

STRAIN ACCOMMODATION ON FOREARC FAULTS: A CASE STUDY ON THE
BEAUFORT RANGE FAULT, AN ACTIVE CRUSTAL FAULT IN THE NORTHERN
CASCADIA FOREARC, VANCOUVER ISLAND, BC, CANADA

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A Dissertation

Submitted in Partial Fulfillment

of the Requirements for the Degree of

Doctor of Philosophy

in Earth Science and Environmental Sustainability

Northern Arizona University

December 2023

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ABSTRACT

STRAIN ACCOMMODATION ON FOREARC FAULTS: A CASE STUDY ON THE BEAUFORT RANGE FAULT, AN ACTIVE CRUSTAL FAULT IN THE NORTHERN CASCADIA FOREARC, VANCOUVER ISLAND, BC, CANADA

EMERSON M. LYNCH

The distribution of permanent deformation on crustal faults in subduction zone forearcs is a product of subduction obliquity, the geometry of the plate interface, and the structure and materials of the overriding plate. Only a portion of the strain accumulated during the interseismic period of the subduction zone seismic cycle is released in large megathrust earthquakes. The remaining accumulated strain is thought to be accommodated through permanent deformation of the upper plate, primarily localized on forearc faults. Therefore, it is important to quantify the magnitude and kinematics of deformation on these forearc faults to understand how the upper plate accommodates tectonic stresses and the hazards these faults pose to the millions of people who live in highly populated coastal regions near plate margins.

The Cascadia Subduction Zone (CSZ) is located off the west coast of the USA and Canada, where the Juan de Fuca plate subducts obliquely under the North American plate. Many faults have been documented throughout the southern and central CSZ forearc, but no active faults have been documented in the northernmost ~ 150 km of the forearc. This may reflect a change in strain accumulation near the USA-Canada border, resulting in minimal permanent deformation in the northernmost CSZ forearc. However, this is inconsistent with the kinematics of the Leech River fault on southern Vancouver Island, and the largest crustal earthquake in Canadian history, the June 23, 1946 M_S 7.3 Vancouver Island earthquake, occurring in the northernmost CSZ forearc.

In this dissertation, I identify the first documented active fault in the northernmost ~ 150 km of the CSZ forearc, the Beaufort Range fault (BRF) on central Vancouver Island, BC, Canada. In Chapter 2, I use lidar and field mapping to identify >35 km of Quaternary fault scarps. Through surficial mapping and topographic surveys, I determine that the BRF has hosted at least three earthquakes since ~ 15 ka. Kinematic inversions of offset geomorphic markers and fault mapping show that the BRF accommodates right-lateral transtension along a steeply NE-dipping fault. This fault geometry and kinematics are similar to those modeled for the 1946 earthquake, suggesting that the BRF is a candidate source fault for this event. In Chapter 3, I refine this chronology using paleoseismic trenching to document five M_W 6.7-7.5 surface-rupturing earthquakes since ~ 11 ka. Recurrence intervals and slip rates on the BRF are non-steady, with shorter recurrence and faster slip rates in the middle and late Holocene. In Chapter 4, I examine recurrence and slip rates on thirteen

northern CSZ forearc faults. I find that recurrence is non-steady on eight of nine faults, and slip rates are non-steady on four of five faults. Faults tend to have long recurrence intervals in the late Pleistocene to early Holocene, and short recurrence intervals in the middle and late Holocene. Furthermore, recurrence and displacement on these forearc faults are inconsistent with periodic, time-predictable, or slip-predictable models of earthquake recurrence.

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Acknowledgements

Land Acknowledgement

This work was conducted on the traditional territory of the Hupačasath, K'ómoks, and *ciisaa?ath* people, and we acknowledge their historical relationships with the land that continue to this day. During my Ph.D., I have lived and worked on land sacred to Massa-adchu-es-et (Massachusetts), Pawtucket, Monacan, *dxk^wsəqk^wəbš* (Suquamish), *dxk^wdəw?abš* (Duwamish), and *bəqəlšut* (Muckleshoot) Nations. Northern Arizona University sits at the base of the San Francisco Peaks, on homelands sacred to Native Americans throughout the region. We honor their past, present, and future generations, who have lived here for millennia and will forever call this place home.

Acknowledgements

Note: Manuscript-specific acknowledgments are found at the end of each corresponding chapter.

Thank you to Johanna Fischl (BU '19), Tatum McLeod (WWU '18), Maximilian Lauch (UVic MS '22), and Megan Caston (UVic MS '21) for their assistance in the field, and undergraduate researchers Kim Hughes (BU '18), Nichole Kulikowski (BU '18), Carina Terry (BU '20), Amanda Binkley (NAU '20), and Emma Martin (NAU '20) for their work on campus. I've also appreciated the opportunity to work with UVic students in the field on the ~~XEOLXELEK~~-Elk Lake project, including Theron Finley, Guy Solomon, and Israporn (Grace) Sethanant.

Cody from the Port Alberni Staples printed and scanned so many different files for us during trenching. Squirrel Chops (Seattle), Late for the Train (Flagstaff), and Season's Yield (Lexington) kept me caffeinated.

Radiocarbon processing was initially started by the late Alice Telka. After her passing, R. Scott Anderson, Darrell Kaufman, Jordon Bright, and Chris Ebert helped me learn how to prepare and process radiocarbon samples.

Thank you to my committee, Darrell Kaufman, Kristin Morell, and Scott Bennett for supporting me in field work, research discussions, confusing radiocarbon results, paper submissions, and finding my phone in the desert. Christine Regalla has been the best advisor I could ask for, through two institutions, a cross-country move (or several), a global pandemic, about six months in the field, and so much more. Thank you

also for cleaning that fish; it was delicious.

Thank you to my friends and classmates at Smith, BU, and NAU, particularly Betsy Cowdery for her help with R. Lexi Riche, Eva Baransky, and Emily Schottenfels have been the best office mates I could ask for, and Lexi continued to put up with me as her roommate for two years. I can't thank Emily enough for her friendship, support, and voice of reason since day 1 (and for making sure I never think of spinach the same way again). Thank you to Jeff Gertler for always reminding me to rest, and for helping me through Honors Physics in high school so I could make force diagrams for the rest of my life.

Thank you to my family for always supporting me: Dad, Keegan, my grandparents, all my aunts and uncles and cousins, and of course my mom. Thank you for everything.

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Preface

The manuscript chapters presented here are written in a format for publication in scientific peer-reviewed journals. The first chapter consists of a general introduction to the entire thesis project and motivation for the study. The second chapter is formatted for publication in *Journal of Geophysical Research: Solid Earth*. On 5 September, 2023, it received three positive reviews with minor to major revisions, but was rejected by the Associate Editor for perceived fit and data management compliance. We plan to complete these revisions in October 2023. The third chapter is formatted for submission to *Seismica*. The fourth chapter is formatted for submission to *Tektonika*. The fifth and final chapter summarizes the overall conclusions of the dissertation. These articles are arranged as chapters in accordance with Northern Arizona University's dissertation guidelines, which results in some redundancy. All figures, tables, and literature cited are presented within the individual chapters.

