47.6°N . Fault B is coincident with an unnamed thrust fault in the Quaternary Faults Database (Schmitt, 2017) but we interpret the offset along this fault to be mostly left-lateral due to the apparent offset of the two anticlines it crosses. The location of the North and South Nitinat Faults are slightly modified

from representation of these faults in the Quaternary Faults Database (Schmitt, 2017) based on our analysis of the GMRT bathymetry in this study. The North Nitinat Fault crosses the frontal thrust, as well as two thrusts at the edge of the active domain, although not much offset of these anticlines is observable on the seafloor. The South Nitinat Fault appears to offset the frontal thrust by 4 km at 47.05°N.

Faults C and D are newly identified left-lateral strike slip faults 20 and 42 km in length, respectively, based on the seafloor geomorphology of anticlines in this area and lineations present in the bathymetry (Fig. 5.4). Neither of these faults is present in the Quaternary Faults Database (Schmitt, 2017) nor in the DOGAMI neotectonic map (Goldfinger et al., 2023). Fault E is modified in our study from initial observations by Beeson et al. (2017) and is the only right-lateral strike-slip fault in our map area. It offsets both the frontal and second anticlines by 2.5 km and has the highest apparent offset of any strike-slip fault in our study.

Just south of Fault D, the Willapa Canyon Fault is mapped at an updated location relative to the Quaternary Faults Database (Schmitt, 2017) due to observed offset along anticlines in both the active and inactive domain on the multibeam bathymetry. South of the Willapa Canyon Fault, there are only two additional strike-slip faults we identify (Fig. 5.4). Fault F is 28 km in length and is updated from an unnamed fault in the Quaternary Faults Database (Schmitt, 2017) based on its offset of several faults in the inactive domain. The southernmost strike-slip fault in our study area is the Wecoma Fault, whose location has been well-established by others (e.g., Appelgate et al., 1992; Tobin et al., 1993).

2.7 Discussion

We document widespread late Quaternary active faulting within the outermost, active domain of the outer wedge in the LVZ, and show that recent fault activity diminishes landward and is nearly nonexistent in the inactive domain. Based on the evidence we have presented above, we propose a conceptual model of Quaternary wedge evo-

lution (Fig. 4.9) in which a concentrated zone of deformation spanning 3-4 individual thrusts and associated thrust sheets forms a continuously seaward-migrating zone of concurrent deformation 15 to 30 km wide; basins landward of this zone subsequently fill with thick sediment and fault activity subsides, creating a zone of no recent fault activity 20 to 40 km wide—the inactive domain.

2.7.1 Deformational domains and along-strike changes

We observe a major transition in outer wedge fault vergence and late Quaternary fault distribution occurring between 46.3°N and 45.5°N , separating the northern LVZ and the southern map area (Fig. 5.4). There is a transition at 46.3°N from predominantly landward vergence in the north to mixed vergence in the south. South of Astoria Canyon, the inactive domain decreases in width suddenly from 32 km at 46°N to 0 km at 45.5°N . The southernmost portion of our map area displays a different morphology from the rest of the LVZ, with the smallest outer wedge width and no region of widespread buried inactive faults (Fig. 4.7a). This contrast in outer wedge characteristics we document aligns with previous work that has designated a major along strike segmentation boundary at 46°N due to a change in the morphotectonic character of the accretionary wedge, such as a decrease in outer wedge width and an increase in outer wedge slope (Watt and Brothers, 2020). This leads us to conclude that the pattern of active and inactive domain faulting we propose here is restricted to the LVZ between 48°N and 45.5°N and does not continue to the south of this area.

We detect only rare and spatially limited recent deformation associated with the inactive domain or in the inner-outer wedge transition. While there are seaward-vergent thrusts coincident with this boundary, notably the two faults between Grays and Astoria Canyons and the fault south of Astoria Canyon, analysis of the sparker seismic data indicates that deformation associated with these faults has decreased over time (Fig. 4.8a, 4.7c). Faults associated with this boundary may appear well-developed in crustal-scale seismic reflection data, indicating a large amount of cu-

mulative slip, but such data sources lack the resolution to discriminate recent (i.e., late Quaternary) slip from longer timescale cumulative deformation. Our analysis suggests that faults at the inner-outer wedge transition in our map area have not experienced late Quaternary slip and thus are inactive at present.

While northwest-southeast oriented strike-slip faulting in this region has been well-documented (e.g., Appelgate et al., 1992; Han et al., 2018; McNeill et al., 1997; Goldfinger et al., 1996), there has been little investigation into the mechanics of faults oriented northeast-southwest, such as Fault E (Beeson et al., 2017) and Fault F. We note that these faults strike at the same orientation as propagator wakes in the downgoing plate (Nedimović et al., 2009), suggesting that strike-slip fault orientation may be influenced by lower plate topography (Fig. 5.4). Differential offset of anticlines along strike-slip faults may illuminate relative timing of fault formation. For example, the offset of the frontal thrust at Fault B (~ 0.5 km) is smaller compared to the offset of the second anticline thrust (~ 1.4 km), indicating that the frontal thrust is more recently formed in this region and there has been less time for strike-slip motion to offset it.

2.7.2 Uncertainties in mapping faults without age constraints

Because the morphology of active fault seafloor deformation depends on a variety of unknown factors, such as the integrated slip rate, the syntectonic sedimentation rate, and fault dip (Chiama et al., 2023), we have not distinguished between surface-breaking or buried active faults. However, we note cases of fault scarps with discrete surface offset, which definitively indicates Holocene and likely coseismic activity (McCalpin and Nelson, 2009). Three faults meet this criterion, all of which are located at the deformation front (Fig. 4.6b, e). The remaining faults in the active domain often produce seafloor fault signatures, but their fault tips do not reach the surface with displacement more than the vertical resolution of the sparker data (< 3.75 m). This difference is likely because the frontal thrust faults are more recently formed rather

than a difference in recency of activity between late Quaternary faults. Additionally, slope basin sediments are composed largely of turbidites and debris flows derived from adjacent anticlines and thus sedimentation rates are dominated by local, non-average effects and may vary widely between individual basins (Barnard, 1978; McAdoo et al., 1997), as well as by the availability of accommodation space within each basin. This variability prevented us from identifying any reflectors which correlate across basins. The confounding effect of variable sedimentation rate means that we cannot quantitatively determine slip rates nor per-event displacements for specific faults without widespread coring and age dating of individual slope basins.

2.7.3 Conceptual model of wedge evolution during the Quaternary

The Cascadia subduction zone, particularly the LVZ, is known for its thick supra-wedge sedimentary cover among subduction zones worldwide. Sediment supply and accumulation rates are heavily influenced by climatic factors such as ice ages; for example, heavy sedimentation during the last glacial period, which ended approximately 11,000 years ago, extensively flooded the accretionary wedge in this region (Barnard, 1978; Underwood, 2002). Glaciation extended to the Juan de Fuca Strait, but glacially derived sediments were deposited largely between the Juan de Fuca and Astoria Canyons (Booth et al., 2003; Hill et al., 2022; Riedel et al., 2018).

Heavy wedge-top sedimentation has been shown to influence accretionary wedge evolution and mechanics. Mechanical modeling indicates that high hinterland sediment input can produce widely spaced thrust sheets with low surface slope (Simpson, 2010), similar to the morphology of the LVZ at Cascadia. Specifically, high slope basin sedimentation tends to suppress underlying fault activity (Fuller et al., 2006), so that deformation either migrates outside of the sedimentation zone or resumes after sediment rate decreases by reactivating or forming out-of-sequence thrusts to reattain critical taper (Mannu et al., 2016, 2017).

Based on these previous findings and our results here, we present a conceptual

model of wedge evolution for the LVZ during the Quaternary (Fig. 4.9). At the end of the Pliocene, the entire accretionary wedge consisted of the present-day inner wedge (Fig. 4.9a). At the ~ 2.58 Ma onset of Pleistocene glaciation, the accretionary wedge began to grow seaward by actively forming the now-inactive domain of the outer wedge through the formation of dominantly landward-vergent faults (Fig. 4.9a; e.g., Adam et al., 2004). Deformation during this time was widespread in the present-day inactive domain, and syndeformational strata were deposited in basins 1 and 2 (Fig. 4.9b). Heavy glacially induced sedimentation caused the deformation front to step seaward rapidly and continuously during the Pleistocene, creating the entire present-day outer wedge (basins 3-5). As deformation migrated outward, faults deactivated in basins 1 and 2 and were buried by thick postdeformational strata (Fig. 4.9c). Deposition of these strata into forearc basins “turns off” the underlying faults, forcing migration of the active domain seaward (Fuller et al., 2006; Mannu et al., 2017; Simpson, 2010). We posit that an active domain consisting of the deformation front accompanied by a ~ 30 km region of active faulting therefore steps seaward continually over time, with faults trailing behind this zone becoming largely inactive. Our documentation of multiple incipient frontal thrusts and surface-breaking scarps at the deformation front suggests that this process is ongoing (e.g., Fig. 4.6b, c, e).

The outer wedge changes character south of Astoria Canyon because glacial sedimentation largely was deposited to its north (Booth et al., 2003; Riedel et al., 2018), and Holocene sedimentation has also been markedly lower south of Willapa Canyon (Barnard, 1978). We posit that the comparatively lower sediment input in our southernmost map area caused the deformation front to step seaward less significantly than in the rest of the LVZ and thus did not create an inactive domain in this region. Basins in the present-day outer wedge were likely not buried as completely and fault activity did not entirely deactivate as in the inactive domain elsewhere, instead remaining active throughout the Pleistocene (Fig. 4.7c, d).

Our conceptual model argues that rapid wedge-top sedimentation is the main con-

trol on atypical morphology and distribution of fault activity of the LVZ at Cascadia, and the key role of sediments in controlling wedge morphology is bolstered by similar observations at other sediment-rich subduction zones and modeling work based on critical wedge theory (e.g., Mannu et al., 2016; Simpson, 2010). For example, fault activity mapping at Makran—which has both high sediment thickness on the incoming plate as well as a high hinterland sediment supply—identified a belt of continuously active seaward-vergent faults close to the deformation front, with activity diminishing landward (Smith et al., 2012), similar to the pattern of landward-vergent faults we document at Cascadia. Along-strike morphotectonic changes at Sumatra are linked with variability in incoming sediment thickness, and the location of landward vergence is correlated with the highest sediment input (McNeill and Henstock, 2014), analogous to the Cascadia case. Wedges with a high hinterland sediment input and which exhibit low taper are supercritical, rather than subcritical, but still stable (Simpson, 2010). This can be ascribed to an increase in vertical stress which in turn reduces differential stress and renders fault slip unfavorable (Simpson, 2010).

2.7.4 Implications for megathrust rupture and tsunami generation

Recent large megathrust events have highlighted the potential for shallow fault slip, both along the décollement and on splay faults within the accretionary wedge. For example, seismic and geological work has established that two splay faults slipped coseismically during the 1964 M_w 9.2 Alaska earthquake (Liberty et al., 2013; Plafker, 1969). Seafloor scarps documented shortly after the 2004 M_w 9.1 Sumatra earthquake indicate that wedge faults either slipped coseismically, or bending moment faulting occurred while the décollement slipped below them (Henstock et al., 2006). Fault scarps at the deformation front created by the M_w 9.1 2011 Tohoku-Oki earthquake have been identified and calculations suggest that up to 120 m of slip was accommodated (Ueda et al., 2023).

In this work, we have documented near-surface late Pleistocene to Holocene deformation

on numerous splay faults offshore Washington and northern Oregon, and previous studies using lower-resolution seismic reflection data have shown that displacement at depth amounts to kilometers of slip (e.g., Adam et al., 2004; MacKay, 1995; Booth-Rea et al., 2008). The observed displacement on active faults in the outer wedge must be formed by either (a) continuous steady creep, (b) rapid coseismic slip, (c) afterslip immediately following a large rupture, or (d) some combination of these processes. The accretionary wedge in this region is known to be either strongly geodetically locked or lying in a stress shadow zone; either way, it is accumulating slip deficit rather than steadily creeping (Lindsey et al., 2021).

Discriminating between slow steady creep during the interseismic period and rapid coseismic slip and afterslip during megathrust rupture in sparker seismic reflection profiles is difficult and likely requires combined detailed bathymetric and structural analysis based on even higher resolution imaging (e.g. AUV bathymetry and chirp). Limited other studies have documented that afterslip may represent 20% to 50% of the documented total slip for megathrust events, the remainder of which is coseismic (e.g., Brooks et al., 2023; Lin et al., 2013). This implies that even if some of the offset visible on seismic data is due to afterslip, at least a portion to a majority of the accumulated slip on each fault is coseismic in nature. Considering these lines of reasoning, we therefore conclude that a large portion of the active domain deformation we show here is likely rapid, coseismic slip that produces seafloor displacement and therefore uplift during megathrust earthquakes. The discrete surface scarps with brittle offset that we document here (e.g., Fig. 4.6b, f) are categorically considered evidence of coseismic slip in terrestrial environments (e.g., Chiama et al., 2023), but even seafloor fold scarps we show would be geomorphic evidence for surface-rupturing earthquakes, and would be sites chosen for paleoseismic investigation were they on land (e.g., McCalpin and Nelson, 2009).

An additional argument favoring coseismic rapid slip events on these splays is that the frictional conditions for shallow slip are met at Cascadia due to the thick

sediment cover and young age of the subducting plate. The $100 - 150^{\circ}\text{C}$ threshold at which a variety of low-grade metamorphic processes occur is thought to correspond to the minimum temperature for earthquake nucleation at subduction zones (Hyndman and Wang, 1993; Moore and Saffer, 2001a). This condition is met at Cascadia above the plate interface and within the accretionary wedge sediments, suggesting that shallow seismogenic behavior on the megathrust and splay faults is possible (Salmi et al., 2017). Recent thermal modeling at Sumatra suggests that shallow diagenetic strengthening of wedge sediments prior to the deformation front due to thick sedimentary cover played a role in driving shallow slip during the 2004 M_w 9.1 event (Hüpers et al., 2017b; Stevens et al., 2021). At Nankai, core samples documenting frictional heating of shallow sediments from coseismic slip along the megathrust and megasplay faults confirm that shallow slip in accretionary wedges is possible (Sakaguchi et al., 2011; Yamaguchi et al., 2011). Earthquakes may in any case slip into shallow regions below the $100 - 150^{\circ}\text{C}$ threshold due to dynamic overshoot, as was likely seen in the 2011 Tohoku-Oki earthquake (e.g. Ide et al., 2011; Sun et al., 2017).

In the past, backward rupture branching was perceived to be improbable, so rupture propagation between faults connected at acute angles was not considered. However, dynamic modeling has found that slip on the megathrust can activate multiple seaward or landward-vergent accretionary wedge faults through dynamic triggering (van Zelst et al., 2022; Xu et al., 2015). Additionally, during the 2023 M_w 7.8 Kahramanmaraş earthquake, bilateral rupture propagated onto faults with variable dips and onto a splay fault through an acute fault bend likely due to a cascading dynamic rupture process, rather than direct triggering (Ding et al., 2023).

While identified active faults in map view are segmented along strike into ~ 30 km sections on average, it is possible that they would rupture together along strike and in concert with the décollement during large megathrust events. Multiple lines of reasoning suggest these faults would not rupture alone and produce their own earthquakes independent of the megathrust, including: they are in the stress shadow

of a locked megathrust (e.g., Lindsey et al., 2021); faults in compliant sediments are not strong enough to initiate coseismic rupture; and very little microseismicity is recorded in this region (e.g., Stone et al., 2018; Morton et al., 2023). There are documented examples of destructive earthquakes in fold-and-thrust belts on land or in subduction zone settings with simultaneous rupture of multiple, separate faults, and/or surface fault rupture on out-of-sequence faults. The 2016 Mw 7.8 Kaikoura earthquake saw rupture of at least 20 interconnected strike-slip and thrust faults with a maximum surface displacement of 12 meters and caused a local tsunami (Kaiser et al., 2017; Litchfield et al., 2018), possibly linked by an underlying décollement fault (Ulrich et al., 2019). The 2008 Mw 7.9 Wenchuan earthquake ruptured at least one out-of-sequence fault while no displacement was recorded on the frontal thrust or other major thrusts in the Longmen Shan fold-and-thrust belt (Hubbard and Shaw, 2009). We suggest future Cascadia megathrust events could similarly involve complex rupture of many splay faults in the outer wedge.

Tsunami excitation by splay faults depends on fault dip, length, and slip magnitude (Wendt et al., 2009). Splay faults imaged in this study are often listric and increase in dip in the near surface, with some approaching vertical as they intersect the seafloor. Similar high-angle near-surface splay faults have been observed using sparker seismic reflection data at other subduction zones, including the Cape Cleare and Patton Bay Faults in the Alaska subduction zone, with 70° dip in the near-surface and up to 39 m-high seafloor scarps (Liberty et al., 2013), but $30 - 40^\circ$ dip overall down to 20 km below the surface (Haeussler et al., 2015a). Faults with steeper near-surface dips have more potential for tsunami excitation because of greater vertical seafloor displacement (e.g., Qiu and Barbot, 2022). Models that have included splay fault rupture at Cascadia have shown that splay fault rupture can double median tsunami height compared to megathrust-only fault rupture (Aslam et al., 2021).

Modeling has shown that the tsunami from the 1700 Cascadia event can be reproduced in Japan by a minimum of an Mw 8.7 rupture, and that such a tsunami would

require a minimum fault source length of only 400 km to be created (Melgar, 2021; Satake et al., 1996). Since there is evidence that splay faults like those documented here may be able to rupture together with the megathrust, linking along strike to create longer rupture lengths, it follows that activation of multiple separate strands of landward-vergent splay faults could contribute to both far-field and local tsunami signal during a megathrust event. We support the inclusion of landward-vergent splay fault rupture in future tsunami scenario models.

Megathrust earthquakes that ruptured splay faults like the Alaska 1964 M_w 9.2, and as suggested for the Tonankai 1944 M_w 8.0 event, have involved long faults located at the inner-outer wedge transition (Moore et al., 2007b; von Huene et al., 2021), sometimes called “megasplays” (Tobin and Kinoshita, 2006a). Because of this, tsunami scenario models that incorporate splay faulting in Cascadia have used a margin-long splay fault located at that boundary (Gao et al., 2018; Witter et al., 2013). We identify only isolated, inactive faults at the inner-outer wedge transition. Our analysis suggests that activation of and slip on landward-vergent splay faults and their associated anticlines in the active domain of the outer wedge is more likely than that of inner-outer wedge boundary-related splay faults.

To synthesize these observations and considerations derived from wedge mechanics, we infer that our results are therefore consistent with a syn-rupture scenario in which the inner wedge and the inactive part of the outer wedge all act as a dominantly rigid upper plate block, translating seaward during megathrust slip without splay activation (Fig. 4.10). The large (potentially 10s of meters) of slip is transferred up the megathrust and distributed onto one to—more likely—several splay thrusts in the outermost, active domain, thereby creating enhanced seafloor displacement and multiple branch rupture over a 15-30 km width of the shallowest portion of the wedge (Fig. 4.10). Our results favor seafloor deformation of not the frontal thrust alone, but rather the entire active domain.

2.8 Conclusions

Using a combination of high-resolution near-surface and lower-resolution crustal-scale seismic reflection data and multibeam bathymetry, we identify and map splay faults in the outer wedge of the landward vergence zone of the Cascadia accretionary wedge over a region spanning ~ 400 km along strike of the largest geodetically locked portion of the megathrust. By conducting stratigraphic analysis of slope basins in the outer wedge, we determine recency of activity of splay faults. We identify two regions of the outer wedge: the active domain, spanning the region within ~ 30 km landward of the deformation front, which contains predominantly recently active landward-vergent splay faults; and the inactive domain, located between the active domain and the inner-outer wedge transition, which contains dominantly inactive, buried landward-vergent splay faults. We develop a conceptual model of continuous seaward migration of a ~ 30 km active deformation zone including the deformation front and several anticlines landward. Based on these results, we infer that, in megathrust events at Cascadia, complex outer wedge deformation including rupture on multiple splay faults is favored, and that slip is strongly partitioned to the active domain of the outer wedge. The potential of slip on multiple, simultaneously activated landward-vergent splay faults close to the deformation front should be accounted for in future tsunami hazard modeling scenarios.

Acknowledgements

Funding for this work was provided by the U.S. Geological Survey Coastal and Marine Hazards and Resources Program, the University of Washington, and the U.S. National Science Foundation through awards 2103713 and 2217468 to Tobin. We thank all of the individuals involved with collection and processing of high-resolution sparker seismic-reflection data, specifically Emily Roland, Jared Kluesner, and Alicia Balster-Gee, as well as the CASIE21 data, including Suzanne Carbotte, Shuoshuo Han, Brian

Boston, and Jeff Beeson. Any use of trade, product, or firm names is for descriptive purposes only and does not imply endorsement by the U.S. Government. James Conrad provided comments on an early version of the manuscript. The manuscript was greatly improved by reviews from Lisa McNeill and Utsav Mannu.

Data and code availability

The seismic data utilized in this study are publicly available. The high-resolution multi-channel sparker seismic data can be found as a USGS data release release (Balster-Gee et al., 2023, USGS field activity 2019-024-FA). The CASIE21 multi-channel seismic reflection data are hosted on Marine Geoscience Data System (Carbotte et al., 2023). Seismic interpretation was conducted using the *Kingdom Suite*.

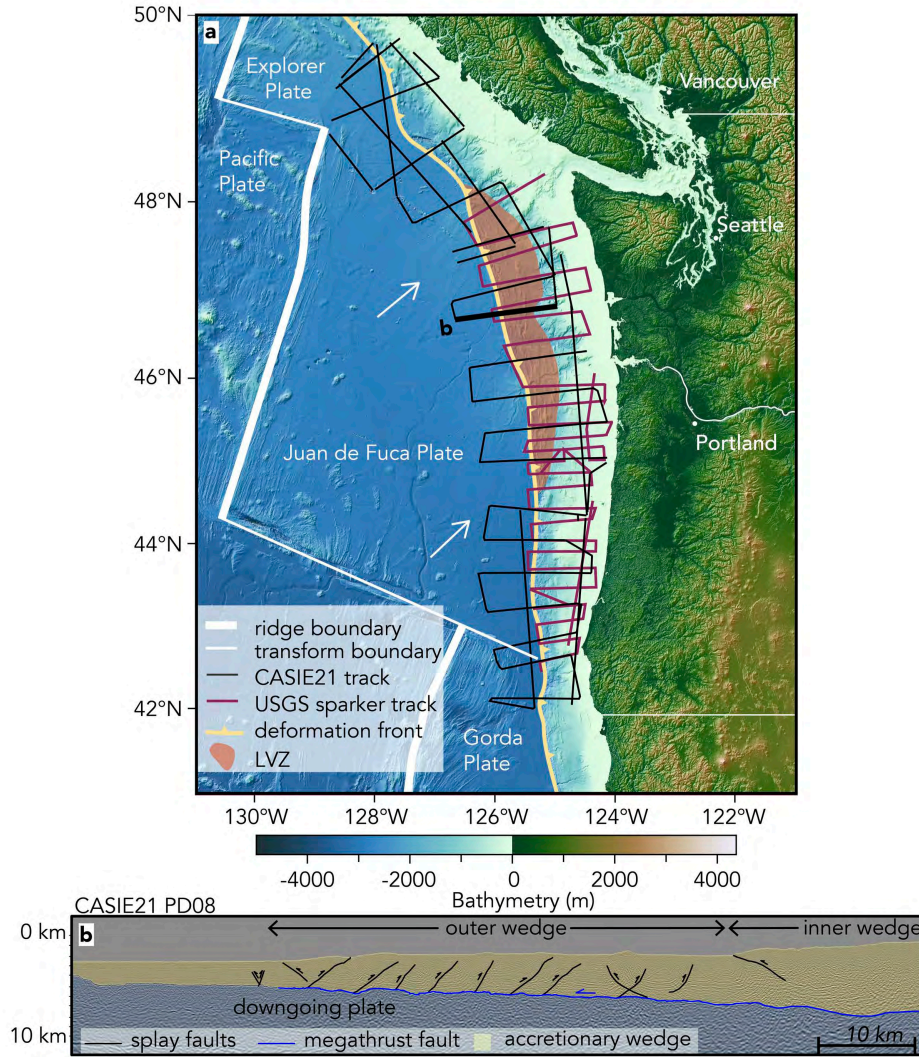


Figure 2.1: **a.** Map of the Cascadia subduction zone with deformation front in yellow. The region where landward-vergent thrusting dominates (LVZ) is shaded pink. Black and pink tracklines are for low-frequency CASIE21 MCS reflection data collected on the *R/V Marcus G. Langseth* and for the high-frequency sparker and chirp taken in 2019 on the *R/V Rachel Carson*, respectively. Incoming plate boundaries shown in white. Relative plate motion between the Juan de Fuca plate and the North American plate indicated by white arrows at rates varying between 42 mm/yr in the north and 34 mm/yr in the south (DeMets et al., 2010). **b.** CASIE21 Line PD08 showing the megathrust fault, splay faults, and the location of the inner and outer accretionary wedge.

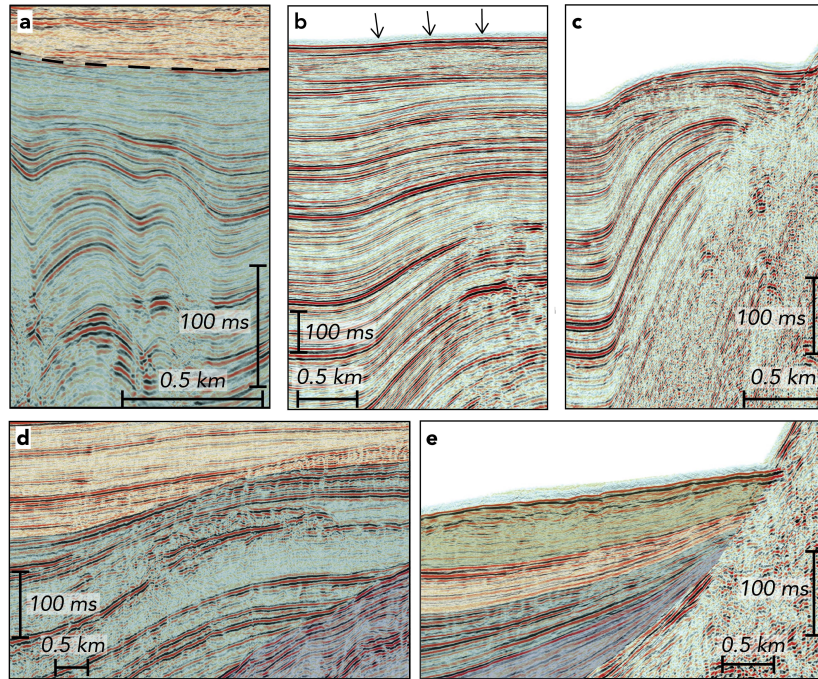


Figure 2.2: Examples of typical deformational and sedimentary characteristics observed in time sections of sparker seismic data in this study. **a.** A currently inactive fold with pre- or syndeformational strata shaded in blue below postdeformational, flat strata in orange. **b.** An active fault-propagation fold which shows offset at the surface highlighted with arrows. **c.** An active fold with a large seafloor signature. **d.** Seismic facies are defined in part by reflector geometries, e.g., where onlap occurs at seismic facies boundaries. Each color in this image represents a distinct seismic facies. **e.** Colored packages thin and onlap onto the anticline on the right boundary of the basin. The topmost green package shows thinning and surface tilting, indicating recent activity of the anticline.

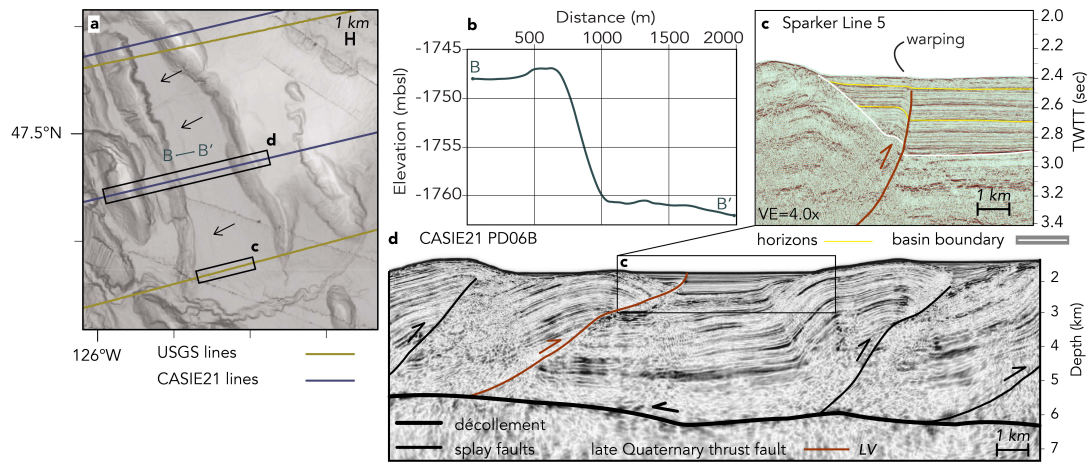


Figure 2.3: Tracing a late Quaternary active thrust to the surface. **a.** Arrows point to fault signature on multibeam bathymetry. **b.** Profile on bathymetry across the fault from B to B' showing a 13 m seafloor fold signature. **c.** Sparker line showing basinal fault that warps uppermost sediment package, indicating late Quaternary activity. VE=4.0x within the sediments using a velocity of 1700 m/s. **d.** Same splay fault identified within CASIE21 Line PD06B as connecting to the décollement.

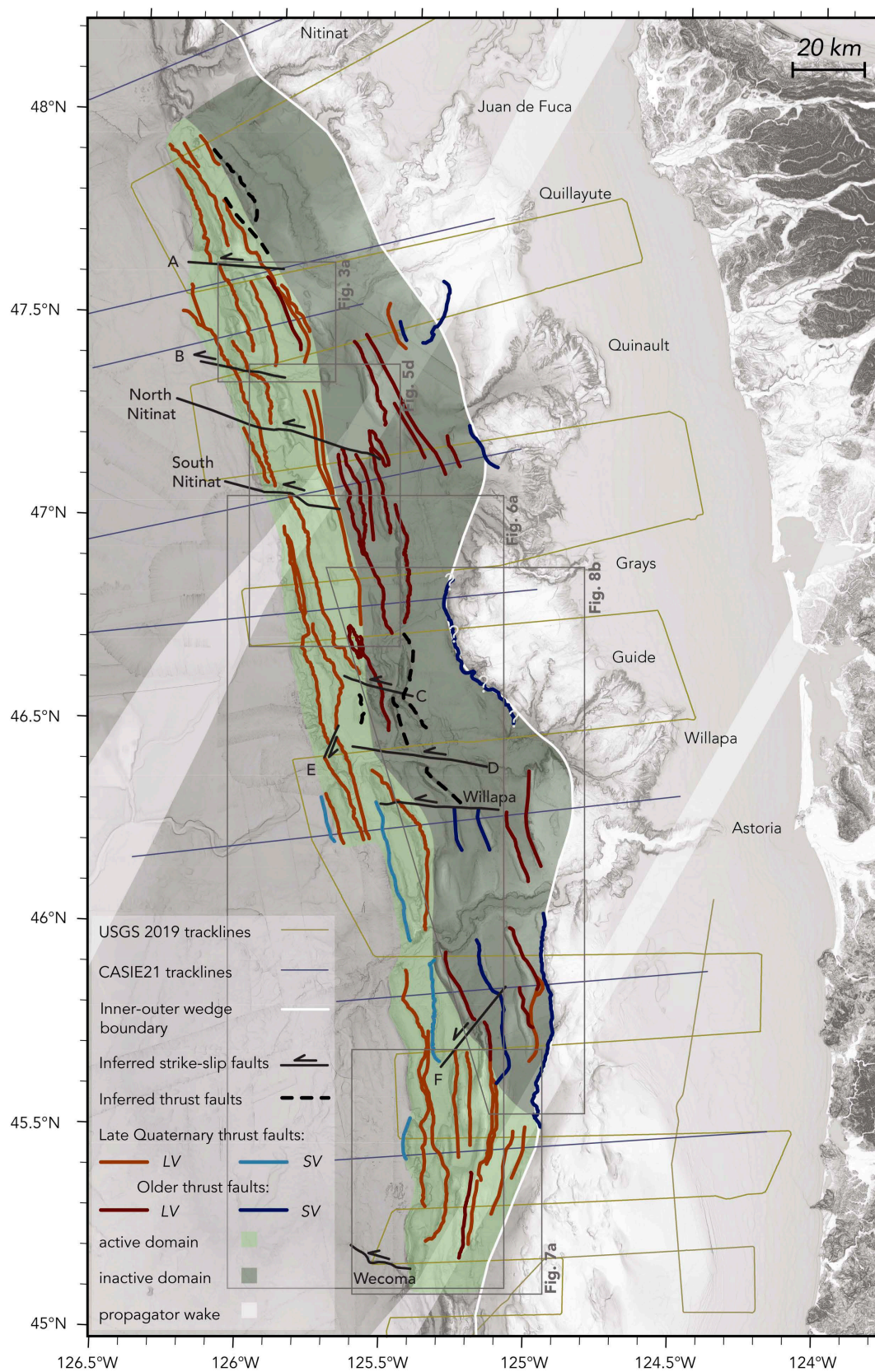


Figure 2.4: Structural map of the outer accretionary wedge in the landward vergence zone. Landward-vergent thrusts are in red and seaward-vergent thrusts in blue, with lighter shade for late Quaternary active thrusts as identified in this study. Inferred strike-slip faults are black lines and inferred thrusts are dashed black. The inner-outer wedge transition is white. Canyons are labeled. The locations of subset maps used in subsequent figures are shown in light grey. Propagator wake shear zones on the downgoing plate shown in white from (Nedimović et al., 2009).

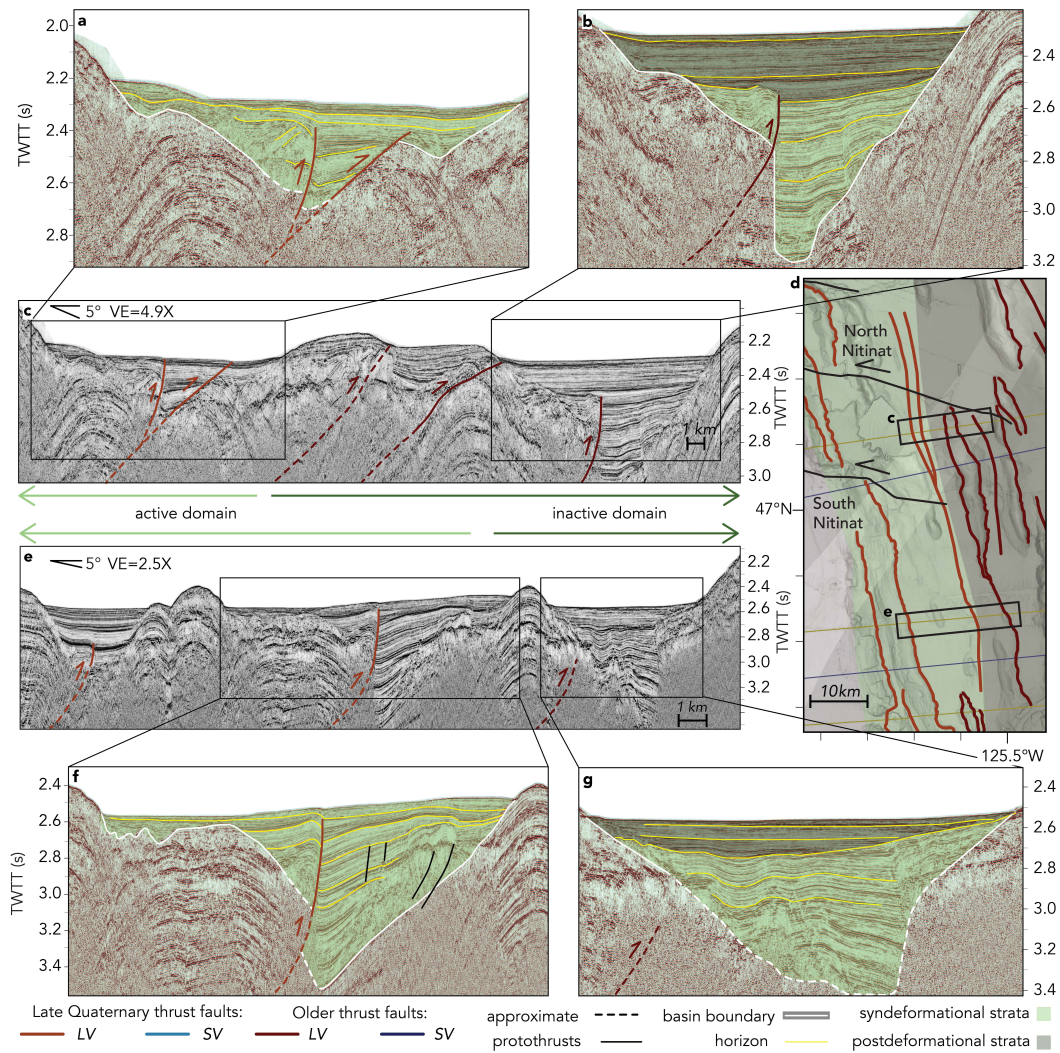


Figure 2.5: Comparison of tectonic stratigraphy across the active-inactive domain boundary. **a.** Detail of active domain basin from panel (c). Two active landward-vergent faults with syndeformational strata. **b.** Detail of inactive domain basin from panel (c). Buried landward-vergent inactive thrust with syndeformational strata in basin bottom and postdeformational strata above. **c.** Sparker profile of wedge across active-inactive domain boundary. VE=4.90x within the sediments using a velocity of 1700 m/s. **d.** Map showing location of A-D, F-G. **e.** Sparker profile of wedge across active-inactive domain boundary. VE=2.48x within the sediments using a velocity of 1700 m/s. **f.** Detail of active domain basin from panel (d). Active late Quaternary fault cuts syndeformational strata. **g.** Detail of inactive domain basin from panel (d). Syndeformational strata showing folding separated from postdeformational strata by an unconformity.

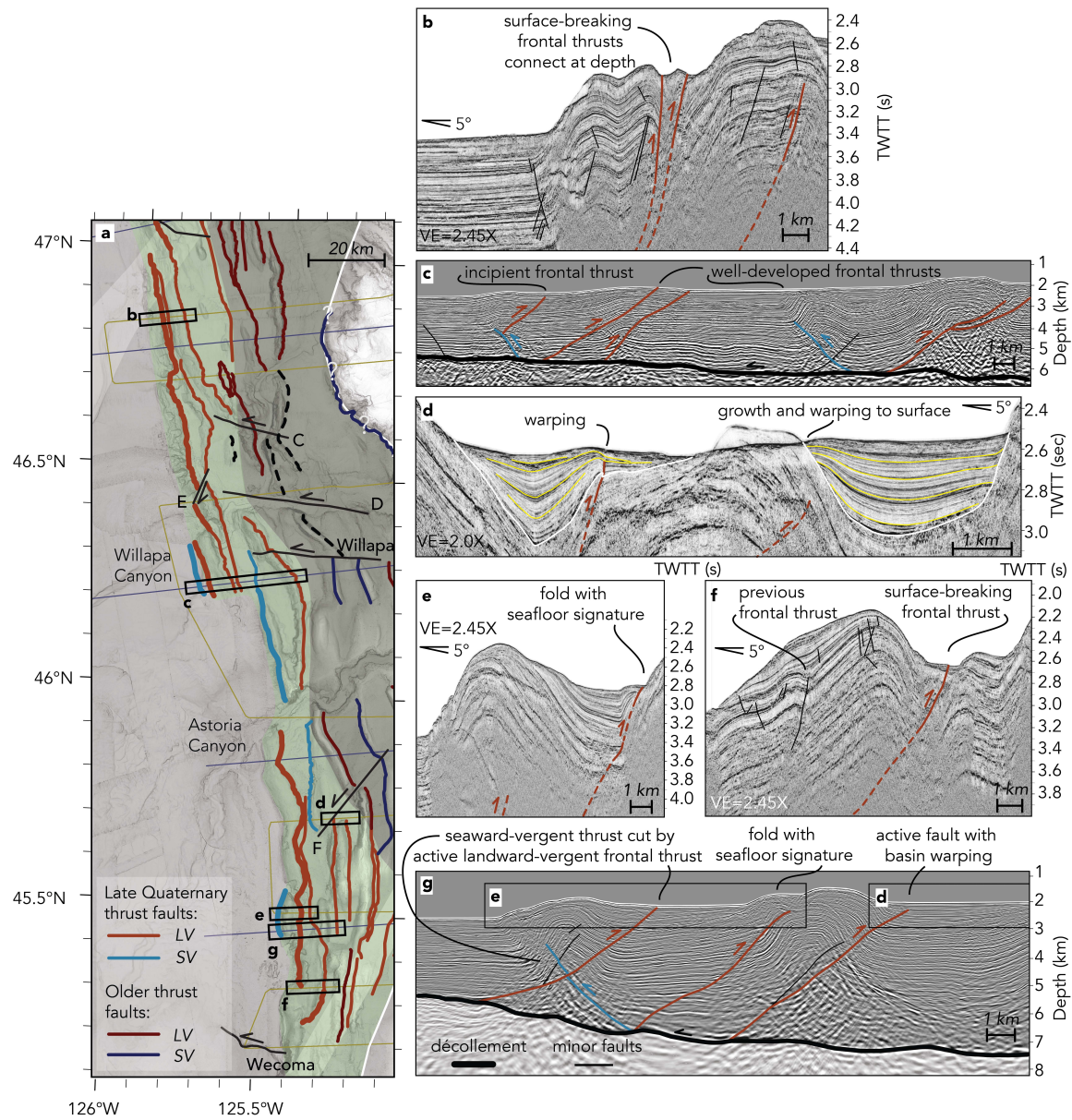


Figure 2.6: Characterization of the active domain along strike. Vertical exaggeration (VE) for all sparker profiles shown is 2.45x within the sediments using a velocity of 1700 m/s, except d which has a VE=2.0x **a.** Map showing the location of b-g. Late Quaternary active thrust faults which act as the frontal thrust in each location are highlighted with bold lines. **b.** Sparker profile. Doubly folded frontal anticline with two active surface-breaking landward-vergent thrusts with another fault-propagation fold imaged seaward. **c.** CASIE21 image. An incipient seaward-vergent frontal thrust is cut by an incipient landward-vergent frontal thrust. The well-developed frontal thrusts from north of this line (landward-vergent) and south of this line (seaward-vergent) are imaged. **d.** Sparker profile. Typical appearance of basins documenting late Quaternary faulting within the active domain. An active landward-vergent thrust is imaged in basin on left and another is inferred under anticline in middle of panel due to presence growth strata in basin nearby. **e.** Sparker profile. Frontal anticline controlled by a landward-vergent frontal thrust with an additional late Quaternary active frontal thrust creating a seafloor fold in the basin behind it. **f.** Sparker profile. Frontal anticline with surface-breaking frontal thrust. **g.** CASIE21 image showing location of basins in d and e. Westward active fault with seafloor warping shown in d appear in the right basin. Landward-vergent frontal thrust and second thrust with seafloor fold signature shown in e appear in the left basin.

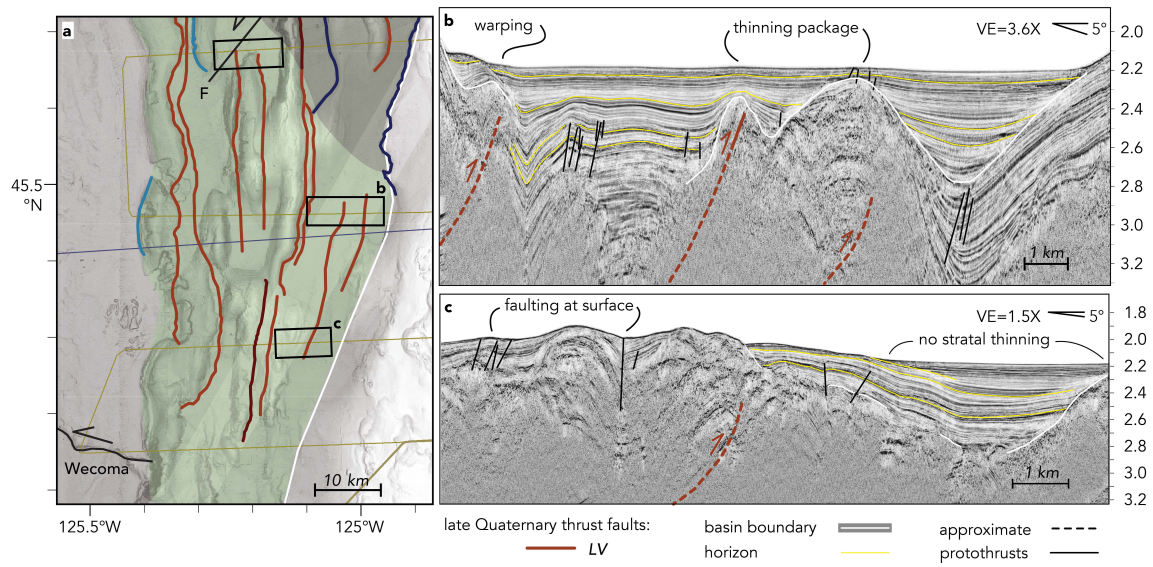


Figure 2.7: The southern portion of the active domain. **a.** Map showing location of b-c. **b.** Wide basin near landward edge of active domain showing near-surface deformation of youngest sediments. $VE=3.64x$ within the sediments using a velocity of 1700 m/s. **c.** Complex active faulting imaged at surface. $VE=1.46x$ within the sediments using a velocity of 1700 m/s.

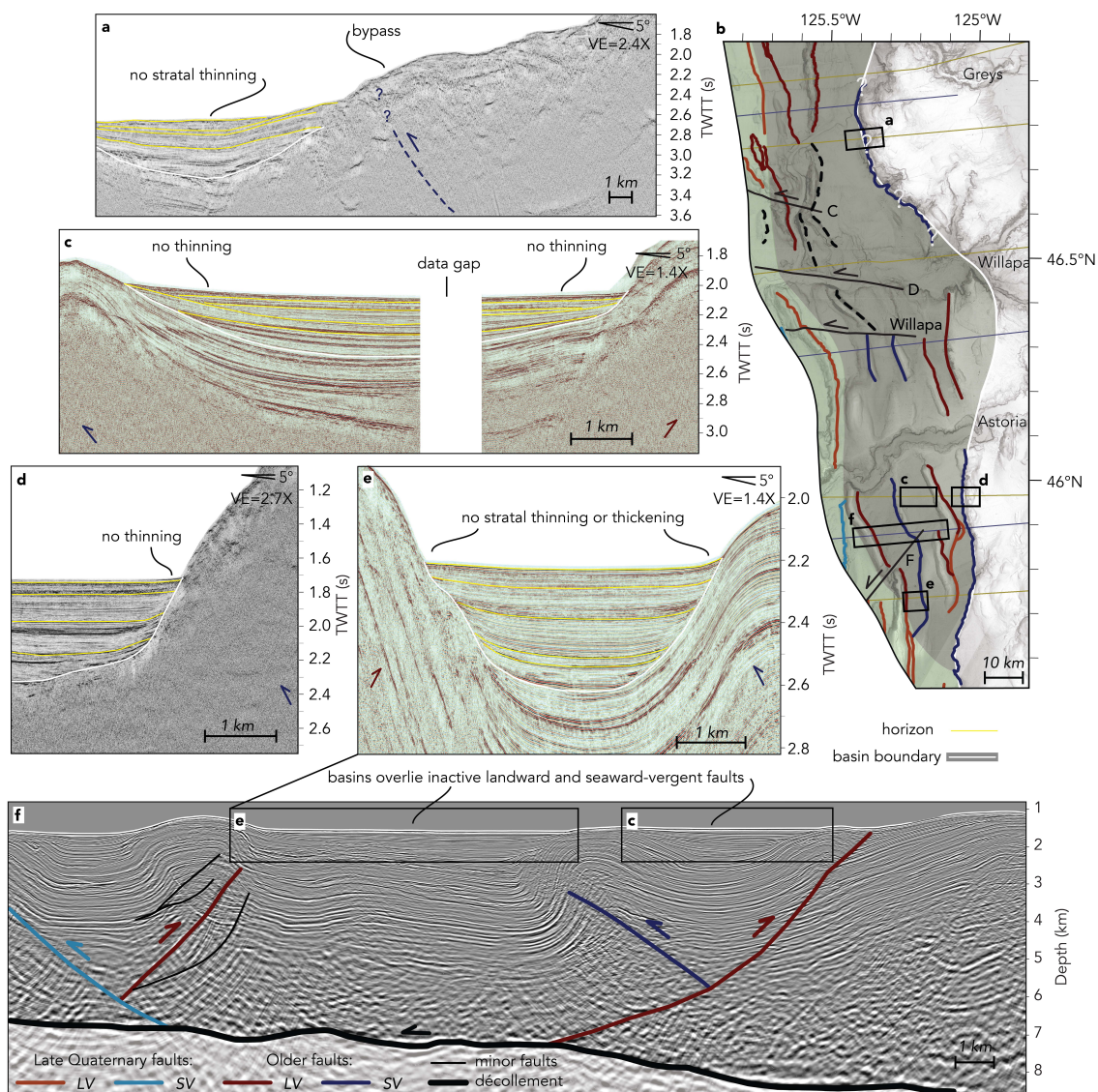


Figure 2.8: The inactive domain. Arrows without faults on annotated sparker profiles indicate the vergence of underlying faults which are not imaged in the sparker data but are imaged in the CASIE21 data. **a.** Sparker profile. Near-surface at location of inner-outer wedge transition showing flat-lying packages nearby. $VE=2.40x$ within the sediments using a velocity of 1700 m/s. **b.** Map showing location of a, c-f. **c.** Sparker profile. Basin with flat-lying near-surface packages above a significant unconformity shows typical appearance of inactive domain basins. $VE=1.40x$ within the sediments using a velocity of 1700 m/s. **d.** Sparker profile. Basin at inner-outer wedge transition showing inactivity of seaward-vergent thrust. $VE=2.67x$ within the sediments using a velocity of 1700 m/s. **e.** Sparker profile. No present-day deformation on adjacent landward-vergent and seaward-vergent faults recorded in basin. $VE=1.41x$ within the sediments using a velocity of 1700 m/s. **f.** CASIE21 image showing location of basins in e and c above inactive mixed vergence splay faults.