Chapter 1

Introduction

A valley leads from this end of Comox Lake, south-westerly to Alberni, a fact which may or may not be significant. There is a possibility that a disturbed fault leads down this valley, but no evidence could be obtained in such wild country.

Hodgson (1946)

1.1 Motivation

How the forearcs of subduction zones deform in response to tectonic stresses has long been an important question both for our understanding of subduction zone processes and the hazards they pose to the millions of people who live in highly populated coastal regions near plate margins (e.g., Fitch, 1972). Subduction zones accumulate strain while locked, and release a portion of that strain during large megathrust earthquakes after interseismic periods that last tens to hundreds of years. Some of the accumulated strain, however, is not released in these large earthquakes, but rather is thought to be accommodated through permanent deformation of the upper plate (e.g., Wang, 2000). This permanent deformation is not evenly distributed throughout the forearc and varies in magnitude and kinematics due to subduction obliquity, the geometry of the plate interface, and the structure and materials of the overriding plate (e.g., ten Brink & Lin, 2004; Escuder-Viruete & Pérez, 2020; McCaffrey, 1992). It is critical to understand where this deformation is localized on individual forearc faults, both to understand the potential drivers of forearc deformation, and to characterize the hazards posed by these faults.

In this study, I focus on the Cascadia Subduction Zone (CSZ), located off the West Coast of the USA and Canada, where the Juan de Fuca plate subducts under the North America plate (Figure 1.1). The CSZ accommodates ~ 43 mm/yr of dextral oblique subduction (Kreemer et al., 2014), and many forearc faults

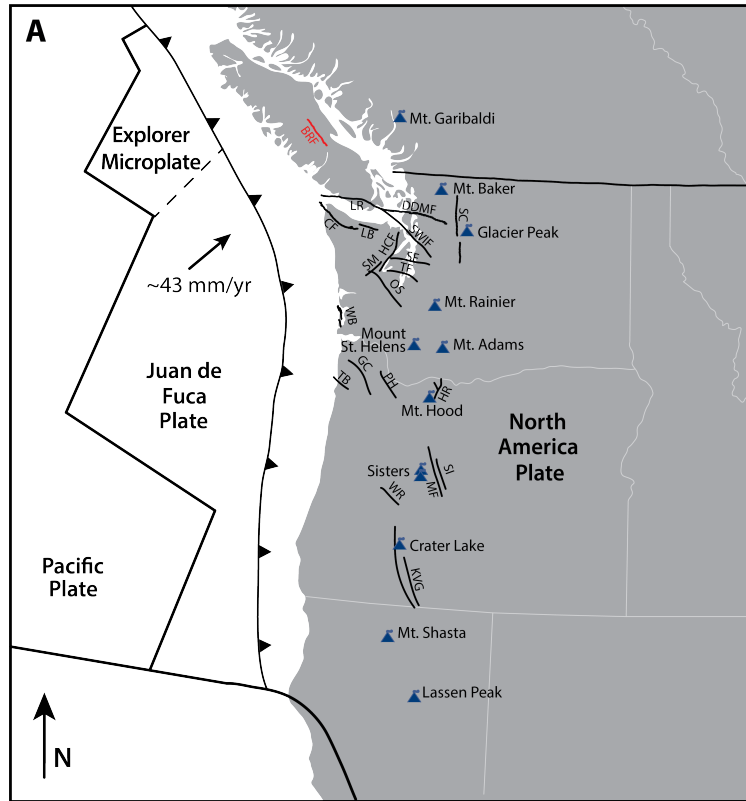


Figure 1.1: Regional tectonic setting of the Cascadia Subduction Zone. BRF – Beaufort Range fault; CF – Calawah fault; CL – Cowichan Lake fault; DDMF – Darrington-Devils Mountain fault; FF – Fulford fault; GC – Gales Creek fault zone; HCF – Hood Canal fault zone; HR – Hood River fault zone; KVG – Klamath Valley graben; LB – Lake Creek-Boundary Creek fault; LR – Leech River fault; MF – Metolius fault zone; OS – Olympia structure; PH – Portland Hills fault; SC – Straight Creek fault; SF – Seattle fault zone; SI – Sisters fault zone; SJ – San Juan fault; SM – Saddle Mountain faults; SWIF – Southern Whidbey Island fault zone; TB – Tillamook Bay fault zone; TF – Tacoma fault; WB – Willapa Bay fault zone; WR – Upper Willamette River fault zone. Quaternary-active faults simplified from U.S. Geological Survey (2018) and Morell et al. (2017), volcanoes after AGI (2003).

have been documented from the Mendocino Triple Junction in northern California to southern Vancouver Island, BC, Canada (Figure 1.1; Delano et al., 2017; Bennett et al., 2017; Finley et al., 2019; Johnson et al., 2004). However, no active faults have been documented in the northernmost ~150 km of the forearc (north of the Leech River fault; Figure 1.1). Some suggest this is because there is no deformation (e.g., Wells et al., 1998; McCaffrey et al., 2007, 2013), in which case there must be a significant change in strain accumulation near the USA-Canada border. However, there is evidence for active deformation in this region, including fault scarps, offset Quaternary deposits and landforms, and several crustal earthquakes.

In this dissertation, I first focus on the Beaufort Range fault (BRF), located on central Vancouver Island (Figure 1.1), to evaluate its Quaternary slip history. The largest crustal earthquake in Canadian history, the June 23, 1946 Ms 7.3 Vancouver Island earthquake, occurred at the northern end of the BRF,

prompting Rogers and Hasegawa (1978) and Slawson and Savage (1979) to suggest that the earthquake may have re-occupied an Eocene structure. Rogers and Hasegawa (1978) relocated the epicenter of the 1946 event, and determined that it likely occurred in the upper plate of the Cascadia Subduction Zone, and may have accommodated right-lateral transtension on a NW-SE-striking, steeply dipping fault, sub-parallel to the BRF. Geodetic surveys by Slawson and Savage (1979) showed deformation consistent with right-lateral transtension on the BRF, and Mathews (1979) described numerous landslides from historical air photos in a swatch parallel to the BRF. Hodgson (1946) and Fyles (1963) both noted the steep front of the Beaufort Range, but no previous studies have identified any surface ruptures from the 1946 event or any other Quaternary earthquake, in part due to the dense forest. High-resolution lidar topographic data recently collected along the Beaufort Range allows us to examine topography in detail despite the ground cover. I use these lidar data, extensive field surveys, and paleoseismic trenching to characterize the length, kinematics, and rupture history of the BRF in order to quantify the amount of strain accommodated in the northernmost Cascadia forearc. I then broaden my scope to faults throughout the northern CSZ forearc, using paleoseismic data from thirteen forearc faults to investigate regional patterns of earthquake timing and recurrence.

1.2 Summary of dissertation research

In Chapter 2, I document the first evidence that the BRF is Quaternary-active, and that there is active strain accumulation in the northernmost Cascadia forearc. I leverage newly available lidar data to map ≥ 60 km of 100s-1000s m-long active fault strands in a zone ~ 500 m wide. Through surficial mapping, I identify offset post-glacial deposits and landforms that indicate the BRF has hosted at least three earthquakes since glacial retreat. I quantify the three-dimensional offset of >60 displaced linear geomorphic piercing lines, and show that the BRF has one of the fastest slip rates in the northern CSZ forearc. I also demonstrate that the BRF accommodates right-lateral transtension, distinct from the compression documented on CSZ forearc faults farther south. The kinematics of the BRF and estimates of per-event displacement are consistent with relocated focal mechanism solutions and geodetic displacements from the 1946 M_S 7.3 Vancouver Island earthquake (e.g., Rogers & Hasegawa, 1978; Slawson & Savage, 1979). These results suggest the BRF may be the source fault for the 1946 M_S 7.3 Vancouver Island earthquake.

Chapter 2 and Appendices A and B are the result of a collaboration with Christine Regalla, Kristin D. Morell, Nicolas Harrichhausen, and Lucinda J. Leonard. Bedrock and Surficial mapping was completed over four field seasons from 2017 to 2019 with coauthors and field assistants Johanna Fischl, and Tatum McLeod. Topographic surveys were collected in the summer and fall of 2018 with Nicolas Harrichhausen,

Lucinda J. Leonard, Scott E.K. Bennett, Johanna Fischl, and Tatum McLeod. Radiocarbon samples were processed by the late Alice Telka at Paleotec Services, Ottawa, ON. I wrote the majority of the R scripts, and adapted components of the R scripts from MATLAB scripts by Christine Regalla. This chapter benefited from discussions with Scott E.K. Bennett, and from significant editing by Christine Regalla and Kristin D. Morell during preparation for submission to *JGR: Solid Earth*.

In Chapter 3, I develop a detailed earthquake history for the BRF and refine the slip rates estimated in Chapter 2. Precise earthquake ages are necessary for estimating recurrence intervals, a key parameter of seismic hazard assessments. Through paleoseismic trenching, I identify five surface-rupturing earthquakes on the BRF since deglaciation ~ 13.6 -11 ka (e.g., Mosher & Hewitt, 2004; Easterbrook, 1992; Clague et al., 1980). The oldest of these earthquakes occurred ~ 10.9 -10.3 ka, followed by a ~ 5 -6 kyr quiet period before a cluster of four earthquakes occurred ~ 5 -2.5 ka. Each earthquake resulted in ~ 0.75 -1.5 m right-lateral oblique displacement. Together with the varying recurrence intervals, closed-interval slip rates increase from ~ 0.2 mm/yr in the early Holocene to >2 mm/yr in the middle and late Holocene. The changing recurrence and slip rates suggest that there may be a tectonic or climatic factor may modulate the timing of rupture on the BRF.

Chapter 3 and Appendix C are the result of a collaboration with Christine Regalla, Kristin D. Morell, Scott E.K. Bennett, Nicolas Harrichhausen, Lucinda J. Leonard, and Edwin Nissen. Walter Langer excavated the trench in June 2019 after trees at the site were cleared by Todd Thompson. Initial trench logs and interpretations were completed collaboratively in the field, with additional discussion with John Clague, Colin Amos, and Liz Schermer. I was assisted in the trench by Maximilian Lauch. Lucinda J. Leonard and Megan Caston completed electrical resistivity tomography surveys that aided our interpretation of the site's structure. A preliminary set of thirteen radiocarbon samples were processed at the UC Irvine Keck AMS facility. I processed the remaining radiocarbon samples at NAU with assistance from R. Scott Anderson and Darrell Kaufman. Jordon Bright and Chris Ebert completed ABA processing, combustion, and graphitization at NAU in the Arizona Climate and Ecosystems lab.

In Chapter 4, I consider whether the non-steady recurrence and slip rates I documented on the BRF occur on other northern CSZ forearc faults. I analyze earthquake timing, displacement, and recurrence for thirteen active faults in the northern CSZ forearc and calculate slip rates for a subset of five faults. I wrote an R script for calculating closed-interval slip rates (available in Appendix D) based on the RISEr algorithm by Zinke et al. (2017) and calculated the likelihood of earthquake age PDF overlap using a script from Regalla et al. (2022). I identify decreasing recurrence intervals and increasing slip rates in the mid-Holocene and propose that this is the result of a regional change in strain accommodation. I also discuss the three primary models for earthquake recurrence used in probabilistic seismic hazard assessment and whether they

are consistent with the behavior of CSZ forearc faults. The research presented in Chapter 4 was a project of my own design and completed by me, aided by discussions with Christine Regalla.

1.3 Additional research contributions

Throughout my graduate work, I have benefited from being part of a large, international group working on forearc faulting on Vancouver Island, BC. With this group, I have participated in paleoseismic trenches on two other faults on Vancouver Island. I participated in paleoseismic trenching for the first time in 2018 as part of a project led by Nicolas Harrichhausen on the Leech River fault. I gained experience with photomosaic processing, trench logging, and radiocarbon collection, which I was later able to put to use in my work for Chapter 3. The results of these trenches were published as “Paleoseismic trenching reveals Late Quaternary kinematics of the Leech River fault: Implications for forearc strain accumulation in northern Cascadia” in *Bulletin of the Seismological Society of America* (Harrichhausen et al., 2021). I returned to Canada in 2021 to collaborate on a trench on the ~~XEOLXELEK~~–Elk Lake fault in metro Victoria, BC. The results of this project are in review at *JGR: Solid Earth* as “Holocene earthquake rupture of the ~~XEOLXELEK~~–Elk Lake fault in the Greater Victoria area, Canada” (Harrichhausen et al., n.d., *in review*).

Additionally, in 2018 I joined our lab group in the Eastern California Shear Zone for work on the Ash Hill fault. We collected drone imagery using our drone, Droney McDroneface, to build high-resolution digital surface models for use in mapping cm-scale fault scarps and offsets. I used the total station skills I developed working on Chapter 2 to survey the locations of ground control points. I learned about optically-stimulated luminescence sampling and collected samples for radiocarbon dating. While these samples ended up being unsuitable for radiocarbon, I learned sieving and flotation techniques I later used in my work for Chapter 3. I also learned that a drying oven is an optimal temperature for sprouting unidentified wind-blown seeds. The results of this project were published as “Late Holocene rupture history of the Ash Hill fault, Eastern California Shear Zone, and the potential for seismogenic strain transfer between nearby faults” in *Earth Surface Processes and Landforms* (Regalla et al., 2022).