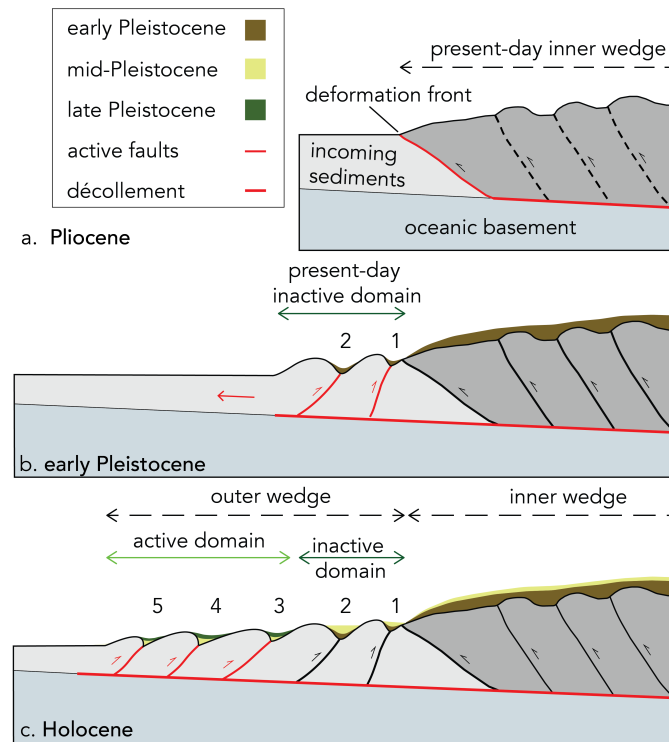


Figure 2.9: Conceptual model of wedge evolution. **a.** In the Pliocene, the present-day inner wedge composed the entire accretionary wedge in the LVZ. **b.** Quaternary glaciation begins at the beginning of the Pleistocene 2.58 Ma increasing the sedimentation rate and causing the outer wedge to build seaward continuously. The present-day inactive domain (basins 1-2) hosts active faults. **c.** Active domain basins 3-5 are formed and inactive domain basins 1-2 become inactive. Fault activity in the active domain continues today.

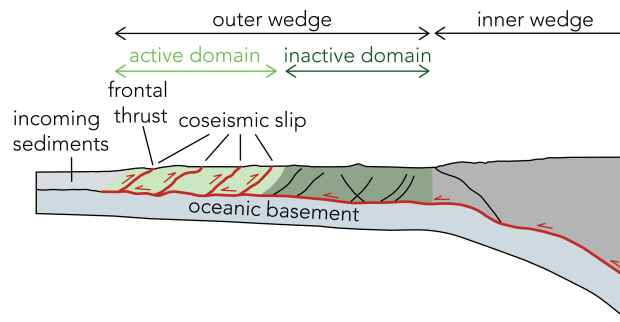


Figure 2.10: Proposed model of coseismic slip in the LVZ during megathrust events. The décollement hosts large slip to the toe of the accretionary wedge, inducing slip on the frontal thrust and other splay faults in the active domain (colored red). The inactive domain faults, including at the inner-outer wedge boundary, do not slip, and the displacement is efficiently transferred up the megathrust to the toe region.

Chapter 3

EVIDENCE FOR A SEDIMENT-HOSTED PLATE BOUNDARY FAULT AT THE CASCADIA SUBDUCTION ZONE

Manuscript in review at *Geochemistry, Geophysics, Geosystems*.

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3.1 Abstract

In accretionary wedges at subduction zones, slip along the megathrust fault, also known as the décollement, is the source of destructive earthquakes and tsunamis. The geology that hosts the fault is a major factor governing rupture mechanics. In the Cascadia subduction zone's landward-vergent outer wedge, previous seismic imaging interpretation work has variably identified the décollement position as both within the sediment section and at the sediment-basement interface. We analyze recently acquired crustal-scale multi-channel seismic data using an analytical structural geological framework, including the application of kinematic fault models and quantitative depth-to-detachment analysis, to produce robust seismic interpretations.

We find that, in all cases, splay faults within the landward vergence zone’s outer wedge most likely sole into stratigraphic levels well within the sedimentary section, positioned on average ~ 500 m above oceanic basement in the active outer wedge and higher beneath the more landward inactive outer wedge. Based on our new interpretation, the décollement dip in the outer wedge of the landward vergence zone is everywhere less than 3° , implying a strong wedge with very low basal friction. The flat, weak megathrust in the landward vergence zone, which is also coincident with the largest patch of the megathrust that is fully locked at Cascadia, suggests that slip may efficiently propagate to the trench in this region. Our findings are consistent with previous work on accretionary wedge structure at other sediment-rich margins, where the megathrust fault has been similarly interpreted as localizing at various levels above, but rarely co-located with, the sediment-basement interface.

3.2 Introduction

Accretionary wedges are submarine fold-and-thrust belts forming between the down-going and overriding plates at subduction zones. The plate boundary fault at the base of the wedge, also known as the megathrust fault or décollement, hosts slip during megathrust earthquakes. Subsidiary splay faults branch from the underlying décollement at higher angles through the wedge and can also slip during megathrust events. Submarine coseismic deformation focused in accretionary wedges is the main source of destructive tsunamis (Wilson and Ma, 2021) and thus characterizing its deformation is paramount to better understanding these hazards.

While a $100 - 150^\circ\text{C}$ isotherm has been commonly identified as the updip limit for frictional locking and coseismic slip at subduction zones (Hyndman and Wang, 1993; Moore and Saffer, 2001b), recent work has shown that both shallow megathrust and splay fault slip outside the thermally-defined seismogenic zone are prevalent during large megathrust earthquakes. The 2011 Mw9.1 Tohoku earthquake caused a paradigm shift in the understanding of shallow slip during megathrust events because

the majority of coseismic rupture occurred above 40 km in depth with an estimated 50 to 120 meters of slip at the trench (Fujiwara et al., 2011; Ide et al., 2011; Ito et al., 2011; Kodaira et al., 2012, 2020; Sun et al., 2017; Ueda et al., 2023), which was previously thought unlikely.

The Cascadia subduction zone is sediment-rich, with one of the thickest incoming plate sediment sequences of subduction zones globally (Heuret et al., 2012) and very high recent supra-wedge clastic sedimentation (e.g., Barnard, 1978; Rohr et al., 2019; Underwood, 2002). Such thick sediments on top of the hot, young subducting Juan de Fuca Plate cause the temperature-defined seismogenic zone (Hyndman and Wang, 1993) to extend all the way to the deformation front (Salmi et al., 2017). However, because the last megathrust earthquake at the Cascadia subduction zone was in 1700 CE (Atwater and Hemphill-Haley, 1997; Satake et al., 1996) and recent seismicity is sparse (Morton et al., 2023; Stone et al., 2018), the expected characteristics of a future megathrust event must be inferred from the available geological and geophysical evidence.

Driven by glacial sedimentation, the accretionary wedge is especially wide, flat, and thick offshore Washington (e.g., Watt and Brothers, 2020), where it is dominated by atypical landward-vergent, or seaward-dipping, splay faults (Fig 1; MacKay, 1995; Seely, 1977). Such systematic landward-vergent faults are not explained by classical accretionary wedge mechanics (Dahlen et al., 1984; Davis et al., 1983) and are globally rare, existing elsewhere in Sumatra and the Niger Delta (Maloney et al., 2010). It has been suggested that the formation of landward-vergent faults requires extremely low coefficients of friction on the décollement (< 0.03 , Cubas et al., 2016) that can only be achieved coseismically through dynamic weakening mechanisms (Di Toro et al., 2011a). At Cascadia, the landward vergence zone (LVZ; Fig. 1A), spanning some ~ 400 km along strike, is also coincident with the largest patch of the megathrust that is fully locked (Li et al., 2018; Lindsey et al., 2021), most likely to the deformation front (DeSanto et al., 2025).

Work at Nankai Trough first documented a large, continuous, and out-of-sequence splay fault known as a megasplay (Moore et al., 2007a; Park et al., 2002; Tobin and Kinoshita, 2006b). This megasplay has been interpreted to localize at the inner-outer wedge transition, also known as the backstop, which is the rheological boundary between younger, less consolidated outer wedge sediments and older, more consolidated inner wedge sediments. However, recent work has shown that the previously-proposed megasplay fault at Cascadia (Witter et al., 2013) is actually inactive to absent along most of the margin (Lucas et al., 2025). Within the LVZ, a zone of active splay faults ~ 30 km in width in the outermost wedge is most likely to slip concurrently with the shallow décollement during megathrust events instead of the proposed megasplay fault (Ledeczi et al., 2024). We refer to this zone as the active domain, and the adjacent portion of the outer wedge where most splay faults have been shown to be currently inactive as the inactive domain, following Ledeczi et al. (2024). Dynamic rupture modeling of splay faults in the outer wedge at Cascadia has further confirmed that rupture along the megathrust to the deformation front is possible and that this slip can readily activate both seaward and landward-vergent splay faults (Biemiller et al., 2025). Turbidite records also indicate that recent coseismic deformation at Cascadia has been focused in the outer wedge (Hill et al., 2026).

Seismic interpretation and drilling at sediment-rich subduction zones have found that the megathrust rarely localizes at the rough sediment-basement contact, instead preferring to form within the sedimentary section in the outer wedge. For example, though subduction zones often exhibit along-strike variations, shallow sediment underthrusting has been interpreted in regions of the Alaska (Cortés-Rivas et al., 2025; von Huene et al., 2021), Barbados (Labaume et al.; Laigle et al., 2013; Shipley et al., 1997), Chile (Tréhu et al., 2019), Hikurangi (Barnes et al., 2010; Ghisetti et al., 2016; Plaza-Faverola et al., 2012; Watson et al., 2019), Makran (Burg, 2018; Grando and McClay, 2007; Kopp et al., 2000), Nankai (Bangs et al., 2009; Moore et al., 2007a), and Sumatra subduction zones (Dean et al., 2010; Frederik et al., 2015; Bradley et al.,

2019).

Thick accretionary wedge sediments allow the plate boundary to localize within the sediments, smoothing basement irregularities and producing more homogeneous fault properties, which allows for larger, more continuous areas of interseismic locking and thus larger earthquakes (Heuret et al., 2012; Ruff, 1989; Wang and Bilek, 2014; Wirth et al., 2022). Therefore, sediment-rich subduction zones, and décollements hosted within the sediments, are inherently linked to increased earthquake hazard. The conditions for rupture initiation, propagation, and arrest all depend critically on the material properties of the megathrust fault’s host rock lithology. Parameters like friction coefficient, pore fluid pressure, elastic properties, and fault roughness can all exhibit fundamental differences in basaltic basement rock, the terrigenous sediments, and at the boundary between the two (Ikari et al., 2018, 2020).

At Cascadia, however, the position of the décollement has been debated in the literature. Previous work focused on detailed seismic interpretation has largely found that the megathrust is localized within the sediments but close to the basement (*sim* 500 m at the deformation front) in the landward vergence zone (Gutscher et al., 1996; MacKay et al., 1992; MacKay, 1995; Webb, 2017). Others have suggested that two décollement levels, one at the basement and one within the sediments at a shallower level with opposing senses of motion, bound a duplex marking the plate boundary fault zone (Adam et al., 2004; Booth-Rea et al., 2008).

Other geophysical work has suggested the outer wedge décollement may instead be localized primarily at the top of crust coincident with the sediment-basement interface (Carbotte et al., 2024; Peterson and Keranen, 2019). Since the first multichannel seismic reflection work at Cascadia (Flueh et al., 1998; MacKay et al., 1992; Tobin et al., 1993, e.g.), improvements in seismic reflection techniques have made it possible to produce vastly more detailed, clearer images of accretionary wedge structure. Carbotte et al. (2024) used the newly collected seismic reflection data from the CASIE21 survey (Fig. 1A) to remap the plate boundary surface for the entire margin, finding

a shallow sediment décollement limited to a ~ 250 km stretch of the margin offshore central to southern Oregon and interpret the decollement at top of crust (TOC) for the rest of the margin. In that study, the plate boundary surface was mapped by visual reflectivity interpretation; however, detailed seismic interpretation of faults in the outer wedge was not a specific focus.

In Carbotte et al. (2024), many splay faults within the landward vergence zone are not interpreted as connecting to the plate boundary, because in many cases it does not seem that they extend to the TOC. Splay faults are similarly interpreted as rooting into wedge sediments above the TOC in the majority of previous work (Gutscher et al., 1996; MacKay et al., 1992; MacKay, 1995; Webb, 2017). Since, by definition, splay faults must develop connected to a décollement which is the plate boundary fault in the accretionary wedge, the fact that there is no conclusive evidence they extend to the TOC suggests that the décollement may lie within the sediments in many areas.

We therefore further analyze the CASIE21 dataset to determine the décollement position beneath the entire landward vergence zone’s outer wedge. We support our reflectivity-based interpretation with two additional analytical methods of structural geology: the application of kinematic fault model construction and quantitative area-depth analysis. Guided by these methods, we carry out a detailed structural interpretation of splay faults and their soling depth. We find that splay faults in the LVZ outer wedge everywhere sole into stratigraphic levels within the sediments; our analysis suggests variable sediment underthrusting within the LVZ, with at minimum 80 m, at maximum 2950 m, and on average 500 m thickness of sediments below the décollement in the active outer wedge. We also recalculate the décollement dip for this new décollement as largely less than 3° , but as low as 0.6° in some cases, suggesting a strong wedge with an extremely weak basal fault. These dip values are smaller than but largely consistent with the prior findings using the CASIE21 dataset (Carbotte et al., 2024; Jiang et al., 2023).

Our results suggest conditions in the shallow landward-vergent accretionary wedge offshore Washington favor earthquake rupture propagation to the deformation front, as has been suggested by previous evidence (e.g., Biemiller et al., 2025; Cubas et al., 2016; Han et al., 2017; Hill et al., 2026; Ledeczi et al., 2024). Moreover, the sediment-hosted décollement we identify here is consistent both with previous work at Cascadia (Adam et al., 2004; Booth-Rea et al., 2008; ?; MacKay et al., 1992; MacKay, 1995; Webb, 2017), as well as evidence from numerous other sediment-rich subduction zones.

3.3 Methods

3.3.1 CASIE21 seismic imaging experiment

We interpret crustal-scale 2-D multichannel seismic reflection data from the Cascadia Seismic Imaging Experiment (CASIE21) collected on the R/V Marcus G. Langseth in 2021 (Carbotte et al., 2023). The data were collected using a 12 to 15 km streamer with a 6600 cu in 36-airgun array source and processed to pre-stack depth migration in collaboration with ION Geophysical. We analyze 7 CASIE21 dip lines, PD06 through PD11, which cover the landward vergence zone and span the accretionary wedge from the incoming plate to the continental shelf (Fig. 1A). The CASIE21 lines image >10 km below seafloor. The seismic reflection profiles in our figures face north with the deformation front at the left side and the inner-outer wedge transition at the right.

Uncertainty in the mapped décollement position arises from two sources: the vertical resolution limits of the CASIE21 seismic data and absolute depth uncertainty introduced by the velocity model. The measured dominant wavelength in the PSDM images at the identified décollement depths is $140 \pm 48\text{m}$, where the range is one standard deviation of the measured wavelengths. Applying the conventional $\lambda/4$ to $\lambda/8$ criterion for the minimum resolvable separation between reflectors gives 18 ± 6 to $35 \pm 12\text{ m}$; if full waveform separation ($\lambda/2$) is required to unambiguously distinguish reflectors, this increases to $70 \pm 24\text{ m}$. The upper end of the ranges is most relevant

where deformation and diminished image quality at depth complicate the wavefield. Separately, assuming a maximum depth uncertainty from the velocity models of 5% (Carbotte et al., 2024, 2023) yields a mean absolute depth uncertainty of 278 ± 32 m at décollement depths. These two sources are independent; combined in quadrature, we estimate a worst-case absolute depth uncertainty of 287 ± 40 m, which is dominated by velocity uncertainty. Although we report depth above the TOC to the nearest 10 meters in the results, all values should be interpreted considering the uncertainties described above.

3.3.2 Structural geological framework

In addition to reflectivity analysis, we apply two established structural-geology based methods to analyze the soling depths of faults in the outer accretionary wedge. We first use established kinematic fault models to evaluate splay fault geometries. Such models have been developed to explain the formation, evolution, and geometry of fault-related folds (Fig. 2, see review in Brandes and Tanner, 2014). The two most basic categories of fault-related folds are fault-bend folds (Suppe, 1983) and fault-propagation folds (Suppe and Medwedeff, 1990). Fault-propagation folds (Fig. 2A) slip incrementally, with cumulative slip decreasing along the fault to the fault tip where deformation is instead consumed by folding (Suppe and Medwedeff, 1990). In fault-bend folding (Fig. 2B), slip is constant along the fault, and the folding is caused by the fault riding over a non-planar surface (Suppe, 1983). While fault-bend folding assumes conservation of area, it is known that real-world fold-and-thrust belts experience significant amounts of internal deformation and/or volume loss (e.g., Moore et al., 2011).

Shear fault-bend folding theory (Fig. 2D) was developed by Suppe et al. (2004) to explain real-world observations of faults with backlimbs that dip less than the fault plane (Fig. 2D). The models depend on an internally deforming décollement layer of finite thickness, rather than a singular fault. Suppe et al. (2004) define

two end-member models. In simple shear fault-bend folds, the décollement layer undergoes bedding-parallel shear, whereas in pure shear fault-bend folds, the layer thickens internally. A continuum between these two end-members is likely to exist. A counterintuitive characteristic of simple shear fault-bend folds is that slip decreases to 0 at the base of the ramp. Because of this, the simple shear fault-bend fold's décollement layer is not technically underlain by a fault; however, the geometric solution in Suppe et al. (2004) does not address what happens internally within the deforming zone, and there is likely heterogeneous deformation within that zone, for example through an anastomosing network of multiple faults.

Since we are interested in the landward vergence zone, the wedge versions of these fold models are more appropriate (Fig. 2C, E); in these cases, the décollement has the same slip sense as the other models but produces a landward-vergent thrust instead. The simple shear fault-bend fold wedge (Fig. 2E) is a more complex case involving a thicker basal deforming layer due to the progressive folding of the footwall ramp (Suppe et al., 2004)

Despite their differences, these models still have one powerful similarity: the axial surface of the syncline intersects the fault itself at the position of the décollement (red dots in Fig. 2). The only fault model that does not meet this criteria is pure shear fault-bend folding (Suppe et al., 2004), in which case the true décollement is shallower than predicted by this intersection; however, this model does not fit the structural style of the faults we assess here. We use this simple geometric relationship to identify the position of the décollement in the landward vergence zone.

The principles above help guide us towards a décollement interpretation which we then independently validate with area-depth analysis. Area-depth or depth-to-detachment analysis (Epard and Groshong, 1993; Groshong et al., 2012; Groshong and Epard, 1994) is a quantitative structural analysis technique used to produce and validate structural interpretations. The method uses the excess area in cross-section of a horizon displaced by a fault from its original horizontal position (Fig. 2F). The

excess area of many horizons compared to their depths produces a linear relationship where a horizon with 0 excess area predicts the position of the décollement (Fig. 2G). We also assume horizons have original horizontality, which is an acceptable assumption given the very small dip of the downgoing plate near the deformation front. There has been some recent work adapting the area-depth method to originally dipping stratigraphy in fold-and-thrust belts (Wang et al., 2018), but we do not apply this updated method here. While the method assumes conservation of area, other assumptions are not required, meaning for example that it is independent of any given kinematic fault model chosen. This method has been widely implemented to produce robust structural interpretations in a variety of geological settings (e.g., Daëron et al., 2007; Doski and McClay, 2022; Gómez de la Peña et al., 2022; Li et al., 2019; Saint-Carlier et al., 2016; Scharer et al., 2004; Shaw et al., 2005).

Here, we utilize the area-depth module in MOVE (PE Limited, 2026). While area-depth analysis is a powerful tool, we can only apply it to select faults where horizons return to an undeformed position between faults; where thrust sheets are widely spaced rather than imbricated; and where there is little backthrusting or deformation that cannot be linked to the primary thrust. For most seismic lines presented here, we only apply this method for the frontal thrust. For shear fault-bend folds, area-depth analysis can predict an artificially shallow décollement. Since simple shear fault-bend folding has been suggested as the best approximation of the structural style of the Cascadia accretionary wedge (Suppe et al., 2004), we use only pre-growth horizons in the upper portion of each fault to avoid the internal deformation within the deforming décollement layer (Ball, 2019) thereby preventing artificially shallow results. It is obvious if a horizon is affected by shear because the measurement falls off the linear trend line given by the other horizons (e.g., Ball, 2019).

While fold-and-thrust belts certainly experience syn-deformational volume loss and compaction, conservation of area is a good approximation at Cascadia because the landward-vergence zone's outer wedge is considered to be particularly strong

and well-compacted (Han et al., 2017; Webb, 2017). Most of the faulting in the wedge is through more consolidated pre-growth stratigraphy, as opposed to weaker growth strata (e.g., Ball, 2019), which is distinct from other subduction zones (e.g., Moore et al., 2011). Cascadia also has an extremely hot, young subducting plate and thick sedimentary package which results in an end-member hot subduction zone environment (Salmi et al., 2017) that drives diagenetic reactions to completion even seaward of the deformation front (Moore et al., 2007b).

We subtract the depth to the TOC of Carbotte et al. (2024) at the position of the fault from its predicted décollement position from area-depth theory, producing a predicted amount of underthrust sediment at each fault. In gaps in the TOC interpretation, we interpolate between the two nearest points. We use area-depth theory only as a guide, and if there is compelling reflectivity evidence for a differing décollement position, we explain the discrepancy. We report the depth values derived from area-depth theory to the nearest 10 meters, but the precision of these values is limited by the resolution and depth uncertainty of the seismic data as described above. We also calculate the basal taper or décollement dip β of the wedge (Davis et al., 1983; Dahlen et al., 1984) for both our identified décollement and that of Carbotte et al. (2024). We approximate the dip as that of a straight line connecting the megathrust fault’s position at the deformation front and at the most seaward point of the inner-outer wedge transition zone as identified in Lucas et al. (2025).

3.4 Results

3.4.1 Applying idealized fault geometries to splay faults at Cascadia

Theoretical fault models can be used to approximate different parts of each fault structure. Shear fault-bend folding theory was developed to explain geometries in submarine accretionary wedges, particularly sediment-rich, low taper wedges like those at Cascadia and Nankai (Suppe et al., 2004). However, it was only previously ap-

plied to seaward-vergent faults offshore Vancouver Island (Suppe et al., 2004). In the examples we show here, we suggest that shear-fault-bend folding is an appropriate structural style for the backlimb of each fault because the horizons dip much less than the fault itself, which is not consistent with other kinds of fault models. Where faults are landward-vergent, we suggest they are wedge-type shear-fault bend folds, which has not been previously used in analysis of this region. Where possible to determine, we find that faults show an increase in slip up-structure (Fig. 2D), which is consistent with shear fault-bend folding theory but is inconsistent with other fault models like fault propagation folding, where slip would decrease to the fault tip (Fig. 2A). The forelimbs of analyzed splay faults in the outer wedge are most consistent with fault-propagation folding because in most cases the forelimbs lack a distinct bend and instead are typically curved.

3.4.2 Refining the position of the megathrust fault

We use a combination of detailed structurally-informed seismic interpretation and area-depth analysis to reexamine the position of the megathrust fault in the CASIE21 survey profiles spanning the LVZ outer wedge (Figs 3-7, Figs S1-S2). We use area-depth analysis as a guide for our seismic interpretation, and we report both the décollement depth it predicts and that which we use in our final models. Where the area-depth analysis seems correct, it still yields décollement depths that are more specific than the resolution of the seismic reflection data allows for, and so our final interpretations can still be ~ 10 s of meters different than these values. However, where there is compelling seismic evidence for a truly different décollement position from the area-depth results, we explain this. We then summarize the interpretations from individual seismic lines to understand spatial changes in sediment subduction and décollement dip throughout the LVZ in Fig. 8.

The widely spaced thrusts in the active domain of PD06 (Fig. 3) are ideal for area-depth analysis and robust seismic interpretation. Thrust fault 1 (TF1) and

TF2 both sole into a reflective décollement (Fig. 3B-C) which area-depth analysis suggests is 300 and 528 m above the TOC (Fig. 3F), respectively; in our final seismic interpretation, the décollement is 330 and 490 m above TOC at these faults. The reflectors under this décollement are horizontal and undeformed at this scale, though they are in some places obscured by what we interpret as fault-plane diffractions (TF1 in Fig. 3B). The décollement we identify varies between 300 and 690 m, and is on average 500 m, above the TOC within the active domain (Fig. 8A-B).

In the inactive domain in PD06, faults and detachment folds root into multiple levels, both a through-going lower level décollement ~ 1 km above the TOC which is continuous with that from the active domain, and an additional upper-level isolated décollement (Fig. 3D-E). The through-going décollement is on average 1020 m (varies from 420 to 1870 m) above the TOC in the inactive domain. The area under the décollement in the inactive domain is not well-imaged and it is unclear if it is deformed. We calculate the dip β of the newly identified décollement as 0.9° (Figs 3A, 8C).

PD06B (Fig. 4) has a décollement very close to the TOC in the outer wedge active domain; it varies between only 80 and 460 m, with an average of 210 m, above the TOC within this zone (Fig. 4B-C; Fig. 8A-B). At the resolution of the seismic reflection imaging, however, the megathrust fault is still not co-located with the sediment-basement boundary; this minimum value is greater than the minimum resolvable separation between reflectors. Area-depth analysis (Fig. 4F) for TF1 and TF2 predicts a décollement 182 and 238 m above TOC, while our final seismic interpretation (Fig. 4B-C) results in a décollement 200 and 320 m above TOC at these faults. This is a very similar interpretation to that in Han et al. (2017) on a nearly coincident seismic line where they interpret 170 m of sediment underthrusting at the frontal thrust.

A sharp step in the décollement level occurs at TF4 on PD06B at the boundary between the active and inactive domains (Fig. 4D). The décollement is at an average of 1010 m above the TOC in the inactive domain (Fig. 8A-B). Because PD06B is a

partial line, and because we have evidence that the megathrust is not through-going between these two domains, we do not calculate the décollement dip β for the outer wedge. Our identified surface seems to rejoin the Carbotte et al. (2024) décollement near the end of the profile, though the imaging is not clear enough to tell whether it is truly coincident with the TOC.

The frontal thrust TF1 at PD07 (Fig. 5) has a bright fault-plane reflection which is very clearly imaged soling into a bright décollement reflector (Fig. 5B-C). While area-depth analysis (Fig. 5F) predicts a décollement only 246 m above the TOC (Fig. 5F), we prefer the bright reflector as the décollement because of the clarity of the imaging at this location. This same reflector is clearly imaged both on the hanging wall side of TF1 in the undeformed section as well as below TF5, where it is 1020 m above TOC (Fig. 5B). This matches with the result of area-depth analysis at TF5 which predicts a décollement 1055 m above TOC (Fig. 5F). The faults between TF1 and TF5 root into a secondary shallower décollement level, which is indicated by the area-depth analysis where the décollement of TF4 is predicted to be the shallowest at 1350 m above TOC (Fig. 5F). Our seismic interpretation shows an upper-level décollement at TF4 at 1350 m above TOC. The through-going décollement at this region is only 820 m above TOC and is therefore clearly the incorrect soling depth for this fault (Fig. 5F). TF2 and TF3 are a more complex mixed-vergence pair which cannot be analyzed with area-depth theory, but which nonetheless clearly root to a shallower secondary detachment level, most likely the same one as TF4. The amount of sediment underthrusting varies from 200 m at the location of TF3-4 to 1090 m in the vicinity of TF5 due to basement topography (Fig. 8A-B).

The detachment fold and additional faults within the inactive domain (Fig. 5D-E) cannot be explained by slip on a décollement as deep as the TOC here; a shallow décollement is required to produce the imaged structures. The surface we interpret is between 200 m and 2110 m above the TOC within the inactive domain. However, this volume of material between the TOC and sediment décollement is not well-imaged

and could be internally imbricated or deformed. This décollement does not appear to rejoin the Carbotte et al. (2024) surface at the inner-outer wedge transition (Fig. 5A). We calculate the outer wedge décollement dip β for our décollement as 1.5° (Fig. 5A, 8C).

TF1 and TF2 imaged in line PD08 are some of the clearest examples of a sediment-hosted detachment in the landward vergence zone (Fig. 6). Both faults have bright fault plane reflections which become listric and merge into the same horizon (Fig. 6B-C). Area-depth analysis predicts the décollement is 320 m above TOC at TF1, but the horizon we identify is instead 490 m above TOC. Reflectors below the identified megathrust are less well-imaged, but appear to be flat-lying at the frontal thrust. The amount of sediment subduction ranges between 270 and 710 m in the active domain and 405 and 1350 m in the inactive domain (Fig. 8A-B). One particularly clear example of a shallow décollement just landward of Quinault Ridge is shown in Fig. 5D-E where shallow faulting and folding truncates at a horizon ~ 1 km above the TOC. In our interpretation, PD08 has the shallowest décollement dip in the LVZ with $\beta = 0.6^\circ$ (Fig. 6A, 8C).

PD09 (Fig. S1) marks the onset of mixed-vergence faulting which renders area-depth analysis and seismic interpretation more difficult. While area-depth analysis would indicate a décollement 440 m above TOC (Fig. S1F), this has greater uncertainties due to the lower coefficient of determination. Therefore, we instead interpret it as a deeper surface that is only 200 m above TOC at TF1 (Fig. S1B-C). This is based on the clear fault-plane reflection from TF2 and TF3 intersecting the same reflector. At TF3, the décollement steps sharply upwards to be consistent with the décollement depth at TF4-TF5. In the active zone, there is between 100 and 970 m of sediment underthrusting (Fig. 8A-B). The inactive zone of the inner wedge is more complex, with mixed-vergence faulting and some faults which do not appear to reach the décollement (Fig. S1D-E). The outer wedge décollement dip β for our décollement is 2.1° (Fig. S1A, 8C).

In both PD10 and PD11, the frontal anticline structure appears to be a complex combination of primarily landward-vergent and some additional seaward-vergent faulting (Fig. 7; Fig. S2). In PD10, area-depth analysis indicates that TF1 has a décollement 1050 m above the TOC (Fig. 7D), but we interpret it slightly deeper at 900 m (Fig. 7C). Just landward of the frontal thrust, two protothrusts, one landward and one seaward-vergent, clearly root into a bright reflector at a similar distance above the TOC (Fig. 7B-C). The outer wedge décollement dip β for our décollement is 2.6° (Fig. 7A; 8C).

At PD11, there is no inactive zone as defined in Ledeczi et al. (2024) because all faults in the outer wedge show near-surface activity (Fig. S2). We don't perform area-depth analysis for the frontal thrust in this region because its backlimb is not well-imaged and it is not straightforward to define continuous horizons. Our seismic interpretation indicates the décollement is 1100 m above TOC at PD11. We find between 690 and 2950 m, and on average 1540 m, of sediment underthrusting across the entire outer wedge in PD11 (Fig. 8A-B). The large increase in sediment subduction in PD11 between 20 and 35 km from the deformation front (Fig. 8A-B) is due to crustal topography in the same location where a sediment-hosted décollement is also interpreted in Carbotte et al. (2024). The décollement we interpret here is therefore only 1300 m above the Carbotte et al. (2024) plate boundary surface in this region. PD11 has the highest outer wedge décollement dip β in the LVZ at 4.5° (Fig. S2A; Fig. 8C).

When taken together, seismic imaging in the landward vergence zone shows ample evidence for a sediment-hosted décollement that is at least 80 meters and at most 2950 m above the TOC (Fig. 8A-B). This interpretation is most well-constrained in the active domain where the total thickness of sediment beneath the décollement is ≤ 1500 m, and on average 500 m, above the TOC for all the seismic profiles (Fig. 8B); however, the variability in the décollement position in the inactive domain is further explored in the discussion.

This identified megathrust surface is near-horizontal and increases in dip as the outer wedge decreases in width towards the south. This means that differences between the dipping Carbotte et al. (2024) TOC and the décollement identified herein often increase from the active domain of the outer wedge to the inner-outer wedge transition (Fig. 8). Our basal dip values are calculated for the entire outer wedge in each line and vary between $0.6 - 4.4^\circ$. These values are somewhat lower than, but ultimately consistent with, those we calculated for the Carbotte et al. (2024) décollement, which range from $1.2 - 5.7^\circ$.

3.5 Discussion

3.5.1 A theoretical framework for seismic interpretation

There is a rich literature on how to extract the full interpretive value from seismic reflection data and create robust and viable structural interpretations. Structural interpretation using kinematic fault models as a theoretical framework that is then further refined by quantitative techniques is much more rigorous than reflectivity-based seismic interpretation alone (e.g., Eichelberger et al., 2015; Shaw et al., 2005). This kind of quantitative approach is common in fold-and-thrust belt analysis (e.g., DeCelles et al., 1991; Ford et al., 1997; ?) and offshore industry-oriented literature (e.g., Bilotti and Shaw, 2005; Corredor et al., 2005; Shaw et al., 1999). Interpretations based on reflectivity alone without an underpinning of analytical structural methods are not necessarily structurally permissible.

Where this kind of analysis has been applied to Cascadia, it has been powerful in revealing new details of accretionary wedge structure, but it is limited to a few studies (Adam et al., 2004; Ball, 2019; Booth-Rea et al., 2008; Gonzalez-Mieres and Suppe, 2006; Suppe et al., 2004; Wang et al., 2018). For example, Gonzalez-Mieres and Suppe (2006) use analysis of a detachment fold and frontal thrust offshore Oregon to confirm a shallow detachment ~ 2.3 km above the oceanic basement and suggest

the possibility of two detachment levels in this region. Suppe et al. (2004) show that shear fault-bend folding is the most appropriate fault model to describe the seaward-vergent Cascadia accretionary wedge offshore Canada. This was further confirmed by structural analysis for the landward vergence zone by Ball (2019). Here, we observe fault geometries which in all cases show backlimbs dipping less than the corresponding faults, and slip increasing along faults, similar to Suppe et al. (2004).

In the theoretical framework of simple shear fault-bend folding, the slip on the fault ramp can be produced solely by shear within the décollement interval rather than at its base, so the bottom of the deforming layer is not required to have localized shear in the form of a fault. However, in order to build the accretionary wedge outwards and have a series of thrusts, the slip must partition between that which is passed forward along the décollement and that which produces the splay fault. To keep the system going, the former must be large relative to the latter. Therefore, our identification of these faults as shear fault-bend folds does not imply the absence of a through-going décollement.

The geometric solution in Suppe et al. (2004) also does not address what happens internally within the deforming zone. In nature, simple shear fault-bend folds more than likely have heterogeneous strain accommodation within the décollement layer. This could be through, for example, anastomosing fault networks contained in deformed *mélange* like seen in the geological record which accumulates in thickness over multiple earthquake cycles (e.g., Rowe et al., 2013; Ledeczi et al., 2025). If so, this could be considered seismic imaging evidence for what has long been observed in the geologic record. The anastomosing structure of megathrust interfaces has also been observed seismologically, with earthquakes sometimes occurring on one fault, and sometimes activating a network of multiple faults (Chalumeau et al., 2024), which could be a mechanism for accumulating a thicker layer of deforming material at the décollement.

3.5.2 *Previous interpretations of the Cascadia megathrust*

Here, we will discuss the main iterations of seismic interpretations of the landward vergence zone based on the seismic imaging available at the time. Sediment-hosted décollements were previously interpreted in MacKay et al. (1992), MacKay (1995), Gutscher et al. (2001), Booth-Rea et al. (2008), and Webb (2017), while a décollement coincident with the oceanic crust in most places in the outer wedge was interpreted in Carbotte et al. (2024) using the new CASIE21 seismic imaging.

MacKay et al. (1992) and MacKay (1995) interpret 400 m of sediment underthrusting beneath the landward-vergent outer wedge (Fig. 9A) based on structural interpretation of seismic line OR-23 from a 1989 survey (Moore et al., 2013). Our interpretation is most similar to Gutscher et al. (2001) who interpret a near-horizontal sediment-hosted décollement 500 m above basement in the outer wedge with increasing amounts of sediment underthrusting landward on line SO108-105 (Fig. 9B). Adam et al. (2004) perform area-depth analysis on the first four landward-vergent faults on seismic line SO 108-103 (Fig. 9C), which is closest to CASIE21 line PD06B. This analysis consistently indicates a décollement level which is at least 500 m above basement within the active domain, consistent with our analysis on line PD06 (Fig. 3). Booth-Rea et al. (2008) also show a near-horizontal, sediment-hosted décollement on line SO108-104 (Fig. 9D; coincident with CASIE21 line PD07, Fig. 4 herein) that facilitates between a few hundred meters of sediment underthrusting at the deformation front up to several km at the inner-outer wedge boundary. While this is the seismic line with the most sediment underthrusting in their interpretation, all lines they interpret have a décollement at least slightly above basement. Webb (2017) similarly interpret at least 100s of meters of underthrust sediments in the landward vergence zone outer wedge (Fig. 9E).

A common feature of the Adam et al. (2004) and the Booth-Rea et al. (2008) interpretations is the postulation of a double-décollement configuration in order to explain

landward vergence and the strain gradient between the surface and the base of the thrust sheet at the frontal thrust. In Adam et al. (2004), the landward-vergent splay faults branch from an upper décollement with a top-to-NE sense of slip, the opposite of what would be expected for the megathrust (Fig. 9C). They invoke a mechanical decoupling of the landward-vergent “passive roof thrust sequence” from the underlying section, a model which directly influenced a similar explanation for landward vergence in Sumatra (Fisher et al., 2007). Booth-Rea et al. (2008)’s interpretation (Fig. 9D) is based on Adam et al. (2004), and they call their interpreted décollement in the inner wedge a “mid-level detachment” and refer to the underthrust sediment section as a duplex, but don’t explicitly show their implied lower décollement. This complex configuration is not structurally supported by the available evidence, including the area-depth analysis presented in Adam et al. (2004) which shows that the sediment-hosted décollement is the primary detachment structure.

We suggest that the strain gradient noted by Booth-Rea et al. (2008) can be explained by shear fault-bend folding theory (Suppe et al., 2004). Additionally, landward-vergent faults do not require a décollement with the same slip sense if the faults are wedges (e.g., Fig. 2E). We now have evidence that slip along a megathrust with the expected slip sense can trigger slip on landward-vergent faults (e.g., Biemiller et al., 2025; Cubas et al., 2016; Ding et al., 2023). Therefore, a lower décollement beneath the underthrusting section in Adam et al. (2004) and Booth-Rea et al. (2008) is neither required nor favored by the evidence, and is not the simplest explanation.

This kind of double décollement is also not interpreted to exist in other areas with landward vergence such as the Niger Delta (Maloney et al., 2010) and Sumatra (Dean et al., 2010; Frederik et al., 2015; Moeremans et al., 2014). Interestingly, Bradley et al. (2019) apply the hinge intersection analysis to map the décollement position in Sumatra and find that the décollement is within the sediments in the seaward-vergent zones, but at the TOC in the landward-vergent region. This work shows that the method we apply here is capable of confirming a décollement at the TOC where

one exists.

Carbotte et al. (2024) and Peterson and Keranen (2019) interpret the décollement dominantly at the top of basement within the LVZ (Fig. 9F), but in neither case is detailed analysis of this position in the outer wedge a focus of those studies. In Carbotte et al. (2024), the megathrust is said to be identified primarily from reflectivity and the geometry and depth extend of splay faults. However, landward-vergent splay faults in both interpretations also do not necessarily reach the décollement (see e.g., Figs 3 and 4 in Peterson & Keranen, 2019 and Fig. 3 in Carbotte et al., 2024), which is not a permissible or retrodeformable structural interpretation. Jiang et al. (in review) comprehensively interpret outer wedge splay faults for the whole margin in the CASIE21 data, but they do not reexamine the position of the décollement in Carbotte et al. (2024). Our purpose in this work is to reexamine this interpretation through a detailed structural geology lens, with multiple independent analytical constraints.

3.5.3 Reinterpretation of the megathrust position

We use three methods to determine the position of the décollement in the outer wedge based on the CASIE21 survey: reflectivity, hinge analysis, and area-depth analysis. Because area-depth analysis does not rely on the assumption of any specific structural style, it serves as an independent confirmation on the décollement position as identified from seismic interpretation alone.

The active domain of the outer wedge

Our interpretations are best constrained in the active domain of the outer wedge (as defined in Ledeczi et al., 2024, see Fig. 1 herein) where area-depth analysis is possible and thrust sheets are widely spaced and well-imaged. We show that, at all CASIE21 lines within the active domain of the LVZ, the décollement is hosted entirely in the sediments, varying between 80 m and 2950 m above TOC, and mainly between ~ 200 m and ~ 1500 m above TOC (Figs. 8A-B). We interpret this megathrust as having a

top-to-the-SW sense of slip, as would be expected for the subduction zone as a whole (Fig. 9F). The landward-vergent faults then branch from this structure as wedge-style faults. There is no requirement nor evidence for multiple décollements with opposing senses of slip as in Adam et al. 2004 and Booth-Rea et al. (2008), or for a fault coincident with the TOC. We have also identified no evidence for duplexing within the package below the décollement, though internal deformation and thickening of the underthrust sediment package may be occurring in some places below the resolution of the imaging.

The inactive domain of the outer wedge

Most previous quantitative structural work, like our area-depth analysis here, focuses only on the less-deformed outermost portion of the wedge where these interpretations are more straightforward to produce (e.g., Adam et al., 2004; Ball, 2019; Gonzalez-Mieres and Suppe, 2006; Suppe et al., 2004). Therefore, our interpretation of the active megathrust in the inactive domain of the outer wedge bears more discussion.

Here, our work strongly supports the interpretation that a shallow, sediment-hosted décollement is responsible for most of the inactive zone deformation. This is the fault that most of the inactive zone splay faults sole into where imaging supports making a determination, thus it has historically hosted a large amount of displacement, though there are isolated additional upper-level décollements in some places. However, what we cannot necessarily constrain is whether (1) this megathrust is the currently active megathrust as identified in the active domain (e.g., Fig. 10A); or (2) this is an abandoned, relict fault structure that would not host present-day megathrust slip (e.g., Fig. 10C). We will address each of these possibilities next.

The simplest explanation, and our preferred one, is that there is one through-going megathrust fault that connects both inactive and active structures (Fig. 10A); this implies that the inactive domain of the wedge would move as a rigid, seaward-translating block during underlying megathrust slip, as proposed in Ledeczi et al.

(2024). In seismic lines like PD06 (Fig. 3) and PD08 (Fig. 6), the décollement identified in these two domains appears connected; there is a relatively thin (usually ≤ 1.5 km) underthrust sediment section that does not seem to allow for another fault below it.

The second possibility is that the inactive domain décollement is a currently inactive structure that has been cut off by megathrust slip on a younger fault structurally below it (Fig. 10C). The only seismic line which explicitly supports this scenario is PD06B (Fig. 4), where a stratigraphically-higher décollement within the inactive domain of the outer wedge is clearly truncated by active splay faulting rooting down to a lower megathrust position. The currently active megathrust must connect to that imaged in the active domain but is not visible in the imaging in the inactive domain (Fig. 10C).

In the other seismic lines with thicker underthrust sediments like PD07 (Fig. 5), such an interpretation may be allowable, but we lack direct evidence for it. The area under our interpreted megathrust in the inactive domain is often less well-imaged, and in some cases appears deformed and thickened, implying that there may be another fault system below this position that is accommodating duplexing or underplating (Fig. 10B). However, if underplating by a structurally lower décollement were occurring, we would expect the surface of the wedge to respond to that perturbation, as predicted by critical wedge theory (Dahlen et al., 1984; Davis et al., 1983). The surface of the inactive domain is flat, even achieving a negative surface slope in some locations (Watt and Brothers, 2020, e.g.), which is at odds with the increased surface slope expected due to underplating.

It is important to note, however, that if there are two décollements, we expect both to have the same, top-to-SW slip sense, unlike what is postulated in Adam et al. (2004). Some deformation within the section below the décollement is also not inconsistent with sediment underthrusting. For example, mechanical decoupling has been documented across the décollement at Barbados, implying that the majority of

plate boundary strain is accommodated by the accretionary wedge, and despite the observation of deformation below the fault, the magnitude of strain accommodated there has been shown to be insignificant (Housen et al., 1996; Shipley et al., 1997)).

While multiple décollements (Fig. 10B) is not our preferred nor the simplest explanation, many accretionary wedges do show evidence for multiple décollement levels (e.g., Noda et al., 2023, and references therein). For example, in the Nankai Trough Kumano 3D survey area, two décollements are interpreted as being segmented by the megasplay fault (e.g., Tobin and Kinoshita, 2006b). In analog models of high basal friction accretionary wedges, alternating cycles of underthrusting and imbrication were observed, promoting the development of multiple décollements (Gutscher et al., 1996, 1998). However, a high basal friction may not be a reasonable assumption for Cascadia’s LVZ (Cubas et al., 2016). Continental fold-and-thrust belt literature has also established that multiple décollements are common and can facilitate duplexing and underplating (Mackay, 2015; Mitra and Boyer, 2020; Vann et al., 1986).

Because structural analysis is much more difficult to perform in the deformed inner wedge, we do not extend our interpretations past the inner-outer wedge transition of Lucas et al. (2025). While the Carbotte et al. (2024) décollement is largely co-located with the TOC in the outer wedge, it also diverges from the TOC close to the inner-outer wedge transition on several profiles (PD06-9). It is therefore plausible that our outer wedge décollement would merge with that sediment décollement at this location. PD06B (Fig. 4) does show evidence for a décollement that may reconnect with the Carbotte et al. (2024) décollement near the inner-outer wedge transition, though imaging is not especially clear in this region and this interpretation is speculative.

3.5.4 Stratigraphic level of the décollement

Ocean drilling conducted during ODP Leg 168 and IODP Expedition 301 (Fisher and Davis, 2000; Fisher et al., 2005) have characterized the stratigraphy on the Juan de Fuca Plate. Previous work synthesizing coring at multiple sites in the Cascadia Basin

identified three lithostratigraphic units that exhibit an upward-coarsening sequence (Su et al., 2000; Underwood and Hoke, 2000). Unit III, adjacent to the oceanic basement, is comprised of basalt talus, diabase sills and hemipelagic mud. Unit II is hemipelagic mudstone. Unit I is subdivided into IB, silty turbidites and hemipelagic mud, and IA, sandier turbidites. Sedimentation rate increased over time and the onset of Nitinat fan deposition occurred at 0.76 Ma within Unit IB (Rohr et al., 2019). Our interpretation places the décollement well below the 0.9 Ma horizon, and potentially below the 1.68 Ma horizon, but the latter has not been correlated to this part of the margin. Therefore, the décollement would either be in Unit IB if soling into levels between 0.9 and 1.68 Ma in age, or in Unit II if below the 1.68 Ma horizon (Lee et al., 2025; Rohr et al., 2019; Shuck et al., 2023; Su et al., 2000). The well-characterized Nankai-Muroto décollement also lies within hemipelagic muds and distal turbidites of the Shikoku Basin (Moore and Saffer, 2001b).

Laboratory friction experiments in Stanislawski et al. (2022) find that both these units are velocity-weakening, or unable to nucleate earthquakes, but suggest they may still support earthquake propagation. However, the sedimentary samples used in that study are far from the accretionary wedge on the incoming plate where temperatures are much cooler (62°C at basement Fisher et al., 2005) due to the thinner sedimentary cover. By the time this material reaches the deformation front where temperatures exceed 150°C (Salmi et al., 2017), shallow diagenetic reactions (e.g., Moore et al., 2007b) are expected to occur which strengthen the sediment and promote velocity-weakening behavior. It is also well-known that the frictional strength of different sedimentary lithologies varies widely, sometimes controlling the décollement position, while in other cases different factors are more important in governing strength (Ikari et al., 2018).

Both Gutscher et al. (2001) and Adam et al. (2004) invoke a basal carbonate sequence below the décollement that is said to deform ductilely and account for landward vergence. However, the amount of carbonate-bearing sediments observed in the

original drilling results is actually very low (Fisher et al., 2007a) compared to the experimental results on carbonates used as an analog in Gutscher et al. (2001).

Our interpretation implies some thickness variations of the underthrust section between the top of crust and active décollement, and therefore that either the décollement does not localize at the same stratigraphic horizon across the margin, or internal thickening and deformation of the zone under the décollement takes place below seismic resolution, or most likely, both of these contribute.

A décollement dipping shallower than the downgoing plate, as interpreted here and in previous work (e.g., Booth-Rea et al., 2008; Gutscher et al., 2001), results in a downdip thickness increase of the underthrust section. This may stem from either basal erosion or an original uneven thickness of sediment deposition on the oceanic plate. If basal erosion is occurring, such that the décollement is cutting through material that was previously accreted to the wedge, that could be one explanation for deformation imaged in the underthrust section. Basal erosion is also consistent with the flat surface slope of the inactive outer wedge (Watt and Brothers, 2020) from critical wedge theory (Dahlen et al., 1984; Davis et al., 1983). This scenario also implies that the fault does not localize at the same stratigraphic horizon everywhere, similar to what has been documented at other subduction zones (Schottenfels et al., 2024, e.g., the Japan Trench). Differences in depositional thickness of sedimentary units due to topography on the downgoing plate is another possibility. Such deposition differences at kilometer spatial scales have been documented at Nankai and suggested to cause lateral heterogeneity in the underthrust sediments and megathrust fault (Ike et al., 2008a).

At Cascadia, the accretionary wedge extends onshore only on the Olympic Peninsula, where the deeply subducted then exhumed units of the Olympic Subduction Complex are exposed (Brandon et al., 1998; Brandon and Calderwood, 1990; Brandon et al., 1992; Tabor and Cady, 1978a,b; Ledeczi et al., 2025). While other exhumed subduction complexes often display deformation involving both sediments and basaltic

units (e.g., Fisher et al., 2021a), the extensively exposed, basally-underplated lower Olympic Subduction Complex is known to be composed of only terrigenous metasediments with the notable absence of basalts or pelagic cherts (Brandon et al., 1998; Tabor and Cady, 1978a,b). A recently identified example of an exhumed Cascadia paleomegathrust fault indicates the plate boundary was hosted entirely within the sediments (Ledeczi et al., 2025). Specifically, the lithologies hosting the fault zone are deep marine turbiditic sandstones and mudstones, with no pelagic carbonates, cherts, nor basalts present (Ledeczi et al., 2025). Therefore, a sediment-hosted décollement is also consistent with the available geological record for this margin.

3.5.5 Sediment-rich wedges and implications for earthquake hazards

The idea that the coseismic slip of shallow megathrusts and accompanying splay faults is possible was vividly demonstrated when the 2011 Tohoku M9.1 earthquake ruptured to the toe of the wedge (Fujiwara et al., 2011; Ide et al., 2011; Ito et al., 2011; Sun et al., 2017; Ueda et al., 2023). There is also mounting evidence for shallow megathrust and splay fault coseismic slip during several instrumentally recorded and well documented megathrust events, including the 1964 Alaska earthquake (DePaolis et al., 2024; Haeussler et al., 2015b; Plafker, 1965; Suleimani and Freymueller, 2020) and the 2004 Mw9.1 Sumatra earthquake (Gulick et al., 2011; Henstock et al., 2006; Ishii et al., 2005; Seeber et al., 2007).

At Cascadia, since the last megathrust earthquake was in the year 1700 (Atwater and Hemphill-Haley, 1997; Satake et al., 1996), this evidence comes from a variety of other sources. For example, Ledeczi et al. (2024) find that splay faults within the outermost “active domain” of the outer wedge show near-surface deformation that is interpreted as coseismic. Watt et al. (in review) present further evidence for coseismic slip of splay faults from high-resolution seismic reflection and bathymetric data. Hill et al. (2026) document mass transport deposits in sediment cores that suggest repeated, coseismic deformation of the outer wedge. Biemiller et al. (2025)

use dynamic rupture models to show that realistic geometries of both landward and seaward-vergent splay faults can and do slip readily along with the megathrust slip to the toe at Cascadia. Cubas et al. (2016) find that the formation of landward-vergent splay faults in particular requires such low coefficients of friction on the megathrust fault (< 0.03) that they must be formed coseismically, therefore essentially serving as geological evidence for coseismic slip. This preponderance of evidence suggests slip along the megathrust fault to the toe of the wedge, likely involving slip on multiple splay faults as well, is a common rupture mode at Cascadia.

Shallow accretionary wedge deformation is the primary control on tsunami size (e.g., Satake and Tanioka, 1999; Wilson and Ma, 2021). Therefore, a shallower megathrust fault position than anticipated may significantly modify and potentially increase the accretionary wedge deformation experienced during megathrust slip. Our work highlights that, where there are multiple alternatives for the position of the megathrust fault, this range of possibilities must be considered in hazard analysis.

Most modeling work at Cascadia (e.g., Biemiller et al., 2025) uses the Slab2 plate interface model (Hayes et al., 2018). Carbotte et al. (2024)’s model is kilometers shallower than the Slab2.0 model in some parts of the LVZ’s outer wedge. When accounting for the new, even shallower megathrust fault position identified herein, these differences from Slab2.0 are compounded. Because the LVZ is also coincident with much of the fully locked region of Cascadia (Li et al., 2018; Lindsey et al., 2021), better understanding the characteristics of the décollement in this area is all the more important.

Comparison to global examples

Our findings are consistent with previous work on outer accretionary wedge structure at other sediment-rich margins, where the megathrust fault has been similarly interpreted as localizing at various levels above, but only rarely co-located with, the sediment-basement interface. For example, except where disrupted by an isolated sub-

ducting seamount, the décollement sits several kilometers above the oceanic crust in the northern, central, and southern Hikurangi margin (Barnes et al., 2010; Ghisetti et al., 2016; Plaza-Faverola et al., 2012; Watson et al., 2019). Though many subduction zones exhibit along-strike heterogeneity, the décollement is also within the sedimentary section in well-studied regions of Barbados (< 1 km Labaume et al.; Laigle et al., 2013; Shipley et al., 1997), Alaska (< 1 km Cortés-Rivas et al., 2025; von Huene et al., 2021), Makran (≤ 5 km Burg, 2018; Grando and McClay, 2007; Kopp et al., 2000), and Nankai (1-2 km Bangs et al., 2009; Moore et al., 2007a). Even at Costa Rica, a sediment-poor subduction zone, the décollement is hosted in the sediments, facilitating ~ 10 to 250 m of sediment subduction (Bangs et al., 2015; Edwards et al., 2018; Kirkpatrick et al., 2020).

Tréhu et al. (2019) find that the rupture zone of the 2010 Mw 8.8 Maule earthquake occurred in a region with a shallow décollement and thick sediment underthrusting, while the rupture was arrested where the megathrust steps down to closer to, but still above, the sediment-basement interface. They suggest that a shallow, sediment-hosted décollement is linked to coseismic slip to the trench. Even with our new analysis, it is true that the décollement is at a higher level, promoting more sediment subduction, in the seaward-vergent zone offshore southern Oregon; however, this is the opposite relationship as observed by Tréhu et al. (2019) in Chile because offshore Washington has both thick sediments and near-full sediment accretion but also is the area likeliest to have large slip to the trench (e.g., Han et al., 2017; Li et al., 2018; Lindsey et al., 2021). The same relationship holds true for Sumatra, where the 2004 Mw9.1 earthquake was confined to the area with the thickest sediments and landward-vergent faults (Gulick et al., 2011; Henstock et al., 2006; Ishii et al., 2005; Seeber et al., 2007). However, in both the 2004 rupture zone and outside of it, the décollement is continuous at around 500 meters above the top of oceanic crust, though with varying amplitude (Dean et al., 2010; Frederik et al., 2015).

Taken together, this work in many subduction zones shows that plate boundary

faults that host the largest earthquakes preferentially are hosted within the elastic sedimentary section as opposed to along the sediment-basement interface or within basaltic crust, as long as an adequate sediment column is present. Our work provides an important constraint on the appropriate choice for physical parameters for dynamic rupture models. Since the strength of sediments in the rupture area is known to strongly control near-trench earthquake and tsunami characteristics (Ulrich et al., 2022), accounting for the lithology of the host rock of the megathrust fault is of the utmost importance. Additionally, while other exhumed subduction complexes often display deformation involving both sediments and basaltic units (e.g., Fisher et al., 2021a), a recently identified example of an exhumed Cascadia paleomegathrust fault indicates the plate boundary was hosted entirely within the sediments (Ledeczi et al., 2025), so this work is also consistent with the geological record for this margin.

Therefore, our work, taken with studies from other margins, would suggest that a first-order influence on earthquake hazards is whether the décollement is hosted in the sediments at all, rather than its specific position and the ratio of sediment subduction to accretion it promotes. This is consistent with global observations that where thick sediments help cover the rough topography of the oceanic crust, smoothing the plate interface, frictional locking and large megathrust earthquakes are favored (Heuret et al., 2012; Ruff, 1989; Wang and Bilek, 2014). The work presented here favors a sediment-hosted megathrust fault system for the entirety of the LVZ.

Critical wedge geometry and basal strength

Critical wedge theory dictates that the wedge taper angle, or the sum of the surface slope α and the décollement dip β , reflects the strength of the wedge and the friction along the décollement (Davis et al., 1983; Dahlen et al., 1984). The décollement dips we calculate for the newly identified décollement are largely below 3° and are all slightly shallower ($\leq 2^\circ$ different) than the Carbotte et al. (2024) décollement. Taper angle increases towards the south as the wedge grows narrower and its surface slope

increases. Watt and Brothers (2020) provided the most recent estimate of these values for Cascadia where they approximate the décollement dip using a slab model with higher uncertainties (Blair et al., 2011) and a larger down-dip area of the wedge to the outer-arc high, deriving values closer to 8° for LVZ.

The basal dip calculated here and in Carbotte et al. (2024) imply a much weaker megathrust from critical wedge theory than in Watt and Brothers (2020) but is more consistent with values reported in previous work (e.g., Cubas et al., 2016; Flueh et al., 1998; Gutscher et al., 2001). Taken with well-established low surface slope of the wedge in this region ($\alpha = 1 - 2^\circ$, e.g., Han et al., 2017; Watt and Brothers, 2020), the low total wedge taper angle requires both a strong wedge and a very weak basal fault. These values would put the LVZ's outer wedge closer to other sediment-rich settings like Barbados, Hikurangi, and Makran (Hu and Wang, 2008; Watt and Brothers, 2020). These values would put the LVZ's outer wedge close to other clastic-sediment dominated wedge settings (Hu and Wang, 2008; Watt and Brothers, 2020) like Barbados (décollement in clays Labaume et al.; Shipley et al., 1997), central and southern Hikurangi (décollement in pelagic sediments Barnes et al., 2010), and Makran (décollement in shale within turbidites Burg, 2018; Schlüter et al., 2002).

The LVZ comprises the entire extent of the very low taper wedge. It has been previously established that landward-vergent faults require extremely low coefficients of friction on the megathrust fault to form (Cubas et al., 2016). Therefore, both critical wedge theory and the existence of landward-vergent faults themselves require a weak basal friction. This condition may be achieved two means: high pore pressure or dynamic friction during earthquake slip.

There is not positive evidence of widespread high pore fluid pressure in the LVZ. The extremely high heat flow should drive shallow dehydrating diagenetic reactions to completion before the deformation front (Salmi et al., 2017) and strengthen the wedge (Han et al., 2017). A similar phenomenon has been observed in the rupture zone of 2004 Mw9.1 Sumatra earthquake where diagenetic strengthening of the wedge

extends the seismogenic zone to the deformation front and drives shallow earthquake rupture (Hüpers et al., 2017b). In this region, a high amplitude negative polarity décollement has been well-documented (Dean et al., 2010) and attributed to these shallow dehydration reactions (Hüpers et al., 2017b). While Cascadia’s LVZ should be a very similar thermal environment to this region, we do not observe evidence for high amplitude negative polarity reflectors at the décollement levels we interpret here. This further suggests that there is no trapped pore pressure below the fault, nor is there a strong impedance contrast between the materials above and below it, consistent with observations of low reflectivity sediments below the 0.9 Ma horizon in Unit IB (Rohr et al., 2019). Any low velocity or high pore pressure zone present is likely too thin to be imaged as a strong reflector.

This is supported by, for example, analysis in Han et al. (2017) which indicates the LVZ’s outer wedge is over-consolidated, and thus mechanically strong, and the basal sediment layer is low in porosity. Peterson and Keranen (2019) document a low porosity and strong basal (1 km thick) sedimentary package in the LVZ. Overpressure ratios are expected to be at maximum 0.1 in the outer wedge and there is no evidence for elevated pore pressures at the décollement (Webb, 2017). While the incoming sediments may be slightly overpressured, fluid likely drains before the frontal thrust and thus does not become trapped within the wedge or at the décollement (Webb, 2017). Previous work has also suggested a regional contrast along the margin where Central Oregon exhibits an anomalously low residual Vp sediment package, interpreted as an undercompacted, fluid-rich package, below a shallow sediment décollement (Han et al., 2017). No evidence for an undercompacted, fluid-rich sediment package was found along the Washington margin.

Because there is no evidence for high pore pressure, the extremely low basal friction values required by the wedge morphology in the LVZ must be achieved through other means. Cubas et al. (2016) also show that landward vergence can only be explained by extremely low values of friction on the megathrust fault of ≤ 0.03 . There are

no known geological materials with intrinsic friction values so low. Therefore, the most likely explanation of required low coefficients of friction is dynamic weakening during earthquake slip (Cubas et al., 2016; Di Toro et al., 2011a). The flat megathrust (e.g., Bletery et al., 2016) in the LVZ furthermore suggests that earthquake rupture propagation here would be favored compared to the higher curvature wedge plate interface to the south.

3.6 Conclusions

In Cascadia’s landward vergence zone (LVZ), the décollement has been interpreted as being hosted within the sediment section (<500 m at the deformation front Gutscher et al., 2001; MacKay et al., 1992; MacKay, 1995; Webb, 2017), as a duplex bound by faults within the sediment section and at the TOC (Adam et al., 2004; Booth-Rea et al., 2008), and as a singular fault coincident with the TOC (Carbotte et al., 2024; Peterson and Keranen, 2019). We apply three methods—observations of reflectivity, as well as two analytical techniques of structural geology—to the newly collected CASIE21 seismic reflection dataset to reexamine the accretionary wedge structure and megathrust fault position.

Building on observations in Booth-Rea et al. (2008) and Suppe et al. (2004), we find that the structural style of the landward-vergent splay faults is most consistent with shear fault-bend fold wedges, in which the folded backlimbs dip less than the fault planes themselves. The forelimbs of these faults are consistent with fault-propagation folding. In simple shear fault-bend folding theory, a deforming volume at the base of the fault is invoked rather than a basal fault for the portion of slip that produces the splay fault. Nevertheless, the majority of slip progresses forward on the underlying décollement fault to build the wedge outwards.

We show that all splay faults within the LVZ sole into sedimentary levels on average ~500 m above oceanic basement in the active outer wedge, and increasing in height above basement landward. The amount of sediment underthrusting varies spa-

tially between 80 and 2.95 km, but at no point is the décollement coincident with the sediment-basement interface in this region. We review examples of other sediment-rich accretionary wedges and find little evidence that décollements are interpreted as located at the TOC, except under special circumstances like seamount subduction. It has been well established that sediment-hosted décollements are smoother faults which can then drive increased interseismic locking and large earthquake ruptures (Heuret et al., 2012; Ruff, 1989; Wang and Bilek, 2014). Therefore, careful seismic interpretation is key to fully understanding the conditions and physical fault zone properties under which rupture may propagate, and therefore the hazards of accretionary wedges.

In the active domain of the outer wedge (as defined in Ledeczi et al., 2024), our newly identified megathrust position is very well constrained by the multiple independent analytical approaches and is the active plate boundary. In the inactive domain, however, it is more ambiguous whether (1) this megathrust is the same fault structure and currently active megathrust as in the active domain; or (2) this is an abandoned, relict fault structure that would not host present-day megathrust slip. In most cases, it is much more likely that this fault connects to the active domain's megathrust because it is close to the TOC and because, in the absence of direct evidence otherwise, it is the simplest explanation (Fig. 10A). In at least one case, the inactive domain's megathrust is an abandoned fault that is cut off by the active domain's throughgoing megathrust, facilitating duplexing and underplating in this zone (Fig. 10C).

It has previously been well-established that the LVZ comprises Cascadia's most highly locked region (Li et al., 2018; Lindsey et al., 2021) and slip to the deformation front in this area is likely (Biemiller et al., 2025; Cubas et al., 2016; Han et al., 2017; Hill et al., 2026; Ledeczi et al., 2024). We find that the amount of underthrust sediment increases from the active domain to the inactive domain of the outer wedge, defining a basal décollement with a much shallower dip than previous estimates.

We calculate the dip of the décollement only in cases where we consider it an active, through-going structure. We find this taper is usually $<3^\circ$ but as low as 0.6° , implying an extremely strong wedge with a weak basal friction. This also defines an increase in sediment underthrusting downdip which could result from basal erosion of the accretionary wedge above or uneven original deposition of sediments on the downgoing plate.

This wedge morphology and weak megathrust can be produced either through high pore fluid pressure or an extremely low (< 0.03 Cubas et al., 2016) friction on the megathrust. Our identified décollement does not always sole into the same stratigraphic level, though it is most likely hosted in hemipelagic muds older than 0.9 Ma (Su et al., 2000; Rohr et al., 2019). It does not seem to have a high amplitude negative polarity reflection, indicating low pore fluid pressure, which is consistent with previous work in the LVZ (Han et al., 2017; Webb, 2017; Peterson and Keranen, 2019). Therefore, we suggest that the extremely low friction required by the presence of landward-vergent faults and a low taper wedge is best explained by dynamic friction during earthquakes (e.g., Di Toro et al., 2011a).

Acknowledgements

Funding for this work was provided by the University of Washington and the U.S. National Science Foundation through awards 2103713 and 2217468 to Tobin. This research was supported by a seed grant to Ledeczi from Cascadia Region Earthquake Science Center (CRESCENT), funded via National Science Foundation Cooperative Agreement EAR-2225286. Use of MOVE (PE Limited) and Kingdon (S&P Global) softwares were permitted by academic licenses to UW.

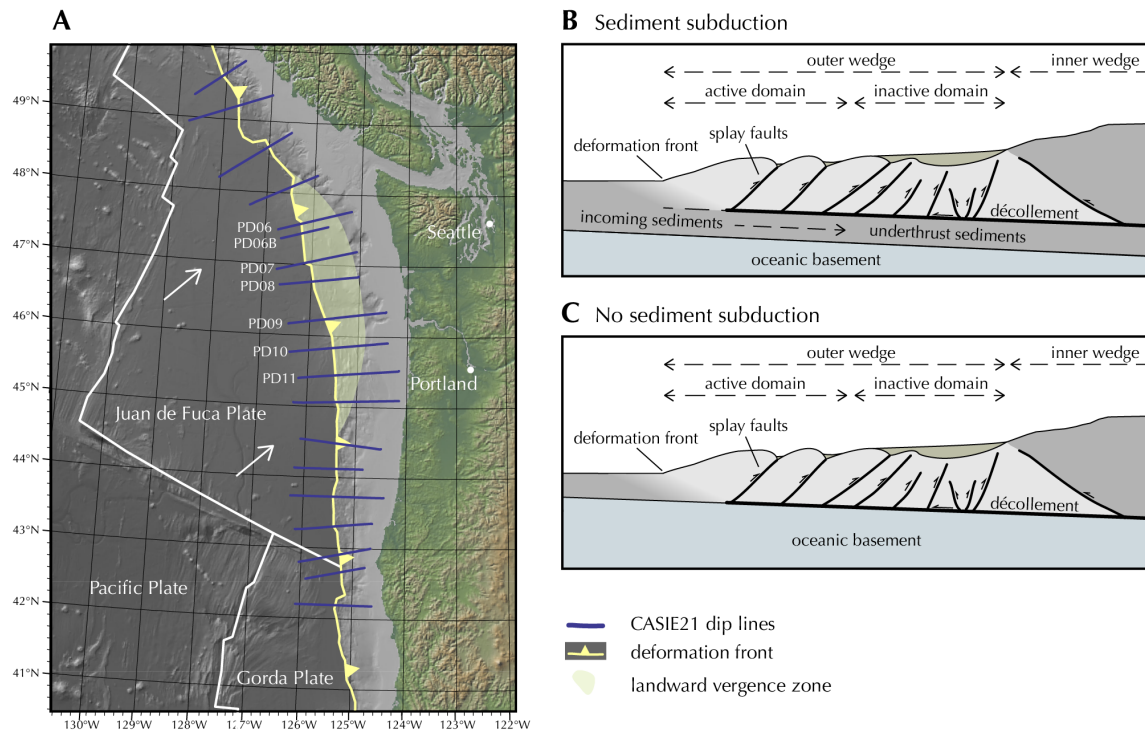


Figure 3.1: **A.** A bathymetric map (GMRT from Ryan et al., 2009a) of the Cascadia subduction zone, where the Juan de Fuca plate subducts at $\sim 3\text{--}4$ mm/yr under the North American plate. The landward vergence zone is shown in green. Dip lines from the CASIE21 survey (Carbotte et al., 2023) are shown in blue. **B.** and **C.** are schematics of landward vergence zone accretionary wedge structure with and without shallow sediment subduction, respectively.

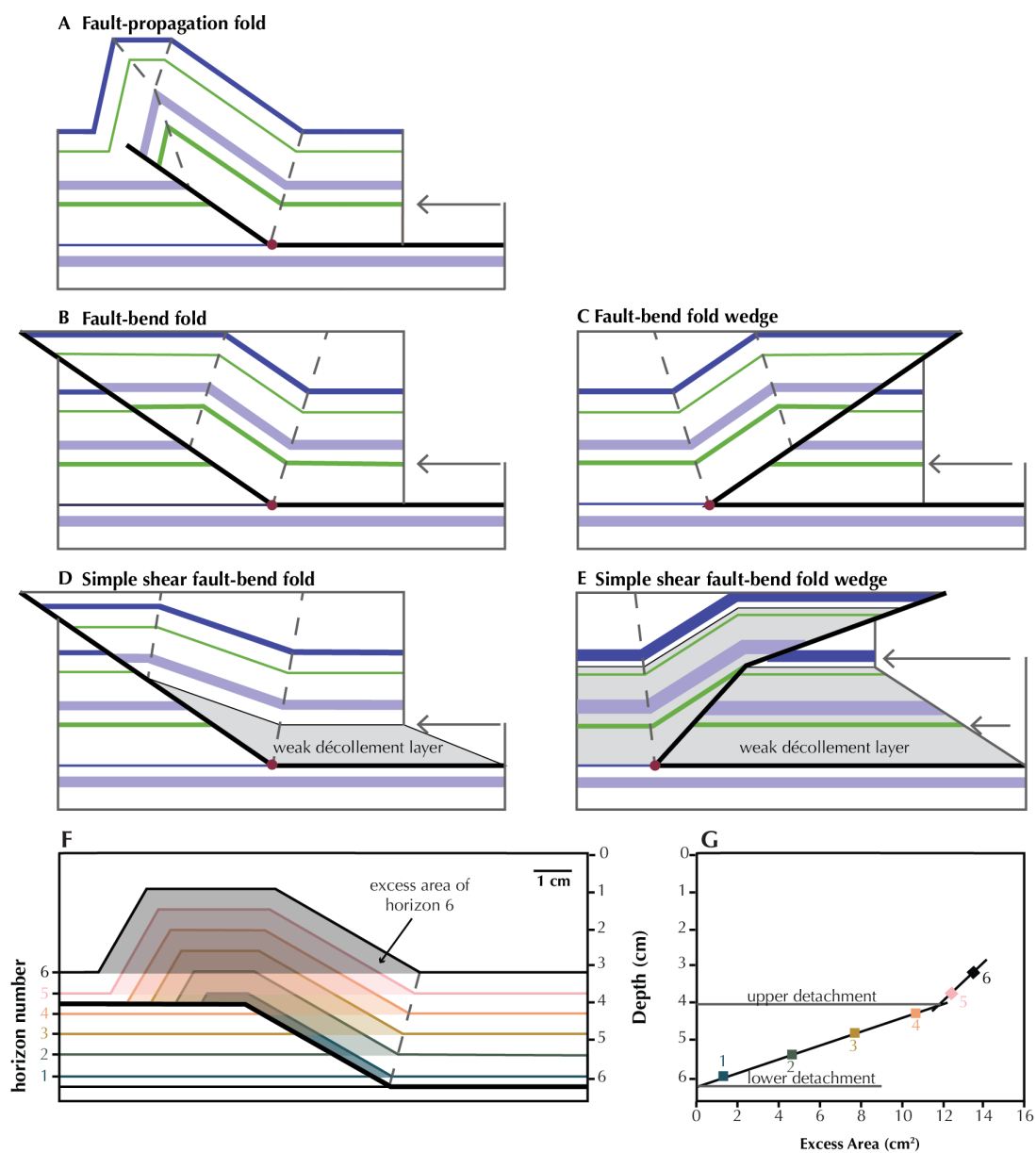


Figure 3.2: Modified after Suppe (1983), Suppe and Medwedeff (1990), and Suppe et al. (2004) In both the **A.** fault-propagation fold and **B.** the fault-bend fold, the backlimb strata are parallel to the fault. **C.** For landward-vergent faults, the fault-bend fold wedge is more appropriate. **D.** In the simple shear fault-bend fold, backlimb strata dip less than the fault due to distributed shearing within the grey décollement layer. **E.** The simple shear fault-bend fold wedge is more complex than in D. and invokes a larger deforming décollement layer. In all models, the synclinal hinge intersects the fault at the position of the décollement (red dot). **F.** Idealized fault-bend fold with two detachment levels. **G.** Demonstration of area-depth method for fault-bend fold in F. The area-depth method correctly predicts both detachment levels. F and G adapted from Groshong et al. (2012).

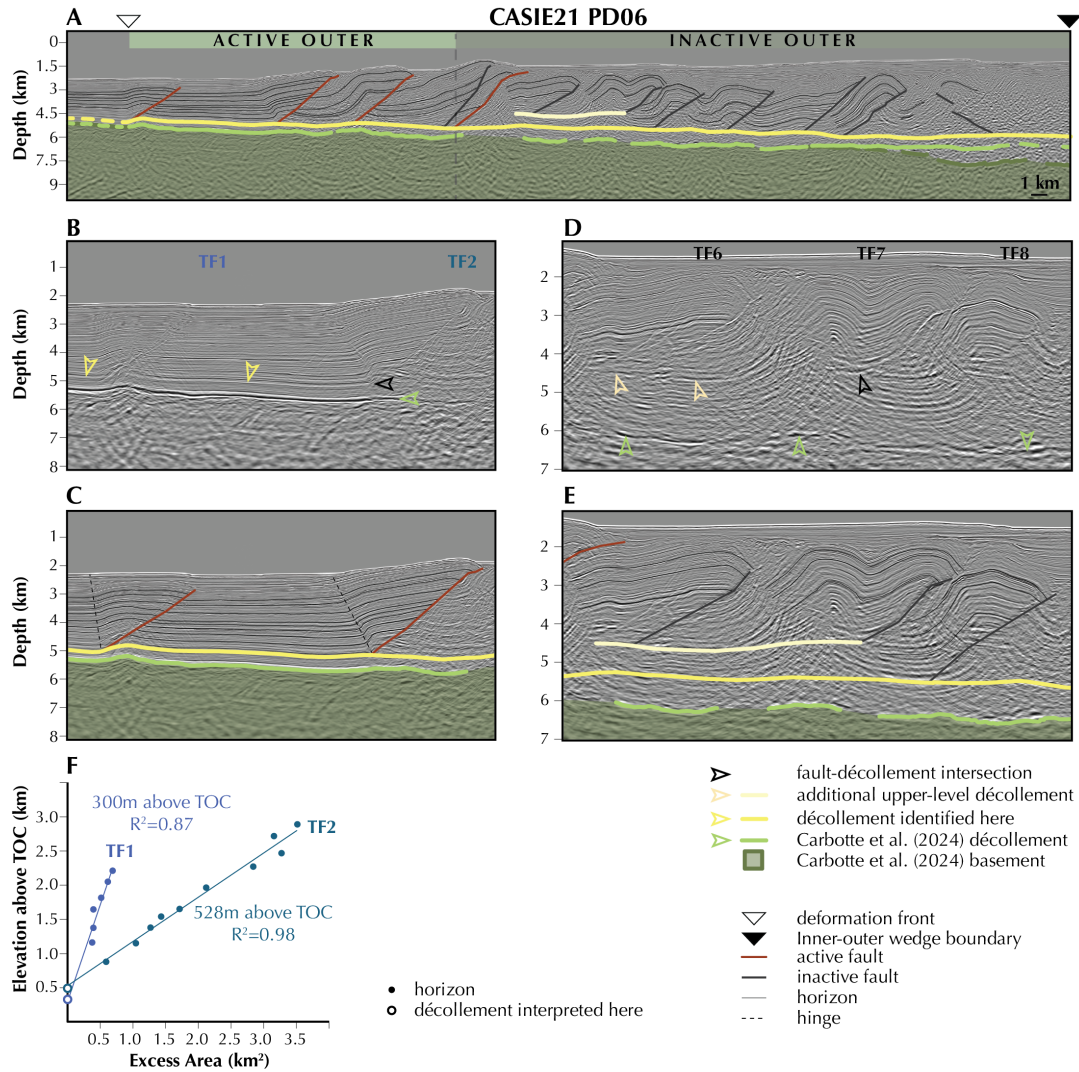


Figure 3.3: CASIE21 PD06. $VE=1.0$ for all panels. **A.** Portion of CASIE21 PD06 from deformation front to inner-outer wedge transition. Décollement dip β for décollement interpreted here and that of Carbotte et al. (2024) is shown. **B.** Uninterpreted and **C.** interpreted image of frontal thrust region. **D.** Uninterpreted and **E.** interpreted image of inactive wedge region. **F.** Area-depth analysis predicts a décollement 300 and 528 m above TOC for TF1 and TF2, respectively. Seismic interpretation in panels A-C shows a décollement 330 and 492 m above TOC for TF1 and TF2.

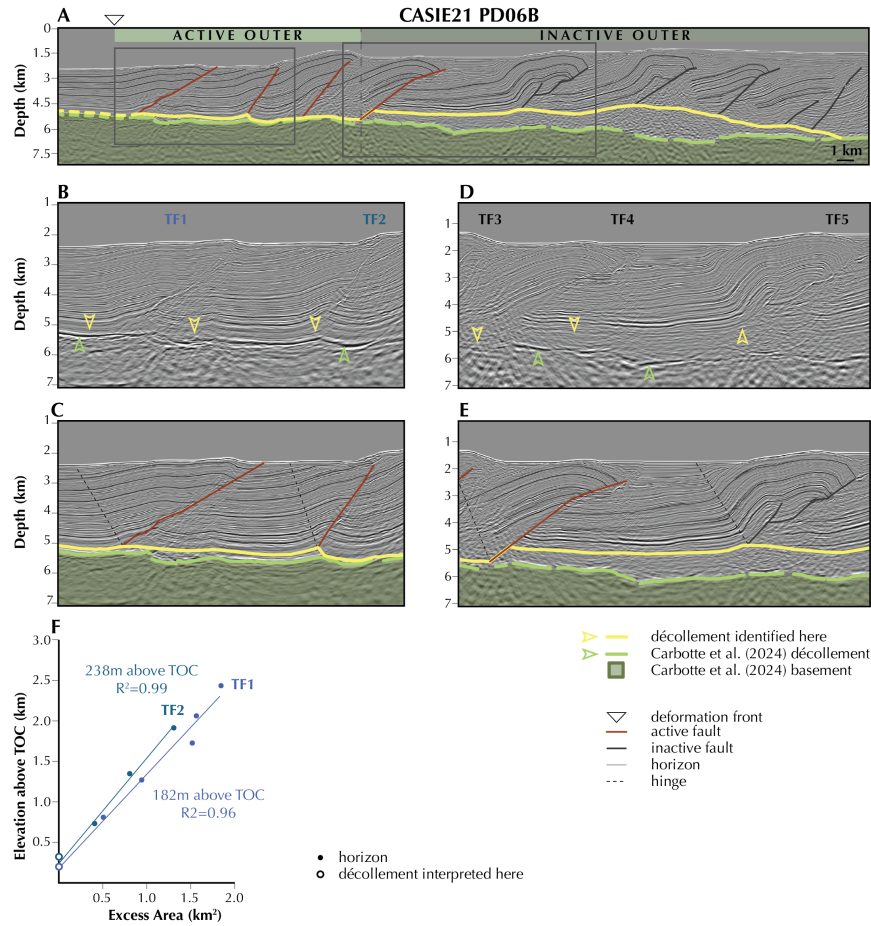


Figure 3.4: CASIE21 PD06B. VE=1.0 for all panels. **A.** Portion of CASIE21 PD06B from deformation front to end of line before the inner-outer wedge transition. **B.** Uninterpreted and **C.** interpreted image of frontal thrust region. **D.** Uninterpreted and **E.** interpreted image of active-inactive domain boundary where the décollement steps up at TF4. **F.** Area-depth analysis predicts a décollement 182 and 238 m above TOC for TF1 and TF2, respectively. Seismic interpretation in panels A-C shows a décollement 201 and 324 m above TOC for TF1 and TF2.

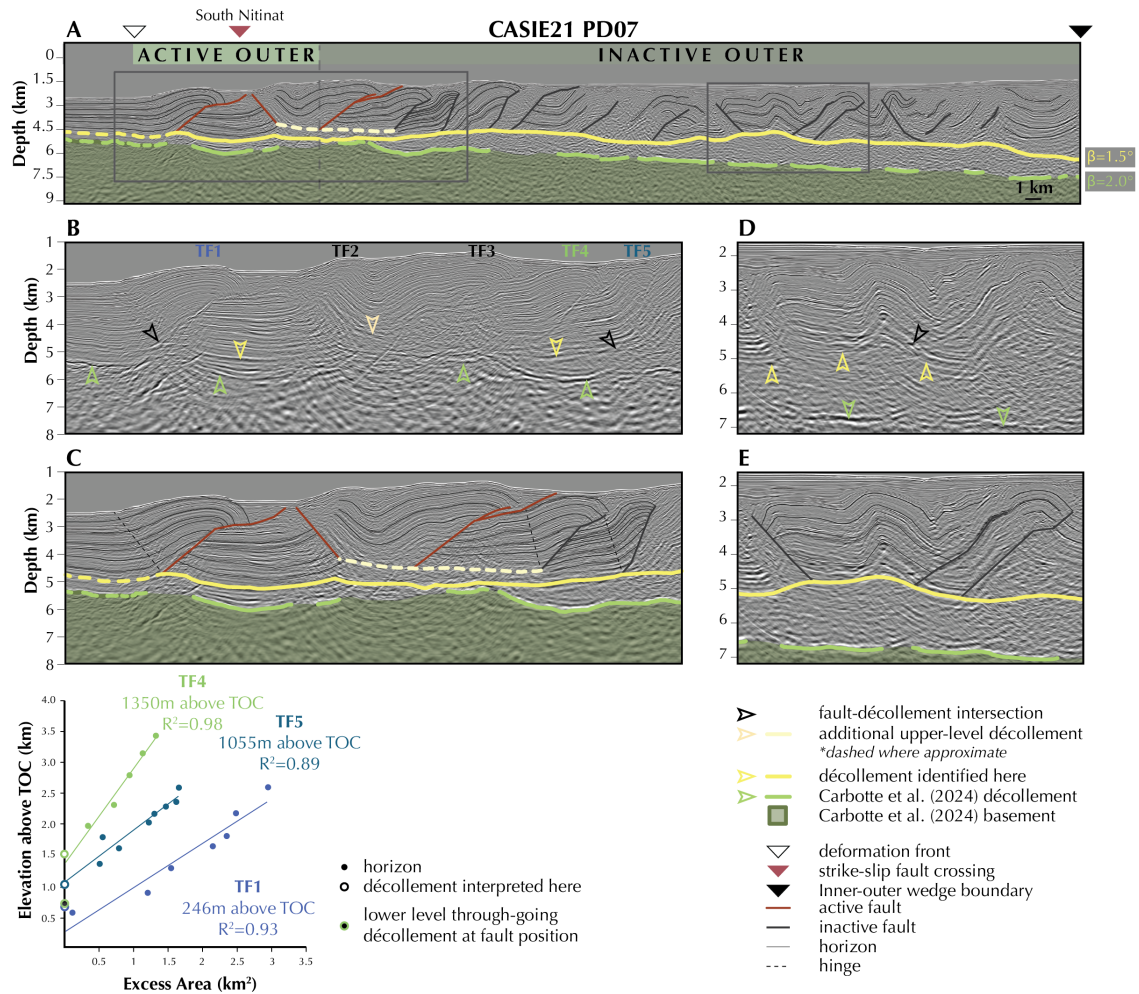


Figure 3.5: CASIE21 PD07. VE=1.0 for all panels. **A.** Portion of CASIE21 PD07 from deformation front to inner-outer wedge transition. Décollement dip β for décollement interpreted here and that of Carbotte et al. (2024) is shown. **B.** Uninterpreted and **C.** interpreted image of active outer wedge. **D.** Uninterpreted and **E.** interpreted image of detachment fold in inactive wedge region. **F.** Area-depth analysis predicts a décollement 246 and 1055 m above TOC for TF1 and TF5, respectively. We interpret a shallower décollement of 708 m above TOC at TF1 based on the clear fault plane reflection shown in B. We follow area-depth analysis for TF5 and pick a décollement 1017 m above TOC in A-C. TF4 soles into a secondary higher décollement level. Area-depth theory predicts a décollement at 1350 m above TOC which is close to our interpretation shown in A-C at 1347 m above TOC. The through-going décollement at this region is only 816 m above TOC (black dot with green outline) and is therefore clearly the incorrect soling depth for this fault.

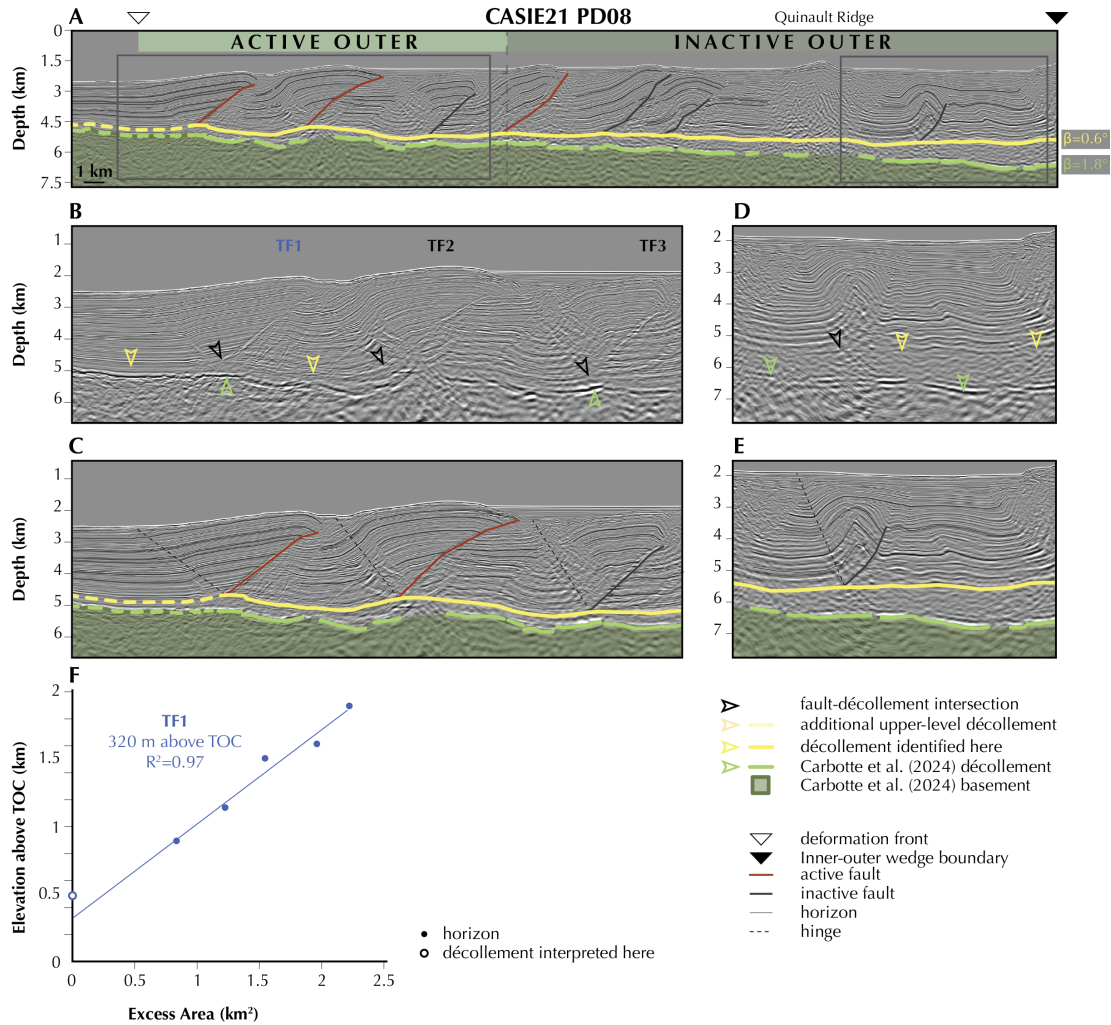


Figure 3.6: CASIE21 PD08. VE=1.0 for all panels. **A.** Portion of CASIE21 PD08 from deformation front to inner-outer wedge transition. Décollement dip β for décollement interpreted here and that of Carbotte et al. (2024) is shown. **B.** Uninterpreted and **C.** interpreted image of frontal thrust region. **D.** Uninterpreted and **E.** interpreted image of inactive wedge region. **F.** Area-depth analysis predicts a décollement 320 m above TOC for TF1. We interpret a shallower décollement of 488 m above TOC at TF1 based on the clear fault plane reflection shown in B.

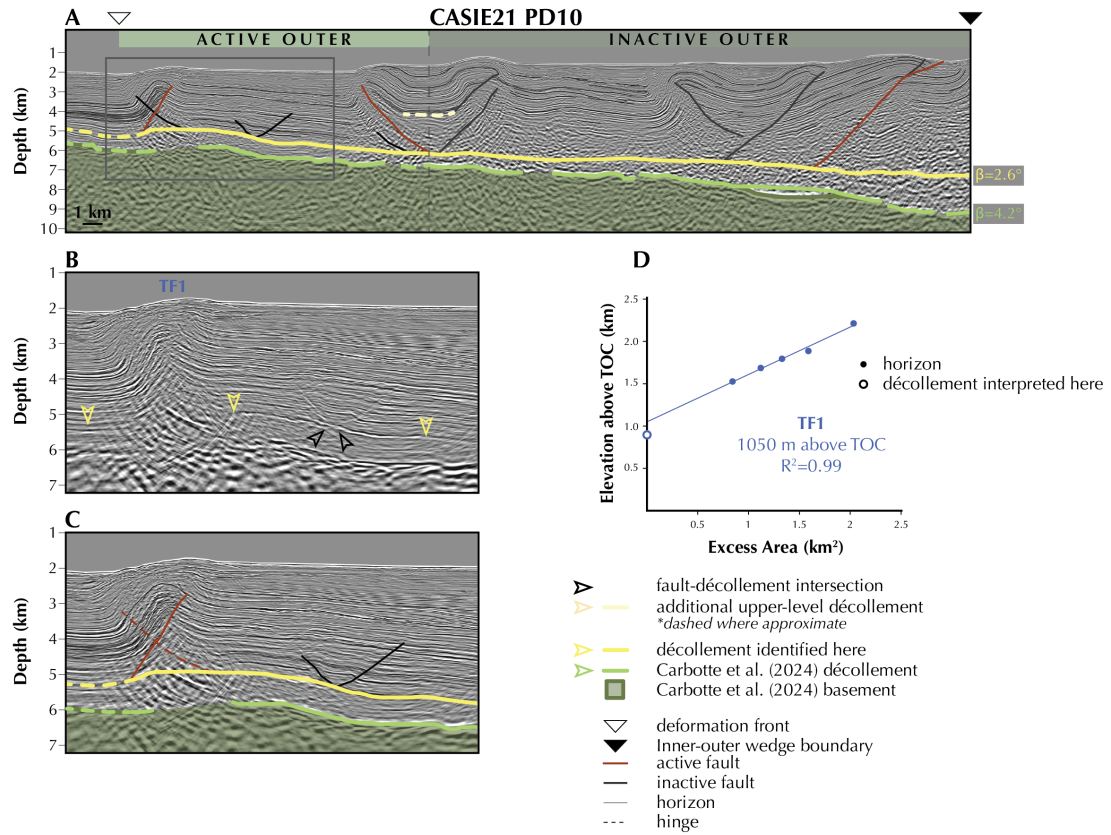


Figure 3.7: CASIE21 PD10. VE=1.0 for all panels. **A.** Portion of CASIE21 PD10 from deformation front to inner-outer wedge transition. Décollement dip β for décollement interpreted here and that of Carbotte et al. (2024) is shown. **B.** Uninterpreted and **C.** interpreted image of frontal thrust region. **D.** Area-depth analysis predicts a décollement 1050 m above TOC for TF1. We interpret a deeper décollement of 896 m above TOC at TF1 based on the clear fault plane reflections of the small conjugate thrusts landward of TF1.

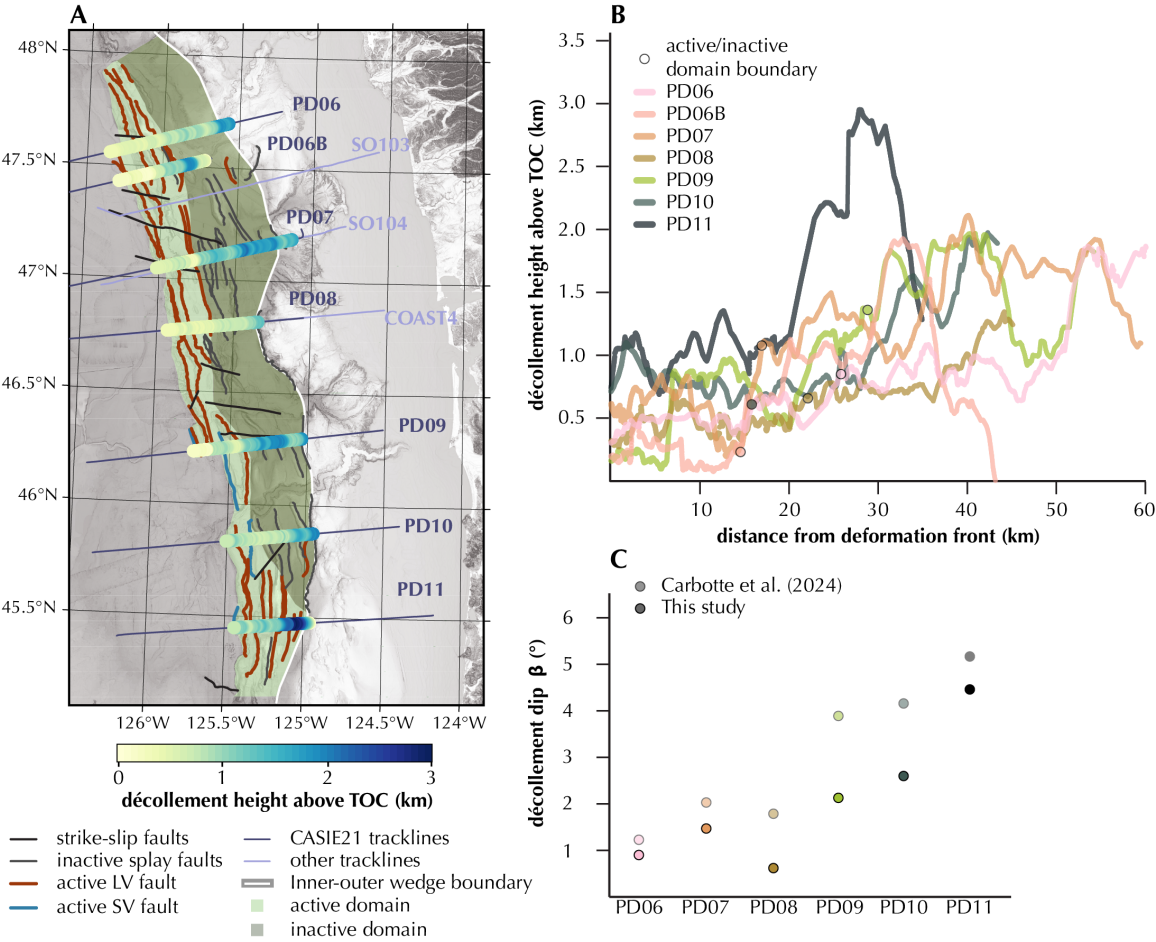


Figure 3.8: **A.** Structural map of the landward vergence zone outer wedge adapted from Ledeczi et al. (2024). The active domain of the wedge, shown in light green, primarily hosts active landward-vergent (red) and some seaward-vergent (blue) splay faults. Strike-slip faults and inactive splay faults are shown in black and grey, respectively. The inner-outer wedge boundary is from Lucas et al. (2025). Tracklines for the CASIE21 dataset, as well as for select profiles from the SO108 (Flueh et al., 1998) and COAST (Holbrook et al., 2012) surveys are shown. The line from the 1989 survey (Moore et al., 2013) shown in Fig. 9A is south of the map area. The colors overlain on each CASIE21 line shows the elevation of our interpreted décollement above the top of oceanic crust (TOC) in Carbotte et al. (2024). Bathymetry utilized in this map are from Beeson et al. (2025) and Dartnell et al. (2023). **B.** Décollement height above TOC in Carbotte et al. (2024) for each seismic line interpreted herein. The dot on each line shows the boundary between the active and inactive domains of the outer wedge, and each profile ends at the inner-outer wedge transition of Lucas et al. (2025). As defined in Ledeczi et al. (2024), there is no inactive outer wedge in PD11. PD06B is a partial line that does not reach the inner-outer wedge boundary. **C.** Dip of the décollement (β , as defined in Dahlen et al., 1984; Davis et al., 1983) within the outer wedge for the Carbotte et al. (2024) plate boundary and the décollement interpreted in this work. Color maps in B and C are from Crameri et al. (2020) and Crameri (2023).

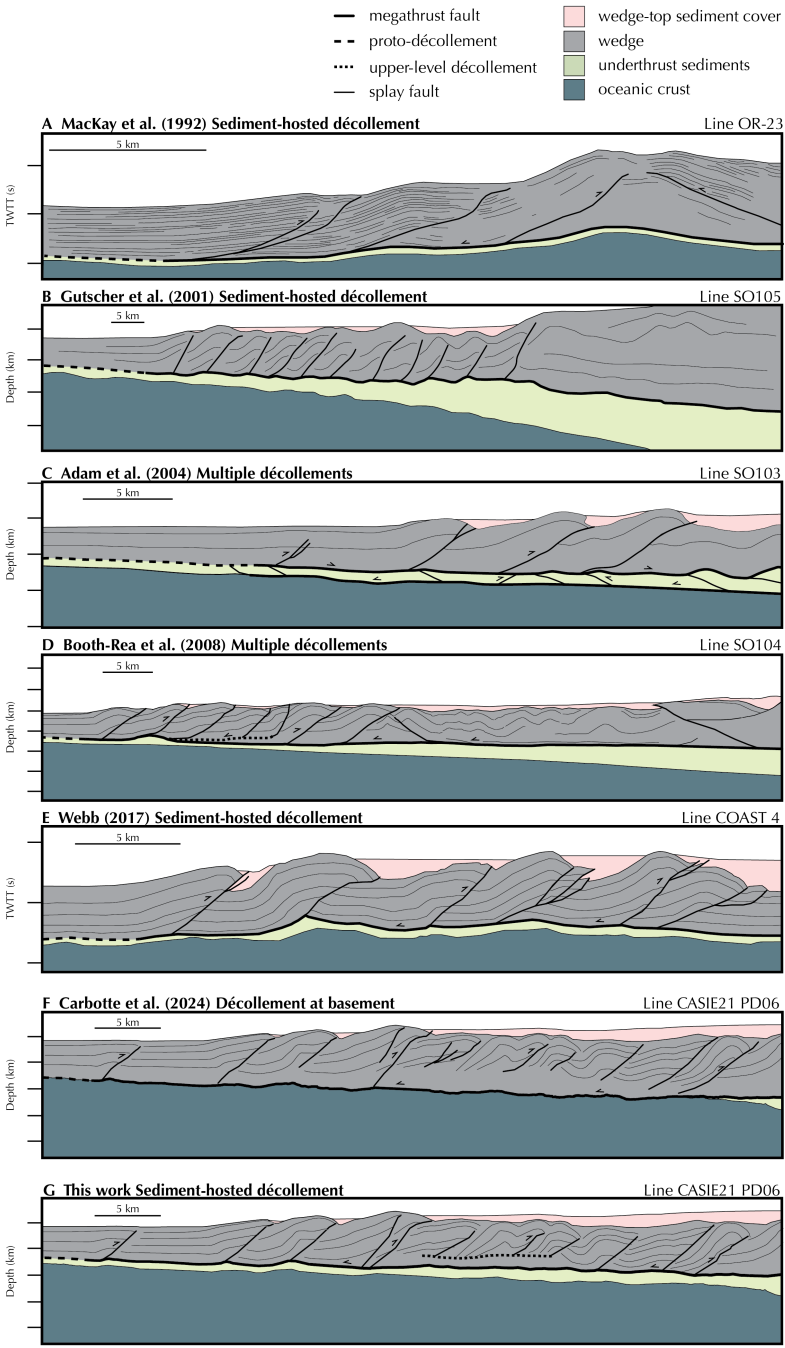


Figure 3.9: Published interpretations of landward vergence zone seismic lines. **A.** MacKay et al. (1992, Fig. 4) interpret 400 m of sediment underthrusting in the outer wedge. The seismic line is in time, which produces the apparent topography in the oceanic crust. This is nearest to CASIE21 Line PD12. $VE \approx 2.0$. **B.** Gutscher et al. (2001, Fig. 2a) interpret a sediment-hosted décollement ~ 500 m above basement near the frontal thrusts. The dip is near-horizontal and the sediment underthrusting increases into the inner wedge. Line SO108-105 is nearest to CASIE21 Line PD08. Original figure modified to make $VE \approx 2.0$. **C.** Adam et al. (2004, Fig. 3b) interpret multiple décollements, one at the oceanic basement and one within the sediments, with opposing slip senses. Line SO108-103 is nearest to CASIE21 PD06B. $VE = 1.0$. **D.** Booth-Rea et al. (2008, Fig. 6)'s shallow sediment-hosted décollement is very similar to our model. Line SO108-104 is coincident with CASIE21 PD07. $VE = 1.0$. **E.** Webb (2017, Fig. 2.7) also interpret a shallow sediment-hosted décollement. Line COAST 4 is coincident with CASIE21 PD08. $VE \approx 2.0$. **F.** Carbotte et al. (2024, Fig. 3a) plate model with Jiang et al. (2023)'s fault interpretations. $VE = 1.0$. **G.** We instead interpret a sediment-hosted décollement. $VE = 1.0$.