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Chapter 2

Late Pleistocene to Holocene transtension in the northern Cascadia forearc: Evidence from surface ruptures along the Beaufort Range fault

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Abstract

The active deformation field in subduction forearcs provides critical information about the stress and strain state of the upper plate and its potential for seismogenesis. However, these properties are challenging to quantify in most subduction systems, and in the northern Cascadia forearc, few faults have been identified that can be used to reconstruct the upper plate deformation field. Here we investigate the slip history of the Beaufort Range fault (BRF) on Vancouver Island. This fault was proposed to host the 1946 M7.3 Vancouver Island earthquake, but no surface rupture or evidence of Quaternary activity has been documented, and the stress and strain conditions that promoted this event are poorly understood. We provide the first evidence that the BRF is active, using newly-collected lidar to map topographic scarps along the fault system and to reconstruct 3D slip vectors from offset geomorphic markers. Quaternary deposits and landforms that show increasing magnitude of displacement with age provide evidence for at least three $M_W \sim 6.5-7.5$ earthquakes since ~ 15 ka, with the most recent event occurring $<3-4$ ka. kinematic slip inversions of geomorphic piercing lines offset by the BRF show that the BRF accommodates right-lateral transtension along a steeply NE-dipping fault. This fault geometry and kinematics are similar to those modeled for the 1946 earthquake, suggesting that the BRF is a candidate source fault for this event. We find that the kinematics of the BRF are consistent over decadal to millennial timescales, suggesting that this portion of the northern Cascadia forearc has accommodated transtension over multiple earthquake cycles.

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Plain Language Summary

Subduction zones, like Cascadia, contain onshore fault networks that can host earthquakes that are dangerous to communities. However in many locations, like Vancouver Island, Canada, we know little about where these faults are and what type and magnitude earthquake they can host (if any). We focus on the Beaufort Range fault (BRF) on Vancouver Island, and show for the first time that the BRF hosted recent earthquakes. Newly-available high-resolution topography data show many scarps, or vertical offsets of the ground surface produced in past earthquakes, along a >40 km zone. Surveys of landforms that have been offset by the BRF show both vertical and horizontal offsets along a near-vertical fault. The nearby 1946 Vancouver Island earthquake had similar vertical and horizontal offsets along a near-vertical fault, suggesting that this earthquake might have happened on the BRF. Our data show there have been >3 large earthquakes on the BRF in the past ~15,000 years, the most recent <3,000-4,000 years ago. The offsets we observe suggest these earthquakes had magnitudes between ~6.5 and 7.5. Future similar earthquakes could cause shaking damage to many nearby communities, including the cities of Port Alberni and Nanaimo, and nearby hydroelectric facilities.

2.1 Introduction

Quantifying the stress state and strain history of subduction zone forearcs is critical for understanding the energy budget of convergent margins (e.g., Huang et al., 2022), the seismic potential and hazard of forearc faults (e.g., Wang et al., 1995; Balfour et al., 2011; Thenhaus & Campbell, 2002), and the evolution of the upper plate during the megathrust seismic cycle (e.g., Regalla et al., 2017; Herman & Govers, 2020). However, stress is notoriously difficult to measure or approximate, and in the northern Cascadia forearc of Vancouver Island, there are several competing models for what controls forearc stress and upper plate deformation (e.g., Mazzotti et al., 2011; Finley et al., 2019; Delano et al., 2017). Quantifying upper plate deformation is also limited in Cascadia because the subduction zone is relatively seismically quiet, limiting our ability to infer stress field data from seismicity. Furthermore, the large locking signal on the plate interface inhibits our ability to isolate Global Navigation Satellite System (GNSS) deformation associated with upper plate faults (e.g., Mazzotti et al., 2011; S. Li et al., 2018), and few active upper plate faults have been identified regionally to date (e.g., Morell et al., 2017).

Although the northern Cascadia region exhibits relatively low rates of instrumental seismicity, this region was also host to the largest onshore historic earthquake in Canada, the M 7.3 1946 Vancouver Island earthquake (Rogers & Hasegawa, 1978; Rogers, 1979; Lamontagne et al., 2018). This earthquake is the largest to have occurred anywhere within the Cascadia subduction zone system, including the megathrust,

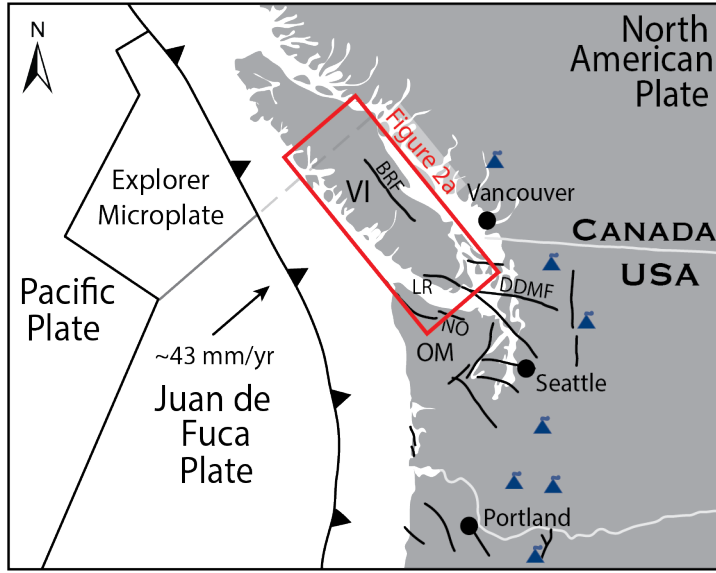


Figure 2.1: Regional tectonic setting showing the location of the Beaufort Range fault (BRF) and other active faults in the Cascadia forearc of Canada and the USA. Juan de Fuca – North America convergence vector after Kreemer et al. (2014). Slab depth contours from Slab2 (Hayes et al., 2018). Active faults in USA after U.S. Geological Survey (n.d.), Leech River fault after Morell et al. (2017), volcanoes after AGI (2003). Red box shows location of Figure 2.2a. DDMF – Darrington-Devils Mountain fault; FZ — fault zone; LR – Leech River fault; NO - North Olympic fault zone; OM - Olympic Mountains; VI - Vancouver Island.

since written historical recordkeeping began (the past ~ 200 yrs). However, despite this earthquake’s size and moderate damage to nearby population centers (Hodgson, 1946; Mathews, 1979; Clague, 1996), the fault that ruptured during the 1946 earthquake remains unknown. In addition, little is known about the current or past stress state and strain field of the crust surrounding this major historical rupture, what upper plate conditions could lead to future ruptures, and if similar events have occurred in the geologic past. Such data are necessary not only to evaluate the seismic potential of forearc faults, but also to determine their deformation rates, kinematics, and relationship to the regional stress field. Yet, no active faults have been identified north of the greater Victoria region to date, including in the region surrounding the approximate epicenter of the 1946 earthquake.

Here, we investigate the kinematics and slip history of the Beaufort Range Fault (BRF), a major fault in the northern Cascadia forearc, to evaluate how forearc strain is accommodated on this structure over decadal to millennial timescales. The BRF is located on central Vancouver Island, near the northern terminus of the Cascadia subduction zone (Figure 2.1). Several researchers proposed that the Beaufort Range fault may have hosted the 1946 rupture, based on the proximity of the epicenter, coseismic slip modeled from geodetic benchmark surveys, and the similarity of the BRF strike to the NW-SE striking nodal plane for the event’s focal mechanism (Rogers & Hasegawa, 1978; Slawson & Savage, 1979). However, no surface ruptures were found by researchers in the days and weeks following the rupture, and it remains unknown whether the BRF hosted the 1946 earthquake, or whether this fault is Quaternary-active or seismogenic.