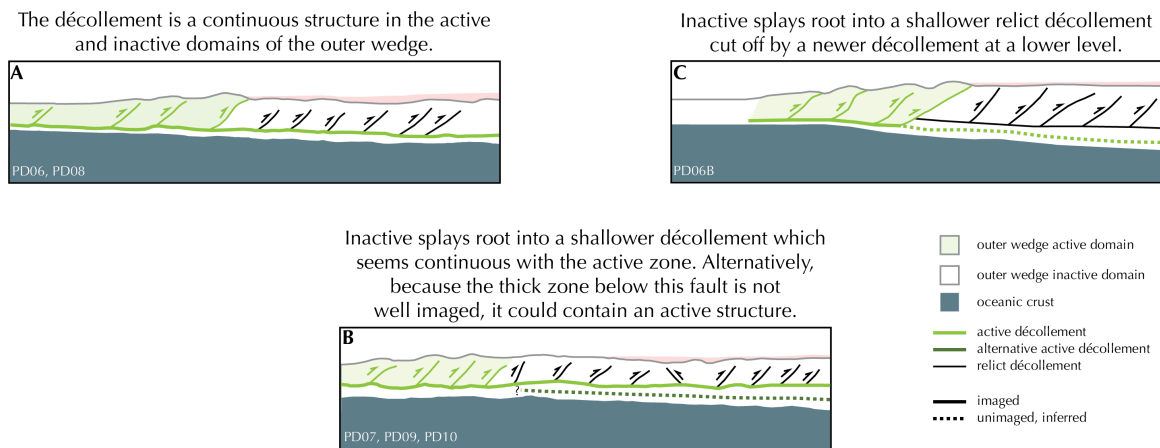


Figure 3.10: Conceptual models of décollement positions within the outer wedge in the landward vergence zone. The active and inactive domains are shown in light green and white, respectively. Oceanic crust is blue.

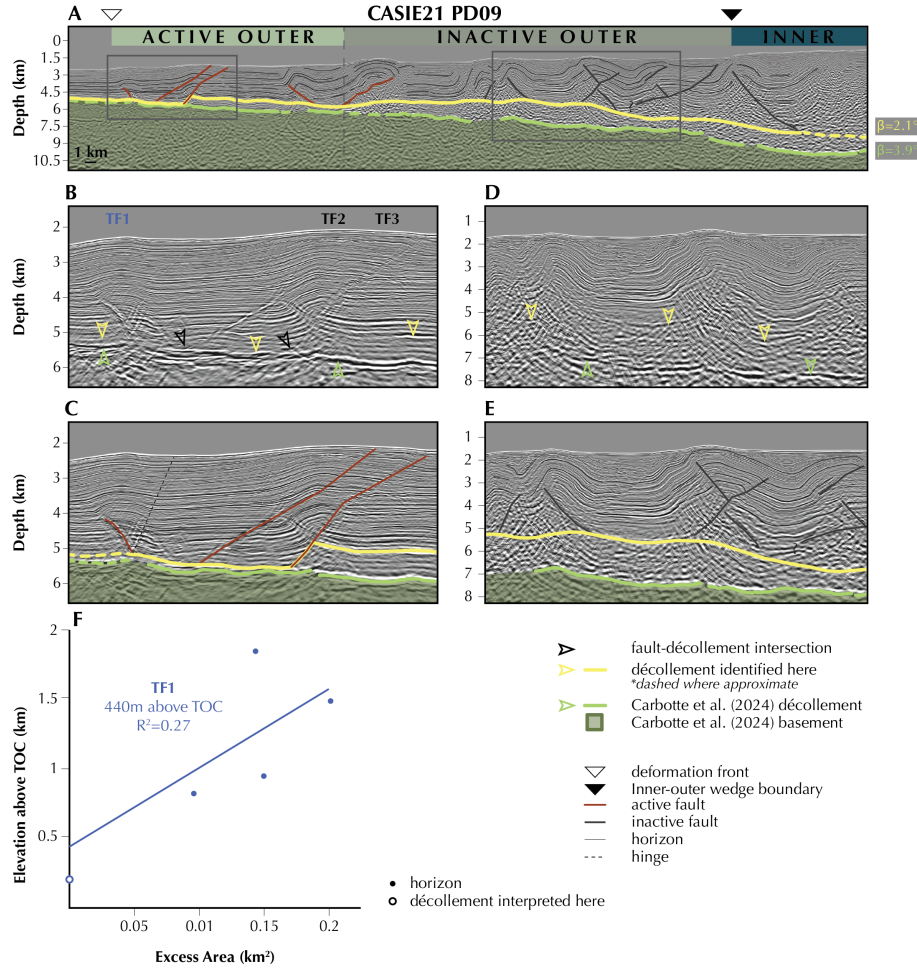


Figure 3.11: Figure S1: CASIE21 PD09. $VE = 1.0$ for all panels. **A.** Portion of CASIE21 PD09 from deformation front to inner wedge. Décollement dip β for décollement interpreted here and that of Carbotte et al. (2024) is shown. **B.** Uninterpreted and **C.** interpreted image of frontal thrust region. **D.** Uninterpreted and **E.** interpreted image of inactive domain. **F.** Area-depth analysis is not highly certain for TF1, but it predicts a décollement 440 m above TOC for TF1. We interpret a deeper décollement of 205 m above TOC at TF1 based on a clearer fault plane reflection from TF2.

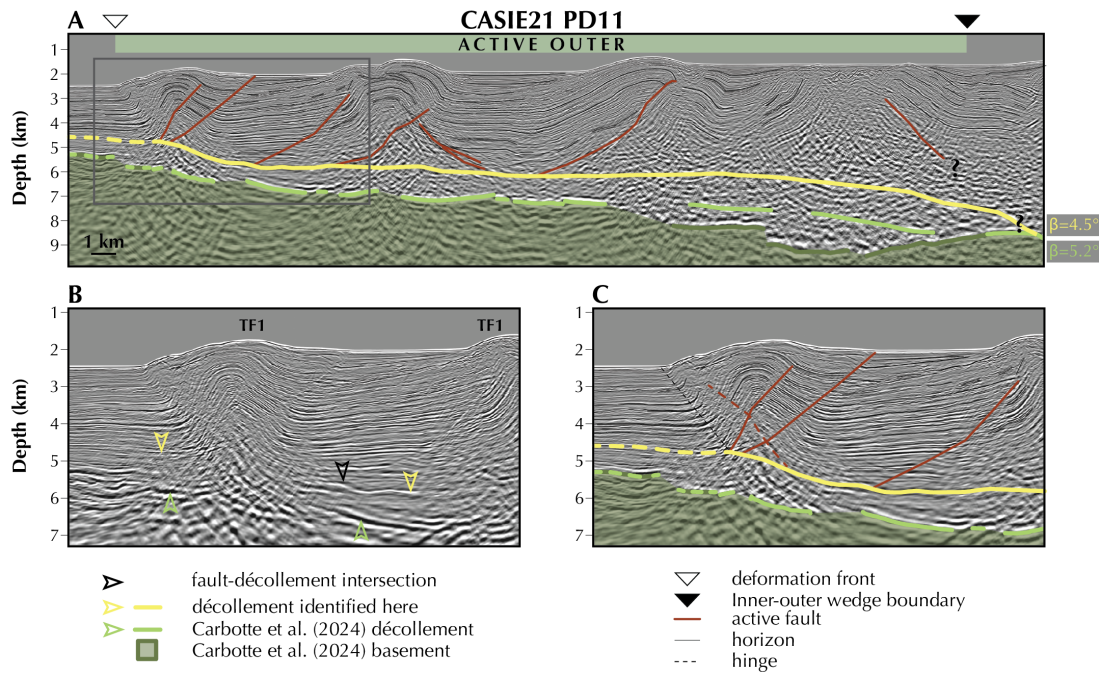


Figure 3.12: Figure S2: CASIE21 PD11. $VE=1.0$ for all panels. **A.** Portion of CASIE21 PD11 from deformation front to inner wedge. Décollement dip β for décollement interpreted here and that of Carbotte et al. (2024) is shown. Note that this calculation is for the outer wedge only so does not take into account the point where the surfaces potentially converge. **B.** Uninterpreted and **C.** interpreted image of frontal thrust region. Because TF1 is not well-imaged, we cannot perform area-depth analysis for it; however, the fault plane reflection at TF2 is sufficiently clear to interpret a décollement above TOC in this region.

Chapter 4

**STRUCTURE AND PROPERTIES OF THE CASCADIA
PLATE INTERFACE: EVIDENCE FROM A NEWLY
DESCRIBED EXHUMED PALEOMEGATHRUST IN THE
OLYMPIC SUBDUCTION COMPLEX**

Manuscript published in *GSA Bulletin* at doi.org/10.1130/B38335.1.

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4.1 Abstract

The Olympic Subduction Complex (OSC) in the central Olympic Mountains is a deeply exhumed continuation of the offshore modern accretionary wedge of the Cascadia subduction zone. Metamorphic grade and thermochronology reveal that the OSC's central core likely accreted by underplating at seismogenic depths. We characterize a previously unknown ~450-m-wide belt of block-in-matrix mélange containing an anastomosing system of fault strands, which in turn include discrete principal slip surfaces consistent with brittle-frictional, likely coseismic, deformation. The lithologies are deep marine turbiditic sandstones and mudstones without other elements of ocean plate stratigraphy. Raman spectroscopy of carbonaceous material refines peak paleotemperatures to 260–280°C, consistent with exhumation from seismogenic zone conditions. We interpret these structures as a composite fault zone that records both fast slip and slow distributed deformation during the seismic cycle through coeval coseismic brittle-frictional and interseismic viscous deformation. The mélange formed by

granular flow, pressure solution, alignment and growth of phyllosilicates, and crystal plastic deformation, while the fault strands are dominated by concentrated cataclasis, brecciation, and veining. We interpret this fault zone as an exhumed paleomegathrust interface, the first direct analog for the modern Cascadia subduction zone. The absence of basalt indicates that the megathrust fault was localized within the incoming plate stratigraphic sequence, facilitating sediment underthrusting, similar to offshore structures observed via seismic reflection imaging in Cascadia and elsewhere today. This exposure presents an exhumed model for plate interface systems hosted entirely in sedimentary rock even at mid-seismogenic depths. It is therefore a revealing potential analog useful in constructing models of seismogenic zone deformation and rupture processes.

4.2 Introduction

Subduction zones are responsible for the world's largest earthquakes, which in turn produce the most devastating tsunamis. Slip along the plate boundary décollement or megathrust fault that forms the base of the accretionary wedge is the source of these destructive events. While geophysical tools and modeling have allowed us to probe active accretionary wedges, direct evidence of the conditions that promote coseismic slip can only be found by examining the plate boundary itself, either through ocean drilling in modern settings or via the study of exhumed fossil subduction zone terranes. Ocean drilling has resulted in significant advances in study of the shallowest plate interfaces (e.g., Moore, 1989; Moore et al., 1995; Tobin and Saffer, 2009; Ujiie et al., 2013; Vannucchi and Tobin, 2000). However, to understand geologic processes and conditions at seismogenic zone depths and deeper, we must turn to exhumed subduction terranes (e.g., Behr and Bürgmann, 2021; Hilley et al., 2022; Kirkpatrick et al., 2021; Rowe et al., 2013).

Given the utility of these terranes, known exhumed paleomegathrust faults are rare, especially those examples that record maximum paleotemperatures between

150°C and 350°C and thus seismogenic zone processes. Of these, most are not directly continuous with present-day subduction, having undergone a major change in tectonic regime since underplating. Most work has focused on a few well-studied examples, such as the Kodiak Accretionary Complex, Alaska, USA; Franciscan Complex, California, USA; the Northern Apennines, Italy; and Shimanto Belt, Japan (e.g., Fisher et al., 2021a; Meneghini et al., 2009; Moore et al., 2007b; Rowe et al., 2013; Ujiie and Kimura, 2014; Vannucchi and Bettelli, 2010). Globally, studies of such exhumed paleomegathrust fault zones have revealed key structural features representative of seismogenic zone deformation. They are characterized by several anastomosing fault strands within a belt of deformed *mélange* comprising total fault thickness up to 500 m, as synthesized in Rowe et al. (2013). Each fault strand in turn contains multiple discrete faults, which host centimeters-thick principal slip surfaces. Many fault strands are contemporaneously active in a fault network, which together represent the plate interface (Rowe et al., 2013). The surrounding tectonic *mélange* is characterized by scaly fabric, consistent foliations and phacoid block orientations, and tabular outcrop geometry (e.g., Festa et al., 2019; Fisher et al., 2021a). The full area of disrupted strata may be up to kilometers in thickness, but does not represent simultaneously active faulting (Rowe et al., 2013).

In this study, we apply geological criteria for identifying paleomegathrust exposures (e.g., Festa et al., 2019; Fisher et al., 2021a; Kirkpatrick et al., 2021) to a locality in the Olympic Subduction Complex (OSC), in the Olympic Mountains, Washington, USA, a deeply exhumed part of the modern Cascadia accretionary prism. We document a well-exposed fault zone exhibiting multiple fault strands anastomosing through a belt of rock with block-in-matrix fabric, which we define as *mélange* (*sensu* Wakita, 2015, Fig. 1 herein). We report new field observations, structural orientations, and peak metamorphic temperatures that together indicate it is likely an exhumed seismogenic zone paleomegathrust fault. We show that localized, discrete brittle structures and non-localized viscous structures coexist and formed coevally, likely through a

combination of coseismic and interseismic processes. This work sheds light on seismic cycle processes at sediment-hosted plate interfaces, with implications for how deformation is accommodated at the modern Cascadia megathrust.

4.3 Geological Setting

Exhumation and exposure of the deep interior and paleomegathrust faults of an accretionary wedge system is relatively rare globally, and most of these systems are exposed because of a major change in tectonic regime such as collision (e.g., Helvetic nappes in the European Alps; Schmid et al., 1996), ridge subduction (e.g., Alaska Chugach terrane; Bradley et al., 2003), or triple junction migration (e.g., Shimanto Belt, Japan; Kimura et al., 2014), rendering the comparison of the exposed rocks to the present-day subduction megathrust indirect.

At Cascadia, by contrast, continuous subduction since the Oligocene has built a large, clastic sediment-dominated accretionary wedge with one of the thickest incoming sediment sections of any subduction zone globally (Heuret et al., 2012). This large wedge underlies the entire continental slope and shelf offshore from Vancouver Island to northern California, but its interior is subaerially exposed onshore only in the Olympic Peninsula of Washington (Fig. 1A; Tabor and Cady, 1978a; Snively, 1987), where it is called the Olympic Subduction Complex (Fig. 1B; OSC of Brandon and Vance, 1992). The offshore accretionary wedge and onshore OSC are structurally continuous (Brandon and Calderwood, 1990; Brandon et al., 1992; Tabor and Cady, 1978b), an interpretation that is bolstered by geophysical evidence that the central Olympic Mountains are underlain by ~ 30 km of low-density and low-seismic velocity, lightly metamorphosed sedimentary rock (Calvert et al., 2011; Merrill et al., 2020; Delph et al., 2021).

Since the 1970s, the outcrop area of the OSC has been understood to be made up entirely of deformed terrigenous clastic turbidites (Rau, 2003; Stewart, 1970; Tabor and Cady, 1978a,b), and interpreted as a suite of frontally offscraped and/or basally

underplated sedimentary packages (Batt et al., 2001; Brandon et al., 1998; Brandon and Calderwood, 1990; Brandon et al., 1992; Clowes et al., 1987; Michel-Wolf et al., 2022). However, it is largely mapped as an internally undifferentiated “structural complex” (see Gerstel and Lingley, 2003). Brandon et al. (1992) first proposed that the OSC be divided into three units—the Coastal, Lower, and Upper OSC—where they hypothesized that the Lower OSC has been exhumed from greater depth and exposed currently in a tectonic window with the Upper OSC overthrust above it. The Lower OSC was deposited on the Juan de Fuca plate at ca. 20 Ma (Brandon et al., 1992; Stewart and Brandon, 2004) and subducted to $\sim 12\text{--}15$ km depth. Based on zircon fission-track results from samples near Mount Tom (~ 2 km from our field site and within the correlative along-strike map unit of Tabor and Cady, 1978a), these rocks then began continuously exhuming beginning ca. 18 Ma (Brandon et al., 1992, 1998). Tabor and Cady (1978a) conducted the only previous geological work west of Mount Olympus in this unit, mapping a major thrust and a “penetratively sheared” unit with partial to complete disruption of original bedding, in the hanging wall of this mapped fault boundary (Figs. 1B and 1C).

The highest topography in the Olympic Mountains corresponds to the most deeply and rapidly exhumed rocks (Brandon et al., 1992; Michel et al., 2018; Michel-Wolf et al., 2022). The region of the Olympic Peninsula is a locale of relatively flat-slab subduction compared to areas both north and south and is undergoing oroclinal bending along with the bend in the Juan de Fuca slab below (Crosson and Owens, 1987; Finley et al., 2019), where repeated underplating has created a bullseye-like uplift and focused exhumation pattern (Brandon et al., 1998; Michel et al., 2018). The net result is that as the modern accretionary wedge grows, subduction underplating transfers previously subducted sedimentary material into the wedge at depths consistent with the location of the deeper part of the locked seismogenic zone in this region of Cascadia, bringing underplated rocks to the surface at geologically rapid rates of ~ 1 km/m.y. (Brandon et al., 1998, 1992; Michel et al., 2018; Pazzaglia and Brandon, 2001), fa-

avorable for preservation of structures formed at or near peak pressure-temperature in the seismogenic zone.

If the Lower OSC is indeed composed of duplex slices that were subducted and underplated, we hypothesized it could contain evidence of structural packages separated by discrete faults representing exhumed examples of major accretionary wedge faults or even the paleomegathrust interface. We have conducted fieldwork in two areas of the Lower OSC near the mapped thrust of Tabor and Cady (1978a) to test this hypothesis (Fig. 1C)—first at Mount Olympus, Panic Peak, and numerous glacially isolated outcrops located on Snow Dome in the unit Twog; then at our primary field area, known as Lakes of the Gods (LOTG), which targets a region close to the mapped fault and in the heart of the zone identified as exhibiting high penetrative shearing in the Tabor and Cady (1978a) unit Twos (Fig. 1C).

4.4 Methods

4.4.1 Structural and Geochemical Analysis

We sought to characterize the geology at these sites from the macro- to microstructural scale. In the field, we made first-order observations, took structural measurements, and performed the geological mapping using StraboField. Using photographs, we created structure-from-motion (SfM) orthomosaics of key outcrops in Agisoft on which we plotted densely spaced measurements to visualize structural relationships. Directional data measured in the field were analyzed using the geologyGeometry library in R (<http://www.joshuadavis.us/software/>; accessed August 2025). Bootstrapping, which produces a 95% confidence interval for the mean of each population, was performed to derive the average and variance of each population of orientations (as applied in Davis and Titus, 2017; Roberts et al., 2019).

We carried out a comprehensive microstructural and microchemical analysis of composition, grain size, and fabric across the range of strain regimes we sampled,

using optical petrography and scanning electron microscope imaging. Energy dispersive X-ray spectroscopy mapping was performed on a ThermoFisher Scientific Apreo variable pressure scanning electron microscope (SEM) at the University of Washington Molecular Analysis Facility, Seattle, Washington, USA. SEM images shown are from the back-scattered electron (BSE) detector at a chamber pressure of 10^{-3} Pa or lower, voltage of 10 kV, and current of 0.8 nA.

4.4.2 *RSCM Thermometry*

We also performed Raman spectroscopy of carbonaceous material (RSCM) to derive maximum metamorphic temperatures for a suite of samples from Snow Dome and LOTG representing a range of lithologies and deformation levels. RSCM measurements were taken on polished thin sections using a Renishaw inVia Qontor confocal Raman microscope in the SciTech facility at Western Washington University, Bellingham, Washington, USA. The analysis was conducted using a 532 nm laser, a $100\times$ objective with a 0.85 numerical aperture, and a grating of 1800 grooves/mm. Spectra were acquired with up to two accumulations of 20 s each using a laser power of 0.5–2 mW, and the background was measured from 400 cm^{-1} to 2200 cm^{-1} . In each sample, we measured 10–15 particles of carbonaceous material that were hosted beneath the surface of transparent mineral phases such as quartz, chlorite, and mica to avoid alteration that may have been caused by sample preparation.

Sample temperatures were estimated using both the calibrations of Kouketsu et al. (2014) and Lünsdorf et al. (2017). The Kouketsu et al. (2014) calibration utilizes the full width at half maximum value of disordered carbon peaks (D1 and D2). Peak-fitting of Raman spectra were performed with the automated routine of Kaneki and Kouketsu (2022), which requires the operator to choose a fitting routine based on the appearance of the spectra before calculation, potentially introducing bias into the results. We applied both Fit F ($\sim 280^\circ\text{C}$) and Fit G ($\sim 300^\circ\text{C}$) to the D1 peaks but found that Fit F produced smaller residuals, so report only that here. The Lünsdorf

et al. (2017) calibration correlates the scaled total area of disordered carbon peaks with temperature and provides a fully automated curve fitting approach. For each sample, we report the mean temperature calculated from the range of temperatures from individually fit spectra in the sample, the within sample heterogeneity as one standard deviation of temperature estimates, and the absolute uncertainty on each mean temperature ($\pm 27^\circ\text{C}$ for the Kouketsu et al., 2014, calibration and $\pm 36^\circ$ for Lünsdorf et al., 2017, calibration).

4.5 Results

4.5.1 Snow Dome

We conducted reconnaissance level fieldwork at Snow Dome, in the Twog unit, which is composed of greenish-gray and black sandstone and semischist in ~50% slate and phyllite as defined by Tabor and Cady (1978a, Fig. 1C herein). There, we identified turbiditic sandstone and mudstone with folded and faulted beds and cut-off tight to isoclinal fold hinges (Fig. 2A). The metapelites have a spaced foliation of slaty parting surfaces and host dismembered sandstone beds that are displaced along discontinuous shear surfaces (Fig. 2B), whereas there is little fabric development in the more competent coarse sandstones, which exhibit intact primary stratigraphic features and bedding such that younging direction is commonly apparent. We identify these rocks as broken formation. Original bedding is pervasively disrupted at the outcrop scale, but we found no block-in-matrix fabric in the Snow Dome or Mount Olympus area. Previous studies by Tabor and Cady (1978a,a) and summarized in Brandon et al. (1998) show that most of the much wider Lower OSC outcrop area (Fig. 1B) is characterized by this level of broken formation disruption of stratigraphy, while block-in-matrix fabrics are rare (Tabor and Cady, 1978b). The Snow Dome area lies ~2 km east of the mapped fault and belt of pervasive shearing on the Tabor and Cady (1978a) map.

4.5.2 *Lakes of the Gods*

We targeted the LOTG site within the Twos unit, described by Tabor and Cady (1978a) as slate and phyllite with ~30% small phacoids of foliated sandstone and semischist. Notably, this unit is mapped as having a high level of “penetrative shearing” compared to other units in the Eastern Olympic Core, as demonstrated by the density of wavy lines in Figure 1C (Tabor and Cady, 1978a). This pattern is described in Tabor and Cady (1978a) as representing “phacoids of foliated sandstone, siltstone, sheared basalt, and fragmental vein quartz in a matrix of slate and phyllite.” At the LOTG site within this unit, we discovered and mapped a previously undescribed, steeply dipping, pervasively sheared fault zone of ~450 m of true structural thickness. Exposure of the bedrock geology approaches 100% across much of the field site, except at the lakes. Throughout this paper, we apply the terminology of Rowe et al. (2013), where the entire deformed area is referred to as the fault zone, which contains meters-wide fault strands, which in turn host discrete faults.

4.5.3 *Lithologies*

The rocks at LOTG, originally deepwater clastic sediments, are composed of three main units: T1, light tan weathering to orange sandstone with ~40% mudstone; T2, black weathering to rusty mudstone with ~40% sandstone; and T3, a coarse, white sandstone, with interbedded siltstone and coarse-grained conglomerates (Fig. 3). Outside of the fault zone to the northeast, T1, T2, and T3 are interbedded at 10 m scale (Fig. 3A). We use the term “turbidite” to refer to interbedded sequences of sandstone and mudstone after Tabor and Cady (1978b) and because characteristic features of turbidites such as graded bedding; sharp, sandy basal contacts; and cross-bedding are evident where deformation fabric is only weakly developed (Fig. 3B). The rocks typically have a phyllosilicate-dominated pelitic matrix, which contains quartz and feldspar-rich sandstone layers and, where disrupted, sandstone lenses. Mixed-

mode calcite and quartz veins are abundant both parallel to and across foliation and show displacement along shear surfaces.

4.5.4 *Fabric*

We characterize the development of structural fabric, or degree of bedding disruption, at the outcrop scale, increasing from Level 0 (L0) to Level 3 (L3). We define L0 as broken formation where outcrop-scale faulted, and more rarely folded, beds are laterally continuous at 10 m scale; original bedding and sedimentary structures are preserved (Figs. 3A and 3B). Dismembered beds and a spaced foliation define L1 (Fig. 3C). L2-L3 have a block-in-matrix fabric, which we term *mélange* (sense of Wakita, 2015). At L2, sandstone beds are broken into centimeter-scale blocks, which we term phacoids, by polished slip surfaces, creating scaly fabric. The most developed fabric is at L3, where sub-centimeter to centimeter-scale phacoids are contained in a fine to ultrafine-grained, penetratively foliated pelitic matrix (Figs. 3D–3F).

Mélange outcrops typically display “pods” with varying levels of fabric development separated by shear surfaces as evidenced by truncated fabric at these boundaries (Fig. 3C). Such shear surfaces commonly host quartz and calcite shear veins (Fig. 3C). Throughout the field area, fabric is more well-developed in finer-grained T2, particularly in the T2-dominated area to the southeast of the Upper Lake. T3 is coarse-grained and thus did not develop fabric, instead outcropping as massive blocks surrounded by *mélange*. Where present, crenulation cleavage overprints scaly fabric in both T1 and T2. Crenulation cleavage is most common in the southeast portion of the fault zone along strike.

4.5.5 *Microstructures*

In rocks at L0 to L1, we observe preserved sedimentary features at the microscopic scale (Figs. 4A and 4B). In the finer-grained mudstones, pencil cleavage—an intersection of bedding and an orthogonal cleavage—is common. Domains with different

grain sizes grade into each other with indistinct boundaries (Fig. 4B). Preserved rip-up clasts (Fig. 4A) and bioclasts (Fig. 4B) are visible. The mudstone matrix contains fine-grained phyllosilicates ($\sim 50 \mu\text{m}$ length; Fig. 4C). In SEM BSE, many of these phyllosilicates are at a high angle to bedding, consistent with sedimentary deposition and later flattening (Fig. 4D). Parallel fractures highlight the bedding at the microscale and show the lack of scaly, anastomosing fabric in these samples (Figs. 4A, 4B, and 4D). In these fractures under BSE, dense (i.e., bright) features are visible, which indicate pressure solution has occurred, resulting in the depletion of soluble and concentration of insoluble elements (see e.g., Chen et al., 2024a). In our samples, fractures are common; although these may be due to sample preparation, they exploit the structural fabric (e.g., cracks showing pencil cleavage in Fig. 4A) and/or contain vein fill or insoluble material, indicating their original presence (Figs. 4B and 4D).

In the *mélange* samples, however, granular flow, cataclasis, foliation development, and pressure solution are the most common deformation mechanisms. Rocks classified as *mélange* display scaly fabric, which is a composite fabric commonly interpreted to form by pressure solution, fracturing, slip along shear planes, and growth of new phyllosilicates (Figs. 4E–4L; Schleicher et al., 2010, 2012; Vannucchi, 2019). In our samples, these scaly fabric seams are commonly dark, anastomosing boundaries in the mudstone matrix, which commonly localize fracturing during sample preparation (e.g., Figs. 4E, 4F, 4H, 4I, and 4L). In SEM BSE, these scaly fabrics again appear bright due to the concentration of insoluble elements left by pressure solution (Figs. 4H and 4L); however, they are more anastomosing in the *mélange* rather than bedding-parallel in the broken formation (Fig. 4D). Phyllosilicates are elongated ($\sim 100\text{s } \mu\text{m}$ in length; Fig. 4K) and highly aligned, indicating growth of these minerals during scaly fabric and foliation development; domains rich in phyllosilicates are separated by dark, ultrafinegrained shear surfaces (Fig. 4K).

Sandstone phacoids contained within this matrix have web structures formed by

initial boudinage and cataclasis (Fig. 4E; Byrne and Fisher, 1987; Fisher and Byrne, 1987; Kimura et al., 2012; Kitamura and Kimura, 2012) and extension veins (Fig. 4F), which are truncated at their boundaries. T2 rocks tend to contain fewer and smaller sandstone phacoids and are dominated instead by the mudstone matrix (Fig. 4I). They have been elongated both by granular flow (Figs. 4E and 4F) and by crystallization of thick veins (Fig. 4I). Quartz in veins shows bulging-related textures, indicating the onset of quartz recrystallization (Fig. 4G). Crenulation cleavage, where present, overprints an earlier foliation (Fig. 4J).

4.5.6 *Fault strands*

The anastomosing nature of the fault strands is highlighted in an SfM orthomosaic that spans the widest fault strand at LOTG (Fig. 5A). The fault strand is ~ 10 m wide and contains blocks of *mélange* cut by prominent tensile fractures at $\sim 90^\circ$ to the foliation (in 3-D). Discrete faults composed of densely foliated cataclasite anastomose around the deformed blocks within the fault strand. Each discrete fault hosts one or multiple 1 mm to 10 cm wide principal slip surfaces within it. The principal slip surface shown in the SfM orthomosaic (Fig. 5A) is localized where the weaker fault strand meets more competent T1 L2 wall rock (Fig. 5B). We define the fault rocks that we observe as a separate lithological unit to highlight their presence in the gullies (Fig. 2A).

The deformed rocks exposed within fault strands are most commonly *mélange* blocks that host later brittle deformation. The fractured outcrop in Figures 5C and 6A has macroscopically and microscopically scaly fabric with superimposed throughgoing tensile fractures (Figs. 6A and 6B). These fractures contain stilpnomelane with basal planes parallel to the fracture edges (Fig. 6C). The principal slip surfaces surrounding such blocks are composed of a variety of fault rocks. Most commonly, they are unfoliated cataclasites, which contain quartz veins and chunks of brecciated wall rock sometimes encased in a soft clayey matrix (Figs. 6D–6F). Dark shear surfaces at

the microscale (Fig. 6E) form by localized cataclasis and particulate flow and are consistent with expected fault motion out of the plane of observation.

Other discrete faults preserve brecciated wall rock hosted in a quartz and calcite matrix (Figs. 7A–7C). Clasts of the wall rock display strong phyllosilicate foliation and scaly fabric (Fig. 7C). These rocks commonly show two or more stages of brecciation, with clasts of brecciated wall rock and veins reworked as new breccias (Fig. 7B). The quartz in the veins surrounding the breccia preserves bulging grain boundaries and other forms of crystal plastic deformation, such as prominent deformation lamellae (Fig. 7C). Following the last stage of brecciation, a chlorite rim grew simultaneously with the quartz vein and surrounded the clasts (Figs. 7B and 7C). In a different fault strand, the principal slip surface is paralleled by an ~15-cm-thick quartz and calcite vein separated by thin, dark, insoluble selvages (Fig. 7D). These seams are phyllosilicate-rich bands of wall rock separating dynamically recrystallized quartz with bulging to incipient subgrain development (Figs. 7E and 7F).

In a fault strand in the northeastern part of the study area that separates the majority of the fault zone from broken formation rock to its northeast, shear localized to an ~30-cm-thick discrete fault containing highly foliated rock with a dark, ultrafine matrix (Fig. 8A). Microscopically, this rock has a highly flattened fabric compared to lithologically similar, nearby mudstones that are outside of the fault, with aligned phyllosilicates creating a smooth, continuous foliation (Fig. 8B). Porphyroclasts of pyrrhotite up to 200 μm wide with strain shadows of quartz are common throughout the sample (Figs. 8B and 8C). This sample has no cataclasis, brecciation, veining, or otherwise microscopically brittle features.

4.5.7 LOTG Fault Zone Macrostructure

At the map scale, tabular outcrops surrounded by numerous anastomosing fault strands several meters wide comprise the scaly fabric of the fault zone (Fig. 9A). These fault strands, which are topographically expressed as distinctive linear gullies

(Fig. 9E), in turn contain localized bands of fault gouge and discrete faults exhibiting numerous apparent individual principal slip surfaces. We mapped at least nine major, continuous fault strands identified on the basis of exposed brittle fault rock at each location, and many additional, small, topographic gullies with little exposure, which we infer are minor fault strands. Lateral “relay” gullies connect the main gully faults and highlight the anastomosing block-in-matrix fabric of the fault zone at the map scale.

To the southeast of the Upper Lake, we map a concentration of L2-L3 T2, which we refer to as the apparently most deformed center of the fault zone (Fig. 9A). Fabric development decreases toward the northeast up to the last fault strand. Beyond this, original sedimentary features such as bedding are preserved and block-in-matrix fabric is absent, such that we classify the rocks as broken formation, or a coherent unit. We interpret this transition as the preserved upper boundary or hanging wall of the fault zone. On the southwest side of the fault zone, block-in-matrix fabric persists beyond the mappable area, so we were not able to identify the fault zone base.

The bedding in the hanging wall broken formation and the foliation and bedding measured in the *mélange* outcrops, compared to the foliations in the fault strands, are similar in orientation—with mean strike and dip measurements of $314^{\circ}/75^{\circ}\text{NE}$ and $320^{\circ}/71^{\circ}\text{NE}$, respectively (Figs. 9B and 9C). However, the variance of orientation in the *mélange* and broken formation is 0.11 compared to 0.05 in the fault strands. Phacoid long axes measured throughout the field area were found to plunge steeply to the east—with an average long axis trend of 100° and plunge of 71° . While they lie within the foliation planes, the phacoids themselves are commonly asymmetric and elongate in the dip direction, indicating top-to-SW tectonic transport by the fault strands and corresponding distributed deformation within the *mélange* itself (Fig. 9D). We measured Y-P-R relations in the field, which confirm the top-to-SW sense of shear (e.g., Cowan and Brandon, 1994).

4.5.8 *Paleotemperature from Raman Spectroscopy of Carbonaceous Materials*

Using the Raman spectroscopy of carbonaceous materials (RSCM) calibrations of Kouketsu et al. (2014) and Lünsdorf et al. (2017), we estimate that the range of sample paleotemperatures at Snow Dome is 244–278°C and 267 – 281°C, and at LOTG the range is 228–290°C and 262–282°C, respectively (Fig. 2F; Table S1). There is no temperature trend across the field area and no observable difference in temperature from across the Tabor and Cady (1978a) fault in unit Two, the LOTG fault zone or the hanging wall broken formation in unit Twos, or the Snow Dome area in unit Twog Tabor and Cady (Fig. 1C for units from 1978a)

Errors on the RSCM method are cited as $\pm 50^\circ\text{C}$ (Beyssac et al., 2002), but are reduced in the individual calibrations used here to $\pm 27^\circ\text{C}$ for Kouketsu et al. (2014) and $\pm 36^\circ\text{C}$ for Lünsdorf et al. (2017). On Figure 2F, we report the standard deviation for the temperatures derived from the Lünsdorf et al. (2017) calibration and the error given by Kaneki and Kouketsu (2022), which accounts for both calibration and measurement uncertainty.

4.6 *Discussion*

4.6.1 *A Modern Look at the Olympic Subduction Complex*

New Temperature Constraints

Previous workers have estimated maximum paleotemperatures for the Lower OSC. Based on optical microscopy, Tabor and Cady (1978b) classified these rocks as prehnite-pumpellyite facies, indicating a temperature range of 250 – 350°C broadly. Brandon et al. (1992) applied fission track to detrital zircons to determine that the OSC underplated ca. 17 Ma at ~ 12 –15 km depth and 240 – 245°C. Further analysis of maximum burial and the geothermal gradient by Brandon et al. (1998) estimated a maximum paleotemperature of 242°C for rocks near the summit of Mount Olympus. Rahl et al. (2005) were the first to use RSCM on rocks from the Olympics and measured a limited

sample set of 11 from the entire peninsula. They found temperatures ranging between 205°C and 295°C for samples in the Lower OSC, with the 205°C measurement taken near the peak of Mount Olympus, ~2 km from our Snow Dome site.

Our RSCM work reveals paleotemperatures for an additional 18 samples from LOTG and four samples from Snow Dome. Using updated calibrations and sample preparation methods to reduce error in our measurements, our samples largely record temperatures between 260°C and 280°C (Fig. 2F; Table S1), slightly higher than previously reported for the Olympic Mountains but consistent with the observed quartz microstructures. Seismogenic zone temperatures are commonly defined as ranging between 150°C and 350°C (e.g., Hyndman and Wang, 1993); thus, our results indicate that LOTG underplated after experiencing seismogenic zone conditions. Our use of updated calibrations may explain the difference in our temperatures from the previous estimates; for example, Lünsdorf et al. (2017) showed that the calibration of Rahl et al. (2005) underestimated temperatures with respect to the ones applied here, especially at temperatures <400°C. Additionally, none of the studies since Tabor and Cady (1978b) used samples taken from west of Mount Olympus or from LOTG.

Contrary to previous work by Aoya et al. (2010) and Moris-Muttoni et al. (2023), our preliminary findings indicate that the RSCM values are not influenced by high strain. Spectra from both little deformed samples and highly deformed cataclasites and fault gouges are similar and have overlapping error bars (Fig. 2F). Aoya et al. (2010) note that their increased values for highly strained rocks are within the $\pm 50^\circ\text{C}$ error range of their applied calibrations, so it is possible any signature of higher temperatures due to strain in our samples are concealed by uncertainty.

LOTG Structure is Unusual in the Olympic Subduction Complex

At Snow Dome in the Twog unit (Fig. 1C), while the rocks ubiquitously display fabric consistent with broken formation, no block-in-matrix fabric was observed. Small faults and folds are abundant, but the primary sedimentary bedding and small-scale

sedimentary structure was commonly present (Fig. 2). Based on descriptions in Tabor and Cady (1978b) and Brandon et al. (1992), the majority of outcrop area of both the Lower and Upper OSC is similar to this region. Pressure-solution foliations are ubiquitous, including intersecting ones forming pencil cleavage, but penetrative scaly shear fabrics are rare to absent from Mount Olympus to the east.

In contrast, the LOTG exposure is the northwesternmost locality for the Tabor and Cady (1978a) map unit “Twos,” which they described as slate and phyllite with “penetrative shearing” with ~30% small phacoids of foliated sandstone and semischist. This lithostructural unit extends more than 30 km farther to the southeast along strike of the fault shown in Figure 1C and is mapped as sheared throughout its length, which is unique in the entire Olympic Core. They did not describe it any further, but Tabor (1975) refers to it as an extensive zone of “breaking and crushing.” We have accessed its northwesternmost mapped location at LOTG and have documented a well-developed fault zone hosted in tectonic *mélange*. We therefore conclude that LOTG is likely one locality in a much more extensive fault zone that may represent the main deeply underplated package exposed in the OSC. While we visited the location of the mapped fault in Tabor and Cady (1978a), which is the western boundary of the Twos unit, we did not find evidence of a specific localized fault structure at the map position. *Mélange* was absent at and to a short distance west of the trace; we suspect that Tabor and Cady (1978a) placed the fault approximately, just west of the belt of penetratively disrupted rocks.

While it has been established since at least the 1990s that the Lower OSC is a deeply exhumed portion of the underplated sedimentary accretionary wedge, none of the previous work differentiates underplated structural packages within it, nor does it identify structural units associated with the tectonic mechanisms of that process. The evidence presented here supports the hypothesis that a portion of the former plate boundary fault zone has been incorporated as the LOTG fault zone, most likely through underplating of duplexes from the subducting sediment section.

4.6.2 LOTG as Tectonic Mélange

Block-in-matrix fabric can form through sedimentary, tectonic, or diapiric processes; for example, sedimentary mélanges can result from mass transport in accretionary wedges (e.g., Cowan, 1985). The synthesis by Festa et al. (2019) defined diagnostic features of these contrasting types of mélange. Sedimentary mélanges are characterized by a highly disordered, isotropic fabric, where there is a random distribution of clasts within the matrix, except close to the basal surface. Clasts can be angular or rounded but are equidimensional. Meanwhile, tectonic mélanges have block-in-matrix fabric that is structurally ordered and consistent with the regional stress field (Festa et al., 2019). These units must contain at least one major fault and record both ductile and brittle deformation processes, forming a scaly fabric. Blocks in the matrix are lenticular or phacoidal in shape.

At LOTG, we document an ~450-m-wide fault zone with at least nine major fault strands contained in a body of mélange. The foliations within the mélange and those in the faults are extremely consistent in orientation (Figs. 9B and 9C). Additionally, in the mélange zones between fault strands, foliation becomes increasingly coplanar at their margins and within the bounding fault strands. Fault strands thus have a foliation variance half of that in the mélange, implying a more planar fabric and a greater amount of shear localization in the former. Phacoids are lenticular in shape; their long axes are consistently oriented parallel to the apparent transport direction both in the fault strands and in the mélange zones (Fig. 2D), which indicates a consistent orientation of non-coaxial stretching strain. All this constitutes a mélange fabric that is consistent with the regional stress field—as inferred from the regional foliations—and suggests a genetic relationship between the fault strands and the mélange. Microscopically, mélange samples also have scaly fabric, which is formed by a combination of brittle (cataclasis, granular flow, veining) and ductile (pressure solution, phyllosilicate growth) processes. All these demonstrate definitively that LOTG is a

mélange zone formed overwhelmingly by tectonic, not sedimentary, processes.

4.6.3 Evidence of a Paleomegathrust Interface

At LOTG, we have discovered and mapped a previously undescribed, steeply dipping, pervasively sheared, fault zone. We have demonstrated that this mélange is tectonic in origin and is genetically related to the fault strands. We observe pressure-solution dominated flattening and foliation development within the mélange that is coplanar with fault strands and parallel to bedding in the hanging wall broken formation (coherent unit). This is consistent with deformation that occurred on a low-angle fault system sub-parallel to bedding. The presence of a well-developed high-strain fault zone of this geometry at LOTG, which reached conditions of 260°C to 280°C (i.e., the seismogenic zone; Fig. 9F) and ~12–15 km depth, indicates that these rocks represent an underplated system rather than frontal accretion at shallower depths, affirming conclusions of earlier work (e.g., Brandon et al., 1998). This interpretation is consistent with originally low-angle layer-parallel shear in a plate boundary fault (e.g., Fisher et al., 2021a; Fisher and Byrne, 1987; Ujiie, 2002), later rotated during exhumation to its present high-angle position (e.g., Kitajima and Saffer, 2012), rather than an intra-wedge fault where bedding, fault strands, and mélange foliations would be much less likely to maintain parallelism through a protracted deformational history. Block-in-matrix fabric can also develop in frontally accreted units, which never experienced seismogenic conditions, due to, for example, competency contrasts between lithologies Bettelli et al. (2003); Remitti et al. (2024); however, both the geometrical argument and the new RSCM temperatures that we present here, in addition to reasoning presented in earlier work Brandon et al. (1998, 1992), exclude this possibility. Furthermore, the uniform (within uncertainty) recorded paleotemperature across the LOTG to Snow Dome study, including outside the fault zone, suggests that it does not represent a major outof-sequence thrust that would juxtapose units of large temperature contrast. We also interpret at least some of the fault strands

as hosting simultaneously active slip, based on the anastomosing macrostructure of the fault zone, where no fault strands exhibit crosscutting relationships with others despite being interlinked (Figs. 3, 4, and 9A). While the foliation is at a high angle in the present day, this is not uncommon in other subduction complexes globally; for example, it is well-documented in the Shimanto Belt (Kitamura et al., 2005).

4.6.4 *The Structure of the Plate Interface*

The geology of exhumed subduction complexes worldwide has contributed to the understanding of subduction megathrust fault structure as a fault zone composed of a braided network of multiple, simultaneously active fault strands hosted in tectonic *mélange*, rather than being a simple interface (Behr and Bürgmann, 2021; Kirkpatrick et al., 2021; Remitti et al., 2007, 2011; Vannucchi et al., 2008). Individual fault strands are $\sim 3\text{--}10$ m in thickness, each hosting thin, centimeter-scale principal slip surfaces, consistent with these proposed models of plate interfaces (e.g., Rowe et al., 2013). Geophysical observations of active systems are beginning to resolve a similar structure. For example, active source seismic reflection data from subduction zones have shown that décollements can commonly have multiple active levels (e.g., Adam et al., 2004), create duplexes to facilitate underplating (Kimura et al., 2010), and display abandoned strands taken over by new ones (Kirkpatrick et al., 2020). Furthermore, an analogous multifault network at the Ecuador megathrust interface has recently been observed a similar scale using high-precision relative location of aftershocks (Chalumeau et al., 2024). Our observations at LOTG add an excellent comparative example of the structure of the plate interface at seismogenic depths, highlighting the fact that multiple, simultaneously active and highly heterogeneous fault strands can host coseismic slip contemporaneously.

In sediment-rich subduction zones, the décollement is commonly localized entirely within the sediments, as observed in portions of Cascadia today (e.g., Carbotte et al., 2024), as well as Nankai, Barbados, and others, which would create the turbidite-

type ocean plate stratigraphy mélange (Wakita, 2015) as observed at LOTG. The absence of basalt at LOTG implies that the underthrusting turbidite section during this time was thick, as was argued for the Kodiak Formation (Fisher and Byrne, 1987). Previous work, in fact, suggests the sediment-basalt mélanges observed at many other subduction complexes, such as Shimanto or the Ghost Rocks Formation at Kodiak, may in some cases have formed partly as a result of pre-subduction basalt intrusion during ridge subduction rather than through subduction-related imbrication of the downgoing plate (Fisher and Byrne, 1987; Meneghini et al., 2009; Rajič et al., 2024). Thus, despite the absence of other elements of ocean plate stratigraphy, which has been included as a diagnostic feature based on some other exhumed examples (e.g., Mugi mélange and Franciscan in Ujiie and Kimura, 2014; Regalla et al., 2018, respectively), we interpret the LOTG fault zone as an exhumed paleomegathrust interface, which represents the first example of its kind from the modern Cascadia subduction zone.

4.6.5 Geologically Recorded Seismogenic Zone Slip Behaviors

Given our constraints that LOTG represents an exhumed paleomegathrust exposure, it is an ideal natural analog for understanding seismogenic zone processes in this tectonic setting. This includes alternating modes of deformation during the seismic cycle. Our samples record prograde metamorphism and deformation with subduction, along with repeated cycles of distributed ductile, likely aseismic deformation alternating with localized brittle, potentially coseismic deformation (Fig. 10).

Interseismic Deformation

At the microscopic scale, we observe a sequence of deformation that facilitates the development of mélange from broken formation, which is consistent with microstructural observations from other subduction complexes. In the broken formation rocks, primary sedimentary features such as bedding (Figs. 4A, 4B, and 4D), rip-up and bio-

clasts (Figs. 4A and 4B), and fine-grained, unaligned phyllosilicates (Figs. 4C and 4D) are present. There is a progressive overprinting of these primary sedimentary features by tectonically produced scaly fabric (e.g., Fig. 4D) and later crenulation cleavage (Fig. 4J). Block-in-matrix fabric forms through the boudinage and cataclastic breakdown of sandstone beds, which preserve web structures and veins within phacoids (Figs. 4E and 4F Byrne and Fisher, 1987; Fisher and Byrne, 1987; Kimura et al., 2012; Kitamura and Kimura, 2012). Sandstone phacoids then underwent granular flow and veining, leading to elongation (Figs. 4E, 4F, and 4I). Scaly fabric in *mélange* forms by distributed, aseismic creep along shear surfaces, which can also act as pressure solution seams (e.g., Vannucchi, 2019); the concentration of dense, insoluble elements is why they appear bright in BSE (Figs. 4D, 4H, and 4L Chen et al., 2024a). Foliation development by growth and alignment of phyllosilicates occurs in conjunction with *mélange* development (Fig. 4K, e.g., Schleicher et al., 2012, 2010).

Thus, we document a clear prograde sequence of deformation from the L0 broken formation in the hanging wall to the *mélange* outcrops from soft sediment deformation; formation of block-in-matrix fabric; slip and pressure solution along scaly fabric and phyllosilicate growth; and finally, crenulation (Figs. 10B–10E). These structures represent interseismic deformation forming along the subduction path when the plate interface is locked but the surrounding wall rock strains.

Potentially Coseismic Deformation

The fault rocks, meanwhile, record overprinting of that ductile deformation by brittle structures (Fig. 10). In some cases, blocks of *mélange* within fault strands are cut by throughgoing fractures filled with stilpnomelane, a phyllosilicate mineral that cannot form in these rocks in the near surface (e.g., Hashimoto and Kanehira, 1975, indicative of pumpellyite facies metamorphism). This indicates fault localization after *mélange* formation but before significant exhumation to lower temperature conditions.

Rocks from the discrete faults also show evidence for repeated cycles of such

ductile and brittle deformation (Fig. 10). For example, wall rock with scaly fabric and well-developed phyllosilicate foliation is brecciated and surrounded by quartz and calcite veins. Clasts of brecciated wall rock and vein material show evidence of additional, later stages of brecciation. The quartz vein material surrounding the breccia shows evidence of bulging microstructures, which likely occurred at or above 280°C (Stipp et al., 2002, though this work was conducted at low strain rates), the highest temperatures experienced by our samples. The static chlorite surrounding the final stage of brecciated clasts likely formed as an alteration product between mudstones and quartz under hot, hydrothermal fluid circulation after deformation concluded. Thus, all brecciation must have occurred after *mélange* formation but before exhumation. The fault with the thick vein paralleling the principal slip surface (Figs. 7D–7F) also shows evidence for bulging recrystallization in the quartz vein material separated by dark shear surfaces. This geometry indicates that the principal slip surface was already present when the vein formed, which also had to occur at the point of maximum subduction due to the observed quartz microstructures. Large, intact quartz veins in principal slip surfaces with evidence of quartz plasticity (e.g., Fig. 7C) are common throughout the field area, which indicate that they likely did not form in the near surface.

In general, fault strand rocks contain a higher concentration of quartz and calcite veins, suggesting they acted as fluid conduits, which cemented the fault rock contained in the gullies and promoted fault healing. When mineralization leads to strain hardening, the fault strand is abandoned, and a new one develops (Meneghini and Moore, 2007). Multiple active fault strands without a single, well-defined fault core at other paleomegathrust interfaces has been attributed to sealing of slip surfaces by vein material and fault healing (Meneghini and Moore, 2007). This enables the widening of the fault zone area.

The fault gouge from the principal slip surfaces itself has a strong cataclastic fabric and is largely unfoliated, but contains slip surfaces consistent with expected

fault motion. The soft clay matrix may be a product of later alteration and surface weathering, but we interpret it as having replaced gouge material that was cataclastic at depth. It is possible that the faults were also reactivated during exhumation in the nearsurface, but the geometry of the fault strands, *mélange*, and the microstructures we observe indicate that fault strand development and slip initially occurred at depth. This reactivation of faults formed at depth has been documented in other *mélanges* (Fisher et al., 2019). The thickness of the principal slip surfaces we observe vary from millimeters to centimeters, even among the same structure (e.g., Fig. 5B versus 6D). However, we were only able to sample the thicker slip surfaces, which are known to show a smaller coseismic temperature rise than thinner slip surfaces along strike of the same faults (Savage and Rowe, 2024). They may also record only the last stage of delocalization during slip, which obliterates the record of an originally thinner principal slip surface (Savage and Rowe, 2024).

The only fault at LOTG that does not show obvious evidence of brittle deformation is a gully in the northeastern part of the fault zone that separates the *mélange* from less-deformed units (Fig. 8). The highly foliated, ultrafine, dark matrix contains porphyroclasts of pyrrhotite with pressure shadows of quartz consistent with top-to-SW shear seen at LOTG (Figs. 8B and 8C). It is possible that since this fault is near the upper boundary of the fault zone, it is the least developed or represents an early stage of fault formation. If so, it may imply that faults at LOTG form by first concentrating ductile shear to a thin zone, which can serve as a place for brittle deformation to later localize as nearby faults “turn off” due to strain hardening (Meneghini and Moore, 2007; Moore and Byrne, 1987). Then, our observations would be consistent with the model presented in Rowe et al. (2013), where multiple but not all fault strands are active at one time, and fault zone thickness builds over time.

While pseudotachylyte is considered the most unequivocal evidence for seismogenic fault slip (Cowan, 1999; Rowe and Griffith, 2015), it is notoriously hard to preserve in the rock record due to its vulnerability to overprinting, and it is difficult

to identify in sedimentary rocks in particular ((Kirkpatrick and Rowe, 2013). Pseudotachylyte veins are typically 1 mm in thickness, which adds to the difficulty in observing them in the field. However, observations from faults that have known slip histories have demonstrated that brittle features beyond pseudotachylyte can accommodate seismic slip rates. Cores from ocean drilling through megasplay and plate boundary faults have shown that seismogenic slip occurs on millimeter-thick surfaces of fault gouge showing distributed fracture and brecciation with preferred orientation of clay minerals (Ujiie and Kimura, 2014). Other paleomegathrust faults thought to record seismogenic slip show evidence of fluidized fault rocks, which later form cataclasites once solidified (Ujiie and Kimura, 2014). Non-foliated cataclasites like those from our fault gouges can be expected to accommodate seismic slip rates at similar temperatures to LOTG (Rowe et al., 2011).

Planar deformation features such as those identified in Figure 7C have been shown to form coseismically in other studies (Hamers and Drury, 2011; Vernooij and Langenhorst, 2005) and are considered potentially earthquake related when identified (Rowe and Griffith, 2015). While we describe dynamic recrystallization in quartz and suggest it indicates that brittle deformation occurred at depth in our fault rocks, some work has shown that localized dynamic recrystallization can occur coseismically as well (e.g., Bestmann et al., 2012).

In this work, we identify evidence of cyclical strain rate variations corresponding to the seismic cycle similar to those documented in other exhumed subduction complexes (e.g., Rowe et al., 2011). Given the geometrical argument that LOTG represents a paleomegathrust interface and the new RSCM temperatures consistent with the seismogenic zone, we consider the brittle fault rocks presented here as evidence for coseismic slip along the plate interface.

4.7 Conclusions

We investigated an area of the Olympic Subduction Complex known as Lakes of the Gods (LOTG), an area that has not seen geological investigation since the 1970s. We performed Raman spectroscopy of carbonaceous materials to derive new paleotemperatures of 260 – 280°C for the Lower Olympic Subduction Complex. This indicates that the LOTG fault zone subducted to the seismogenic zone conditions at ~12–15 km depth before beginning exhumation. We documented the structure of the LOTG fault zone, where there are numerous centimeter-scale principal slip surfaces in multiple meter-scale fault strands, all contained in an accumulated deformed zone up to hundreds of meters thick.

We show the microstructural evidence for coeval ductile and brittle deformation in the rocks at LOTG. The *mélange* develops through granular flow, pressure solution, alignment and growth of phyllosilicates, and crystal plastic deformation, all of which develop scaly fabric, during interseismic periods. The fault strands show brittle deformation superimposed on the scaly fabric including cataclasis, brecciation, and veining, all of which potentially represent coseismic deformation. Quartzofeldspathic rocks at these temperature and pressure conditions are expected to be frictionally unstable and therefore prone to dominantly seismic stick-slip mechanics (Hyndman and Wang, 1993; Moore and Saffer, 2001b); moreover, modern clastic-rich subduction zone megathrusts globally are observed to slip mostly in very large earthquakes (Ruff, 1989). We conclude that brittle deformation in these faults would almost certainly take place via rapid, seismic slip rather than slow, aseismic processes. The principal slip surfaces within the fault strands were fluid conduits, which may have led to the strengthening and widening of the fault zone over time, forming the braided network of faults we observe today. We conclude that the LOTG fault zone specifically represents a paleomegathrust interface, which was localized in the sediments when formed and provides a blueprint for the structure of megathrust interfaces, including, but not

limited to, Cascadia.

Acknowledgements

We gratefully acknowledge Olympic National Park for permitting this research, and specifically Matt Dubeau and Bill Baccus for their continued support. This work was greatly improved by assistance from and discussions with field assistants Sophie Johnson and Peter Lindquist. Part of this work was conducted at the Molecular Analysis Facility, a National Nanotechnology Coordinated Infrastructure site at the University of Washington, which is supported in part by the National Science Foundation (grant NNCI-1542101), the University of Washington, the Molecular Engineering & Sciences Institute, the Clean Energy Institute, and the National Institutes of Health. Funding for some of the field and lab work was provided by the Geological Society of America Graduate Student Research Grant program and by StraboSpot to A. Ledeczi. We gratefully acknowledge support from the Paros Endowed Chair in Seismology and Geohazards to H. Tobin. This work was improved by comments from reviewers Paola Vannucchi, Christie Rowe, and Donald Fisher, and Associate Editor Stacia Gordon.

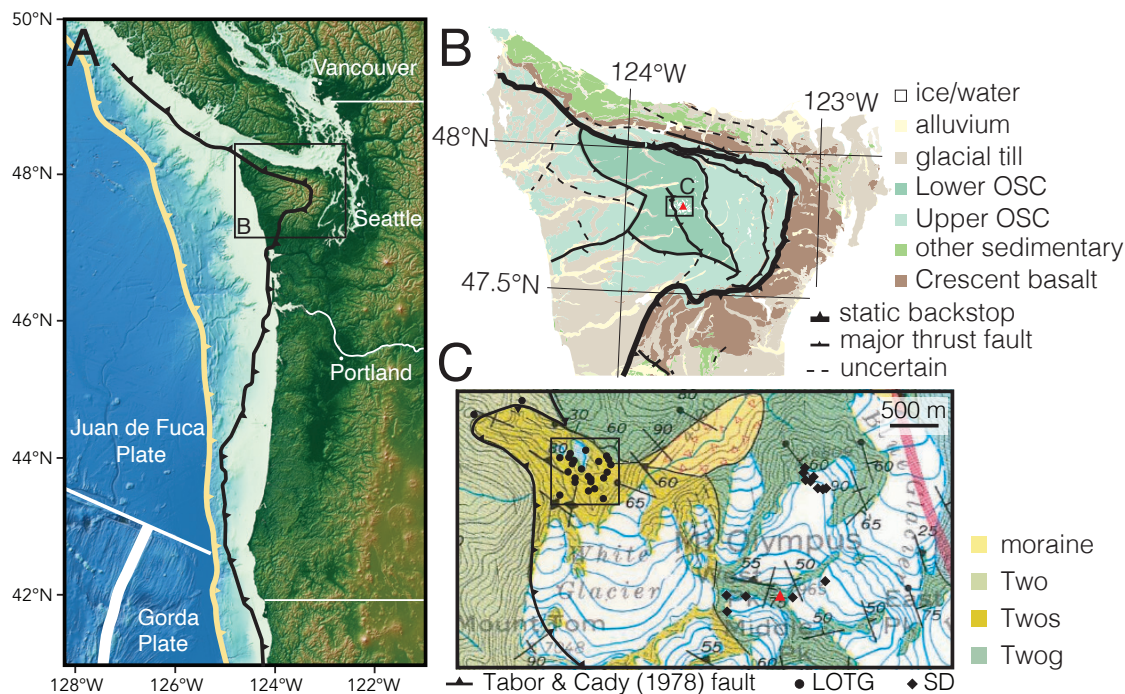


Figure 4.1: **(A)**. Bathymetric map of the Cascadia subduction zone, where the Juan de Fuca plate subducts beneath North America. Yellow barbed line is the deformation front; black barbed line is the crystalline continental basement, or static backstop from Watt and Brothers (2020). The area between these lines is the modern Cascadia accretionary wedge, exposed on land only in the Olympic Peninsula of Washington. Box shows location of part B. **(B)**. Simplified geologic map of the Olympic Peninsula after Gerstel and Lingley (2003) and Brandon et al. (1998) showing the tectonic window into the Lower Olympic Subduction Complex (OSC) in the high Olympics. Other units include alluvium, glacial till, Upper OSC, other Eocene to Miocene sedimentary units, and the Crescent basalt. The Tabor and Cady (1978a) thrust is visible in the middle of the Lower OSC unit. Box shows location of part C and red triangle is the location of Mount Olympus. **(C)**. Excerpt from Tabor and Cady (1978a) image showing samples collected at the Lakes of the Gods (LOTG) and Snow Dome (SD) field sites and location of the Tabor and Cady (1978a) mapped fault. Two—Tertiary Western Olympic lithic assemblage; Twos—slate and phyllite with <30% small phacoids of foliated sandstone and semischist; Twog—greenish-gray and black sandstone and semischist in <50% slate and phyllite. Box shows location of Figure 9A, and red triangle is the location of Mount Olympus. All sample photos are available on StraboSpot and are archived at osf.io (<https://doi.org/10.17605/OSF.IO/JRH37>).

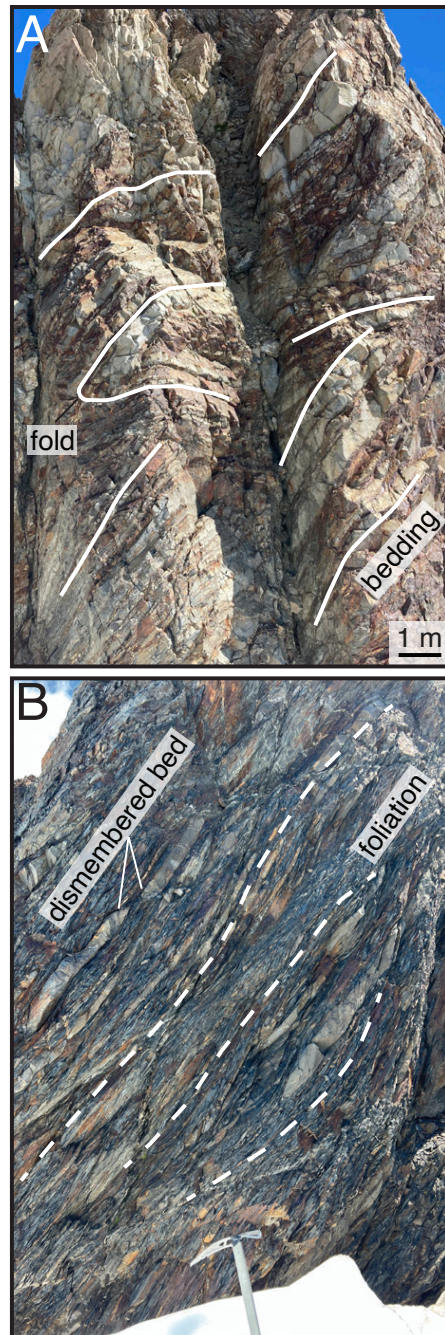


Figure 4.2: Typical appearance of broken formation at Snow Dome. (A). Interbedded sandstone and mudstone showing folded bedding at the outcrop scale. (B). Mudstone-dominated unit with spaced foliation and dismembered sandstone beds, which show offset along shear surfaces. Ice axe is 70 cm in length.

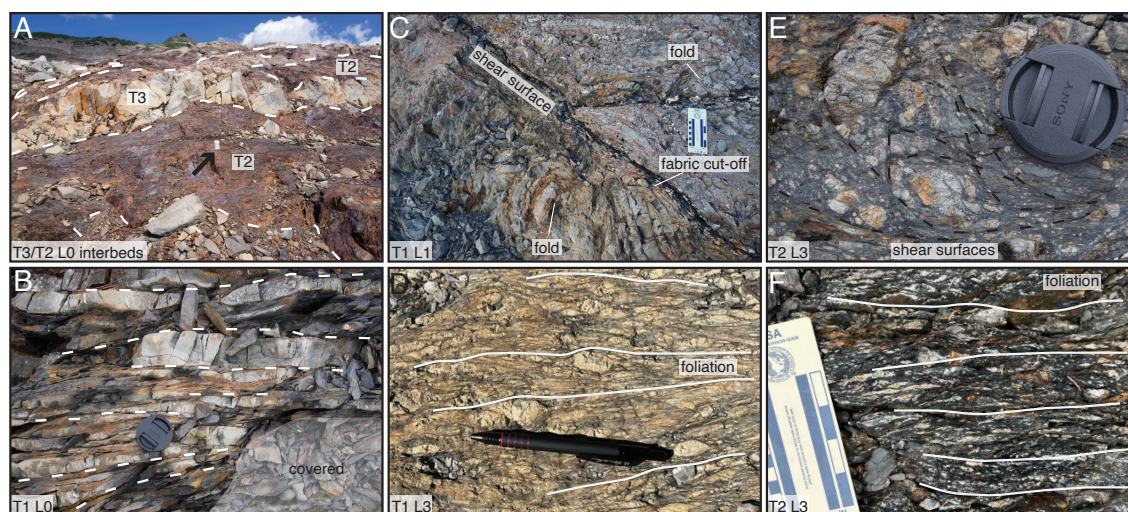


Figure 4.3: Typical appearance of lithological units at Lakes of the Gods. See text for description of units and fabric development levels. Dashed white lines, dashed black lines, and solid white lines indicate bedding contacts, shear surfaces, and foliation, respectively. Scale object in parts A, C, and F is 16 cm in length; lens cap in parts B and E is 7 cm in diameter; and pen in part D is 15 cm in length. **(A)**. Interbedded L0 T2 and T3. Contacts are depositional, but beds are thinned, thickened, and folded at the outcrop scale. Arrow points to scale in middle ground. **(B)**. T1 L0 turbidite showing intact bedding at the outcrop scale. **(C)**. T1 L1 showing dismembered folds and discontinuous beds with fabric cut at shear surfaces. **(D)**. Glacially polished outcrop of T1 L3, showing scaly fabric wrapping around sandstone, forming a penetrative foliation. **(E)**. Glacially polished outcrop of T2 L3 showing sub-centimeter scale phacoids separated by dark, curving, intersecting shear surfaces. **(F)**. T2 L3, where sub-centimeter scale phacoids are highly aligned parallel to the foliation.

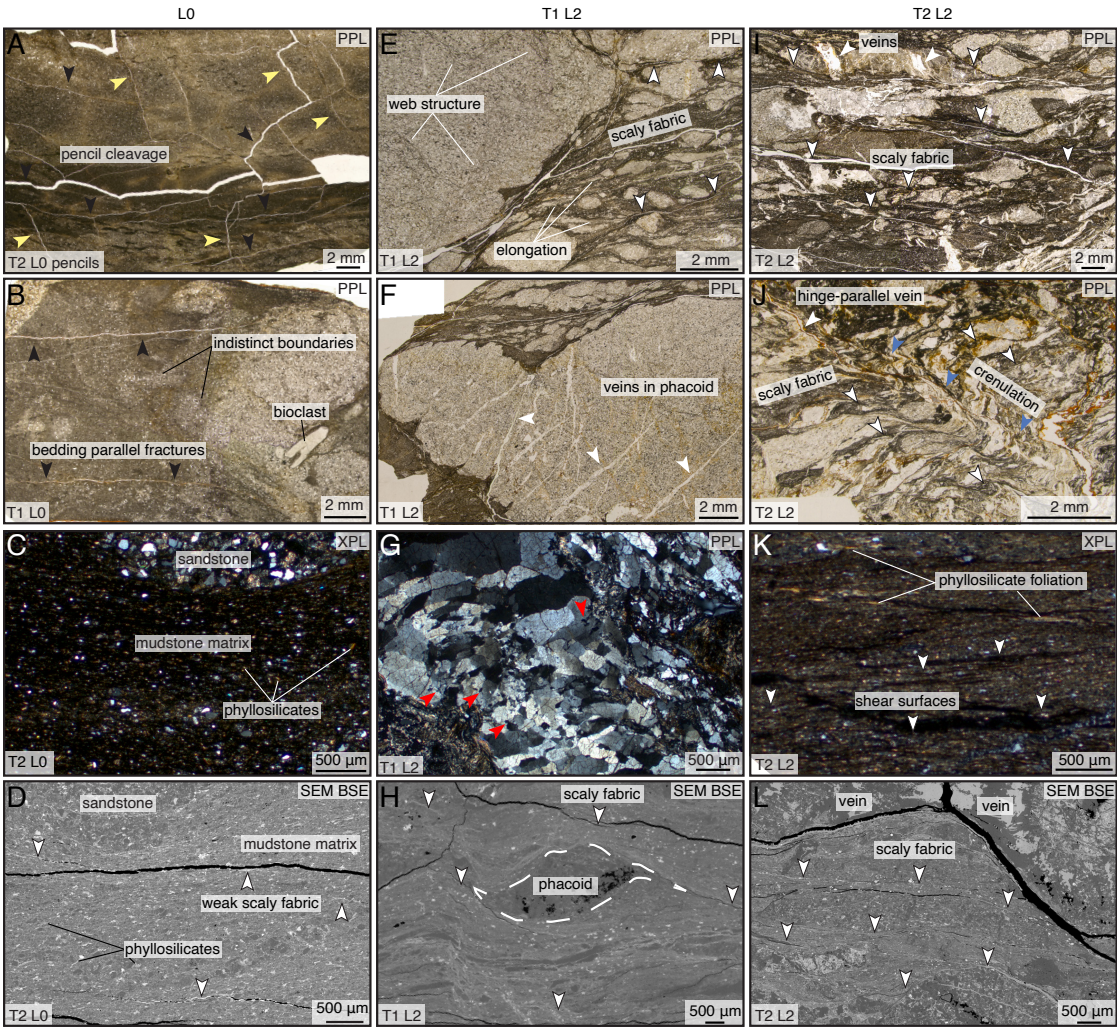


Figure 4.4: Microstructures of broken formation (column 1) and T1 L2 and T2 L2 mélange (columns 2 and 3). All images are displayed so that foliation or bedding is horizontal. Bedding, veins, and scaly fabric are highlighted by black, white, and white with black outline arrows, respectively. L0 column: **(A)** Plane-polarized light (PPL) image of T2 L0 sample showing pencil cleavage. The intersecting cleavage is highlighted by yellow arrows. **(B)** PPL image of T1 L0 sample, which displays sedimentary clasts, sandstone with indistinct boundaries—interpreted as soft sediment deformation—and bedding-parallel fractures. **(C)** Cross-polarized light (XPL) image of a T2 L0 sample showing compositional layering without foliation development or alignment of phyllosilicates. Phyllosilicate grains are small ($\sim 50 \mu\text{m}$ length) compared to part K. **(D)** Scanning electron microscope (SEM) backscattered electron (BSE) image of T2 L0 sample showing bedding parallel fractures with little enrichment of dense (i.e., bright) elements, indicating weak scaly fabric development at this deformation level. Phyllosilicates are relatively unaligned, with some at a high angle to the bedding. T1 L2 column: **(E)** PPL image of T1 L2 sample showing sandstone phacoids in mudstone matrix separated by scaly fabric. Open cracks form along and highlight the scaly fabric. Weakly developed web structure is visible in the large sandstone phacoid. Smaller phacoids show elongation through granular flow. **(F)** PPL image of T1 L2 sample showing large sandstone phacoid containing early-stage quartz veins and surrounded by shear surfaces. **(G)** XPL image of deformed quartz vein in T1 L2 sample. Bulging quartz grains highlighted by red arrows are prevalent, indicating dynamic recrystallization. **(H)** SEM BSE image of T1 L2 sample showing large sandstone phacoid outlined by white dashed line surrounded by bright scaly fabrics, providing evidence for pressure solution along boundaries. T2 L2 column: **(I)** PPL image of T2 L2 sample dominated by mudstone matrix with quartz veins and fewer sandstone phacoids than in T1 L2. Thick shear veins accommodate extension. **(J)** PPL image of T2 L2 sample with crenulation cleavage (blue arrows) folding original scaly fabric. **(K)** XPL image of T2 L2 sample showing prominent ultrafine-grained, black shear surfaces separating regions of mudstone with highly aligned, elongate ($\sim 100\text{s } \mu\text{m}$ length) phyllosilicate grains defining the foliation as highlighted by parallel illumination. **(L)** SEM BSE image of T2 L2 sample showing bright scaly fabrics within mudstone matrix. All photomicrographs are available on StraboSpot and are archived at osf.io (<https://doi.org/10.17605/OSF.IO/8ZWXB>).

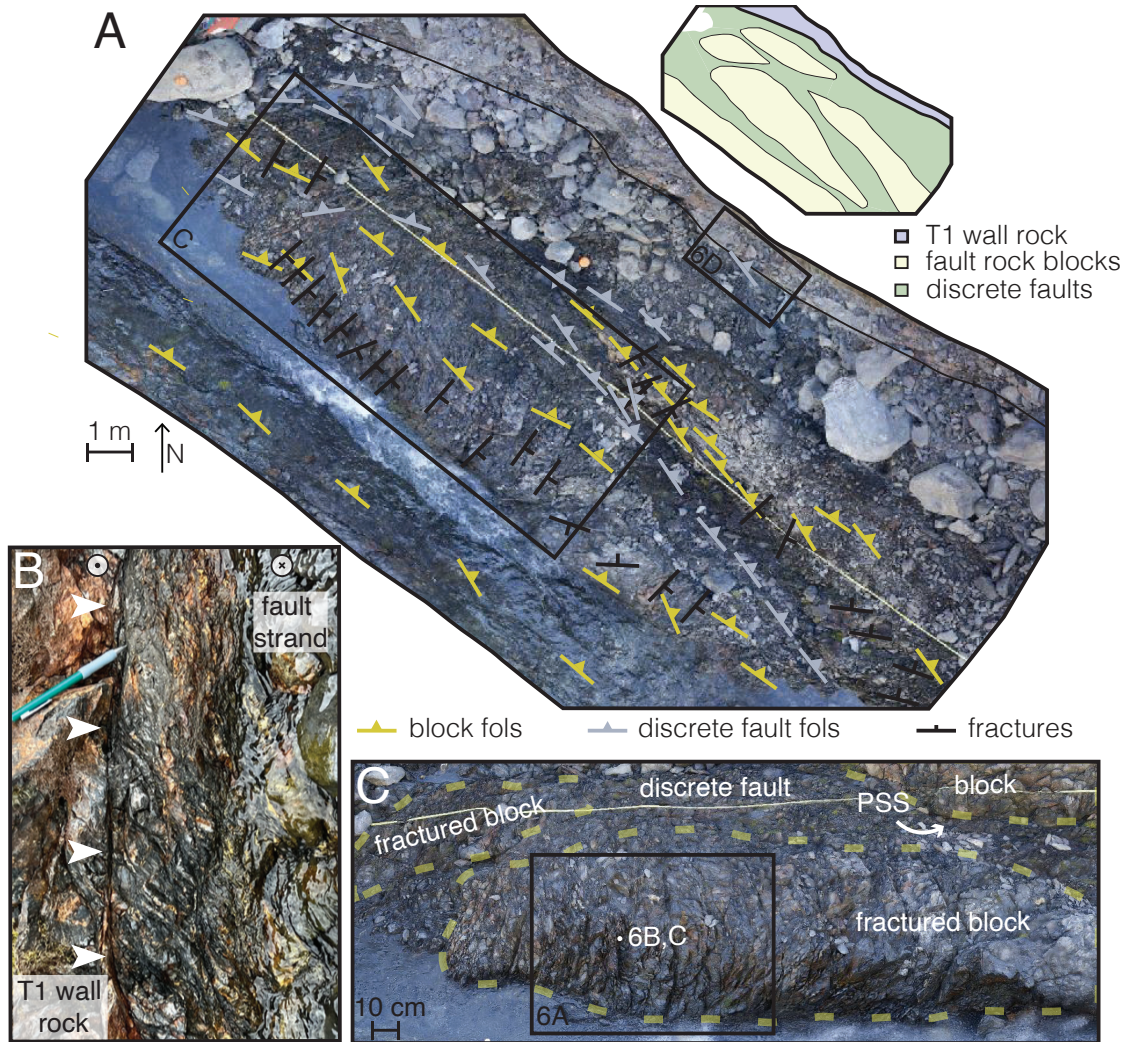


Figure 4.5: **(A)** Structure-from-motion (SfM) orthomosaic facing down of fault strand showing blocks of brittle fault rock, yellow, surrounded by discrete faults, green. Foliation measurements show how discrete fault fabric anastomoses around brittle fault rock blocks. Fracture orientations, in black, are at $\sim 90^\circ$ to foliation (in 3-D). Boxes show location of Figure 5C and Figure 6D. **(B)** Foliated fault gouge cut by thin principal slip surfaces (PSS; white arrows) in fault strand farther south of SfM orthomosaic. **(C)** Portion of model facing east showing fractured blocks. Box shows location of photo in Figure 6A and sample in Figures 6B and 6C. fols—foliations.

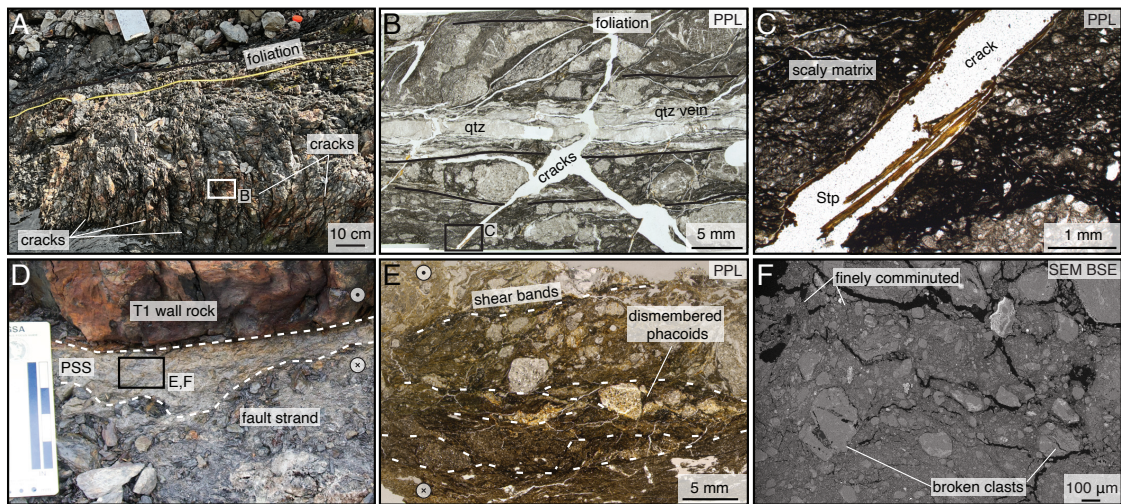


Figure 4.6: Fault rock characteristics. **(A)** Fractured block shown in Figure 5C. Box shows location of sample in part B. Foliation is parallel to the yellow tape measure in the background; cracks crosscut the foliation. **(B)** Plane-polarized light (PPL) image of sample in part A displaying scaly fabric cut by foliation-parallel quartz (qtz) vein and same cracks shown at the macroscale in part A. **(C)** PPL image of stilpnomelane (stp) growing in these fractures. **(D)** Principal slip surfaces (PSS; outlined with white dashed line) composed of fine-grained cataclasite separating T1 wall rock from foliated cataclasites in the fault strand shown in Figure 5A. Fault movement is hanging wall out of the plane. Box shows location of sample in parts E and F. Scale object is 16 cm in length. **(E)** PPL image of cataclastic fault gouge with shear surfaces dashed in white and dismembered sandstone phacoids. **(F)** Scanning electron microscope backscattered electron (SEM BSE) image showing broken and finely comminuted clasts in a fine-grained matrix.

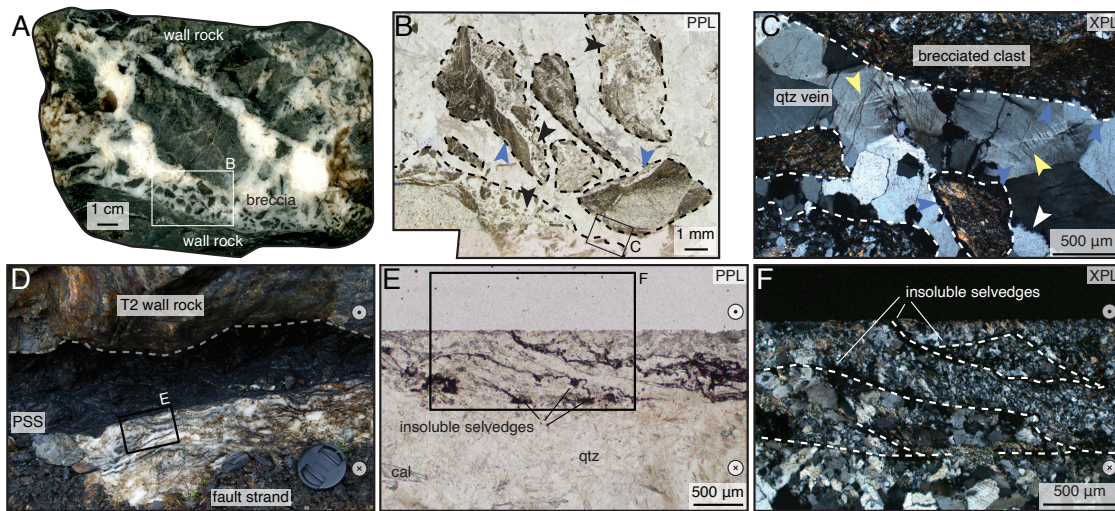


Figure 4.7: Examples of fault rocks with vein material. **(A)** Hand sample of brecciated fault rock with clasts of wall rock (dark) surrounded by quartz (light gray) and calcite (white) vein material. **(B)** Plane-polarized light (PPL) image of sample in part A. Dashed lines indicate clasts created during “final stage” of brecciation, in some cases with previously brecciated clasts inside, highlighted with black arrows. Blue arrows point to chlorite growing on outer edge of all “final stage” clasts. **(C)** Cross-polarized light (XPL) image of brecciated clasts surrounded by chlorite (blue arrows) and quartz vein material with deformation bands (yellow arrows) and bulging microstructures (white arrow). See Figure S2 (see text footnote 1) for a photomicrograph of the chlorite. **(D)** Dark, foliated, fine-grained cataclasite principal slip surfaces (PSS) paralleled by 15-cm-thick quartz vein separating T2 hanging wall rock (gray dashed line) on top from fault strand. Fault motion is hanging wall out of the plane. Box shows location of sample in part E. Scale object is a camera lens cap (7 cm in diameter). **(E)** PPL image of quartz (qtz) and calcite (cal) vein material separated by insoluble selvages. **(F)** Same photomicrograph in part E in XPL image showing dynamic recrystallization of quartz vein material between the insoluble selvages highlighted by white dashed lines.

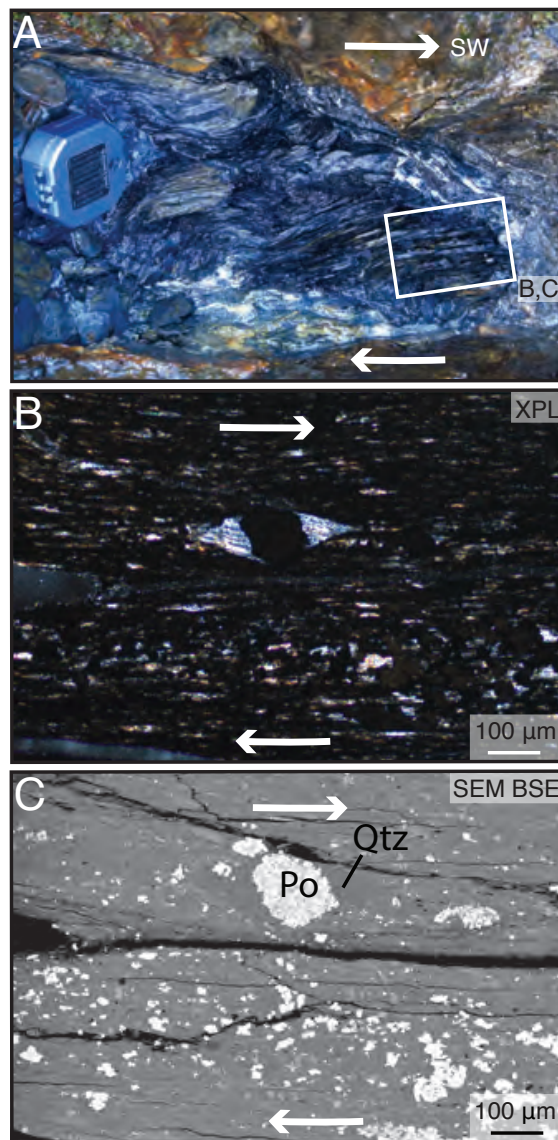


Figure 4.8: The gully between the fault zone and coherent unit shows no evidence of brittle fault rock. **(A)** The fault strand in gully 9 is composed entirely of an ~30-cm-wide discrete fault containing foliated rock with an ultrafine-grained, black matrix. Motion of fault is top to the right. Box shows location of sample in parts B and C. **(B)** Cross-polarized light (XPL) image of dark, fine-grained, highly foliated rock containing porphyroclasts of pyrrhotite (po) with strain shadows of quartz (qtz). **(C)** Scanning electron microscope backscattered electron (SEM BSE) image. All bright spots are porphyroclasts of pyrrhotite.

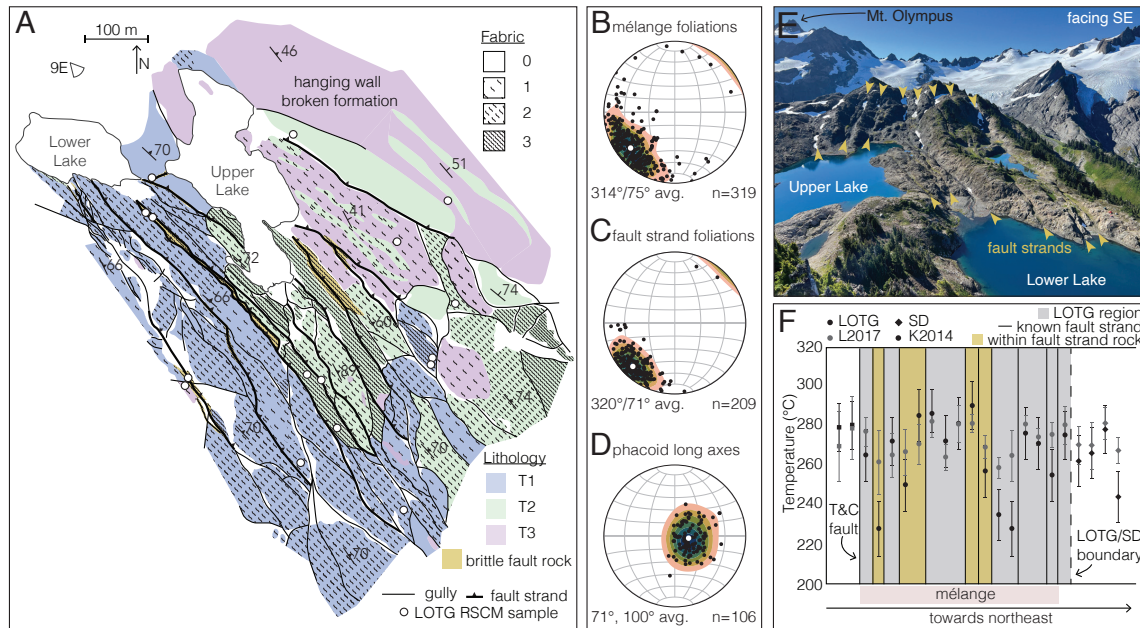


Figure 4.9: **(A)** Geologic map of the Lakes of the Gods (LOTG) fault zone. Colors and superimposed patterns represent lithology and fabric, respectively. Dark black lines with barbs represent known gully fault strands, while thinner black lines are gullies without exposed fault rock. Representative foliations and bedding attitudes shown are a subset of over 400 measurements taken. Location where photo E was taken is shown. Stereoplots of **(B)** foliation poles in mélangé outcrops, **(C)** in fault strands, and **(D)** lineations of phacoid long axes within foliation planes. **(E)** Photo with fault strands annotated. **(F)** Raman spectroscopy of carbonaceous material (RSCM) results for LOTG (circle), Snow Dome (SD; diamond), and western side of the Tabor and Cady (1978a, T&C) fault (square) using the Kouketsu et al. (2014), black, and Lünsdorf et al. (2017), gray, calibrations. Locations of LOTG samples are shown in part A, and full results and coordinates of all samples are provided in Table S1 (see text footnote 1). Error bars are one standard deviation of the measurements from each sample. Vertical lines are fault strands separating samples, and the dashed line is the unknown boundary between LOTG and SD. The gray shaded area represents samples from LOTG, while the pink bar is samples within the LOTG mélangé zone (i.e., excluding the hanging wall broken formation). Yellow bars are samples taken within brittle fault rock in fault strands.

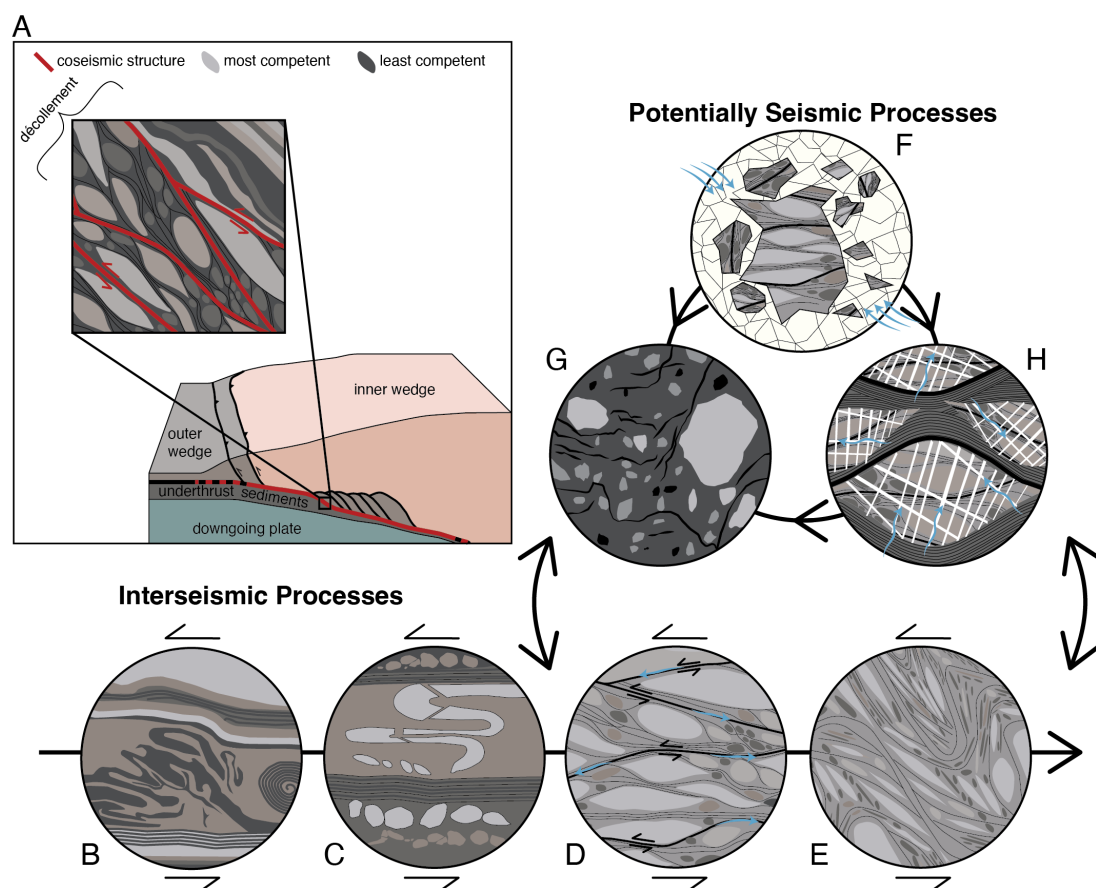


Figure 4.10: **(A)** Proposed model of Lakes of the Gods fault zone formation, where the sediment-hosted décollement forms a web of faults, which slip coseismically surrounded by interseismically formed mélangé. Prograde, interseismic deformation is a sequence of **(B)** soft sediment deformation, **(C)** formation of block-in-matrix fabric, **(D)** slip and pressure solution along shear surfaces and phyllosilicate growth, and **(E)** crenulation. Meanwhile, potentially seismic processes like **(F)** brecciation and veining, **(G)** cataclasis, and **(H)** localization overprint and cycle with interseismic processes.

Chapter 5

OFF-FAULT MÉLANGE FORMATION ACCOMMODATES ONLY A SMALL FRACTION OF STRAIN IN THE SUBDUCTION SEISMOGENIC ZONE

Manuscript in review at *Earth and Planetary Science Letters*.

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5.1 Abstract

Exhumed accretionary complexes provide a window into the architecture of the subduction megathrust interface, but coexistence and overprinting of brittle and viscous deformation structures makes it difficult to disentangle the specific mechanisms hosting various slip modes. Within these exposures, interseismic strain between megathrust earthquakes is accommodated by mélangé zones up to kilometers in thickness. Because mélangé dominates the rock record, it is often the focus of geological investigations, in contrast to the narrow fault zones that host localized seismic slip. Pressure solution is invoked as a significant mechanism of interseismic strain accommodation in mélangé zones within the subduction seismogenic zone, and its efficiency is commonly said to control the downdip limit of seismogenic slip and amount of coupling on the plate interface. However, observations from modern highly-locked subduction

zones show that megathrust earthquakes account for the vast majority of total plate displacement. Here, we apply chemical and magnetic methods to a recently described Cascadia subduction zone paleomegathrust fault exposure to constrain the spatial and structural partitioning of strain accommodation across a seismogenic zone megathrust interface among the *mélange*, discrete faults, and coherent wall rocks. Both volume loss and distributed frictional sliding occur in the *mélange* but accommodate orders of magnitude less shear strain than plate convergence requires. Our work suggests that, even though *mélange* fabrics are volumetrically dominant in the rock record, the majority of plate displacement accommodated by the subduction seismogenic zone is instead taken up on discrete faults therein.

5.2 Introduction

In seismological and modeling approaches, the subduction plate boundary is often treated as a discrete, planar fault. However, the exhumed rock record has shown that the seismogenic megathrust is actually an interconnected network of simultaneously active faults contained in a much wider, tabular body of deformed *mélange* rock up to kilometers in maximum thickness (Rowe et al., 2013). Recently, seismological observations have also resolved earthquakes on an anastomosing system of active braided faults in a zone 100s of meters thick at the plate interface in Ecuador (Chalumeau et al., 2024), corroborating this geological evidence. Exhumed seismogenic megathrust interfaces preserve evidence of brittle and viscous processes on fault strands and *mélange* host rocks. The partitioning of plate boundary motion into interseismic viscous strain versus coseismic, and potentially aseismic, localized elastic strain release is important for understanding the accumulation of slip deficit over time, and by extension, seismic hazards.

The pervasive anastomosing foliation of *mélange* rocks is known as scaly fabric, comprising competent phacoids (i.e., blocks) that are surrounded by an incompetent matrix containing shear surfaces (e.g., Byrne, 1984; Fagereng and Sibson, 2010; Van-

nucchi and Bettelli, 2010, and references therein). Scaliness is a feature of *mélange* fabrics that is produced by a variety of processes operating at different conditions. In the shallowest subduction zone plate boundary faults accessed via ocean drilling, scaly fabric is produced primarily through the collapse of open clay networks into discrete slip planes, which in turn become abandoned due to strain hardening, promoting the formation of a wider deformed area (Moore and Byrne, 1987; Moore et al., 1986, e.g.,). Pressure solution does not operate to a significant degree at shallow conditions below 150°C (Moore et al., 2007b).

Deeper, along the plate interface in the seismogenic zone, scaly fabric is formed by both brittle and viscous processes, primarily through pressure solution and phyllosilicate alignment, along with other mechanisms such as distributed frictional sliding along numerous discrete surfaces, particulate flow, and fracturing (e.g., Vannucchi, 2019). At seismogenic zone conditions during the interseismic period, both distributed (viscous) deformation via pressure solution and frictional sliding accommodates strain in the *mélange* host rock of the plate interface. However, quantifying the relative contribution of these two mechanisms in the formation and progressive evolution of scaly fabric is challenging because shear surfaces lack offset markers, and volume loss estimates from *mélanges* vary wildly (Ague, 1991; Chen et al., 2024a; Kawabata et al., 2007, e.g., 10 to 50% in). Ultimately, the amount of strain that is required to produce *mélange* fabrics is not well understood (Vannucchi, 2019, e.g.,).

Geological and geophysical views are often at odds because *mélange* is volumetrically dominant at exhumed subduction complexes compared to discrete brittle faults that accommodate seismic slip; however, at highly-locked modern subduction zones, large megathrust earthquakes account for the majority of displacement (Avouac, 2015). Earthquake-related strain is not always visible in the rock record because the geological evidence for seismic slip is often overprinted or unclear (Cowan, 1999; Rowe and Griffith, 2015; Fisher et al., 2021a). Thus, even structures known to be exhumed from the seismogenic zone are often interpreted in terms of slow slip or

aseismic deformation (Condit et al., 2025b; Kirkpatrick et al., 2021, e.g.). Here, we consider the plate boundary *mélange* zone as a shear zone in which discrete displacement and viscous deformation contribute to the total finite strain which we assume is controlled by the amount of relative plate motion that must be taken up across the plate boundary. We refer to deformation operating continuously as viscous, in contrast to the episodic short-lived seismic and/or transient aseismic slip on discrete faults.

Fundamentally opposing views of the role and formation of *mélange* during subduction at seismogenic zone conditions have been put forth and remain controversial. The view which is most consistent with geophysical observations, such as slip budgets derived from geodetic locking models and earthquake slip inversions, is that a small amount of interseismic strain is accommodated on *mélange* zones 100s of meters in thickness, while seismogenic slip that facilitates the majority of relative plate motion occurs on thin faults (Ujiie and Kimura, 2014; Rowe et al., 2013; Fisher et al., 2021a; Fisher and Hirth, 2024, e.g.), sometimes accompanied by relatively minor afterslip or aseismic frictional slip on faults. Processes such as localized frictional slip on discrete faults and scaly surfaces can occur aseismically; however, in the context of clastic-sedimented dominated megathrusts, frictional slip is typically interpreted as dominated by seismic processes (Ruff, 1989; Wang and Bilek, 2014; Wirth et al., 2022, e.g.). In such models, pressure solution-driven aseismic creep in *mélange* zones accommodates a small portion of the relative plate slip, thereby reducing the portion which is released as an elastic response during megathrust earthquakes (Fisher et al., 2021b). This view suggests that the downdip limit of the seismogenic zone may occur when pressure solution is able to accommodate the full plate rate aseismically (Condit et al., 2022; Fisher and Hirth, 2024).

On the other end of the spectrum, workers have suggested that frictional-viscous flow driven by pressure solution is the primary deformation mechanism in the seismogenic zone (Fagereng and den Hartog, 2016). This work argues that increased effi-

ciency of pressure solution can explain why some subduction zones or other kinds of plate boundaries deform dominantly aseismically at lower-than-expected temperatures (Fagereng and den Hartog, 2016; Gratier et al., 2011). Some authors have suggested that large enough amounts of strain are recorded in *mélange* to accommodate plate motion on their own, and any faults preserved in the rock record are a result of underplating, not active plate interface slip (Raimbourg et al., 2019). These views (Fagereng and den Hartog, 2016; Raimbourg et al., 2019) are, however, fundamentally inconsistent with geophysical observations that megathrust earthquakes dominate the interplate displacement at most modern highly-locked subduction zones.

Here, we address this discrepancy by interpreting the deformation recorded in a particularly well-exposed exhumed paleomegathrust fault from the Olympic subduction complex in Cascadia (Ledeczi et al., 2025) and comparing it to an estimate of the amount of displacement that the rocks might have experienced during subduction. We show that the low amount of strain recorded by the *mélange* fabrics indicates that viscous deformation in these rocks would have only been able to accommodate on the order of 0.1% of the plate displacement from the subduction process, while the remainder must instead be accommodated almost entirely on discrete surfaces, potentially during seismic slip. This is consistent with geophysical observations of highly-locked subduction zones that host megathrust earthquakes, such as the modern Cascadia subduction zone (Li et al., 2018; Lindsey et al., 2021, e.g.,) and many others (Avouac, 2015, e.g.,). Our results can help inform assumptions in geodetic locking models that about how much and through what mechanism off-fault deformation is accommodated. Furthermore, it provides new context for the geological interpretation of subduction fault complexes globally.