In this paper, we undertake a field-based tectonogeomorphic investigation to evaluate the seismogenic potential of the BRF and to determine its slip history and kinematics with respect to historical seismicity

and regional tectonics. We exploit a well-preserved set of offset paleochannels on the southwestern flank of the Beaufort Range, visible in recently acquired bare-earth lidar Digital Elevation Models (DEMs), to demonstrate that the BRF is a highly active, right-lateral transtensional fault that has hosted multiple surface-rupturing earthquakes throughout the Quaternary. We find evidence for at least three late Pleistocene to Holocene earthquakes along the BRF, with surface ruptures extending >40 km, consistent with paleo-earthquake magnitudes of ~ 6.5 to 7.5 . While these data do not constrain the age of the most recent surface-rupturing event, our results do suggest that the most recent event occurred in the past ~ 3 - 4 kyr. We find that paleoseismic earthquakes along the BRF have kinematics similar to the 1946 Vancouver Island earthquake, suggesting that the BRF is a candidate host fault for this event. Finally, the similarities of the BRF deformation field and P- and T-axes derived from its slip over decadal to millennial timescales, suggest the stresses that lead to permanent deformation in this portion of the northern Cascadia forearc have been relatively consistent over multiple earthquake cycles.

2.2 Background

2.2.1 Tectonic Setting

The BRF is located in the northern forearc of the Cascadia subduction zone, where the Juan de Fuca plate subducts under the North American plate at a rate of ~ 43 mm/yr (DeMets et al., 2010; Kreemer et al., 2014). The fault is positioned ~ 150 km north of the Olympic Mountains, and ~ 60 km south of the onshore projection of the Nootka fault zone, the northern end of the Juan de Fuca slab (Figure 2.1). Active faults that accommodate forearc strain have been recognized along most of the Cascadia subduction zone south of the Olympic Mountains (e.g., Figure 2.1; Brocher et al., 2001; Goldfinger et al., 1992; Liberty et al., 2003; Personius et al., 2003; Sherrod et al., 2004; Kelsey et al., 2008; R. E. Wells et al., 2020; Horst et al., 2021; Bennett et al., 2017), and north of the Olympic Mountains (e.g., Figure 2.1; Schermer et al., 2021; Morell et al., 2017, 2018; Harrichhausen et al., 2021). However, no active faults have been identified in the northern 150-300 km of the forearc on Vancouver Island. It remains unclear if and how the slip accommodated by these southern faults is translated farther north and what role the BRF may play in accommodating forearc strain.

The BRF occurs along the southwestern flank of the Beaufort Range, near the city of Port Alberni on Vancouver Island (Figure 2.2a). The Beaufort Range consists of a ~ 70 km-long, ~ 5 - 10 km-wide set of peaks, whose elevations range from 1000 to 1600 masl. The range is asymmetric, with a gently sloping, glacially scoured northeastern flank that slopes toward the Strait of Georgia, and a steep (up to 35°) southwestern flank that slopes toward the Alberni Valley (Figure 2.2b). The BRF strikes NW-SE, following the southwestern

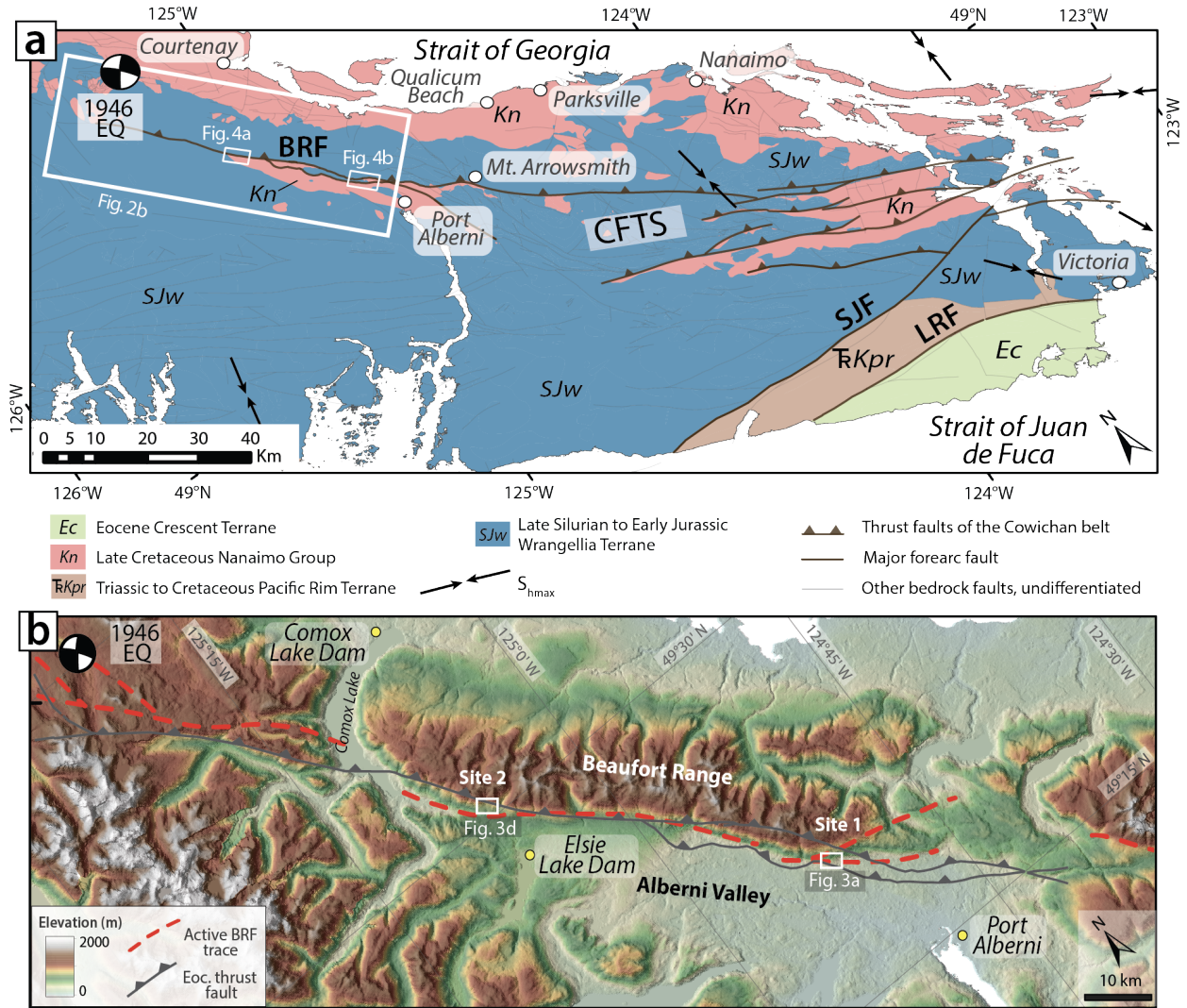


Figure 2.2: Geologic and geomorphic setting of the Beaufort Range fault. **A:** Simplified geologic map of southern Vancouver Island showing major lithologic units, thrust faults of the Cowichan fold and thrust system (CFTS), and other forearc faults. The epicenter of the 1946 M 7.3 Vancouver Island earthquake is shown by the focal mechanism (Rogers & Hasegawa, 1978). Maximum horizontal stress directions after Balfour et al. (2011). Bedrock geology after the British Columbia Geological Survey compilation by Cui et al. (2017). BRF—Beaufort Range fault. LRF—Leech River fault. SJF—San Juan fault. **B:** Hillshaded SRTM DEM showing the topography of the Beaufort Range and Alberni Valley, the locations of hydroelectric dams, the trace of the Eocene bedrock Beaufort Range thrust fault (in legend), and a simplified inferred trace of the active BRF (in legend) based on the locations of mapped scarps (Figure A.1).