5.3 Geologic Context

The extensive study of exhumed paleomegathrust exposures, such as those in the Shikama Belt and Kodiak Complex, has revealed connections between strain as recorded

in rocks and the kinematics and dynamics of plate boundary processes (Fisher et al., 2021a; Kimura et al., 2012; Kitamura and Kimura, 2012; Rowe et al., 2005; Ujiie et al., 2007, e.g.). These exposures are typically characterized by a block-in-matrix *mélange* fabric hosting a series of anastomosing discrete brittle faults (Rowe et al., 2013). Coseismic slip during earthquakes is localized onto thin cataclastic zones, for example along roof thrusts, while interseismic deformation is primarily accommodated via pressure solution in the surrounding *mélange*.

The Olympic Structural Complex (OSC) is a deeply exhumed and subaerially-exposed part of the Cascadia subduction zone exposed on the Olympic Peninsula in Washington state, USA (Brandon et al., 1998; Tabor and Cady, 1978b, e.g.). The outcrop area of the central Olympic mountain core (lower OSC of Brandon et al., 1998) is composed exclusively of prehnite-pumpellyite grade clastic metasediments including hemipelagite to turbidite mudstones and sandstones, interpreted to have been initially carried down along a megathrust then underplated into the accretionary wedge. In the lower OSC, recent work has revealed a previously unknown example of a clastic-sediment-hosted exhumed paleomegathrust known as Lakes of the Gods shear zone (LOTG; Ledeczi et al., 2025). We use shear zone here to denote the entire width of the fault system, including both brittle, discrete structures and distributed *mélange* fabrics. This shear zone consists of 10 major anastomosing faults contained in a 450 m wide *mélange* zone structurally underlying a less deformed coherent unit (Fig. 1A). The coherent wall rock unit (*sensu* e.g., Kitamura et al., 2005) above the shear zone is a broken formation in which outcrop-scale faulted and folded beds are laterally continuous at the 10 m scale; primary sedimentary deformation structures and macroscopically bedded sequences are common. Ledeczi et al. (2025) document coplanar foliations and bedding in the coherent unit, the *mélange*, and the fault rocks, suggesting the *mélange* is tectonic in origin and is genetically related to the fault strands. This also implies original subhorizontality of the faults and, as concluded in that study, is consistent with the interpretation that the LOTG shear zone is a

paleomegathrust recording plate interface deformation. Furthermore, geothermometry indicates this paleomegathrust experienced mid-seismogenic zone temperatures peaking at 250-280°C (Ledeczi et al., 2025).

Microstructural observations show that the *mélange* formed interseismically via pressure solution, alignment and growth of phyllosilicates, granular flow, and crystal plastic deformation (Ledeczi et al., 2025). The fault strands (Fig. 1A) preserve evidence of widespread cataclasis, brecciation, and veining (Ledeczi et al., 2025). Localized, frictional displacement on discrete surfaces, breccias, and cataclasites are not necessarily diagnostic of exclusively coseismic slip (Rowe and Griffith, 2015, e.g.); however, microstructural observations in Ledeczi et al. (2025) document evidence that these features are likely coseismic. Additionally, the direct correlative relationship of the LOTG locality with the largest highly-locked asperity at the modern Cascadia subduction zone (Lindsey et al., 2021, e.g.) suggests that coseismic slip was a primary deformation mechanism in the faults. Here, we use LOTG as a natural laboratory to understand strain partitioning in the three structural domains of the subduction interface: the coherent unit, *mélange* host rocks, and fault rocks.

5.4 Methods

We apply analysis of field-measured fabrics, anisotropy of magnetic susceptibility (AMS), electron probe microanalysis (EPMA), and bulk rock elemental analysis to samples along a fault-strike perpendicular transect spanning the outcrop exposure. The anisotropy of magnetic susceptibility (AMS) is mathematically defined as a second-rank tensor with six independent matrix elements, which geometrically can be represented as an ellipsoid defined by the maximum (K1, or the magnetic lineation), intermediate (K2), and minimum axes (K3, or the pole to the magnetic foliation), or principal magnetic susceptibilities. The magnitude of these principal axes can be used to define parameters like the corrected anisotropy degree P_j (Jelinek, 1981), which can be related with the intensity of the preferred mineral alignment within the

sample, or the deviation of the magnetic ellipsoid from a sphere. A Flinn diagram (Flinn, 1956) is used to illustrate the shape of the magnetic ellipsoids by comparing the magnetic lineation ($K1/K2$) and the magnetic foliation ($K2/K3$). It has been widely demonstrated that AMS may often correlate with the finite strain geometry of the rock (Borradaile and Henry, 1997; Hrouda, 1982; Parés, 2015). In deformed sedimentary rocks, AMS reflects mineral alignment that may be produced by combinations of original depositional and compactional fabrics and the bulk viscous strain from processes such as grain rotation, kinking, deformation banding, particulate flow, and pressure solution (e.g., Parés, 2015). AMS does not capture slip on discrete surfaces activated during brittle deformation. Therefore, it reflects solely the distributed deformation experienced by these rocks. In subduction zone settings, the deformation recorded by AMS is a combination of depositional, compactional, and tectonic fabrics, the latter of which is typically inferred to represent solely interseismic deformation (Housen et al., 1996; Kitamura and Kimura, 2012; Parés, 2015).

We also use elemental maps and quantitative spot data from EPMA to perform a chemical mass balance calculation to quantify the amount of volume loss via pressure solution and compare derived strain rates with those expected for plate convergence. The total strain due to pressure solution is constrained by comparing the enrichment of fluid-immobile Ti and the depletion of fluid-mobile Si along shear surfaces compared to the adjacent unaltered wall rock in each sample (Chen et al., 2024a). Bulk rock compositions allow us to determine whether those microscopic trends are applicable at the field scale. Finally, we utilize constitutive relationships, including the Fisher and Hirth (2024) flow law for pressure solution, to provide a qualitative comparison and contextualize our results. Detailed methodology is provided in Appendix A.

5.5 Results

5.5.1 AMS

Ledeczi et al. (2025) presented structural data for LOTG, including attitudes of *mélange* and fault foliations, bedding, and long axes of phacoids, which are elongate in the transport direction. We compare that data to results derived from new AMS data for the three units analyzed in this study: the *mélange*, faults, and coherent rocks. Magnetic mineralogy experiments (Appendix A.2) indicate lithological variation is unlikely to cause the AMS variation between the three units because their mineralogy is all mainly paramagnetic (i.e., phyllosilicate-rich) with only minor to negligible contributions from ferromagnetic minerals. Therefore, we interpret that the magnetic fabric is likely representative of the structural fabric (Borradaile et al., 1987).

For the coherent unit samples, the magnetic lineation (K1) and magnetic foliation poles (K3) have predominantly oblate AMS ellipsoids, indicating that the magnetic minerals reflect initial deposition and compaction. For sites located in the *mélange* and fault rock units the confidence ellipses around the mean magnetic lineation (K1) overlap with the long axes of phacoids in these rocks, which are elongate in the transport direction of the shear zone (Fig. 1B-C; Ledeczi et al., 2025). The confidence ellipses of the poles to magnetic foliation (K3) in the *mélange* and fault units also overlap with poles to tectonic foliation and bedding planes in the *mélange*, as well as with poles to tectonic foliation planes in the fault strands (Fig. 1B-C). In each case, there is a $\leq 12^\circ$ difference between the mean magnetic axes orientations and the mean of the corresponding structural data.

While the orientations of the AMS are similar between the structural units, the shape and degree of anisotropy of the AMS ellipsoids varies between them. The *mélange* units have higher degree of anisotropy (Pj Jelinek, 1981) than the fault and coherent unit samples. The ellipsoid shapes (Fig. 1D and A3) vary from more

oblate-to-triaxial shapes in the coherent unit, to shapes that have more pronounced variation between prolate and oblate shapes in the fault rocks, with similar but slightly more oblate shapes for the *mélange* unit. The average ellipsoid shapes for both of the more-deformed units fall nearer to the plane-strain line (Fig. 1D). In weakly deformed sedimentary rocks, the AMS can represent both the primary bedding and superimposed tectonic fabrics (Housen et al., 1993; Parés and van der Pluijm, 2002), producing a composite fabric. For these units, the most likely effect of these composite AMS fabrics would be to have the K1 orientation fixed to the intersection of the bedding and cleavage fabrics. The ellipsoid shapes will be less affected, because bedding and foliation are coplanar at LOTG (Fig. 1B Ledeczi et al., 2025).

5.5.2 *Chemistry*

Ledeczi et al. (2025) presented microstructural evidence for prograde, interseismic *mélange* development occurring through soft sediment deformation, boudinage, then pressure solution (Fig. 2A, G, M). Here, we use microscale chemical mapping to compare the degree of scaly fabric development between the three deformational domains in our study area: the coherent rocks, the *mélange*, and the fault rocks. Previous work suggests that while Si is progressively depleted along shear surfaces due to pressure solution, Ti is similarly enriched because it is insoluble and conserved in many geological settings (Ague, 2017; Chen et al., 2024a; Evans and Chester, 1995; Kawabata et al., 2007; Kimura et al., 2013). Fe concentrates along foliation surfaces through alignment and growth of Fe-bearing minerals along those same surfaces (Chen et al., 2024a; Kawabata et al., 2007).

Our EPMA maps of the coherent rocks show little to no scaly fabric development, evidence for pressure solution, or foliation, consistent with previous observations in Ledeczi et al. (2025) (Fig. 2B-F). Fault rocks have more scaly fabric development, but with minimal evidence for pressure solution or alignment of phyllosilicates (Fig. 2H-L). *Mélange* samples have the most well-developed, anastomosing scaly fabrics,

evidence for pressure solution, and foliation development (Fig. 2M-R).

Scanning electron microscope backscattered electron (SEM BSE) images at higher magnifications show that coherent rocks host only rare phyllosilicates that are often unaligned with the fabric (Fig. 2F). In the *mélange*, phyllosilicates are prevalent and aligned with and elongated within foliation planes (Fig. 2L). However, the phyllosilicate grains do not form an interconnected network and are instead isolated. Within localized cataclastic zones, phyllosilicates in fault rocks do not have a visible preferred orientation (Fig. 2R). This is consistent with previous microstructural work in these rocks (Ledeczi et al., 2025).

Point measurements by EPMA allow us to quantify compositional differences between shear surfaces and host material and perform Ti mass balance calculations to estimate volume strains accommodated by shear surface development (Chen et al., 2024a). Local volume strain along a single shear surface from EPMA spot measurements is high and increases from coherent unit to *mélange* to faults (Fig. 3A). Accounting for the scaly fabric area in thin section (Fig. 3B), the total volume loss (Fig. 3C) is highest in the *mélange* ($\sim 17\%$), followed by the faults ($\sim 11\%$) and coherent unit ($\sim 2\%$). Bulk rock chemistry also indicates that Si abundance decreases while Ti and Fe abundance increase with apparent deformation, from coherent unit to faults to *mélange*, respectively (Fig. 3D-F).

We combine our estimates of volume strain with geodynamic constraints from the modern Cascadia subduction zone to calculate strain rates. The maximum temperature for these rocks is $\sim 280^\circ\text{C}$ (Ledeczi et al., 2025), which is consistent with previous estimates from Brandon et al. (1998). Cascadia thermal models (Salmi et al., 2017) indicate the modern plate interface intersects the $\sim 280^\circ\text{C}$ isotherm at ~ 80 km horizontal distance from the trench. At Cascadia, the temperature at the modern trench is $\sim 150^\circ\text{C}$, which is also approximately the minimum temperature needed for pressure solution to occur (Moore et al., 2007). Given a plate rate of 4 cm/year (DeMets et al., 2010) and the current extremely shallow ($< 10^\circ$) plate dip,

the footwall rocks would require ~ 2 m.y. to traverse the ~ 80 km from the trench. We divide the average volume loss percent of each unit in Fig. 3C by 2 m.y., giving strain rates of $2.7\text{e-}15$, $1.7\text{e-}15$, and $2.3\text{e-}16$ for the mélange, fault rocks, and coherent unit, respectively.

Our mass balance calculations estimate the volume strain associated with pressure solution. In calculating strain rates from this, we assume that this local volume strain, paired with redistribution of the mass mobilized in that volume strain, contributes to a bulk shear strain at the scale of the plate boundary shear zone. This assumption is consistent with the result that average AMS ellipsoids lie near the plane-strain line (Fig. 1D). While it is a simplifying assumption, this framework provides an upper bound on the simple shear strain rate required to accommodate plate motion through an active shear zone via pressure solution.

This calculation is also sensitive to shear zone thickness. It is likely that the thickness of the shear zone accumulates over time (Rowe et al., 2013), so we consider reasonable minimum (250 m) and maximum (1000 m, double the thickness of the modern LOTG exposure) shear zone widths for the simultaneously active portion. For 250 and 1000 m wide shear zones, strain rates of $5.0\text{e-}12$ and $1.3\text{e-}12$, respectively, are necessary to accommodate the expected plate motion. Both of these values are three orders of magnitude faster than the strain rates that we estimate could be accommodated by pressure solution.

5.5.3 *Constitutive relations and rheology*

We applied constitutive relations, including both frictional relations and flow laws, to assess the slip behaviors these rocks may have hosted. We first use the Fisher and Hirth (2024) pressure solution flow law as an end-member test to evaluate the differential stress required to accommodate all deformation via pressure solution. Fisher and Hirth (2024) developed the flow law specifically for Cascadia, and here, we further constrain key parameters such as shear zone thickness and grain size using observa-

tions from LOTG, the exhumed equivalent of the Cascadia subduction interface. As described above, modern Cascadia plate rates (DeMets et al., 2010), produce shear zone strain rates on the order of $e-12$

The strength of pressure solution, which is a diffusive mechanism, strongly depends on grain size, which is assumed to represent the diffusion length scale (Fisher and Hirth, 2024, d). Based on our microstructural observations (Fig. 2) and in previous work (Ledeczi et al., 2025), grain sizes ranging between 5 and 15 μm are most representative for these rocks, with 10 μm as a reasonable average. To accommodate all plate motion deformation via pressure solution (Fisher and Hirth, 2024) at a 10 μm grain size and maximum LOTG temperatures ($\sim 280^\circ\text{C}$, Ledeczi et al., 2025), leaving no slip deficit would require >2000 MPa of differential stress or >1000 MPa equivalent shear stress (Fig. 4C). Pressure solution is also sensitive to temperature, requiring higher stresses to drive the same strain rates at lower temperatures, so these stress estimates represent minimum values given by the highest T conditions recorded by these rocks. Furthermore, the differential stress and equivalent shear stress are reduced to ~ 250 MPa and >100 MPa, respectively, when assuming the minimum grain size of 5 μm . Reasonable stresses during megathrust earthquakes are lower than needed to accommodate the full plate rate via pressure solution (Fig. 4C).

Pore pressure (P_f) along the seismogenic décollement exerts a key influence on fault mechanics, but its value is hard to determine quantitatively. Taken together, arguments from fault mechanics and rheological modeling, coupled with direct borehole data, seismic velocity analysis, and geologic observations all suggest the seismogenic zone is typically only moderately overpressured but flanked by regions of near-lithostatic P_f in the updip and downdip aseismic regions (Kitajima and Saffer, 2012; Saffer and Tobin, 2011; Tobin and Saffer, 2009). LOTG rocks are also relatively vein-poor (Chen et al., 2024b; Ledeczi et al., 2025) compared to other paleosubduction complexes at similar temperatures (e.g., Chen et al., 2024a; Rowe et al., 2009). Therefore, we select $\lambda = 0.7$ as a plausible likely value of the pore pressure at the

plate interface, whereas near-lithostatic Pf values are likely not appropriate for this setting.

We plot interseismic frictional sliding on phyllosilicates as $\mu = 0.3$ to approximate the relation used in Fagereng and den Hartog (2016) for 0 to 300°C (Fig. 4B). This shows that, at all reasonable strain rates and pore pressures, frictional sliding on phyllosilicates is easier to activate than accommodating the plate rate via the Fisher and Hirth (2024) flow law. However, our microstructural observations reveal a lack of interconnected phyllosilicate layers (Fig. 2L). Additionally, significant deformation of the phyllosilicates in the bulk rock volume would result in much stronger AMS fabrics, indicating that phyllosilicate friction did not accommodate much deformation and is unlikely to control the strength of the mélange. Therefore, the relations for phyllosilicate friction represent a lower limit on the strength of LOTG rocks.

5.6 Discussion

Comparing different methods of quantifying deformation in these rocks helps refine our understanding of the interactions between strain, slip, and volume loss in the plate boundary shear zone. The mélange has the best-developed grain-scale mineral fabrics as demonstrated by both microstructural and AMS observations, with both the faults and coherent units exhibiting weaker development of scaly fabric. The same holds true for estimates of total volume loss, with highest volume loss in the mélange, lower in the fault rocks, and least in the coherent unit (Fig. 3C). Scaly shear surfaces have locally higher degrees of volume loss (Fig. 3A), indicating that volume loss is spatially variable.

The AMS fabrics record the geometry and development of mineral fabrics within the whole rock volume, so differences in fabrics among the three structural units reflect the progressive deformation the rocks experienced. The AMS ellipsoid shapes for the deformed rocks are also influenced by initially-developed weakly oblate fabrics of the coherent units. The trend to proportionately more prolate shapes and higher

degrees of anisotropy in the fault and mélange units is consistent with progressive tectonic fabric development primarily via mineral realignment and new mineral growth (Hirt et al., 2004; Parés, 2004) (Hirt et al., 2004; Parés, 2004). Volume loss via coaxial shortening leads to more oblate ellipsoids on a Flinn-type diagram (Housen and Kanamatsu, 2003; Parés, 2004; Ramsay and Wood, 1973).

The average AMS ellipsoid shapes fall slightly within the oblate region near the plane-strain line of the Flinn plot (Fig. 1D and S3), while the ellipsoid shapes from the individual samples are variable between sites and also within sites in some cases (Fig S3). On average, this suggests that the AMS results are consistent with deformation under constant-volume strain conditions with only minor volume loss. However, the variations in fabric intensity (Pj) and ellipsoid shape likely reflect local variations of mineral alignment and relatively minor volume loss in individual samples. The relatively low Pj values suggest weak penetrative strain in the bulk of these rocks. Taken together, this is consistent with initially oblate fabrics being modified by tectonic deformation towards more prolate fabrics falling near the plane-strain line (Fig. 1D).

5.6.1 *The nature of scaly fabric*

Scaly fabric can form at many different conditions through multiple geological mechanisms. Active accretionary prisms accessed via drilling commonly host a thin zone of scaly fabric at the megathrust interface (Fagereng et al., 2019; Kirkpatrick et al., 2026, 2015; Labaume et al.; Lundberg and Moore, 1986; Prior and Behrmann, 1990; Vannucchi, 2019), though coseismic slip is usually thought to occur on discrete surfaces within or at the boundaries of the scaly volume (Tobin et al., 2001). These observations mean that scaly fabric can form without pressure solution at all before its onset at 150°C (Moore et al., 2007b).

In lithified rocks within the seismogenic zone of the plate interface, a thicker zone of scaly mélange surrounds the megathrust interface (Rowe et al., 2013). At these

conditions, scaly fabric is typically interpreted as forming mainly through pressure solution and sliding on phyllosilicates along shear surfaces due to distributed simple shear during aseismic creep (Vannucchi, 2019). The relative component of pressure solution versus distributed sliding that contributes to the overall fabric development, however, is not well understood.

Our results here show that at seismogenic zone conditions, pressure solution requires much more stress than frictional sliding on phyllosilicates to accommodate plate strain rates (Fig. 4). However, microstructural observations provide a constraint on the applicability of the frictional constitutive relation as expressed in Fagereng and den Hartog (2016) to the bulk rock. We observe a lack of interconnected phyllosilicate layers (Fig. 2L), implying phyllosilicate sliding is likely limited.

The AMS data corroborates that there is little preferred mineral alignment, and thus penetrative strain, recorded in these fabrics. Vannucchi et al. (2003) review scaly fabrics that are known to form primarily through distributed simple shear versus pressure solution; notably, there are not many examples of dominantly pressure solution-derived scaly fabrics. Fisher and Hirth (2024) propose that scaly fabric forms primarily via dissolution and reprecipitation of silica, without sliding on shear surfaces. Instead, while both pressure solution and sliding on shear surfaces do operate at these conditions, we suggest that the scaly fabric produced by these mechanisms likely records very little strain from either process.

5.6.2 Off-fault pressure solution contributes little strain

As opposed to the more localized EPMA measurements, bulk rock chemistry considers the sample's entire composition, so it can capture pressure solution-driven chemical changes that may not be visible at the EPMA scale, especially, for example, if later cataclasis in fault rocks breaks up shear surfaces. Therefore, the bulk rock chemistry results (Fig. 3D-F) confirm that fault rocks have experienced less pressure solution than *mélange* rocks over their lifetime. This implies the lower Pj , shape anisotropy,

and strain in fault rocks compared to the *mélange* samples is not due to *mélange* development and later cataclasis, but rather the partitioning of strain and dominant deformation mechanisms between these units, with explicitly localized mechanisms operating in the faults and pressure solution and other viscous mechanisms in the *mélange*. Because these rocks formed under seismogenic zone conditions, slip in the faults likely occurs on discrete surfaces which is not recorded by these methods. This result is similar to that found at the toe of the modern Barbados accretionary wedge (Housen et al., 1996), where AMS of drill core samples indicates weakly aligned mineral fabrics in the scaly fabric of the megathrust fault of an active subduction zone.

Our chemistry results (Fig. 2), microstructural observations (Fig. 3), and constitutive relations (Fig. 4) show that, while *mélange* is volumetrically dominant at paleosubduction complexes, it represents a negligibly small component of the net plate displacement at seismogenic depths, which must be accommodated via other mechanisms. Of these mechanisms, we suggest that brittle mechanisms presumed to be coseismic during earthquakes are most likely. While localized, frictional displacement on discrete surfaces and cataclastic fabrics are not in themselves diagnostic of coseismic slip (Rowe and Griffith, 2015, e.g.), studies of the slip budget for highly locked megathrusts show a consensus that the majority of total plate motion is taken up in seismic slip (Avouac, 2015; Pacheco et al., 1993) plus relatively smaller afterslip and slow transients. Based on microstructural evidence discussed in detail in Ledeczi et al. (2025) and guided by the correspondence of the lithology and other features of the LOTG shear zone with the modern strongly-locked Cascadia megathrust offshore, we infer that the brittle-frictional localized structural features we document have formed largely through seismic slip rather than aseismically.

Even in the *mélange*, which we show hosts the most pressure solution, the pressure solution strain rates we calculate are on the order of $e-15$. Meanwhile, the strain rate required to accommodate the plate rate through reasonable shear zone widths is orders of magnitude greater, around $e-12$. The constitutive relations (Fig. 4) we use further

show that pressure solution as the primary method of deformation in the seismogenic zone is not possible, contrary to the view that pressure solution in *mélange* may accommodate significant amounts of plate displacement within the seismogenic zone (Fagereng and den Hartog, 2016; Raimbourg et al., 2019).

The view that *mélange* fabrics are volumetrically dominant but accommodate little strain, while much strain is localized along thin cataclastic fault zones within it, is consistent with work from other paleosubduction complexes (Fisher et al., 2021a; Rowe et al., 2005, 2013; Ujiie et al., 2007, e.g.,). For example, AMS studies conducted in the Shimanto belt also found K1 axes parallel to the transport direction and noted an increase of the degree of anisotropy with increasing P-T conditions in the *mélange* (Kitamura and Kimura, 2012), consistent with the view that the efficiency of pressure solution increases with temperature (Fisher and Hirth, 2024). Chen et al. (2024a) find that total volume loss in Kodiak and Shimanto *mélanges* of various P-T conditions are all $\leq 10\%$.

Locking models are key to understand the slip rate deficit distribution and likelihood of future megathrust earthquakes in present-day subduction zones. These models often assume that all deformation is accommodated on the megathrust fault, and that wall rock deformation is fully recoverable, or that permanent plastic deformation around the fault is negligible relative to the elastic rebound cycle (e.g., Li et al., 2018; Lindsey et al., 2021). Our results here indicate that pressure solution, a significant mechanism accommodating interseismic deformation of the plate interface, is limited at seismogenic depths at the Cascadia subduction zone, a result that lends support to the validity of such assumptions. Furthermore, we document that on-fault aseismic pressure solution is less significant than in the surrounding rocks, suggesting that the Cascadia megathrust at these depths may have been interseismically locked in the past. This is consistent with models which predict near-full locking of the seismogenic zone at the predicted correlative maximum depths and location of LOTG in the modern Cascadia subduction zone (e.g., Li et al., 2018; Lindsey et al., 2021).

5.7 *Conclusions and implications for the seismic cycle*

Measurements of strain in the exhumed rock record are a challenge to conduct. We applied a coupled chemical and magnetic fabric approach to reveal how the accommodation of strain via pressure solution varies both spatially and structurally across the diversities of rocks exposed along the megathrust interface within the seismogenic zone for a well-exposed example from Cascadia. Our work shows that interseismically, frictional sliding and pressure solution on shear surfaces produces scaly fabrics distributed throughout the wall rock of the plate interface, while the faults within this volume do not deform in this manner. Pressure solution during this time is not volumetrically significant, and is minor in comparison to displacement during other processes, such as megathrust earthquakes. During the coseismic period, slip localizes on mm-scale principal slip surfaces within fault strands while the surrounding *mélange* does not deform.

This work suggests that for exhumed rocks from seismogenic depths, *mélange* rocks may be volumetrically dominant and most visible in the field, yet may record little deformation associated with accommodating relative plate motions (Fig. 5). The apparently large strains required to disrupt stratigraphy and produce *mélange* fabrics still reflect strain rates that are minuscule relative to strains and the strain rates necessitated by plate motions. Significant plate-displacement deformation is instead recorded by thin, localized zones of fault rocks such as cataclasites, consistent with the view of Fisher and Hirth (2024). Pressure solution cannot accommodate enough strain to contribute significantly to plate motion at this setting, contrary to previous suggestions (Fagereng and den Hartog, 2016; Raimbourg et al., 2019). Finally, this work lends support to extant assumptions for the amount and mechanism of off-fault deformation in present-day locking models of subduction zones.

Acknowledgements

We gratefully acknowledge Olympic National Park, Matt Dubeau, and Bill Baccus for permitting this research. We thank Sophie Johnson, Anette von der Handt, Juliet Crider, and Nick Roberts for assistance in collecting and understanding the data. We thank Josh Davis and Sarah Titus and the geologyGeometry library in R <http://www.joshuadavis.us/software/>. Part of this work was conducted at the Molecular Analysis Facility, a National Nanotechnology Coordinated Infrastructure site at the University of Washington, which is supported in part by the National Science Foundation (grant NNCI-1542101), the University of Washington, the Molecular Engineering & Sciences Institute, the Clean Energy Institute, and the National Institutes of Health. We gratefully acknowledge support from the Paros Endowed Chair in Seismology and Geohazards to H. Tobin.

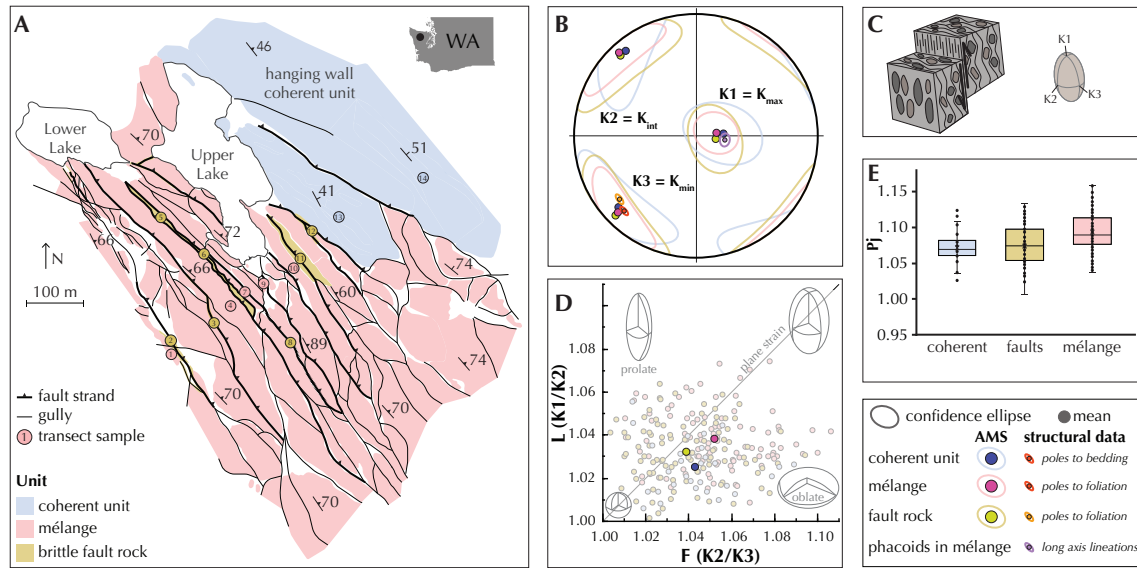


Figure 5.1: **A.** Map of Lakes of the Gods (LOTG) adapted from (Ledeczi et al., 2025). **B.** Stereoplot of mean axes and confidence ellipses for structural (Davis and Titus, 2017; Roberts et al., 2019) and AMS (Jelínek, 1978) data. Structural data includes foliation poles in mélangé outcrops and bedding poles in the coherent unit; foliation poles in fault strands; and lineations of phacoid long axes within foliation planes in mélangé. This can be compared to AMS data from the corresponding units. **C.** Cartoon showing how AMS ellipsoid axes relate to field structural orientations. **D.** Flinn plot of AMS data from all specimens (transparent) and arithmetic mean of all values for each unit (bold). **E.** The degree of anisotropy Pj (Jelinek, 1981) for each unit. Whiskers represent minimum and maximum dataset values or 1.5x the interquartile range where outliers exist.

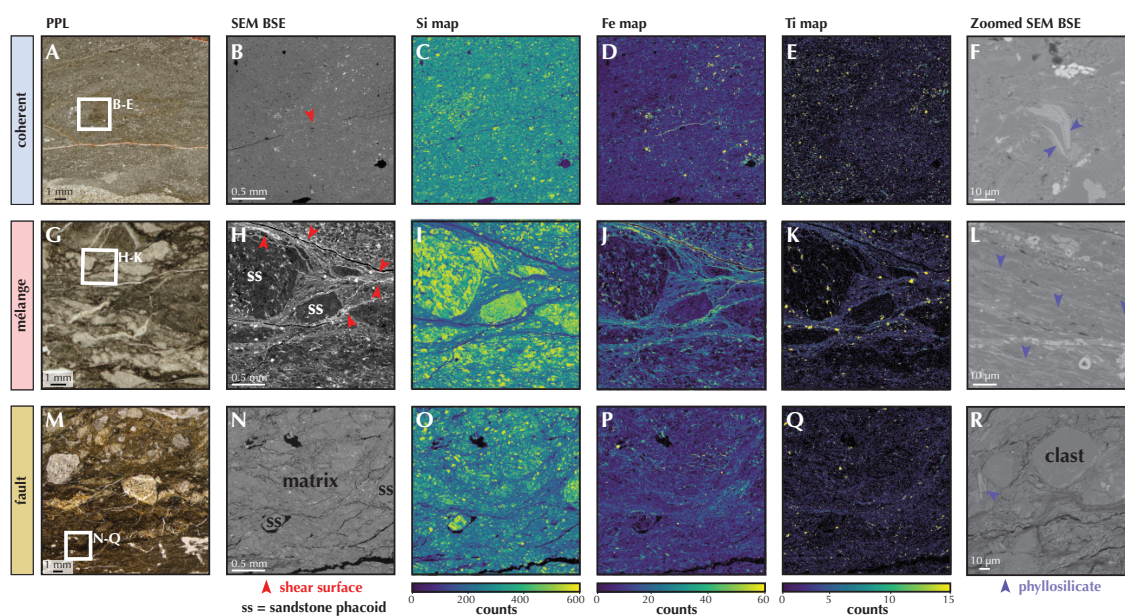


Figure 5.2: Typical microstructures and EPMA maps of coherent (row 1), mélange (row 2), and fault (row 3) rocks. Column 1 are plane polarized light (PPL) images. Columns 2-5 are SEM BSE images and Si, Fe, and Ti maps of the same area. Column 6 shows higher-magnification SEM BSE images. Bedding/foliation is horizontal in all images.

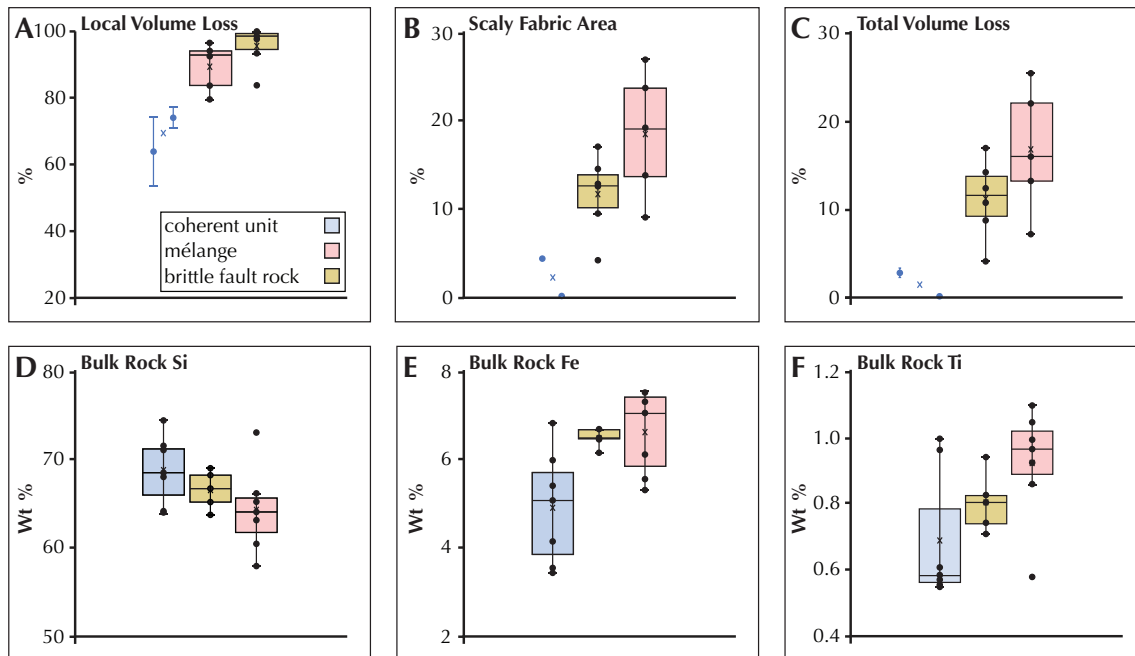


Figure 5.3: **A-C.** report volume loss calculated using elemental maps. Each box and whisker plot represents the combined results from all samples of each unit along the transect. **A.** Average volume loss associated with individual shear surfaces. **B.** Percent area of scaly fabric in each EPMA map. **C.** Total volume loss in each elemental map area. **D-F.** report bulk rock chemistry of **D.** Si, **E.** Fe, and **F.** Ti for samples throughout field area. Whiskers represent minimum and maximum dataset values or 1.5x the interquartile range where outliers exist.

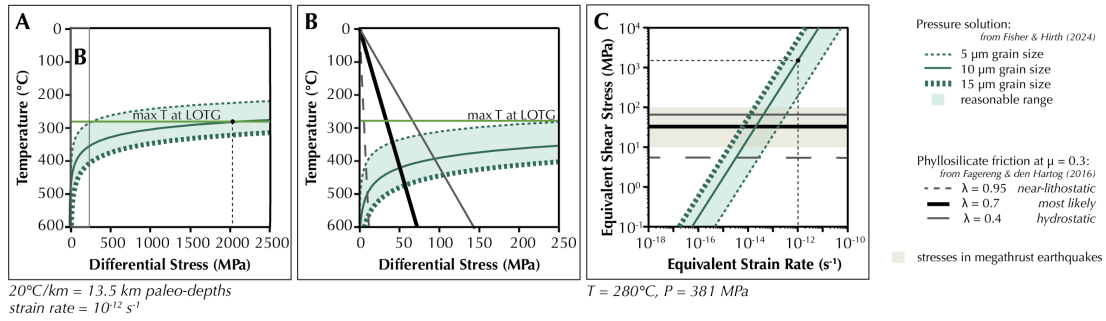


Figure 5.4: **A.** Differential stress required to reach plate motion strain rates using the Fisher and Hirth (2024) pressure solution flow law for grain sizes between 5 and 15 μm . Following Fisher and Hirth (2024), we use $Q = 50 \text{ kJ/mol}$ and $A = 1.5e - 6$. Maximum T at LOTG=280°C from Ledeczi et al. (2025) and additional RSCM herein (Fig. S1). **B.** Crustal yield strength profile for interseismic frictional sliding on phyllosilicates from Fagereng and den Hartog (2016). Frictional sliding on phyllosilicates is weaker than pressure solution at all λ . **C.** Differential stress is converted to equivalent shear stress following Paterson and Olgaard (2000). Equivalent strain rate versus equivalent shear stress for pressure solution and phyllosilicate friction for $T = 270^\circ\text{C}$ and $P = 381 \text{ MPa}$ using 2680 kg/m^3 and 13.5 km depth (Brandon et al., 1998). Tan region represents reasonable stresses during megathrust earthquake slip.

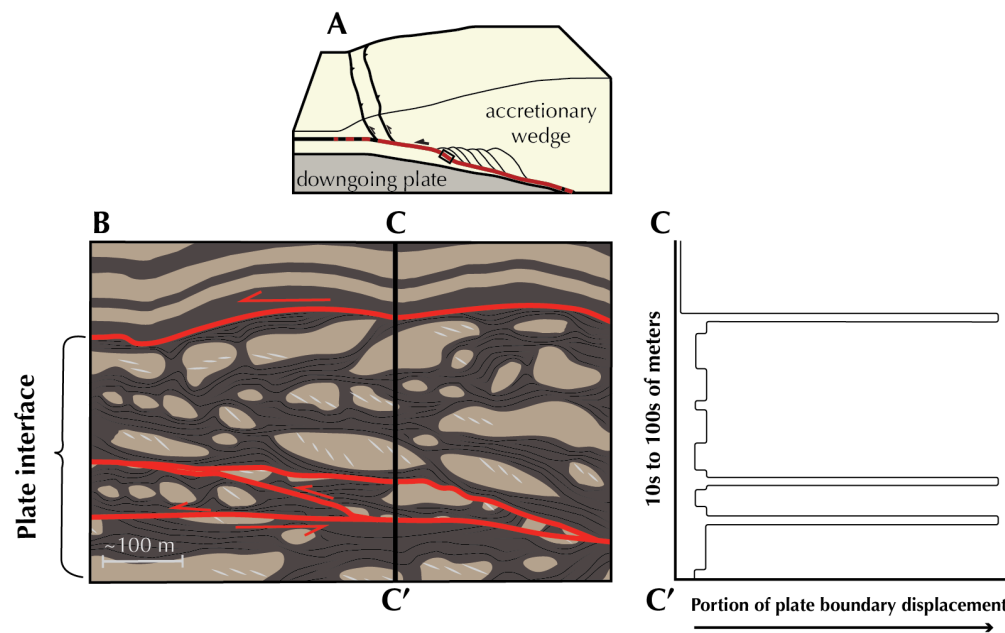


Figure 5.5: **A.** Generalized subduction zone with location of A in box. **B.** Schematic of the plate interface with coseismic slip localized on red faults contained in the mélangé host rock. Coherent units sit atop the plate interface. **C.** Portion of plate boundary displacement accommodated along a transect of the plate interface shear zone, with most deformation concentrated in faults. Within the mélangé, the matrix records more strain than the competent phacoids. Coherent units record relatively low strain.

Chapter 6

CONCLUSIONS

The work in this Dissertation uses a wide-ranging, interdisciplinary set of techniques to better constrain accretionary wedge deformation and megathrust earthquake processes at the Cascadia subduction zone. Here, I summarize these findings and discuss their implications for future work.

In *Chapter 2*, I use multiple resolutions of seismic reflection imaging to identify the most recently active splay faults in the landward vergence zone and explain the implications of these findings for megathrust earthquakes. I identify a belt of active faults located within ~ 30 km of the deformation front that show evidence for recent activity and suggest they slip concurrently with the shallow megathrust during megathrust earthquakes (Ledeczi et al., 2024); meanwhile, faults further landward are covered by thick sediments (Lucas et al., 2025).

This work raises questions about splay fault activity at sediment-rich subduction zones more globally. It seems that sediment-rich subduction zones such as Hikurangi (e.g., Coffey et al., 2021), Nankai (e.g., Lackey et al., 2020; Schottenfels and Regalla, 2021), Sumatra, and Makran also show evidence for a system of active splay faults in the outermost wedge rather than a singular megasplay-style fault at the backstop. For example, widespread faulting, folding, and mass wasting of the accretionary wedge following the 2004 Mw9.1 Sumatra earthquake was attributed to shallow coseismic megathrust rupture and activation of the frontal splay fault (Gulick et al., 2011; Henstock et al., 2006; Ishii et al., 2005; Seeber et al., 2007). Makran shows a similar ‘belt’ of active faults within tens of km of the deformation front, with a zone of inactivity further landward (Smith et al., 2012). These observations are also consistent with

both theoretical (Dahlen et al., 1984; Davis et al., 1983) and numerical (Mannu et al., 2016; Fuller et al., 2006; Simpson, 2010) models of accretionary wedges. However, this mode of splay fault activity has not been fully investigated at all sediment-rich subduction zones (e.g., Alaska; von Huene et al., 2021).

In *Chapter 3*, I apply structural geological techniques to newly-collected crustal-scale seismic reflection data to settle a debate about the position of the megathrust fault, specifically whether it localizes at the top of crust or within the sedimentary column. My work suggests that within the landward vergence zone’s active outer wedge, the megathrust fault is on average ~ 500 m above, and at no point coincident with, the top of crust. This newly refined décollement surface produces a shallow décollement dip, implying a weak basal fault below a strong wedge based on critical wedge theory (Dahlen et al., 1984; Davis et al., 1983), which is consistent with other geophysical observations (e.g., Webb, 2017; Han et al., 2017). I then review seismic reflection data from other subduction zones, which suggests that the megathrust fault usually prefers to localize within the sediments, rather than at the rough sediment-basement interface, when there are adequate sediments available. These new results suggest that dynamic rupture and other numerical earthquake models should account for shallow, seismogenic megathrust as being hosted within the sediments, which influence both the fault zone friction and the wall rock elasticity, in turn fundamentally impacting rupture characteristics.

In *Chapter 4* and *Chapter 5*, we focus on an exhumed paleomegathrust fault in the Olympic Subduction Complex, the first of its kind identified for the Cascadia subduction zone. We characterize its structure and properties, and present evidence that it is a paleomegathrust, in *Chapter 4*. This work shows that the Cascadia megathrust interface was also localized entirely within the sediments in the past (Ledeczi et al., 2025). In *Chapter 5*, I then investigate how strain partitions spatially across the paleomegathrust interface, quantifying how interseismic strain varies between discrete faults, the surrounding mélange, and the coherent unit above the fault. I find that

the *mélange* records very little strain—about 0.1% of plate boundary displacement. This indicates that the majority of strain is localized along faults in rocks that are volumetrically minor in the rock record.

Based on modern geophysical observations, we can estimate that a plate interface exposure like LOTG should have accommodated more than 10,000 large earthquakes during its lifetime. However, what constitutes geological evidence of coseismic slip has been highly debated for decades (Cowan, 1999; Rowe & Griffith, 2015). While frictionally melted rock, or pseudotachylyte, is considered the evidentiary gold standard, it is rarely preserved—or potentially rarely generated, with other kinds of fault rocks perhaps more commonly representing fault slip (Kirkpatrick et al., 2009). The missing strain signal we identify in *Chapter 5* specifically calls into question strict standards for fault rocks and how we can improve the identification of earthquakes signatures in the rock record.

BIBLIOGRAPHY

- Adam J., Klaeschen D., Kukowski N. and Flueh E. (2004) Upward delamination of cascadia basin sediment infill with landward frontal accretion thrusting caused by rapid glacial age material flux. *Tectonics*, **23**.
- Ague J.J. (1991) Evidence for major mass transfer and volume strain during regional metamorphism of pelites. *Geology*, **19**, 855.
- Ague J.J. (1994) Mass transfer during barrovian metamorphism of pelites, south-central connecticut; i, evidence for changes in composition and volume. *American Journal of Science*, **294**, 989–1057.
- Ague J.J. (2017) Element mobility during regional metamorphism in crustal and subduction zone environments with a focus on the rare earth elements (REE). *American Mineralogist*, **102**, 1796–1821.
- Aitchison J. (1992) On criteria for measures of compositional difference. *Mathematical Geology*, **24**, 365–379.
- Aoya M., Kouketsu Y., Endo S., Shimizu H., Mizukami T., Nakamura D. and Wallis S. (2010) Extending the applicability of the Raman carbonaceous-material geothermometer using data from contact metamorphic rocks. *Journal of Metamorphic Geology*, **28**, 895–914.
- Appelgate B., Goldfinger C., MacKay M.E., Kulm L.D., Fox C.G., Embley R.W. and Meis P.J. (1992) A left-lateral strike-slip fault seaward of the Oregon convergent margin. *Tectonics*, **11**, 465–477.

- Aslam K.S., Thomas A.M. and Melgar D. (2021) The effect of fore-arc deformation on shallow earthquake rupture behavior in the Cascadia Subduction Zone. *Geophysical Research Letters*, **48**.
- Atwater B.F. and Hemphill-Haley E. (1997) Recurrence intervals for great earthquakes of the past 3,500 years at northeastern Willapa Bay, Washington.
- Avouac J.P. (2015) From geodetic imaging of seismic and aseismic fault slip to dynamic modeling of the seismic cycle. *Annual Review of Earth and Planetary Sciences*, **43**, 233–271.
- Ball S. (2019) Compaction-constrained structural restoration of the outer wedge in the Cascadia subduction zone, M.S.]. University of Wisconsin, Madison.
- Balster-Gee A.F., Miller N.C., Watt J.T., Roland E.C., Kluesner J.W., Heller S.J., Hart P.E., Sliter R.W., Myers E.K., Wyland R.M., Marcuson R.K., Johnson C., Nichols A.R., Pszczola K. and Williams C. (2023) High-resolution multichannel sparker seismic-reflection and chirp sub-bottom data acquired along the Cascadia margin during USGS field activity 2019-024-FA.
- Bangs N., Moore G., Gulick S., Pangborn E., Tobin H., Kuramoto S. and Taira A. (2009) Broad, weak regions of the nankai megathrust and implications for shallow coseismic slip. *Earth and Planetary Science Letters*, **284**.
- Bangs N.L., McIntosh K.D., Silver E.A., Kluesner J.W. and Ranero C.R. (2015) Fluid accumulation along the costa rica subduction thrust and development of the seismogenic zone. *Journal of Geophysical Research: Solid Earth*, **120**, 67–86.
- Barnard W.D. (1978) The Washington continental slope: Quaternary tectonics and sedimentation. *Marine Geology*, **27**, 79–114.
- Barnes P.M., Lamarche G., Bialas J., Henrys S., Pecher I., Netzeband G.L., Greinert J., Mountjoy J.J., Pedley K. and Crutchley G. (2010) Tectonic and geological

framework for gas hydrates and cold seeps on the hikurangi subduction margin, new zealand. *Marine Geology*, **272**.

Batt G.E., Brandon M.T., Farley K.A. and Roden-Tice M. (2001) Tectonic synthesis of the Olympic Mountains segment of the Cascadia wedge, using two-dimensional thermal and kinematic modeling of thermochronological ages. *Journal of Geophysical Research: Solid Earth*, **106**, 26731–26746.

Beeson J., Dartnell P. and Watt J.T. (2025) Composite multibeam bathymetry surface of the northern cascadia margin offshore washington.

Beeson J.W., Goldfinger C. and Fortin W.F. (2017) Large-scale modification of submarine geomorphic features on the Cascadia accretionary wedge caused by catastrophic flooding events. *Geosphere*, **13**, 1713–1728.

Behr W.M. and Bürgmann R. (2021) What’s down there? the structures, materials and environment of deep-seated slow slip and tremor. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, **379**.

Bestmann M., Pennacchioni G., Nielsen S., Göken M. and De Wall H. (2012) Deformation and ultrafine dynamic recrystallization of quartz in pseudotachylite-bearing brittle faults: A matter of a few seconds. *Journal of Structural Geology*, **38**, 21–38.

Bettelli G., Vannucchi P. et al. (2003) Structural style of the offscraped Ligurian oceanic sequences (Early and Late Cretaceous) of the Northern Apennines: new hypothesis concerning the development of mélange block-in-matrix fabric. *Journal of Structural Geology*, **25**, 373–390.

Beyssac O., Goffé B., Chopin C. and Rouzaud J. (2002) Raman spectra of carbonaceous material in metasediments: a new geothermometer. *Journal of metamorphic Geology*, **20**, 859–871.

- Biemiller J., Gabriel A., Staisch L., Ulrich T., Dunham A., Wirth E., Watt J., Lucas M.C. and Ledeczi A. (2025) Structural controls on splay fault rupture dynamics during cascadia megathrust earthquakes. *AGU Advances*, **6**.
- Bilotti F. and Shaw J.H. (2005) Deep-water Niger Delta fold and thrust belt modeled as a critical-taper wedge: The influence of elevated basal fluid pressure on structural styles. *AAPG Bulletin*, **89**, 1475–1491.
- Blair J., McCrory P., Oppenheimer D. and Waldhauser F. (2011) A Geo-referenced 3D model of the Juan de Fuca Slab and associated seismicity.
- Bletery Q., Thomas A.M., Rempel A.W., Karlstrom L., Sladen A. and De Barros L. (2016) Mega-earthquakes rupture flat megathrusts. *Science*, **354**, 1027–1031.
- Booth D.B., Troost K.G., Clague J.J. and Waitt R.B. (2003) The Cordilleran Ice Sheet, Elsevier, p. 17–43.
- Booth-Rea G., Klaeschen D., Grevemeyer I. and Reston T. (2008) Heterogeneous deformation in the Cascadia convergent margin and its relation to thermal gradient. *Tectonics*, **27**.
- Borradaile G. and Henry B. (1997) Tectonic applications of magnetic susceptibility and its anisotropy. *Earth-Science Reviews*, **42**, 49–93.
- Borradaile G., Keeler W., Alford C. and Sarvas P. (1987) Anisotropy of magnetic susceptibility of some metamorphic minerals. *Physics of the Earth and Planetary Interiors*, **48**, 161–166.
- Bradley D.C., Kusky T.M., Haeussler P.J., Goldfarb R.J., Miller M.L., Dumoulin J.A., Nelson S.W. and Karl S.M. (2003) Geology of a transpressional orogen developed during ridge-trench interaction along the North Pacific margin, chap. Geologic signature of early Tertiary ridge subduction in Alaska, Geological Society of America.

- Bradley K., Qin Y., Carton H., Hananto N., Villanueva-Robles F., Leclerc F., Shengji W., Tapponier P., Sieh K. and Singh S. (2019) Stratigraphic control of frontal décollement level and structural vergence and implications for tsunamigenic earthquake hazard in sumatra, indonesia. *Geochemistry, Geophysics, Geosystems*, **20**, 1646–1664.
- Brandes C. and Tanner D.C. (2014) Fault-related folding: A review of kinematic models and their application. *Earth-Science Reviews*, **138**, 352–370.
- Brandon M.T. and Calderwood A.R. (1990) High-pressure metamorphism and uplift of the Olympic subduction complex. *Geology*, **18**, 1252–1255.
- Brandon M.T., Roden-Tice M.K. and Garver J.I. (1998) Late Cenozoic exhumation of the Cascadia accretionary wedge in the Olympic Mountains, northwest Washington State. *Geological Society of America Bulletin*, **110**, 985–1009.
- Brandon M.T., Vance J.A. et al. (1992) Tectonic evolution of the Cenozoic Olympic subduction complex, Washington State, as deduced from fission track ages for detrital zircons. *American Journal of Science*, **292**, 565–565.
- Brooks B.A., Goldberg D., DeSanto J., Ericksen T.L., Webb S.C., Nooner S.L., Chadwell C.D., Foster J., Minson S., Witter R., Haeussler P., Freymueller J., Barnhart W. and Nevitt J. (2023) Rapid shallow megathrust afterslip from the 2021 M8.2 Chignik, Alaska earthquake revealed by seafloor geodesy. *Science Advances*, **9**.
- Burg J.P. (2018) Geology of the onshore makran accretionary wedge: Synthesis and tectonic interpretation. *Earth-Science Reviews*, **185**, 1210–1231.
- Byrne D.E., Davis D.M. and Sykes L.R. (1988) Loci and maximum size of thrust earthquakes and the mechanics of the shallow region of subduction zones. *Tectonics*, **7**, 833–857.

- Byrne T. (1984) Early deformation in mélangé terranes of the Ghost Rocks Formation, Kodiak islands, Alaska, Geological Society of America.
- Byrne T. and Fisher D. (1987) Episodic growth of the Kodiak convergent margin. *Nature*, **325**, 338–341.
- Calvert A.J., Preston L.A. and Farahbod A.M. (2011) Sedimentary underplating at the Cascadia mantle-wedge corner revealed by seismic imaging. *Nature Geoscience*, **4**, 545–548.
- Carbotte S., Han S., Boston B. and Canales J. (2023) Processed pre-stack depth-migrated seismic reflection data from the 2021 CASIE21 multi-channel seismic survey (MGL2104).
- Carbotte S.M., Boston B., Han S., Shuck B., Beeson J., Canales J.P., Tobin H., Miller N., Nedimovic M., Tréhu A. et al. (2024) Subducting plate structure and megathrust morphology from deep seismic imaging linked to earthquake rupture segmentation at Cascadia. *Science Advances*, **10**, eadl3198.
- Caulet J. (1995) Proceedings of the Ocean Drilling Program, 146 Part 1 Scientific Results, Ocean Drilling Program.
- Chalumeau C., Agurto-Detzel H., Rietbrock A., Frietsch M., Oncken O., Segovia M. and Galve A. (2024) Seismological evidence for a multifault network at the subduction interface. *Nature*, **628**, 558–562.
- Chen T., Smye A., Fisher D., Hashimoto Y., Raimbourg H. and Famin V. (2024a) Quantifying interseismic volume strain from chemical mass-balance analysis in tectonic mélanges. *Geochemistry, Geophysics, Geosystems*, **25**.
- Chen T.W., Tobin H. and Ledeczi A. (2024b) Veins in the Olympic Subduction Complex: Implications for the stress field in the paleo-Cascadia accretionary wedge, in: Geological Society of America Abstracts, vol. 56, p. 402176.

- Chiama K., Chauvin B., Plesch A., Moss R. and Shaw J.H. (2023) Geomechanical modeling of ground surface deformation associated with thrust and reverse-fault earthquakes: A distinct element approach. *Bulletin of the Seismological Society of America*, **113**, 1702–1723.
- Clowes R.M., Brandon M.T., Green A., Yorath C., Brown A.S., Kanasewich E. and Spencer C. (1987) LITHOPROBE—southern Vancouver Island: Cenozoic subduction complex imaged by deep seismic reflections. *Canadian Journal of Earth Sciences*, **24**, 31–51.
- Coffey G.L., Savage H.M., Polissar P.J., Meneghini F., Ikari M.J., Fagereng Å., Morgan J.K. and Wang M. (2021) Evidence of seismic slip on a large splay fault in the hikurangi subduction zone. *Geochemistry, Geophysics, Geosystems*, **22**, e2021GC009638.
- Condit C.B., French M.E., Hayles J.A., Yeung L.Y., Chin E.J. and Lee C.A. (2022) Rheology of metasedimentary rocks at the base of the subduction seismogenic zone. *Geochemistry, Geophysics, Geosystems*, **23**.
- Condit C.B., Poulaki E.M., Lindquist P.C., Nichols C.I., Ferrell M.E., Odlum M.L., Hoover W.F. and Guevara V.E. (2025a) Fluid-mediated deformation leads to weakening, strengthening, and block-in-matrix structures during prograde subduction mélange formation. *Geology*, **53**, 869–874.
- Condit C.B., Poulaki E.M., Lindquist P.C., Nichols C.I., Ferrell M.E., Odlum M.L., Hoover W.F. and Guevara V.E. (2025b) Fluid-mediated deformation leads to weakening, strengthening, and block-in-matrix structures during prograde subduction mélange formation. *Geology*, **53**, 869–874.
- Corredor F., Shaw J.H. and Bilotti F. (2005) Structural styles in the deep-water fold and thrust belts of the niger delta. *AAPG Bulletin*, **89**.

- Cortés-Rivas V., Shillington D.J., Lizarralde D. and Mark H. (2025) Interactions between megathrust behavior and forearc deformation in the andean segment of the aleutian subduction zone, offshore alaska, usa. *Geology*, **53**, 301–305.
- Cowan D.S. (1985) Structural styles in Mesozoic and Cenozoic mélanges in the western Cordillera of North America. *Geological Society of America Bulletin*, **96**, 451–462.
- Cowan D.S. (1999) Do faults preserve a record of seismic slip? a field geologist's opinion. *Journal of Structural Geology*, **21**, 995–1001.
- Cowan D.S. and Brandon M.T. (1994) A symmetry-based method for kinematic analysis of large-slip brittle fault zones. *American Journal of Science*, **294**.
- Crameri F. (2023) Scientific colour maps.
- Crameri F., Shephard G.E. and Heron P.J. (2020) The misuse of colour in science communication. *Nature Communications*, **11**.
- Crosson R. and Owens T. (1987) Slab geometry of the Cascadia subduction zone beneath Washington from earthquake hypocenters and teleseismic converted waves. *Geophysical Research Letters*, **14**, 824–827.
- Cubas N., Souloumiac P. and Singh S.C. (2016) Relationship link between landward vergence in accretionary prisms and tsunami generation. *Geology*, **44**, 787–790.
- Dahlen F.A., Suppe J. and Davis D. (1984) Mechanics of fold-and-thrust belts and accretionary wedges: Cohesive coulomb theory. *Journal of Geophysical Research: Solid Earth*, **89**, 10087–10101.
- Dartnell P., Watt J.T., Hill J.C. and Conrad J.E. (2023) Composite multibeam bathymetry surface and data sources of the central cascadia margin offshore oregon.
- Davis D., Suppe J. and Dahlen F.A. (1983) Mechanics of fold-and-thrust belts and accretionary wedges. *Journal of Geophysical Research: Solid Earth*, **88**, 1153–1172.

- Davis J.R. and Titus S.J. (2017) Modern methods of analysis for three-dimensional orientational data. *Journal of Structural Geology*, **96**, 65–89.
- Daëron M., Avouac J. and Charreau J. (2007) Modeling the shortening history of a fault tip fold using structural and geomorphic records of deformation. *Journal of Geophysical Research: Solid Earth*, **112**.
- Dean S.M., McNeill L.C., Henstock T.J., Bull J.M., Gulick S.P.S., Austin J.A., Bangs N.L.B., Djajadihardja Y.S. and Permana H. (2010) Contrasting décollement and prism properties over the sumatra 2004–2005 earthquake rupture boundary. *Science*, **329**.
- DeCelles P.G., Gray M.B., Ridgway K.D., Cole R.B., Pivnik D.A., Pequera N. and Srivastava P. (1991) Controls on synorogenic alluvial-fan architecture, beartooth conglomerate (palaeocene), wyoming and montana. *Sedimentology*, **38**, 567–590.
- Delph J.R., Thomas A.M. and Levander A. (2021) Subcretionary tectonics: Linking variability in the expression of subduction along the Cascadia forearc. *Earth and Planetary Science Letters*, **556**, 116724.
- DeMets C., Gordon R.G. and Argus D.F. (2010) Geologically current plate motions. *Geophysical Journal International*, **181**, 1–80.
- DePaolis J.M., Dura T., Witter R.C., Haeussler P.J., Bender A., Curran J.H. and Corbett D.R. (2024) Repeated coseismic uplift of coastal lagoons above the patton bay splay fault system, montague island, alaska, usa. *Journal of Geophysical Research: Solid Earth*, **129**.
- DeSanto J.B., Schmidt D.A., Zumberge M., Sasagawa G. and Chadwell C.D. (2025) Near full locking on the shallow megathrust of the central cascadia subduction zone revealed by gnss-acoustic. *Earth and Planetary Science Letters*, **665**.

- Di Toro G., Han R., Hirose T., De Paola N., Nielsen S., Mizoguchi K., Ferri F., Cocco M. and Shimamoto T. (2011a) Fault lubrication during earthquakes. *Nature*, **471**, 494–498.
- Di Toro G., Han R., Hirose T., De Paola N., Nielsen S., Mizoguchi K., Ferri F., Cocco M. and Shimamoto T. (2011b) Fault lubrication during earthquakes. *Nature*, **471**, 494–498.
- Ding X., Xu S., Xie Y., Van den Ende M., Premus J. and Ampuero J.P. (2023) The sharp turn: Backward rupture branching during the 2023 Mw 7.8 Kahramanmaraş (Türkiye) earthquake. *Seismica*, **2**.
- Donovan J.J., Moy A., von der Handt A., Gainsforth Z., Maner J.L., Nachlas W. and Fournelle J. (2023) a new method for dead time calibration and a new expression for correction of wds intensities for microanalysis. *Microscopy and Microanalysis*, **29**, 1096–1110.
- Doski J.A. and McClay K. (2022) Tectono-stratigraphic evolution, regional structure and fracture patterns of the zagros fold-thrust belt in the duhok region, kurdistan, northern iraq. *Tectonophysics*, **838**.
- Editor (1986) Hyphenation exception log. *TUGboat*, **7**, 145.
- Edwards J.H., Kluesner J.W., Silver E.A., Brodsky E.E., Brothers D.S., Bangs N.L., Kirkpatrick J.D., Wood R. and Okamoto K. (2018) Corrugated megathrust revealed offshore from costa rica. *Nature Geoscience*, **11**, 197–202.
- Eichelberger N.W., Hughes A.N. and Nunns A.G. (2015) Combining multiple quantitative structural analysis techniques to create robust structural interpretations. *Interpretation*, **3**, SAA89–SAA104.
- Epard J.L. and Groshong R.H. (1993) Excess area and depth to detachment. *AAPG Bulletin*, **77**, 1291–1302.

- Evans J.P. and Chester F.M. (1995) Fluid-rock interaction in faults of the San Andreas system: Inferences from San Gabriel fault rock geochemistry and microstructures. *Journal of Geophysical Research: Solid Earth*, **100**, 13007–13020.
- Fagereng and den Hartog S.A.M. (2016) Subduction megathrust creep governed by pressure solution and frictional–viscous flow. *Nature Geoscience*, **10**, 51–57.
- Fagereng, Savage H., Morgan J., Wang M., Meneghini F., Barnes P., Bell R., Kitajima H., McNamara D., Saffer D., Wallace L., Petronotis K. and LeVay L. (2019) Mixed deformation styles observed on a shallow subduction thrust, Hikurangi margin, New Zealand. *Geology*, **47**, 872–876.
- Fagereng and Sibson R.H. (2010) Mélange rheology and seismic style. *Geology*, **38**, 751–754.
- Festa A., Pini G.A., Ogata K. and Dilek Y. (2019) Diagnostic features and field-criteria in recognition of tectonic, sedimentary and diapiric mélanges in orogenic belts and exhumed subduction-accretion complexes. *Gondwana Research*, **74**, 7–30.
- Finley T., Morell K., Leonard L., Regalla C., Johnston S.T. and Zhang W. (2019) Ongoing oroclinal bending in the Cascadia forearc and its relation to concave-outboard plate margin geometry. *Geology*, **47**, 155–158.
- Fisher A., Wheat C., Becker K., Davis E., Jannasch H., Schroeder D., Dixon R., Pettigrew T., Meldrum R., MacDonald R. et al. (2005) Scientific and technical design and deployment of long-term seafloor observatories for hydrogeologic and related experiments, iodp expedition 301, eastern flank of Juan de Fuca ridge, in: *Proc. IODP— Volume*, vol. 301, p. 2.
- Fisher A.T. and Davis E.E. (2000) An introduction to the scientific results of leg 168, in: *Proceedings of the Ocean Drilling Program, Scientific Results*, vol. 168, College Station, TX, USA: Ocean Drilling Program, pp. 3–5.

- Fisher D. and Byrne T. (1987) Structural evolution of underthrust sediments, Kodiak Islands, Alaska. *Tectonics*, **6**, 775–793.
- Fisher D., Mosher D., Austin J.A., Gulick S.P., Masterlark T. and Moran K. (2007a) Active deformation across the sumatran forearc over the december 2004 mw9.2 rupture. *Geology*, **35**, 99.
- Fisher D., Mosher D., Austin Jr J.A., Gulick S.P., Masterlark T. and Moran K. (2007b) Active deformation across the sumatran forearc over the december 2004 mw9. 2 rupture. *Geology*, **35**, 99–102.
- Fisher D.M. and Hirth G. (2024) A pressure solution flow law for the seismogenic zone: Application to Cascadia. *Science Advances*, **10**, eadi7279.
- Fisher D.M., Hooker J.N., Smye A.J. and Chen T.W. (2021a) Insights from the geological record of deformation along the subduction interface at depths of seismogenesis. *Geosphere*, **17**, 1686–1703.
- Fisher D.M., Hooker J.N., Smye A.J. and Chen T.W. (2021b) Insights from the geological record of deformation along the subduction interface at depths of seismogenesis. *Geosphere*, **17**, 1686–1703.
- Fisher D.M., Tonai S., Hashimoto Y., Tomioka N. and Oakley D. (2019) K-Ar dating of fossil seismogenic thrusts in the Shimanto accretionary complex, southwest Japan. *Tectonics*, **38**, 3866–3880.
- Fisher M.A., Flueh E.R., Scholl D.W., Parsons T., Wells R.E., Trehu A., Brink U.t. and Weaver C.S. (1999) Geologic processes of accretion in the Cascadia subduction zone west of Washington State. *Journal of Geodynamics*, **27**, 277–288.
- Flinn D. (1956) On the deformation of the Funzie conglomerate, Fetlar, Shetland. *The Journal of Geology*, **64**, 480–505.

- Flueh E.R., Fisher M.A., Bialas J., Childs J.R., Klaeschen D., Kukowski N., Parsons T., Scholl D.W., ten Brink U., Tréhu A.M. and Vidal N. (1998) New seismic images of the Cascadia subduction zone from cruise SO108 — ORWELL. *Tectonophysics*, **293**, 69–84.
- Ford M., Williams E.A., Artoni A., Vergés J. and Hardy S. (1997) Progressive evolution of a fault-related fold pair from growth strata geometries, sant llorenç de morunys, se pyrenees. *Journal of Structural Geology*, **19**, 413–441.
- Frederik M.C.G., Gulick S.P.S., Austin J.A., Bangs N.L.B. and Udrekx (2015) What 2-D multichannel seismic and multibeam bathymetric data tell us about the North Sumatra wedge structure and coseismic response. *Tectonics*, **34**, 1910–1926.
- Fujiwara T., Kodaira S., No T., Kaiho Y., Takahashi N. and Kaneda Y. (2011) The 2011 Tohoku-oki earthquake: Displacement reaching the trench axis. *Science*, **334**, 1240–1240.
- Fuller C.W., Willett S.D. and Brandon M.T. (2006) Formation of forearc basins and their influence on subduction zone earthquakes. *Geology*, **34**, 65.
- Gao D., Wang K., Insua T.L., Sypus M., Riedel M. and Sun T. (2018) Defining megathrust tsunami source scenarios for northernmost Cascadia. *Natural Hazards*, **94**, 445–469.
- Gerstel W. and Lingley W. (2003) Geologic map of the Mount Olympus 1:100,000 quadrangle.
- Ghisetti F.C., Barnes P.M., Ellis S., Plaza-Faverola A.A. and Barker D.H.N. (2016) The last 2 myr of accretionary wedge construction in the central hikurangi margin (north island, new zealand): Insights from structural modeling. *Geochemistry, Geophysics, Geosystems*, **17**.

- Goldfinger C., Beeson J., Romsos C. and Patton J. (2023) Neotectonic map of the Cascadia margin, (open-file report o-23-05)., Oregon Department of Geology and Mineral Industries.
- Goldfinger C., Kulm L.D., Yeats R.S., Appelgate B., MacKay M.E. and Moore G.F. (1992) Transverse structural trends along the Oregon convergent margin: Implications for Cascadia earthquake potential and crustal rotations. *Geology*, **20**, 141.
- Goldfinger C., Kulm L.D., Yeats R.S., Hummon C., Huftile G.J., Niem A.R. and McNeill L.C. (1996) Oblique Strike-Slip Faulting of the Cascadia Submarine Forearc: The Daisy Bank Fault Zone off Central Oregon, American Geophysical Union, p. 65–74.
- Goldfinger C., Kulm L.D., Yeats R.S., McNeill L. and Hummon C. (1997) Oblique strike-slip faulting of the central Cascadia submarine forearc. *Journal of Geophysical Research: Solid Earth*, **102**, 8217–8243.
- Gonzalez-Mieres R. and Suppe J. (2006) Relief and shortening in detachment folds. *Journal of Structural Geology*, **28**, 1785–1807.
- Goossens M., Mittelbach F. and Samarin A. (1994) *The L^AT_EX Companion*, Addison-Wesley.
- Grando G. and McClay K. (2007) Morphotectonics domains and structural styles in the makran accretionary prism, offshore iran. *Sedimentary Geology*, **196**, 157–179.
- Gratier J.P., Richard J., Renard F., Mittempergher S., Doan M.L., Di Toro G., Hadizadeh J. and Boullier A.M. (2011) Aseismic sliding of active faults by pressure solution creep: Evidence from the San Andreas Fault Observatory at Depth. *Geology*, **39**, 1131–1134.
- Groshong R.H. and Epard J.L. (1994) The role of strain in area-constant detachment folding. *Journal of Structural Geology*, **16**, 613–618.

- Groshong R.H., Withjack M.O., Schlische R.W. and Hidayah T.N. (2012) Bed length does not remain constant during deformation: Recognition and why it matters. *Journal of Structural Geology*, **41**, 86–97.
- Gulick S.P., Bangs N.L., Moore G.F., Ashi J., Martin K.M., Sawyer D.S., Tobin H.J., Kuramoto S. and Taira A. (2010) Rapid forearc basin uplift and megasplay fault development from 3D seismic images of Nankai margin off Kii Peninsula, Japan. *Earth and Planetary Science Letters*, **300**, 55–62.
- Gulick S.P.S., Austin J.A., McNeill L.C., Bangs N.L.B., Martin K.M., Henstock T.J., Bull J.M., Dean S., Djajadihardja Y.S. and Permana H. (2011) Updip rupture of the 2004 Sumatra earthquake extended by thick indurated sediments. *Nature Geoscience*, **4**, 453–456.
- Gutscher M., Kukowski N., Malavieille J. and Lallemand S. (1998) Episodic imbricate thrusting and underthrusting: Analog experiments and mechanical analysis applied to the alaskan accretionary wedge. *Journal of Geophysical Research: Solid Earth*, **103**, 10161–10176.
- Gutscher M.A., Klaeschen D., Flueh E. and Malavieille J. (2001) Non-coulomb wedges, wrong-way thrusting, and natural hazards in Cascadia. *Geology*, **29**, 379.
- Gutscher M.A., Kukowski N., Malavieille J. and Lallemand S. (1996) Cyclical behavior of thrust wedges: Insights from high basal friction sandbox experiments. *Geology*, **24**, 135.
- Gómez de la Peña L., R. Ranero C., Gràcia E., Booth-Rea G., Azañón J.M., Tinivella U. and Yelles-Chaouche A. (2022) Evidence for a developing plate boundary in the western mediterranean. *Nature Communications*, **13**.
- Haeussler P.J., Armstrong P.A., Liberty L.M., Ferguson K.M., Finn S.P., Arkle J.C.

- and Pratt T.L. (2015a) Focused exhumation along megathrust splay faults in Prince William Sound, Alaska. *Quaternary Science Reviews*, **113**, 8–22.
- Haeussler P.J., Armstrong P.A., Liberty L.M., Ferguson K.M., Finn S.P., Arkle J.C. and Pratt T.L. (2015b) Focused exhumation along megathrust splay faults in prince william sound, alaska. *Quaternary Science Reviews*, **113**, 8–22.
- Hamers M. and Drury M. (2011) Scanning electron microscope-cathodoluminescence (SEM-CL) imaging of planar deformation features and tectonic deformation lamellae in quartz. *Meteoritics & Planetary Science*, **46**, 1814–1831.
- Han S., Bangs N.L., Carbotte S.M., Saffer D.M. and Gibson J.C. (2017) Links between sediment consolidation and cascadia megathrust slip behaviour. *Nature Geoscience*, **10**, 954–959.
- Han S., Carbotte S.M., Canales J.P., Nedimović M.R. and Carton H. (2018) Along-trench structural variations of the subducting Juan de Fuca Plate from multichannel seismic reflection imaging. *Journal of Geophysical Research: Solid Earth*, **123**, 3122–3146.
- Hashimoto M. and Kanehira K. (1975) Some petrological aspects on stilpnomelane in glaucophanitic metamorphic rocks. *The Journal of the Japanese Association of Mineralogists, Petrologists and Economic Geologists*, **70**, 377–387.
- Hayes G.P., Moore G.L., Portner D.E., Hearne M., Flamme H., Furtney M. and Smoczyk G.M. (2018) Slab2, a comprehensive subduction zone geometry model. *Science*, **362**, 58–61.
- Henstock T.J., McNeill L.C. and Tappin D.R. (2006) Seafloor morphology of the Sumatran subduction zone: Surface rupture during megathrust earthquakes? *Geology*, **34**, 485.

- Heuret A., Conrad C., Funiciello F., Lallemand S. and Sandri L. (2012) Relation between subduction megathrust earthquakes, trench sediment thickness and upper plate strain. *Geophysical Research Letters*, **39**.
- Hill J.C., Brothers D.S., Craig B.K., ten Brink U.S., Chaytor J.D. and Flores C.H. (2017) Geologic controls on submarine slope failure along the central U.S. Atlantic margin: Insights from the Currituck Slide Complex. *Marine Geology*, **385**, 114–130.
- Hill J.C., Watt J.T. and Brothers D.S. (2022) Mass wasting along the Cascadia subduction zone: Implications for abyssal turbidite sources and the earthquake record. *Earth and Planetary Science Letters*, **597**, 117797.
- Hill J.C., Watt J.T., Paull C.K., Caress D.W., Brothers D.S., Arizmendi K., Gwiazda R., Kluesner J., Lundsten E., Nieminski N.M., Padgett J.S., Paduan J.B. and Snyder G. (2026) Widespread abyssal turbidites record megathrust earthquake-triggered landslides and coseismic deformation in the cascadia subduction zone. *Science Advances*, **12**.
- Hilley G., Brodsky E., Roman D., Shillington D., Brudzinski M., Behn M. and Tobin H. (2022) SZ4D implementation plan. SZ4D Implementation Plan.
- Hirt A.M., Lowrie W., Lüneburg C., Lebit H. and Engelder T. (2004) Magnetic and mineral fabric development in the Ordovician Martinsburg Formation in the central Appalachian fold and thrust belt, Pennsylvania. *Geological Society, London, Special Publications*, **238**, 109–126.
- Holbrook W.S., Kent G., Keranen K., Johnson H.P., Trehu A., Tobin H., Caplan-Auerbach J. and Beeson J. (2012) Cascadia fore arc seismic survey: Open-access data available. *Eos, Transactions American Geophysical Union*, **93**, 521–522.
- Housen B.A. and Kanamatsu T. (2003) Magnetic fabrics from the Costa Rica margin:

- sediment deformation during the initial dewatering and underplating process. *Earth and Planetary Science Letters*, **206**, 215–228.
- Housen B.A., Richter C. and van der Pluijm B.A. (1993) Composite magnetic anisotropy fabrics: experiments, numerical models and implications for the quantification of rock fabrics. *Tectonophysics*, **220**, 1–12.
- Housen B.A., Tobin H.J., Labaume P., Leitch E.C., Maltman A.J. and Party O.D.P.L..S.S. (1996) Strain decoupling across the décollement of the Barbados accretionary prism. *Geology*, **24**, 127–130.
- Hrouda F. (1982) Magnetic anisotropy of rocks and its application in geology and geophysics. *Geophysical Surveys*, **5**, 37–82.
- Hu Y. and Wang K. (2008) Coseismic strengthening of the shallow portion of the subduction fault and its effects on wedge taper. *Journal of Geophysical Research: Solid Earth*, **113**.
- Hubbard J. and Shaw J.H. (2009) Uplift of the Longmen Shan and Tibetan plateau, and the 2008 Wenchuan ($M = 7.9$) earthquake. *Nature*, **458**, 194–197.
- Hyndman R.D. and Wang K. (1993) Thermal constraints on the zone of major thrust earthquake failure: The Cascadia Subduction Zone. *Journal of Geophysical Research: Solid Earth*, **98**, 2039–2060.
- Hüpers A., Torres M.E., Owari S., McNeill L.C., Dugan B., Henstock T.J., Milliken K.L., Petronotis K.E., Backman J., Bourlange S., Chemale F., Chen W., Colson T.A., Frederik M.C.G., Guérin G., Hamahashi M., House B.M., Jeppson T.N., Kachovich S., Kenigsberg A.R., Kuranaga M., Kutterolf S., Mitchison F.L., Mukoyoshi H., Nair N., Pickering K.T., Pouderoux H.F.A., Shan Y., Song I., Vannucchi P., Vrolijk P.J., Yang T. and Zhao X. (2017a) Release of mineral-bound water prior to subduction tied to shallow seismogenic slip off sumatra. *Science*, **356**, 841–844.

- Hüpers A., Torres M.E., Owari S., McNeill L.C., Dugan B., Henstock T.J., Milliken K.L., Petronotis K.E., Backman J., Bourlange S., Chemale F., Chen W., Colson T.A., Frederik M.C.G., Guèrin G., Hamahashi M., House B.M., Jeppson T.N., Kachovich S., Kenigsberg A.R., Kuranaga M., Kutterolf S., Mitchison F.L., Mukoyoshi H., Nair N., Pickering K.T., Pouderoux H.F.A., Shan Y., Song I., Vannucchi P., Vrolijk P.J., Yang T. and Zhao X. (2017b) Release of mineral-bound water prior to subduction tied to shallow seismogenic slip off Sumatra. *Science*, **356**, 841–844.
- Ide S., Baltay A. and Beroza G.C. (2011) Shallow dynamic overshoot and energetic deep rupture in the 2011 Mw 9.0 Tohoku-oki earthquake. *Science*, **332**, 1426–1429.
- Ikari M.J., Kopf A.J., Hüpers A. and Vogt C. (2018) Lithologic control of frictional strength variations in subduction zone sediment inputs. *Geosphere*, **14**, 604–625.
- Ikari M.J., Wilckens F.K. and Saffer D.M. (2020) Implications of basement rock alteration in the nankai trough, japan for subduction megathrust slip behavior. *Tectonophysics*, **774**, 228275.
- Ike T., Moore G.F., Kuramoto S., Park J., Kaneda Y. and Taira A. (2008a) Variations in sediment thickness and type along the northern philippine sea plate at the nankai trough. *Island Arc*, **17**, 342–357.
- Ike T., Moore G.F., Kuramoto S., Park J.O., Kaneda Y. and Taira A. (2008b) Variations in sediment thickness and type along the northern philippine sea plate at the nankai trough. *Island Arc*, **17**, 342–357.
- Ishii M., Shearer P.M., Houston H. and Vidale J.E. (2005) Extent, duration and speed of the 2004 Sumatra–Andaman earthquake imaged by the Hi-Net array. *Nature*, **435**, 933–936.
- Ito Y., Tsuji T., Osada Y., Kido M., Inazu D., Hayashi Y., Tsushima H., Hino R.

- and Fujimoto H. (2011) Frontal wedge deformation near the source region of the 2011 Tohoku-oki earthquake. *Geophysical Research Letters*, **38**.
- Jelinek V. (1981) Characterization of the magnetic fabric of rocks. *Tectonophysics*, **79**, T63–T67.
- Jelínek V. (1978) Statistical processing of anisotropy of magnetic susceptibility measured on groups of specimens. *Studia Geophysica et Geodaetica*, **22**, 50–62.
- Jiang D., Han S., Bangs N.L., Carbotte S.M., Boston B., Tobin H.J., Beeson J.W., Canales J.P. and Nedimovic M.R. (2023) Characterizing outer wedge structure along the cascadia subduction margin using casie21 seismic data, in: *AGU Fall Meeting Abstracts*, vol. 2023, pp. T53E–0195.
- Kaiser A., Balfour N., Fry B., Holden C., Litchfield N., Gerstenberger M., D’Anastasio E., Horspool N., McVerry G., Ristau J., Bannister S., Christophersen A., Clark K., Power W., Rhoades D., Massey C., Hamling I., Wallace L., Mountjoy J., Kaneko Y., Benites R., Van Houtte C., Dellow S., Wotherspoon L., Elwood K. and Gledhill K. (2017) The 2016 Kaikōura, New Zealand, earthquake: Preliminary seismological report. *Seismological Research Letters*, **88**, 727–739.
- Kaneki S. and Kouketsu Y. (2022) An automatic peak deconvolution method for Raman spectra of terrestrial carbonaceous material for application to the geothermometers of Kouketsu et al.(2014). *Island Arc*, **31**, e12467.
- Kawabata K., Tanaka H. and Kimura G. (2007) Mass transfer and pressure solution in deformed shale of accretionary complex: Examples from the Shimanto Belt, southwestern Japan. *Journal of Structural Geology*, **29**, 697–711.
- Kimura G., Hamahashi M., Okamoto S., Yamaguchi A., Kameda J., Raimbourg H., Hamada Y., Yamaguchi H. and Shibata T. (2013) Hanging wall deformation of

- a seismogenic megasplay fault in an accretionary prism: The Nobeoka Thrust in southwestern Japan. *Journal of Structural Geology*, **52**, 136–147.
- Kimura G., Hashimoto Y., Kitamura Y., Yamaguchi A. and Koge H. (2014) Middle miocene swift migration of the TTT triple junction and rapid crustal growth in southwest Japan: A review. *Tectonics*, **33**, 1219–1238.
- Kimura G., Yamaguchi A., Hojo M., Kitamura Y., Kameda J., Ujiie K., Hamada Y., Hamahashi M. and Hina S. (2012) Tectonic mélange as fault rock of subduction plate boundary. *Tectonophysics*, **568**, 25–38.
- Kimura H., Takeda T., Obara K. and Kasahara K. (2010) Seismic evidence for active underplating below the megathrust earthquake zone in Japan. *Science*, **329**, 210–212.
- Kirkpatrick J., Shipton Z. and Persano C. (2009) Pseudotachylytes: Rarely generated, rarely preserved, or rarely reported? *Bulletin of the Seismological Society of America*, **99**, 382–388.
- Kirkpatrick J.D., Edwards J.H., Verdecchia A., Kluesner J.W., Harrington R.M. and Silver E.A. (2020) Subduction megathrust heterogeneity characterized from 3D seismic data. *Nature Geoscience*, **13**, 369–374.
- Kirkpatrick J.D., Fagereng and Shelly D.R. (2021) Geological constraints on the mechanisms of slow earthquakes. *Nature Reviews Earth & Environment*, **2**, 285–301.
- Kirkpatrick J.D. and Rowe C.D. (2013) Disappearing ink: How pseudotachylytes are lost from the rock record. *Journal of Structural Geology*, **52**, 183–198.
- Kirkpatrick J.D., Rowe C.D., Ujiie K., Moore J.C., Regalla C., Remitti F., Toy V., Wolfson-Schwehr M., Kameda J., Bose S. and Chester F.M. (2015) Structure

- and lithology of the Japan Trench subduction plate boundary fault. *Tectonics*, **34**, 53–69.
- Kirkpatrick J.D., Savage H.M., Regalla C., Shreedharan S., Ross C., Okuda H., Nicholson U., Ujiie K., Hackney R., Conin M., Pei P., Satolli S., Zhang J., Fulton P., Ikari M., Kodaira S., Maeda L., Okutsu N., Toczko S., Eguchi N., Bellanova P., Brown C., Brunet M., Castillo M.R., Chang Y.C., Doan M.L., Everard J., Fintel A., Ford J., Fukuchi R., Gough A., Guo H., Gürer D., Hagino M., Hamada Y., Hosono H., Jurado M., Ijiri A., Ishikawa T., Iwai M., Jeppson T., Kamiya N., Kanamatsu T., LaPlante A., Lin W., Miyakawa A., Morono Y., Nakamura Y., Pizer C., Rasbury T., Robertson R., Schaible K., Sone H., Sun C., Uchida T., Vannucchi P., Yamaguchi A., Yamamoto Y. and Yoshimoto T. (2026) Extreme plate boundary localization promotes shallow earthquake slip at the Japan Trench. *Science*, **391**, 489–493.
- Kitajima H. and Saffer D.M. (2012) Elevated pore pressure and anomalously low stress in regions of low frequency earthquakes along the Nankai Trough subduction megathrust. *Geophysical Research Letters*, **39**.
- Kitamura Y. and Kimura G. (2012) Dynamic role of tectonic mélange during inter-seismic process of plate boundary mega earthquakes. *Tectonophysics*, **568–569**, 39–52.
- Kitamura Y., Sato K., Ikesawa E., Ikehara-Ohmori K., Kimura G., Kondo H., Ujiie K., Onishi C.T., Kawabata K., Hashimoto Y., Mukoyoshi H. and Masago H. (2005) Mélange and its seismogenic roof décollement: A plate boundary fault rock in the subduction zone—An example from the Shimanto Belt, Japan. *Tectonics*, **24**.
- Kluesner J., Brothers D., Hart P., Miller N. and Hatcher G. (2018) Practical approaches to maximizing the resolution of sparker seismic reflection data. *Marine Geophysical Research*, **40**, 279–301.

Knuth D.E. (1984) The T_EX book, Addison-Wesley.

Knuth D.E. (1986a) T_EX: The Program, Addison-Wesley.

Knuth D.E. (1986b) Computer Modern Typefaces, Addison-Wesley.

Knuth D.E. (1986c) The Metafont book, Addison-Wesley.

Kodaira S., Fujiwara T., Fujie G., Nakamura Y. and Kanamatsu T. (2020) Large coseismic slip to the trench during the 2011 Tohoku-oki earthquake. *Annual Review of Earth and Planetary Sciences*, **48**, 321–343.

Kodaira S., No T., Nakamura Y., Fujiwara T., Kaiho Y., Miura S., Takahashi N., Kaneda Y. and Taira A. (2012) Coseismic fault rupture at the trench axis during the 2011 Tohoku-oki earthquake. *Nature Geoscience*, **5**, 646–650.

Kopp C., Fruehn J., Flueh E., Reichert C., Kukowski N., Bialas J. and Klaeschen D. (2000) Structure of the makran subduction zone from wide-angle and reflection seismic data. *Tectonophysics*, **329**, 171–191.

Kopp H. and Kukowski N. (2003) Backstop geometry and accretionary mechanics of the Sunda margin. *Tectonics*, **22**.

Kouketsu Y., Mizukami T., Mori H., Endo S., Aoya M., Hara H., Nakamura D. and Wallis S. (2014) A new approach to develop the Raman carbonaceous material geothermometer for low-grade metamorphism using peak width. *Island Arc*, **23**, 33–50.

Labat P., Maltman A., Bolton A., Tessier D., Ogawa Y. and Takizawa S. (2000) Scalp fabrics in sheared clays from the décollement zone of the Barbados accretionary prism.

- Lackey J., Regalla C. and Moore G. (2020) Tectonic influences on trench slope basin development via structural restoration along the outer nankai accretionary prism, southwest japan. *Geochemistry, Geophysics, Geosystems*, **21**, e2020GC009038.
- Laigle M., Becel A., de Voogd B., Sachpazi M., Bayrakci G., Lebrun J.F. and Evain M. (2013) Along-arc segmentation and interaction of subducting ridges with the lesser antilles subduction forearc crust revealed by mcs imaging. *Tectonophysics*, **603**.
- Lamport L. (1994) *L^AT_EX: A Document Preparation System*, 2nd ed., Addison-Wesley.
- Ledeczi A., Lucas M., Tobin H., Watt J. and Miller N. (2024) Late quaternary surface displacements on accretionary wedge splay faults in the cascadia subduction zone: Implications for megathrust rupture. *Seismica*, **2**.
- Ledeczi A.M., Tobin H.J., Chen T.W., Mulcahy S.R. and Condit C.B. (2025) Structure and properties of the Cascadia plate interface: Evidence from a newly described exhumed paleomegathrust in the Olympic Subduction Complex. *Geological Society of America Bulletin*, **138**, 1453–1468.
- Lee M.K., Carbotte S.M., Han S., Shuck B., Gurun P., Boston B., Nedimovic M., Tobin H., Norvell B. and Spinelli G. (2025) Anomalous sediment consolidation and alteration from buried incoming plate seamounts along the cascadia margin. *Geochemistry, Geophysics, Geosystems*, **26**, e2024GC011949.
- Li S., Wang K., Wang Y., Jiang Y. and Dosso S.E. (2018) Geodetically inferred locking state of the Cascadia megathrust based on a viscoelastic earth model. *Journal of Geophysical Research: Solid Earth*, **123**, 8056–8072.
- Li T., Chen Z., Chen J., Thompson Jobe J.A., Burbank D.W., Li Z., He X., Zheng W., Zhang P. and Zhang B. (2019) Along-strike and downdip segmentation of the

- pamir frontal thrust and its association with the 1985mw6.9 wuqia earthquake. *Journal of Geophysical Research: Solid Earth*, **124**.
- Liberty L.M., Finn S.P., Haeussler P.J., Pratt T.L. and Peterson A. (2013) Megathrust splay faults at the focus of the Prince William Sound asperity, Alaska. *Journal of Geophysical Research: Solid Earth*, **118**, 5428–5441.
- Lin Y.N., Sladen A., Ortega-Culaciati F., Simons M., Avouac J., Fielding E.J., Brooks B.A., Bevis M., Genrich J., Rietbrock A., Vigny C., Smalley R. and Socquet A. (2013) Coseismic and postseismic slip associated with the 2010 Maule Earthquake, Chile: Characterizing the Arauco Peninsula barrier effect. *Journal of Geophysical Research: Solid Earth*, **118**, 3142–3159.
- Lindsey E.O., Mallick R., Hubbard J.A., Bradley K.E., Almeida R.V., Moore J.D.P., Bürgmann R. and Hill E.M. (2021) Slip rate deficit and earthquake potential on shallow megathrusts. *Nature Geoscience*, **14**, 321–326.
- Litchfield N.J., Villamor P., Dissen R.J.V., Nicol A., Barnes P.M., A. Barrell D.J., Pettinga J.R., Langridge R.M., Little T.A., Mountjoy J.J., Ries W.F., Rowland J., Fenton C., Stirling M.W., Kearse J., Berryman K.R., Cochran U.A., Clark K.J., Hemphill-Haley M., Khajavi N., Jones K.E., Archibald G., Upton P., Asher C., Benson A., Cox S.C., Gasston C., Hale D., Hall B., Hatem A.E., Heron D.W., Howarth J., Kane T.J., Lamarche G., Lawson S., Lukovic B., McColl S.T., Madugo C., Manousakis J., Noble D., Pedley K., Sauer K., Stahl T., Strong D.T., Townsend D.B., Toy V., Williams J., Woelz S. and Zinke R. (2018) Surface rupture of multiple crustal faults in the 2016 Mw7.8 Kaikōura, New Zealand, earthquake. *Bulletin of the Seismological Society of America*, **108**, 1496–1520.
- Lucas M.C., Ledeczi A.M., Tobin H.J., Carbotte S.M., Watt J.T., Han S., Boston B. and Jiang D. (2025) No evidence for an active margin-spanning megasplay fault at the cascadia subduction zone. *Seismica*, **2**.

- Lundberg N. and Moore J.C. (1986) Macroscopic structural features in Deep Sea Drilling Project cores from forearc regions, Geological Society of America, p. 13–44.
- Lünsdorf N.K., Dunkl I., Schmidt B.C., Rantitsch G. and von Eynatten H. (2017) Towards a higher comparability of geothermometric data obtained by Raman spectroscopy of carbonaceous material. part 2: A revised geothermometer. *Geostandards and Geoanalytical Research*, **41**, 593–612.
- MacKay M.E. (1995) Structural variation and landward vergence at the toe of the Oregon accretionary prism. *Tectonics*, **14**, 1309–1320.
- MacKay M.E., Moore G.F., Cochrane G.R., Casey Moore J. and Kulm L.D. (1992) Landward vergence and oblique structural trends in the Oregon margin accretionary prism: Implications and effect on fluid flow. *Earth and Planetary Science Letters*, **109**, 477–491.
- Mackay P.A. (2015) The role of fluid pressure in contractional systems: examples from the southern Canadian Rocky Mountains. Geological Society, London, Special Publications, **421**, 69–82.
- Maloney D., Davies R., Imber J., Higgins S. and King S. (2010) New insights into deformation mechanisms in the gravitationally driven Niger Delta deep-water fold and thrust belt. *AAPG Bulletin*, **94**, 1401–1424.
- Mannu U., Ueda K., Willett S.D., Gerya T.V. and Strasser M. (2016) Impact of sedimentation on evolution of accretionary wedges: Insights from high-resolution thermomechanical modeling. *Tectonics*, **35**, 2828–2846.
- Mannu U., Ueda K., Willett S.D., Gerya T.V. and Strasser M. (2017) Stratigraphic signatures of forearc basin formation mechanisms. *Geochemistry, Geophysics, Geosystems*, **18**, 2388–2410.

- McAdoo B.G., Orange D.L., Screatton E., Lee H. and Kayen R. (1997) Slope basins, headless canyons, and submarine palaeoseismology of the Cascadia accretionary complex. *Basin Research*, **9**, 313–324.
- McCalpin J.P. and Nelson A.R. (2009) *Introduction to Paleoseismology*, Elsevier, p. 1–27.
- McNeill L.C., Goldfinger C., Kulm L.D. and Yeats R.S. (2000) Tectonics of the Neogene Cascadia forearc basin: Investigations of a deformed late Miocene unconformity. *Geological Society of America Bulletin*, **112**, 1209–1224.
- McNeill L.C. and Henstock T.J. (2014) Forearc structure and morphology along the Sumatra-Andaman subduction zone. *Tectonics*, **33**, 112–134.
- McNeill L.C., Piper K.A., Goldfinger C., Kulm L.D. and Yeats R.S. (1997) Listric normal faulting on the Cascadia continental margin. *Journal of Geophysical Research: Solid Earth*, **102**, 12123–12138.
- Melgar D. (2021) Was the January 26th, 1700 Cascadia earthquake part of a rupture sequence? *Journal of Geophysical Research: Solid Earth*, **126**.
- Meneghini F., Marroni M., Moore J., Pandolfi L. and Rowe C. (2009) The processes of underthrusting and underplating in the geologic record: Structural diversity between the Franciscan Complex (California), the Kodiak Complex (Alaska) and the Internal Ligurian Units (Italy). *Geological Journal*, **44**, 126–152.
- Meneghini F. and Moore J.C. (2007) Deformation and hydrofracture in a subduction thrust at seismogenic depths: The Rodeo Cove thrust zone, Marin Headlands, California. *Geological Society of America Bulletin*, **119**, 174–183.
- Merrill R., Bostock M.G., Peacock S.M., Calvert A. and Christensen N.I. (2020) A double difference tomography study of the Washington forearc: Does Siletzia

- control crustal seismicity? *Journal of Geophysical Research: Solid Earth*, **125**, e2020JB019750.
- Michel L., Ehlers T.A., Glotzbach C., Adams B.A. and Stübner K. (2018) Tectonic and glacial contributions to focused exhumation in the Olympic Mountains, Washington, USA. *Geology*, **46**, 491–494.
- Michel-Wolf L., Ehlers T.A. and Bendick R. (2022) Transitions in subduction zone properties align with long-term topographic growth (Cascadia, USA). *Earth and Planetary Science Letters*, **580**, 117363.
- Mitchum R.M., Vail P.R. and Sangree J.B. (1977) Seismic Stratigraphy and Global Changes of Sea Level: Stratigraphic Interpretation of Seismic Reflection Patterns in Depositional Sequences, American Association of Petroleum Geologists.
- Mitra G. and Boyer S.E. (2020) Duplexes as slip transfer zones: Application to the himalayan fold-thrust belt. *Journal of Asian Earth Sciences*, **190**, 104214.
- Moeremans R., Singh S.C., Mukti M., McArdle J. and Johansen K. (2014) Seismic images of structural variations along the deformation front of the andaman–sumatra subduction zone: Implications for rupture propagation and tsunamigenesis. *Earth and Planetary Science Letters*, **386**, 75–85.
- Moore G., Cochran G., Trehu A. and Moore C. (2013) Multi-channel seismic shot data from the cascadia region of the eastern north pacific ocean acquired during the r/v geotide expedition gt8909 (1989).
- Moore G.F., Bangs N.L., Taira A., Kuramoto S., Pangborn E. and Tobin H.J. (2007a) Three-dimensional splay fault geometry and implications for tsunami generation. *Science*, **318**, 1128–1131.

- Moore G.F., Saffer D., Studer M. and Costa Pisani P. (2011) Structural restoration of thrusts at the toe of the nankai trough accretionary prism off shikoku island, japan: Implications for dewatering processes. *Geochemistry, Geophysics, Geosystems*, **12**.
- Moore J., Roeske S., Cowan D., Lundberg N., Gonzales E., Schoonmaker J. and Lucas S. (1986) Scaly fabrics from deep sea drilling project cores from forearcs, structural fabrics in deep sea drilling project cores from forearcs. *Geological Society of America Memoirs*, 75–88.
- Moore J., Shipley T., Goldberg D., Ogawa Y., Filice F., Fisher A., Jurado M.J., Moore G., Rabaute A., Yin H. et al. (1995) Abnormal fluid pressures and fault-zone dilation in the Barbados accretionary prism: Evidence from logging while drilling. *Geology*, **23**, 605–608.
- Moore J.C. (1989) Tectonics and hydrogeology of accretionary prisms: role of the décollement zone. *Journal of Structural Geology*, **11**, 95–106.
- Moore J.C. and Byrne T. (1987) Thickening of fault zones: A mechanism of melange formation in accreting sediments. *Geology*, **15**, 1040–1043.
- Moore J.C., Rowe C. and Meneghini F. (2007b) *How Accretionary Prisms Elucidate Seismogenesis in Subduction Zones*, Columbia University Press, p. 288–315.
- Moore J.C. and Saffer D. (2001a) Updip limit of the seismogenic zone beneath the accretionary prism of southwest Japan: An effect of diagenetic to low-grade metamorphic processes and increasing effective stress. *Geology*, **29**, 183.
- Moore J.C. and Saffer D. (2001b) Updip limit of the seismogenic zone beneath the accretionary prism of southwest Japan: An effect of diagenetic to low-grade metamorphic processes and increasing effective stress. *Geology*, **29**, 183–186.
- Moris-Muttoni B., Raimbourg H., Champallier R., Augier R., Lahfid A., Le Trong E. and Di Carlo I. (2023) The effect of strain on the crystallinity of carbonaceous

- matter: Application of Raman spectroscopy to deformation experiments. *Tectonophysics*, **869**, 230126.
- Morton E.A., Bilek S.L. and Rowe C.A. (2023) Cascadia Subduction Zone fault heterogeneities from newly detected small magnitude earthquakes. *Journal of Geophysical Research: Solid Earth*, **128**.
- Nedimović M.R., Bohnenstiehl D.R., Carbotte S.M., Pablo Canales J. and Dziak R.P. (2009) Faulting and hydration of the Juan de Fuca plate system. *Earth and Planetary Science Letters*, **284**, 94–102.
- Niemeijer A., Di Toro G., Griffith W.A., Bistacchi A., Smith S.A. and Nielsen S. (2012) Inferring earthquake physics and chemistry using an integrated field and laboratory approach. *Journal of Structural Geology*, **39**, 2–36.
- Noda A., Graveleau F., Witt C., Chanier F. and Vendeville B. (2023) Accretion cycles, structural evolution, and thrust activity in accretionary wedges with various décollement configurations: Insights from sandbox analog modeling. *Journal of Geophysical Research: Solid Earth*, **128**.
- Pacheco J.F., Sykes L.R. and Scholz C.H. (1993) Nature of seismic coupling along simple plate boundaries of the subduction type. *Journal of Geophysical Research: Solid Earth*, **98**, 14133–14159.
- Park J.O., Tsuru T., Kodaira S., Cummins P.R. and Kaneda Y. (2002) Splay fault branching along the nankai subduction zone. *Science*, **297**, 1157–1160.
- Parés J.M. (2004) How deformed are weakly deformed mudrocks? insights from magnetic anisotropy. *Geological Society, London, Special Publications*, **238**, 191–203.
- Parés J.M. (2015) Sixty years of anisotropy of magnetic susceptibility in deformed sedimentary rocks. *Frontiers in Earth Science*, **3**.