topographic range front for >40 km (Figure 2.2b).

2.2.2 Eocene slip along the Beaufort Range thrust fault

The BRF has been previously mapped as an Eocene bedrock fault that places the Late Triassic Karmutsen Formation basalts that form the peaks of the Beaufort Range over the Cretaceous Nanaimo Group sediments

that underlie the Alberni Valley (Figure 2.2a, Figure A.1; Yorath et al., 1985; England & Calon, 1991). Geologic mapping, balanced cross sections, and LITHOPROBE seismic reflection profiles suggest this bedrock thrust fault dips NE, at 45° to sub-vertical (Yorath et al., 1985; Clowes et al., 1987). Geologic maps depict the BRF as an along-strike projection of the frontal thrust fault of the Cowichan Fold and Thrust System (CFTS), located ~40 km along strike to the southeast of the BRF (Cui et al., 2017; England & Calon, 1991). Low-temperature thermochronology data indicating exhumation at ~50-40 Ma suggest the thrust faults of the CFTS, including the BRF, initially formed during the Eocene accretion of the Pacific Rim and Crescent terranes (England, 1990; England & Calon, 1991; England et al., 1997).

2.2.3 Glacial history

The Beaufort Range and Alberni Valley experienced two major phases of glaciation during the last glacial period. The region was inundated by the south-southwestward flowing Cordilleran continental ice sheet during the Fraser stage glaciation (~25-12 ka; Fyles, 1963; Alley & Chatwin, 1979). Then, during the retreat of the ice sheet, the Alberni Valley was occupied by a southeastward flowing valley glacier that produced streamlined landforms and associated glacial deposits (Mosher & Hewitt, 2004; Easterbrook, 1992; Clague & James, 2002). Existing maps document sub-glacial till, colluvial, and alluvial deposits that extend to an elevation of ~300 m along the range front (Fyles, 1963). These deposits have been correlated to the last glacial maximum at ~13.6-11 ka, based on ages from marine shells, peat, and wood in glaciomarine deposits in the Strait of Juan de Fuca and along the eastern coast of Vancouver Island (e.g., Clague et al., 1980; Easterbrook, 1992). However, there has been limited surficial mapping of the Beaufort Range front, and no deposits in the Alberni Valley region have been directly dated. We expand and refine these mapping data to constrain the ages of deposits offset by scarps and evaluate the Quaternary activity of the BRF.

2.2.4 Possible association of the BRF with the 1946 Vancouver Island earthquake

Although post-Eocene deformation has not been previously documented along the BRF, several researchers proposed that the Beaufort Range fault may have hosted the 1946 M 7.3 Vancouver Island earthquake. The earthquake epicenter was located at the northern tip of the BRF at a depth of <30 km, and focal mechanism solutions contain a NW-SE striking nodal plane sub-parallel to the BRF (Figure 2.2; Rogers & Hasegawa, 1978). These data led Rogers and Hasegawa (1978) to propose that the 1946 earthquake may have been a right-lateral oblique event hosted by the BRF (Figure 2.2). Geodetic surveys of a triangulation network before and after the event suggest ~1-2.5 m of right-lateral oblique slip along a steeply NE dipping

(70°) fault. While multiple ground surface failures and slumps have been identified around the Beaufort Range associated with the 1946 event (Mathews, 1979; Clague, 1996), no fault-related surface ruptures associated with the 1946 event were ever discovered.

2.3 Methods

Our methodological approach is motivated by newly available lidar bare-earth elevation models along the surface trace of the Beaufort Range fault that reveal a series of topographic scarps that suggest the fault has accommodated Quaternary offset (Figure 2.2b). These scarps, clearly visible in bare-earth lidar DEMS (Figure 2.3), occur in en echelon arrays of 1-6 sets, each ~100-500 m long, and spaced 10s to 100s of meters apart. The majority of well-preserved scarps are located near the base of the range—20-100 m above the valley floor, or 500-870 m below the range crest—and strike sub-parallel to the trend of the southwestern flank of the Beaufort Range front (~290-320°; Figure 2.2b, Figure A.1). Our initial observations of the lidar data suggested these scarps exhibit apparent right-lateral and SW-side-up 1-10-m scale displacement of a network of V-shaped paleochannels with paired offset sharp-crested interfluvies. Given the glacial history of the region, we surmised that these channels may be no older than the time of ice retreat, and therefore the offset channels may record Holocene fault displacement.

Based on these initial observations, we undertook detailed field-based mapping and topographic surveying of faults and offset landforms to determine the geometry of the fault networks potentially associated with these scarps, the relative ages of offset deposits, the magnitude of potential offset, and the associated kinematics of fault slip.

2.3.1 Mapping

We completed surficial and bedrock mapping to: 1) identify earthquake-generated fault scarps along the BRF, 2) determine the relative ages of Quaternary deposits offset by surface ruptures, and 3) determine if active fault strands re-occupied inherited bedrock faults or shear zones. Identifying fault-related deformation (e.g., fault scarps) in dateable Quaternary sediments is essential for characterizing the slip history of active faults (e.g., Van Der Woerd et al., 2002; Zinke et al., 2017; Hatem et al., 2017; Regalla et al., 2022), but dense temperate rainforest limits exposures and accessibility of offset Quaternary deposits in the study area. Thick soils and dense vegetative cover limit bedrock exposures to road cuts, logging roads, quarries, and stream channels, and obscure many Quaternary landforms beneath the forest canopy. However, these fault-related landforms are well-resolved in the newly available lidar point clouds collected along the BRF.

We used bare-earth lidar data, satellite imagery, and historical air photos to map potentially earthquake-

generated fault surface ruptures (scarps) within a ~ 100 km-long swath area extending from Mt. Arrowsmith to the Forbidden Plateau (Figure A.1). Lidar point cloud data were collected by Terra Remote Sensing, and TimberWest and Island Timberlands logging companies provided ground returns. The lidar point clouds contained an average of ~ 1.2 - 1.4 ground returns per square meter. We gridded these data into a 0.5 m DEM and generated topographic derivatives such as hillshade, standard deviation, and slope maps to aid in mapping. We additionally used satellite imagery (Google Earth Pro, 2016) and British Columbia provincial government historical air photos from 1947 and 1952 to evaluate any anthropogenic modification of key sites, including past roads, railroads, and logging trails.