- Parés J.M. and van der Pluijm B.A. (2002) Evaluating magnetic lineations (AMS) in deformed rocks. *Tectonophysics*, **350**, 283–298.
- Paterson M. and Olgaard D. (2000) Rock deformation tests to large shear strains in torsion. *Journal of structural Geology*, **22**, 1341–1358.
- Pazzaglia F.J. and Brandon M.T. (2001) A fluvial record of long-term steady-state uplift and erosion across the Cascadia forearc high, western Washington State. *American Journal of Science*, **301**, 385–431.
- Peterson D.E. and Keranen K.M. (2019) A high wave speed basal sedimentary layer identified from seismic imaging of the plate boundary in central cascadia. *Journal of Geophysical Research: Solid Earth*, **124**.
- Plafker G. (1965) Tectonic deformation associated with the 1964 alaska earthquake: The earthquake of 27 march 1964 resulted in observable crustal deformation of unprecedented areal extent. *Science*, **148**.
- Plafker G. (1969) Tectonics of the March 27, 1964, Alaska earthquake.
- Plaza-Faverola A., Klaeschen D., Barnes P., Pecher I., Henrys S. and Mountjoy J. (2012) Evolution of fluid expulsion and concentrated hydrate zones across the southern hikurangi subduction margin, new zealand: An analysis from depth migrated seismic data. *Geochemistry, Geophysics, Geosystems*, **13**.
- Posamentier H.W., Paumard V. and Lang S.C. (2022) Principles of seismic stratigraphy and seismic geomorphology i: Extracting geologic insights from seismic data. *Earth-Science Reviews*, **228**, 103963.
- Prior D.J. and Behrmann J.H. (1990) Thrust-related mudstone fabrics from the Barbados forearc: A backscattered scanning electron microscope study. *Journal of Geophysical Research: Solid Earth*, **95**, 9055–9067.

- Qiu Q. and Barbot S. (2022) Tsunami excitation in the outer wedge of global subduction zones. *Earth-Science Reviews*, **230**, 104054.
- Rahl J.M., Anderson K.M., Brandon M.T. and Fassoulas C. (2005) Raman spectroscopic carbonaceous material thermometry of low-grade metamorphic rocks: Calibration and application to tectonic exhumation in Crete, Greece. *Earth and Planetary Science Letters*, **240**, 339–354.
- Raimbourg H., Famin V., Palazzin G., Yamaguchi A., Augier R., Kitamura Y. and Sakaguchi A. (2019) Distributed deformation along the subduction plate interface: The role of tectonic mélanges. *Lithos*, **334–335**, 69–87.
- Rajič K., Raimbourg H., Famin V. and Moris-Muttoni B. (2024) The origin of tectonic mélanges from the Kodiak complex and Shimanto Belt and its implication for subduction interface processes. *Earth and Planetary Science Letters*, **648**, 119085.
- Ramos M.D., Liberty L.M., Haeussler P.J. and Humphreys R. (2022) Upper-plate structure and tsunamigenic faults near the Kodiak Islands, Alaska, USA. *Geosphere*, **18**, 1474–1491.
- Ramsay J.G. and Wood D.S. (1973) The geometric effects of volume change during deformation processes. *Tectonophysics*, **16**, 263–277.
- Rau W. (2003) Geology of the Washington coast between Point Grenville and the Hoh River.
- Regalla C.A., Rowe C.D., Harrichhausen N., Tarling M.S. and Singh J. (2018) Geology and Tectonics of Subduction Zones: A Tribute to Gaku Kimura, chap. Styles of underplating in the Marin Headlands terrane, Franciscan complex, California, Geological Society of America.

- Remitti F., Bettelli G. and Vannucchi P. (2007) Internal structure and tectonic evolution of an underthrust tectonic mélange: The Sestola-Vidiciatico tectonic unit of the Northern Apennines, Italy. *Geodinamica Acta*, **20**, 37–51.
- Remitti F., Festa A., Nirta G., Barbero E. and Mittempergher S. (2024) Role of folding-related deformation in the seismicity of shallow accretionary prisms. *Nature Geoscience*, **17**, 600–607.
- Remitti F., Vannucchi P., Bettelli G., Fantoni L., Panini F. and Vescovi P. (2011) Tectonic and sedimentary evolution of the frontal part of an ancient subduction complex at the transition from accretion to erosion: The case of the Ligurian wedge of the northern Apennines, Italy. *Bulletin*, **123**, 51–70.
- Riedel M., Côté M.M., Urlaub M., Geersen J., Scholz N.A., Naegeli K. and Spence G.D. (2018) Slope failures along the deformation front of the Cascadia margin: linking slide morphology to subduction zone parameters. Geological Society, London, Special Publications, **477**, 47–67.
- Roberts N.M., Tikoff B., Davis J.R. and Stetson-Lee T. (2019) The utility of statistical analysis in structural geology. *Journal of Structural Geology*, **125**, 64–73.
- Rohr K.M.M., King H., Riedel M. and Schmidt U. (2019) From mid-plate to subduction zone: stratigraphy of the northeast Juan de Fuca Plate, offshore British Columbia.
- Ross W.C., Halliwell B.A., May J.A., Watts D.E. and Syvitski J. (1994) Slope readjustment: A new model for the development of submarine fans and aprons. *Geology*, **22**, 511.
- Rowe C.D. and Griffith W.A. (2015) Do faults preserve a record of seismic slip: A second opinion. *Journal of Structural Geology*, **78**, 1–26.

- Rowe C.D., Meneghini F. and Moore J.C. (2009) Fluid-rich damage zone of an ancient out-of-sequence thrust, Kodiak Islands, Alaska. *Tectonics*, **28**.
- Rowe C.D., Meneghini F. and Moore J.C. (2011) Geology of the Earthquake Source: A Volume in Honour of Rick Sibson, chap. Textural record of the seismic cycle: Strain-rate variation in an ancient subduction thrust, Geological Society of America.
- Rowe C.D., Moore J.C., Meneghini F. and McKeirnan A.W. (2005) Large-scale pseudotachylytes and fluidized cataclasites from an ancient subduction thrust fault. *Geology*, **33**, 937–940.
- Rowe C.D., Moore J.C. and Remitti F. (2013) The thickness of subduction plate boundary faults from the seafloor into the seismogenic zone. *Geology*, **41**, 991–994.
- Ruff L.J. (1989) Do trench sediments affect great earthquake occurrence in subduction zones? *Pure and Applied Geophysics*, **129**, 263–282.
- Ryan W.B.F., Carbotte S.M., Coplan J.O., O'Hara S., Melkonian A., Arko R., Weis-
sel R.A., Ferrini V., Goodwillie A., Nitsche F., Bonczkowski J. and Zemsky R.
(2009a) Global multi-resolution topography synthesis. *Geochemistry, Geophysics, Geosystems*, **10**.
- Ryan W.B.F., Carbotte S.M., Coplan J.O., O'Hara S., Melkonian A., Arko R., Weis-
sel R.A., Ferrini V., Goodwillie A., Nitsche F., Bonczkowski J. and Zemsky R.
(2009b) Global multi-resolution topography synthesis. *Geochemistry, Geophysics, Geosystems*, **10**.
- Saffer D.M. and Tobin H.J. (2011) Hydrogeology and mechanics of subduction zone forearcs: Fluid flow and pore pressure. *Annual Review of Earth and Planetary Sciences*, **39**, 157–186.
- Saint-Carlier D., Charreau J., Lavé J., Blard P.H., Dominguez S., Avouac J.P. and Wang S. (2016) Major temporal variations in shortening rate absorbed along a large

- active fold of the southeastern tianshan piedmont (china). *Earth and Planetary Science Letters*, **434**, 333–348.
- Sakaguchi A., Chester F., Curewitz D., Fabbri O., Goldsby D., Kimura G., Li C.F., Masaki Y., Screaton E.J., Tsutsumi A., Ujiie K. and Yamaguchi A. (2011) Seismic slip propagation to the updip end of plate boundary subduction interface faults: Vitrinite reflectance geothermometry on Integrated Ocean Drilling Program NanTroSEIZE cores. *Geology*, **39**, 395–398.
- Salmi M.S., Johnson H.P. and Harris R.N. (2017) Thermal environment of the Southern Washington region of the Cascadia subduction zone. *Journal of Geophysical Research: Solid Earth*, **122**, 5852–5870.
- Satake K., Shimazaki K., Tsuji Y. and Ueda K. (1996) Time and size of a giant earthquake in Cascadia inferred from Japanese tsunami records of January 1700. *Nature*, **379**, 246–249.
- Satake K. and Tanioka Y. (1999) Sources of tsunami and tsunamigenic earthquakes in subduction zones. *Pure and Applied Geophysics*, **154**, 467–483.
- Savage H.M. and Rowe C.D. (2024) Localization and delocalization during seismic slip. *Geophysical Research Letters*, **51**, e2024GL110058.
- Scharer K., Burbank D., Chen J., Weldon R., Rubin C., Zhao R. and Shen J. (2004) Detachment folding in the southwestern tian shan–tarim foreland, china: shortening estimates and rates. *Journal of Structural Geology*, **26**, 2119–2137.
- Schleicher A., van der Pluijm B. and Warr L. (2010) Nanocoatings of clay and creep of the San Andreas fault at Parkfield, California. *Geology*, **38**, 667–670.
- Schleicher A.M., Van Der Pluijm B. and Warr L. (2012) Chlorite-smectite clay minerals and fault behavior: New evidence from the San Andreas Fault Observatory at Depth (SAFOD) core. *Lithosphere*, **4**, 209–220.

- Schlüter H., Prexl A., Gaedicke C., Roeser H., Reichert C., Meyer H. and Von Daniels C. (2002) The makran accretionary wedge: sediment thicknesses and ages and the origin of mud volcanoes. *Marine Geology*, **185**, 219–232.
- Schmalzle G.M., McCaffrey R. and Creager K.C. (2014) Central Cascadia subduction zone creep. *Geochemistry, Geophysics, Geosystems*, **15**, 1515–1532.
- Schmid S.M., Pfiffner O.A., Froitzheim N., Schönborn G. and Kissling E. (1996) Geophysical-geological transect and tectonic evolution of the Swiss-Italian Alps. *Tectonics*, **15**, 1036–1064.
- Schmitt R.G. (2017) Quaternary Faults web application.
- Schottenfels E., Regalla C. and Nakamura Y. (2024) Influence of outer-rise faults on shallow décollement heterogeneity and sediment flux at the japan trench. *Seismica*, **3**.
- Schottenfels E.R. and Regalla C.A. (2021) Bathymetric signatures of submarine fore-arc deformation: A case study in the nankai accretionary prism. *Geochemistry, Geophysics, Geosystems*, **22**, e2021GC010050.
- Seeber L., Mueller C., Fujiwara T., Arai K., Soh W., Djajadihardja Y. and Cormier M. (2007) Accretion, mass wasting, and partitioned strain over the 26 Dec 2004 Mw9.2 rupture offshore Aceh, northern Sumatra. *Earth and Planetary Science Letters*, **263**, 16–31.
- Seely D.R. (1977) The significance of landward vergence and oblique structural trends on trench inner slopes, American Geophysical Union, p. 187–198.
- Shakespeare W. (1946) Hamlet, F.S. Crofts & Co., Inc., NY, act I, Scene 3, Lines 70–72, are apropos.

- Shaw J.H., Bilotti F. and Brennan P.A. (1999) Patterns of imbricate thrusting. *Geological Society of America Bulletin*, **111**, 1140–1154.
- Shaw J.H., Connors C. and Suppe J., editors (2005) *Seismic Interpretation of Contractional Fault-Related Folds*, American Association of Petroleum Geologists.
- Shipley T., Ogawa Y., Blum P., and Bahr J., editors (1997) *Proceedings of the Ocean Drilling Program, 156 Scientific Results*, Ocean Drilling Program.
- Shuck B., Babendreier C., Han S., Carbotte S.M., Beeson J.W., Boston B., Tobin H.J. and Lee M. (2023) A new regional stratigraphic framework for the cascadia oceanic basin, in: *AGU Fall Meeting Abstracts*, vol. 2023, pp. T53E–0197.
- Silver E.A. (1972) Pleistocene tectonic accretion of the continental slope off Washington. *Marine Geology*, **13**, 239–249.
- Simpson G.D. (2010) Formation of accretionary prisms influenced by sediment subduction and supplied by sediments from adjacent continents. *Geology*, **38**, 131–134.
- Smith G., McNeill L., Henstock T.J. and Bull J. (2012) The structure and fault activity of the makran accretionary prism. *Journal of Geophysical Research: Solid Earth*, **117**.
- Snively P. (1987) *Geology and Resource Potential of the Continental Margin of Western North America and Adjacent Ocean Basins—Beaufort Sea to Baja California*, chap. Tertiary geologic framework, neotectonics, and petroleum potential of the Oregon-Washington Continental Margin, Circum-Pacific Council for Energy and Mineral Resources.
- Spivak, M.D., Ph.D. (1985) *PCT_EX Manual*, Personal T_EX, Inc., CA.
- Spivak, M.D., Ph.D. (1986) *The Joy of T_EX*, American Mathematical Society, RI.

- Stanislawski K., Roesner A. and Ikari M.J. (2022) Implications for megathrust slip behavior and pore pressure at the shallow northern cascadia subduction zone from laboratory friction experiments. *Earth and Planetary Science Letters*, **578**, 117297.
- Stevens D.E., Henstock T.J. and McNeill L.C. (2021) Evolution of the thermal and dehydration state of sediments entering the north Sumatra Subduction Zone. *Geochemistry, Geophysics, Geosystems*, **22**.
- Stewart R. (1970) Petrology, metamorphism and structural relations of graywackes in the western Olympic Peninsula, Washington, Ph.D. thesis, Stanford University, Stanford, CA.
- Stewart R.J. and Brandon M.T. (2004) Detrital-zircon fission-track ages for the “Hoh Formation”: Implications for late Cenozoic evolution of the Cascadia subduction wedge. *Geological Society of America Bulletin*, **116**, 60–75.
- Stipp M., Stünitz H., Heilbronner R. and Schmid S.M. (2002) The eastern Tonale fault zone: a ‘natural laboratory’ for crystal plastic deformation of quartz over a temperature range from 250 to 700°C. *Journal of structural geology*, **24**, 1861–1884.
- Stone I., Vidale J.E., Han S. and Roland E. (2018) Catalog of offshore seismicity in Cascadia: insights into the regional distribution of microseismicity and its relation to subduction processes. *Journal of Geophysical Research: Solid Earth*, **123**, 641–652.
- Strasser M., Moore G.F., Kimura G., Kitamura Y., Kopf A.J., Lallemand S., Park J.O., Screatton E.J., Su X., Underwood M.B. et al. (2009) Origin and evolution of a splay fault in the Nankai accretionary wedge. *Nature Geoscience*, **2**, 648–652.
- Su X., Baumann K. and Thiede J. (2000) Calcareous nannofossils from leg 168: biochronology and diagenesis, in: *Proceedings of the Ocean Drilling Program, Scientific Results*, vol. 168, Ocean Drilling Program College Station, TX, pp. 39–49.

- Suleimani E. and Freymueller J.T. (2020) Near-field modeling of the 1964 alaska tsunami: The role of splay faults and horizontal displacements. *Journal of Geophysical Research: Solid Earth*, **125**.
- Sun T., Wang K., Fujiwara T., Kodaira S. and He J. (2017) Large fault slip peaking at trench in the 2011 Tohoku-oki earthquake. *Nature Communications*, **8**.
- Suppe J. (1983) Geometry and kinematics of fault-bend folding. *American Journal of science*, **283**, 684–721.
- Suppe J., Chou G.T. and Hook S.C. (1992) Rates of folding and faulting determined from growth strata, Springer Netherlands, p. 105–121.
- Suppe J., Connors C.D. and Zhang Y. (2004) Shear fault-bend folding.
- Suppe J. and Medwedeff D.A. (1990) Geometry and kinematics of fault-propagation folding. *Eclogae Geologicae Helvetiae*, **83**, 409–454.
- Tabor R. (1975) *Guide to the Geology of Olympic National Park*, University of Washington Press.
- Tabor R. and Cady W. (1978a) *Geologic map of the Olympic Peninsula, Washington*.
- Tabor R. and Cady W. (1978b) *The structure of the Olympic Mountains, Washington: analysis of a subduction zone*. US Geological Survey Professional Paper 1033.
- Tarling D. and Hrouda F. (1993) *Magnetic anisotropy of rocks*, Springer Science & Business Media.
- Tobin H., Kimura G. and Kodaira S. (2019) Processes governing giant subduction earthquakes: IODP drilling to sample and instrument subduction zone megathrusts. *Oceanography*, **32**, 80–93.

- Tobin H., Vannucchi P. and Meschede M. (2001) Structure, inferred mechanical properties, and implications for fluid transport in the décollement zone, Costa Rica convergent margin. *Geology*, **29**, 907–910.
- Tobin H.J. and Kinoshita M. (2006a) NanTroSEIZE: The IODP Nankai Trough Seismogenic Zone Experiment. *Scientific Drilling*, **2**, 23–27.
- Tobin H.J. and Kinoshita M. (2006b) Nantroseize: The iodp nankai trough seismogenic zone experiment. *Scientific Drilling*, **2**, 23–27.
- Tobin H.J., Moore J.C., Mackay M.E., Orange D.L. and Kulm L.D. (1993) Fluid flow along a strike-slip fault at the toe of the Oregon accretionary prism: Implications for the geometry of frontal accretion. *Geological Society of America Bulletin*, **105**, 569–582.
- Tobin H.J. and Saffer D.M. (2009) Elevated fluid pressure and extreme mechanical weakness of a plate boundary thrust, Nankai Troughsubduction zone. *Geology*, **37**, 679–682.
- Tréhu A.M., Hass B., de Moor A., Maksymowicz A., Contreras-Reyes E., Vera E. and Tryon M.D. (2019) Geologic controls on up-dip and along-strike propagation of slip during subduction zone earthquakes from a high-resolution seismic reflection survey across the northern limit of slip during the 2010 mw 8.8 maule earthquake, offshore chile. *Geosphere*, **15**, 1751–1773.
- Ueda H., Kitazato H., Jamieson A., Bond T., Cardigos S., Funaki M., Maroni P.J., Nanbu H., O’Callaghan J.M., Onishi T., Pedersen S.W., Roperez J., Tsuruzono H., Watanabe H.K. and Yasuda T. (2023) The submarine fault scarp of the 2011 Tohoku-oki earthquake in the Japan Trench. *Communications Earth & Environment*, **4**.

- Ujiie K. (2002) Evolution and kinematics of an ancient décollement zone, mélange in the Shimanto accretionary complex of Okinawa Island, Ryukyu Arc. *Journal of Structural Geology*, **24**, 937–952.
- Ujiie K. and Kimura G. (2014) Earthquake faulting in subduction zones: Insights from fault rocks in accretionary prisms. *Progress in Earth and Planetary Science*, **1**, 7.
- Ujiie K., Tanaka H., Saito T., Tsutsumi A., Mori J.J., Kameda J., Brodsky E.E., Chester F.M., Eguchi N., Toczko S., 343 E. and Scientists T. (2013) Low coseismic shear stress on the Tohoku-oki megathrust determined from laboratory experiments. *Science*, **342**, 1211–1214.
- Ujiie K., Yamaguchi H., Sakaguchi A. and Toh S. (2007) Pseudotachylytes in an ancient accretionary complex and implications for melt lubrication during subduction zone earthquakes. *Journal of Structural Geology*, **29**, 599–613.
- Ulrich T., Gabriel A.A., Ampuero J.P. and Xu W. (2019) Dynamic viability of the 2016 Mw 7.8 Kaikōura earthquake cascade on weak crustal faults. *Nature Communications*, **10**.
- Ulrich T., Gabriel A.A. and Madden E.H. (2022) Stress, rigidity and sediment strength control megathrust earthquake and tsunami dynamics. *Nature Geoscience*, **15**, 67–73.
- Underwood M.B. (2002) Strike-parallel variations in clay minerals and fault vergence in the Cascadia subduction zone. *Geology*, **30**, 155.
- Underwood M.B. and Hoke K.D. (2000) Composition and provenance of turbidite sand and hemipelagic mud in the northwestern cascadia basin, in: *Proceedings of the Ocean Drilling Program, Scientific Results*, vol. 168, Ocean Drilling Program College Station, TX, pp. 51–65.

- University of Washington G.S. (2012) Format guidelines for theses and dissertations.
- van Zelst I., Rannabauer L., Gabriel A. and van Dinther Y. (2022) Earthquake rupture on multiple splay faults and its effect on tsunamis. *Journal of Geophysical Research: Solid Earth*, **127**.
- Vann I., Graham R. and Hayward A. (1986) The structure of mountain fronts. *Journal of Structural Geology*, **8**, 215–227.
- Vannucchi P. (2019) Scaly fabric and slip within fault zones. *Geosphere*, **15**, 342–356.
- Vannucchi P. and Bettelli G. (2010) Myths and recent progress regarding the Argille scagliose, Northern Apennines, Italy. *International Geology Review*, **52**, 1106–1137.
- Vannucchi P., Maltman A., Bettelli G. and Clennell B. (2003) On the nature of scaly fabric and scaly clay. *Journal of Structural Geology*, **25**, 673–688.
- Vannucchi P., Remitti F. and Bettelli G. (2008) Geological record of fluid flow and seismogenesis along an erosive subducting plate boundary. *Nature*, **451**, 699–703.
- Vannucchi P. and Tobin H. (2000) Deformation structures and implications for fluid flow at the Costa Rica convergent margin, ODP sites 1040 and 1043, Leg 170. *Journal of Structural Geology*, **22**, 1087–1103.
- Vergés J., Burbank D.W. and Meigs A. (1996) Unfolding: An inverse approach to fold kinematics. *Geology*, **24**, 175.
- Vernooij M.G. and Langenhorst F. (2005) Experimental reproduction of tectonic deformation lamellae in quartz and comparison to shock-induced planar deformation features. *Meteoritics & Planetary Science*, **40**, 1353–1361.
- von Huene R., Miller J.J. and Krabbenhoef A. (2021) The Alaska Convergent Margin backstop splay fault zone, a potential large tsunami generator between the frontal prism and continental framework. *Geochemistry, Geophysics, Geosystems*, **22**.

- Wakita K. (2015) OPS mélange: A new term for mélanges of convergent margins of the world. *International Geology Review*, **57**, 529–539.
- Wang K. and Bilek S.L. (2014) Invited review paper: Fault creep caused by subduction of rough seafloor relief. *Tectonophysics*, **610**, 1–24.
- Wang K. and Hu Y. (2006) Accretionary prisms in subduction earthquake cycles: The theory of dynamic Coulomb wedge. *Journal of Geophysical Research: Solid Earth*, **111**.
- Wang W., Yin H., Jia D., Wu Z., Wu C. and Zhou P. (2018) Calculating detachment depth and dip angle in sedimentary wedges using the area–depth graph. *Journal of Structural Geology*, **107**, 1–11.
- Watson S.J., Mountjoy J.J., Barnes P.M., Crutchley G.J., Lamarche G., Higgs B., Hillman J., Orpin A.R., Micallef A., Neil H., Mitchell J., Pallentin A., Kane T., Woelz S., Bowden D., Rowden A.A. and Pecher I.A. (2019) Focused fluid seepage related to variations in accretionary wedge structure, hikurangi margin, new zealand. *Geology*, **48**, 56–61.
- Watt J.T. and Brothers D.S. (2020) Systematic characterization of morphotectonic variability along the Cascadia convergent margin: Implications for shallow megathrust behavior and tsunami hazards. *Geosphere*, **17**, 95–117.
- Webb S. (2017) interaction of structure and physical properties in accretionary wedges: Examples from the cascadia and nankai trough subduction zones—uwdc - uw-madison libraries.
- Wendt J., Oglesby D.D. and Geist E.L. (2009) Tsunamis and splay fault dynamics. *Geophysical Research Letters*, **36**.

- Wilson A. and Ma S. (2021) Wedge plasticity and fully coupled simulations of dynamic rupture and tsunami in the Cascadia Subduction Zone. *Journal of Geophysical Research: Solid Earth*, **126**.
- Wirth E.A., Sahakian V.J., Wallace L.M. and Melnick D. (2022) The occurrence and hazards of great subduction zone earthquakes. *Nature Reviews Earth & Environment*, **3**, 125–140.
- Witter R.C., Zhang Y.J., Wang K., Priest G.R., Goldfinger C., Stimely L., English J.T. and Ferro P.A. (2013) Simulated tsunami inundation for a range of Cascadia megathrust earthquake scenarios at Bandon, Oregon, USA. *Geosphere*, **9**, 1783–1803.
- Wood N., Peters J., Sheehan A. and Bausch D. (2025) National population exposure and evacuation potential in the united states to earthquake-generated tsunami threats. *International Journal of Disaster Risk Reduction*, **123**, 105511.
- Xu S., Fukuyama E., Ben-Zion Y. and Ampuero J.P. (2015) Dynamic rupture activation of backthrust fault branching. *Tectonophysics*, **644–645**, 161–183.
- Yamaguchi A., Sakaguchi A., Sakamoto T., Iijima K., Kameda J., Kimura G., Ujiie K., Chester F.M., Fabbri O., Goldsby D., Tsutsumi A., Li C.F. and Curewitz D. (2011) Progressive illitization in fault gouge caused by seismic slip propagation along a megasplay fault in the Nankai Trough. *Geology*, **39**, 995–998.

Appendix A

SUPPLEMENTAL MATERIAL TO CHAPTER 5

We collected 60 oriented samples at 13 outcrops along a fault-strike perpendicular transect of the field area (Fig. 1A). Samples for AMS were not collected at site 12 (according to the numbering scheme in Figs 1, S1-3) due to limited exposure. Of these samples, 5 were *mélange* (100 specimens), 7 were fault (95 specimens), and 2 were coherent (29 specimens) rocks. These samples yielded 224 total specimens for AMS in addition to oriented thin sections at each outcrop. We performed Raman spectroscopy of carbonaceous materials (S1) to determine maximum paleotemperatures at each outcrop, AMS and magnetic mineralogy experiments (S2), and chemical mapping using EPMA (S3). Bulk rock chemistry (S4) was performed at samples not along the transect but throughout the field area. The RSCM, AMS, EPMA, and bulk rock chemistry data, as well as the code for the pressure solution and constitutive relations calculations, are available on Zenodo.

A.1 Raman spectroscopy of carbonaceous materials

We performed Raman spectroscopy of carbonaceous material (RSCM) to derive maximum metamorphic temperatures for each outcrop along our fault-parallel transect in a targeted comparison between these units. RSCM measurements were taken on polished thin sections using a Renishaw inVia Qontor confocal Raman microscope in the Sci-Tech facility at Western Washington University, Bellingham, Washington, USA following the methods described in Ledeczi et al. (2025). That work established maximum metamorphic temperatures of $\sim 260 - 280^\circ\text{C}$ for the LOTG region of the OSC. Preliminary findings indicated that there were no significant differences in the

temperatures measured in the more coherent broken formation, the mélange, or the fault rocks.

Applying the RSCM calibration of Lünsdorf et al. (2017), we estimate that the range of sample paleotemperatures at LOTG is $253 - 282^{\circ}\text{C}$. Our results here underscore those in Ledeczi et al. (2025) that these samples show no systematic deviation of temperatures with brittle or ductile fabric development. The RSCM method has errors of $\pm 50^{\circ}\text{C}$ (Beyssac et al., 2002), which are reduced in the calibration and thermometer of Lünsdorf et al. (2017) to $\pm 36^{\circ}\text{C}$. On Fig. S1, we report the standard deviation of the measurements taken in each thin section.

While in our previous work (Ledeczi et al., 2025), we use both thermometers of Kouketsu et al. (2014) and Kaneki and Kouketsu (2022) and Lünsdorf et al. (2017), here we only employ the latter. The peak fitting approach of Kaneki and Kouketsu (2022) only uses a small background window near the carbonaceous material peaks and treats the background as linear, while Lünsdorf et al. (2017) models the full width of the measured without any assumption for the shape of the baseline beneath the carbonaceous material peaks. Where the two fits conflict in (Ledeczi et al., 2025), this is usually due to the sample having a curved background, causing the K14 routine to underestimate the true temperature. Therefore, the best values for temperatures derived at LOTG, combining the results of (Ledeczi et al., 2025) and those herein, is $253 - 282^{\circ}\text{C}$. Contrary to some previous work (Aoya et al., 2010; Moris-Muttoni et al., 2023), our findings indicate that our RSCM values are not influenced by strain intensity. Spectra from both coherent samples and highly deformed cataclasites and fault gouges are similar and have overlapping error bars (Fig. S1 herein, Ledeczi et al., 2025). Aoya et al. (2010) note that their increased values for highly strained rocks are within the $\pm 50^{\circ}\text{C}$ error range of their applied calibrations, so it is possible any signature of higher temperatures due to strain in our samples are concealed by uncertainty. Also, we do not have samples of concentrated principal slip surfaces like those measured in Moris-Muttoni et al. (2023), so this may impact our record of this

signal.

A.2 AMS and magnetic mineralogy

Low-field AMS measurements were collected on an AGICO KLY-3S Susceptometer at the Pacific Northwest Paleomagnetism Laboratory at Western Washington University. Magnetic susceptibility was measured using the rotating sample head, with a measurement made every 0.8 degrees, in three orthogonal planes, and from those data the AMS matrix with eigenvalues and eigenvectors corresponding to the three axes of the AMS ellipsoid was determined. A geographic correction was applied to the AMS results in order to link the AMS axes to field sample orientations (Table 1). The data confidence ellipse and average axes for each unit was analyzed using Jelínek (1978) statistics with Anisoft 6. The structural data presented in Ledeczi et al. (2025) were analyzed using Joshua Davis’s geologyGeometry library in R <http://www.joshuadavis.us/software/> and via the methods described in Davis and Titus (2017) and Roberts et al. (2019). AMS axes for each of the specimens in each of the units in geographic coordinates are shown in Fig. S2.

The trend of plunge of the average of the phacoid long axes at LOTG is 71, 100 (Ledeczi et al., 2025). The trend and plunge of the average K1 axis values for the mélange and faults is 76.2, 081.8 and 76.8, 098.7, yielding angular distances from the phacoid long axes average of 7.25° and 5.81°, respectively. The trend and plunge of the average K3 axis in the mélange is 11.2, 225.7 and the average mélange foliation strike and dip is 314/75 (Ledeczi et al., 2025), a distance of 4.14°. The trend and plunge of the average K3 axis in the faults is 7.9, 225.1 and the average fault foliation strike and dip is 320/71 (Ledeczi et al., 2025), a distance of 12.07°.

We analyzed the magnetic mineralogies of representative samples to ensure the observed AMS was independent to lithological variation between the three units. We grouped the bulk susceptibilities by unit (Fig. S4) and there is no significant difference, indicating a first-order mineralogical similarity.

We also measured the temperature-dependent magnetic susceptibility on the KLY-3 susceptometer coupled to a CSL cryostat through low-temperature experiments at -192° to 0°C (Fig. S5) and coupled to a CS3 furnace for high-temperature experiments at 40° to 700°C (Fig. S6). The experiments were conducted under Argon to minimize oxidation. The bulk magnetic susceptibility was recorded every $1.5 - 1.8^{\circ}\text{C}$ during the low-temperature experiments, and during both heating and cooling back to room temperature during the high-temperature experiments in order to detect potential mineralogical changes that could be triggered by the heating. All results shown are corrected to the signal of the empty holder and the bulk susceptibility of the nearest AMS specimen.

The samples display a concave hyperbolic curve in the low-temperature experiments (Fig. S5), as well as at lower temperatures in the high-temperature experiments (Fig. S6), indicating a dominant paramagnetic mineralogy in all three units, which in the studied lithologies we interpret are the phyllosilicate minerals. During the high-temperature experiments, samples typically experienced oxidation as indicated by the very different magnetic susceptibility results between the heating run and the following cooling run of each experiment. Many samples showed drops in magnetic susceptibility at characteristic temperatures of $\sim 400 - 500^{\circ}\text{C}$ and $\sim 550 - 590^{\circ}\text{C}$ (Fig. S6), which we interpret to be associated with the Curie temperatures of titanomagnetite and magnetite, respectively.

Hysteresis loops were also performed on powdered samples on a MicroMag 3900 Vibrating Sample Magnetometer. This entails measuring the magnetic moment of a sample when applying a field of increasing then decreasing intensity in one direction, then the same in the opposite direction. The maximum applied field intensity was 1 T. Results show a positive slope where the two halves of the loop are superimposed (Fig. S7A); this indicates that all measured samples, regardless of unit, are dominated by paramagnetic minerals. When the paramagnetic effect is subtracted from the results, the hysteresis loops show a very low coercivity and magnetic saturation,

which confirms that any ferromagnetic material present is likely (Ti-) magnetite, as well as noisy behavior, potentially suggesting that the ferromagnetic fraction can be likely considered negligible in concentration (Fig. S7B). This is notable because ferromagnetic minerals can dominate the magnetic signal even in low concentrations and may not be representative of the structural fabric (e.g., Borradaile et al., 1987). Repeat experiments were performed on many of these samples to confirm that noise was not due to instrumental or methodological factors.

A.3 Chemical analyses

We collected both quantitative spot analyses and qualitative chemical maps on the JEOL JXA-iHP200F field emission electron microprobe at the University of British Columbia. The microprobe has five wavelength-dispersive spectrometers and dual Bruker 30 MM2 energy dispersive X-ray detectors. For the spot analyses, we used an accelerating voltage of 15 kV, beam current of 20 nA, and the beam diameter was 5 microns. Elements were acquired using analyzing crystals LIFL for Fe $k\alpha$, Mn $k\alpha$, Ti $k\alpha$, PETL for K $k\alpha$, Ca $k\alpha$, TAP for Si $k\alpha$, Al $k\alpha$, TAPL for Na $k\alpha$, Mg $k\alpha$. All X-ray intensities were acquired and processed within the Probe for EPMA software (Probe Software Inc). The on and off-peak counting time was 15 seconds for Si $k\alpha$, Mg $k\alpha$, Fe $k\alpha$, Ca $k\alpha$, Al $k\alpha$, Mn $k\alpha$, Na $k\alpha$, K $k\alpha$, and 30 seconds for Ti $k\alpha$. Linear background fits were used for all elements. Unknown and standard intensities were corrected for deadtime using the Super High Precision (six terms) correction method (Donovan et al., 2023). Reference materials used for calibration included Albite (SPI), NaAlSi₃O₈ for Si $k\alpha$, Al $k\alpha$, Na $k\alpha$, Bustamite (SPI), (Mn,Ca)SiO₃ for Mn $k\alpha$, Diopside (SPI), MgCaSi₂O₆ for Ca $k\alpha$, Mg $k\alpha$, Hematite (SPI), Fe₂O₃ for Fe $k\alpha$, Rutile (SPI), TiO₂ for Ti $k\alpha$, and Orthoclase (SPI), KAlSi₃O₈ for K $k\alpha$. Standard intensities were corrected for standard drift over time on an element-by-element basis.

The total amount of volume strain due to pressure solution can be constrained by comparing the enrichment of fluid-immobile elements like Ti and the depletion of

fluid-mobile elements like Si along shear surfaces. The volume loss due to pressure solution was calculated via the method of Chen et al. (2024a), which is based on previous work (Ague, 1991, 1994; Aitchison, 1992). This assumes that the mass of Ti is constant during mass transfer which causes an increase in concentration along the scaly fabrics. Thin sections are XZ planes which should capture one of two maximum directions of shortening due to foliation-parallel pressure solution (as assumed by Ague, 1991, and which seems to be a valid assumption for these rocks). We slightly modified the Chen et al. (2024a) method, which uses spot analyses from LA ICP-MS, to the data collected here via EPMA. The ERR % reported from EPMA is relative error in percent, not weight percent, as calculated from counting statistics. When there is a low element abundance, the counting statistics are poor, but when there is a higher element abundance the errors are much better. However, the low elemental abundance samples can be important and valid, such as for when Ti is low in wall rock due to low pressure solution. When the Monte Carlo sampling method employed in CChen et al. (2024a) uses draws with high relative errors, it can lead to a noisy mean with a high uncertainty. Thus, we sample in log space instead which avoid this issue and provides a more robust uncertainty calculation. Additionally, rather than using pairs of points inside and outside a single scaly fabric, we instead calculate the volume loss between each two pairs of points on the EPMA map, providing a more robust estimate. We validated our code by reproducing the pressure solution values calculated in Chen et al. (2024a) and reported in their supplement using this method.

First, we calculated the individual volume strain along a single scaly fabric as an average of the 20 point measurements. The area of the scaly fabric was then calculated using image analysis in ImageJ. We estimated the scaly fabric area on both Fe and Si EPMA maps and use the median of the two percent values as the area and report the standard deviation as the error on Fig. 3B. Finally, we multiplied these values together to derive the total volume loss at the map scale and propagated the errors of the two previous calculations.

A.4 Bulk rock chemistry

Major and trace elements were analyzed via the ThermoARL Advant'XP+ sequential X-ray fluorescence spectrometer housed at Washington State University. Representative 3 g cuts of whole samples excluding any weathering surfaces were sent for powdering.

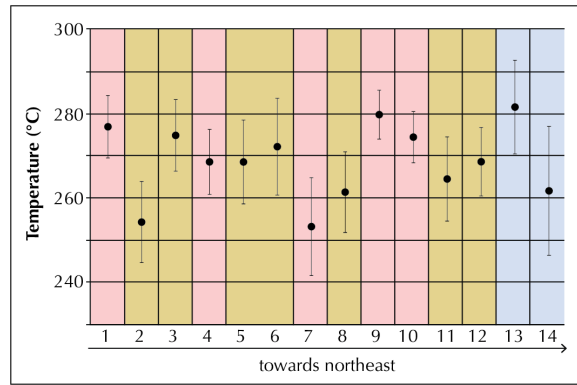


Figure A.1: RSCM data at samples from the transect using the calibration of Lünsdorf et al. (2017). Error bars show the standard deviation of the measurements taken in each thin section. Color scheme follows the figures from the remainder of the paper.

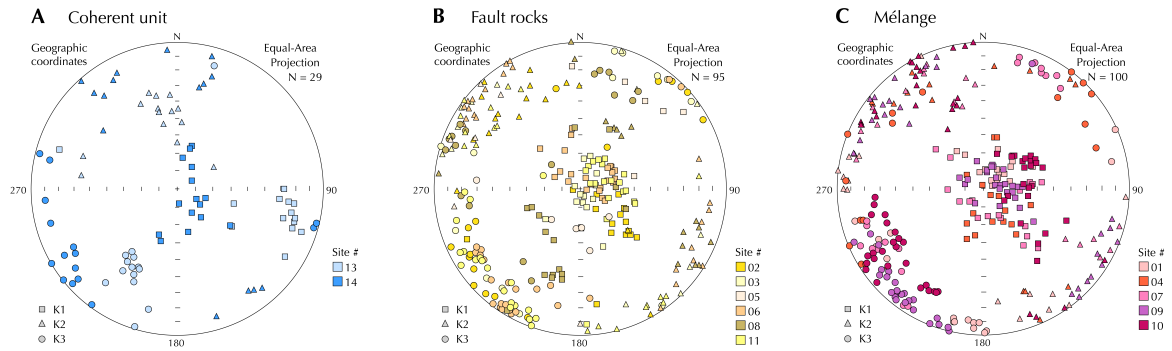


Figure A.2: AMS axes for the **A** coherent, **B** fault, and **C** mélange rocks. Square, triangle, and circle symbols are K1, K2, and K3 axes, respectively. The site numbers correspond to those on the map in Fig. 1A. This is the same data as shown with confidence ellipses in Fig. 2A of the main text.

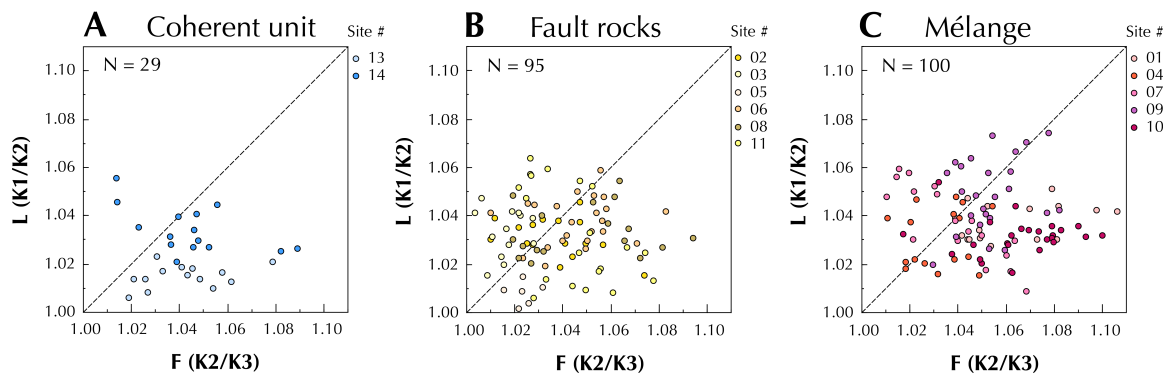


Figure A.3: The site numbers correspond to those on the map in Fig. 1A.

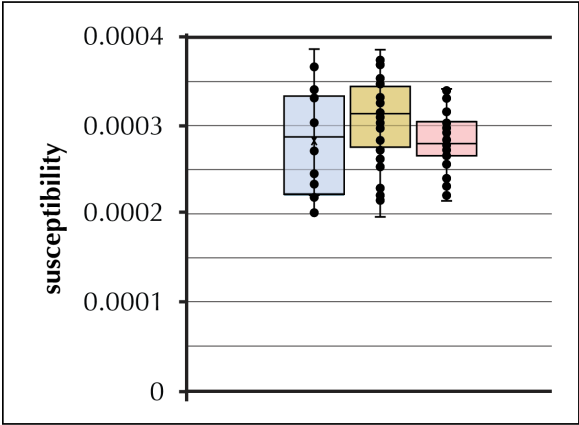


Figure A.4: Bulk susceptibilities (SI volume units) from the three units.

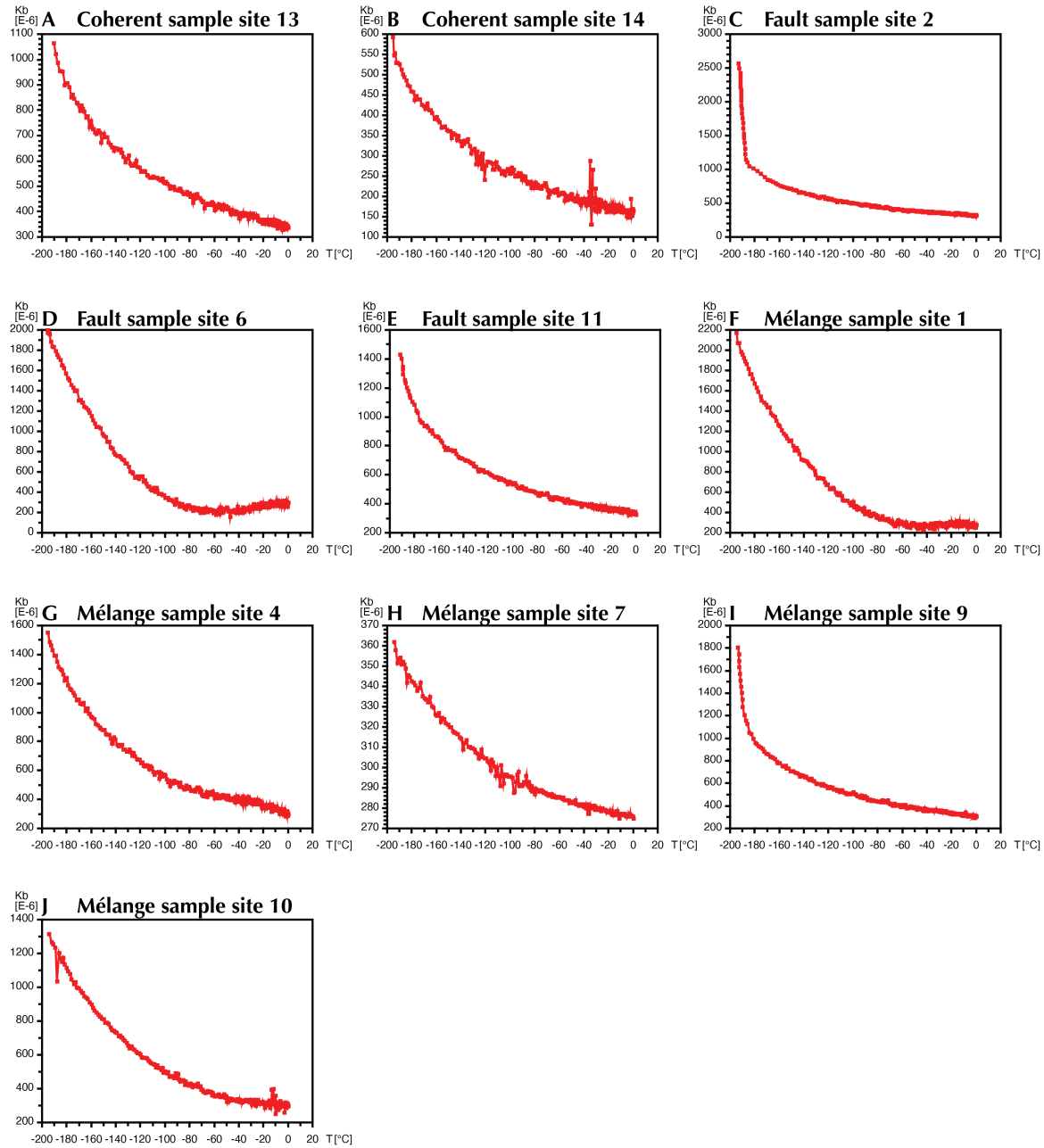


Figure A.5: Low temperature experiments, run from -192° to 0°C . Experiments are corrected to the susceptibility of the empty vessel used in the experiments and to the bulk susceptibility of the nearest AMS specimen. The site numbers correspond to those on the map in Fig. 1A.

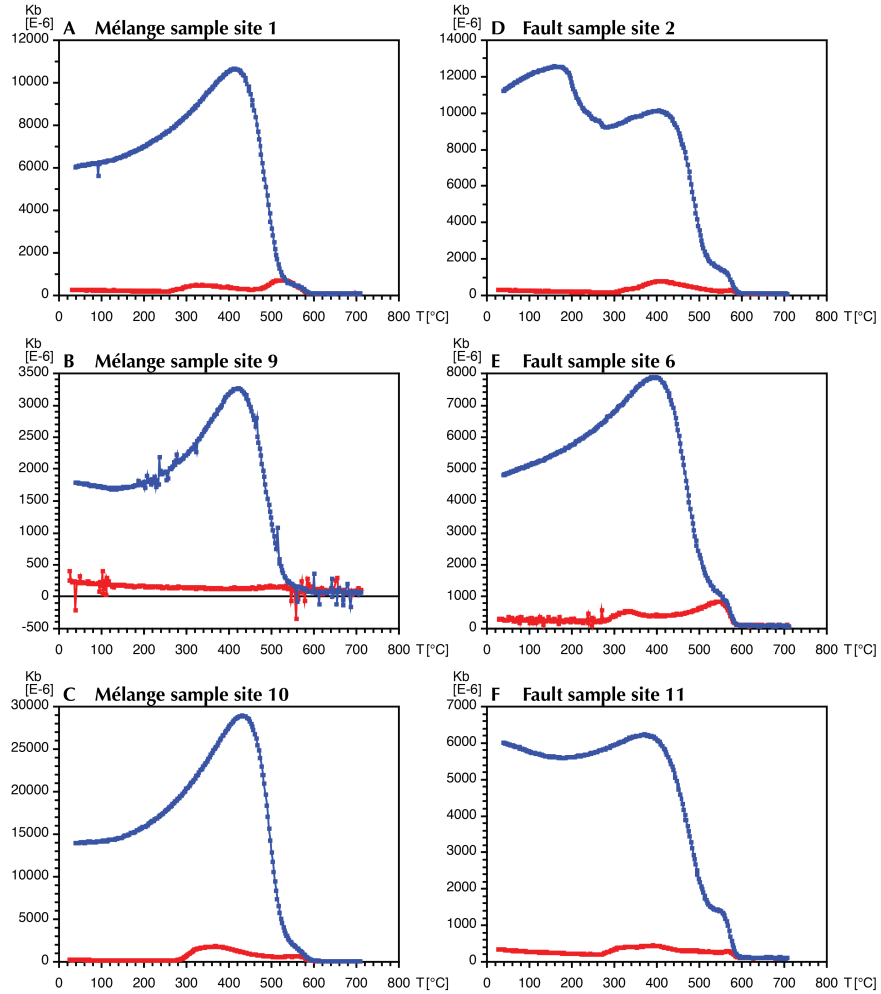


Figure A.6: Temperature vs. magnetic susceptibility curves for high-temperature experiments for mélange **A-C** and fault **D-F** samples, run from 40 $^{\circ}\text{C}$ to 700 $^{\circ}\text{C}$ (heating runs, red) and back to 40 $^{\circ}\text{C}$ (cooling runs, blue). Experiments are corrected to the susceptibility of the empty vessel used in the experiments and to the bulk susceptibility of the nearest AMS specimen. The site numbers correspond to those on the map in Fig. 1A.

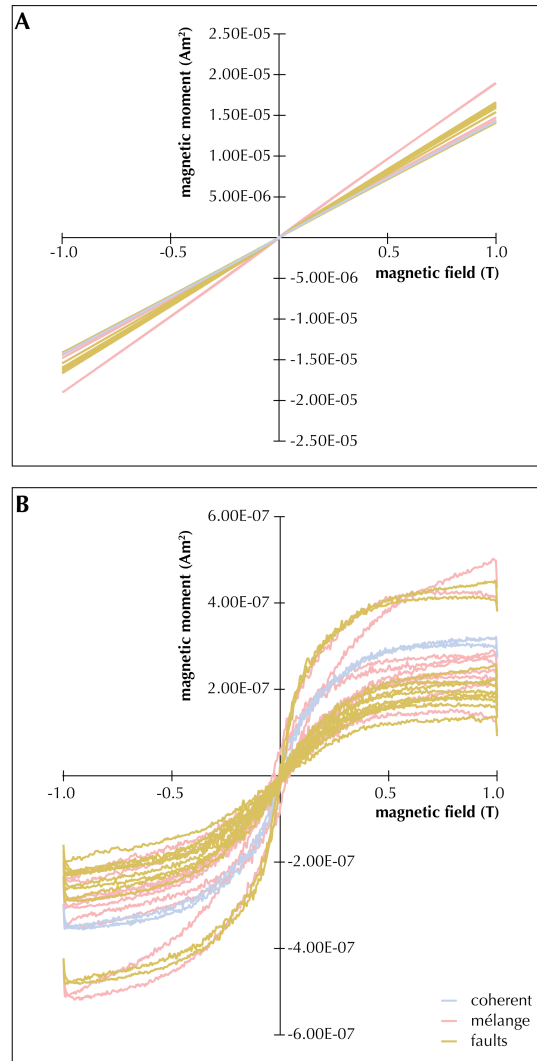


Figure A.7: Hysteresis loops of samples show **A** samples from all units are dominated by the paramagnetic signal, and **B** the moment of the remanent ferromagnetic fraction is small with the paramagnetic signal subtracted. The ferromagnetic fraction has very low coercivity after subtracting the ferromagnetic signal.