We visited each accessible remotely-mapped scarp to confirm they were tectonically-generated features

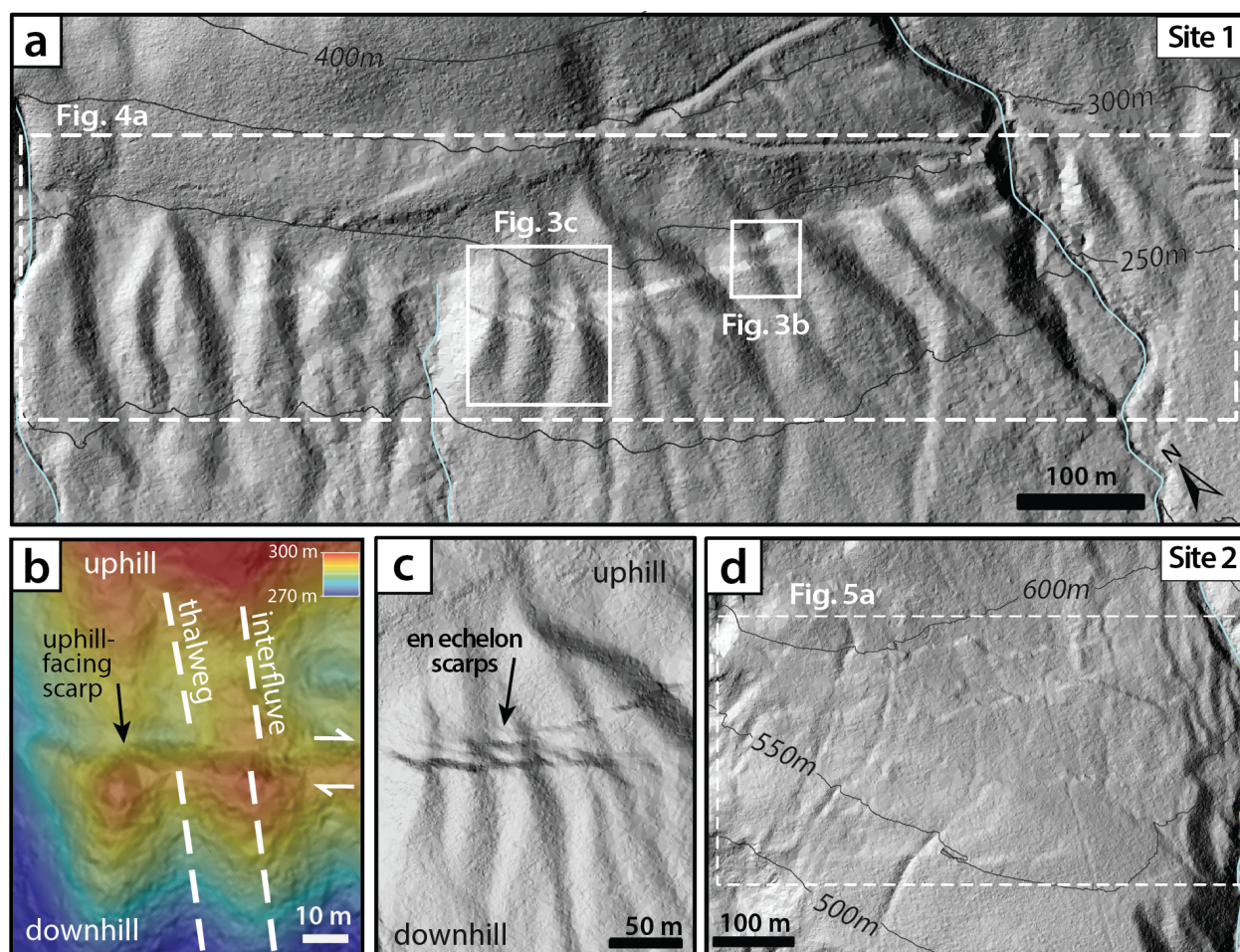


Figure 2.3: Examples of tectonic fault scarps visible in hillshaded bare-earth lidar DEMs. **A:** Unannotated DEM of Site 1 showing a network of en echelon fault scarps offsetting a series of abandoned channels and interfluvies. **B:** Example of an uphill-facing scarp developed on a till-mantled hillside. The scarp offsets a channel thalweg and adjacent interfluvial crests both vertically (downhill-side-up) and right-laterally. **C:** Example of en echelon array of scarps at Site 1. **D:** Unannotated DEM of Site 2 showing a network of right-laterally sheared channels. Examples of non-tectonic landforms are presented in Figure A.2.

(i.e., not related to slumping, etc.). Criteria used to distinguish fault scarps from other features include whether the features are linear, continuous over >50-100 m length scales, are cut across topography, and if they offset hillslopes, abandoned channels, or interfluves (Figure 2.3b-c). We took care to distinguish potentially fault-related scarps from landforms produced by glacial deposition or scour, anthropogenic disturbance, gravitational failure, or differential erosion (see Section A.2 and Figure A.2).

We then completed highly detailed and more focused field mapping, at a scale of 1:3000, of Quaternary deposits and bedrock units in two ~6 km by ~2 km regions (Sites 1 and 2) that each contain a high density of fault scarps (Figures 2.1, 2.2). Surficial mapping was completed based on field and lidar-based observations of surface topography, roughness, morphology, and inset and burial relationships, accompanied by detailed lithologic descriptions of each Quaternary unit. We used these observations to create a local Quaternary stratigraphy that allowed us to determine the relative ages of units offset by faults. Bedrock mapping was completed using outcrops exposed in road cuts, streams, and quarries. We measured the structural orientations of fault planes, slickenlines, foliation fabrics, and fractures within the principal shear zones and damage zones, where exposed.

2.3.2 Quantifying fault slip

2.3.2.1 Topographic surveys of offset landforms

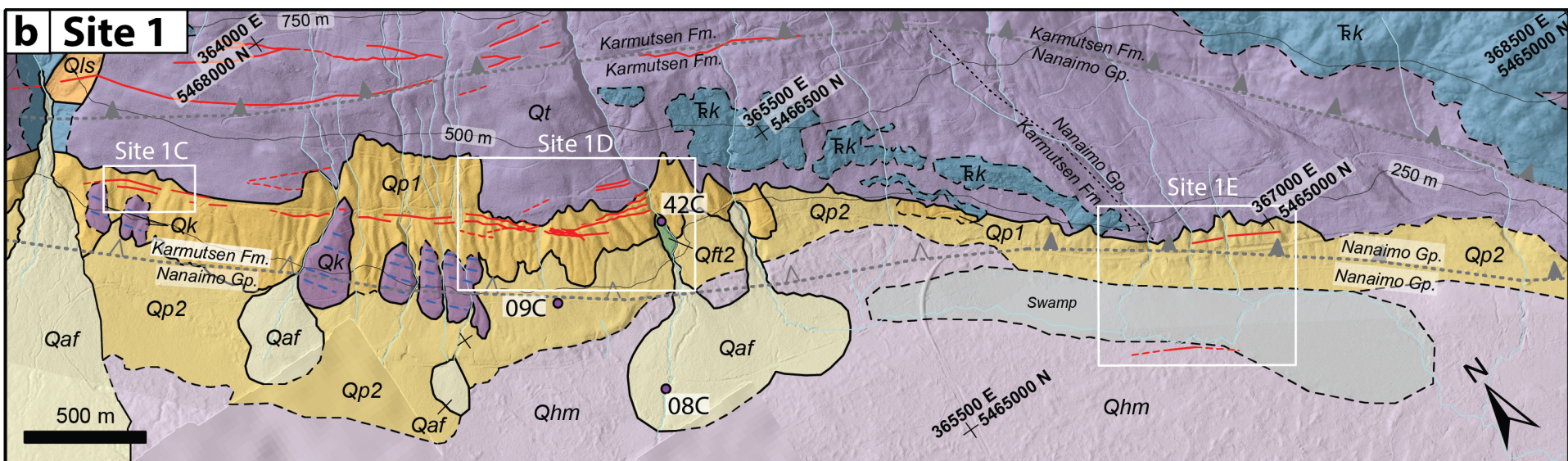
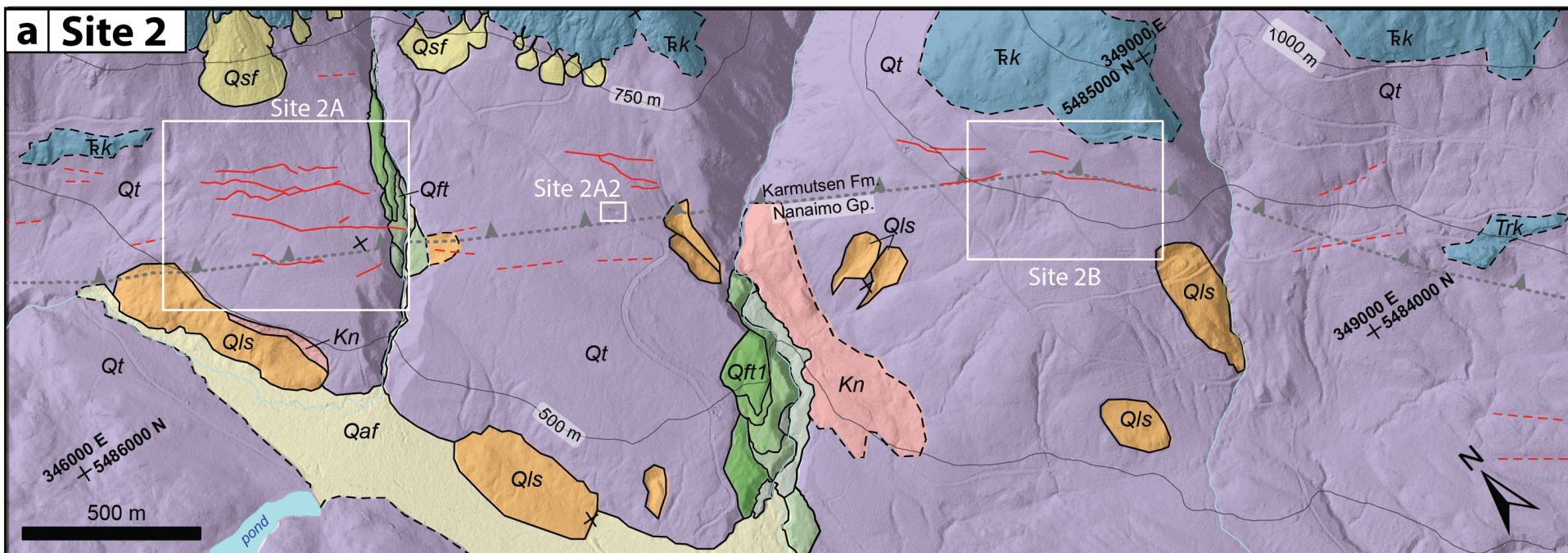
We collected topographic survey data across fault scarps at 64 locations at Sites 1 and 2 to determine the attitudes of fault planes associated with fault scarps, and to quantify the vertical and lateral offset of displaced Quaternary deposits and landforms (Figures 2.5 and 2.6). These data included 58 surveys of offset geomorphic piercing lines where the three-dimensional oblique-slip vector could be calculated (Figure 2.7), and 6 additional “straight-line” profiles used to calculate the vertical component of displacement in locations where geomorphic piercing lines were absent (Figures 2.5, 2.6, A.4, and Dryad data repository Lynch et al., 2023). Surveys were collected with a Nikon XS and Spectra Precision Focus 6 total station, which yielded more continuous topographic data than the lidar DEMs which had non-uniform return spacing and included some false ground returns.

Our primary survey targets were a series of abandoned channels and interfluves at Sites 1 and 2 whose axes intersect fault scarps at near-orthogonal angles, that serve as piercing lines from which fault slip vectors can be reconstructed. Topographic surveys of these landforms followed either the channel thalweg or the interfluve crest. In locations where channels and interfluves are absent, we collected linear profiles with trends perpendicular to the fault scarp. For each profile, total station survey data were collected every ~0.5-1 m, to a distance of >20 m uphill and downhill of each fault scarp (Figure 2.7). Along survey transects

where a geomorphic piercing line extended for less than 20 m (e.g., between closely-spaced fault strands), we collected a minimum of 3 survey points, with an average of 11 points. We complemented these ground surface elevation profiles with six additional topographic profiles extracted from lidar DEMs in a portion of Site 2A where thick forest cover and uneven topography prevented total station surveys of offset abandoned channels.

2.3.2.2 Reconstructing oblique fault displacement

We used the topographic survey data to reconstruct both the magnitude and orientation of the slip vector at each surveyed location where a geomorphic piercing line intersected an individual fault plane. In order to calculate a slip vector, the local orientation of the fault plane must be known. No outcrop exposures of fault planes in Quaternary deposits were present in the field area, but we were instead able to reconstruct the local strike and dip of the fault plane associated with mapped scarps using a modified three-point problem approach. In this approach, we assumed the midpoint, or inflection point, of a fault scarp represents the most likely intersection of the fault plane with the surface. We surveyed scarp midpoints at a range of elevations (~ 4 -12 m elevation range) and determined fault strike and dip through linear regression of a plane through the surveyed scarp midpoints using all surveyed data along a single continuous fault strand segment (3-17 points per regression). We used these data to determine a representative fault dip for each scarp segment, using the average dip from all regressions at Site 1 or Site 2, and a representative fault strike given by the local strike of each fault strand or segment. Because fault dips determined from surveys of degraded scarp faces over small elevation ranges may underestimate true fault dip, we allowed our model reconstructions to permit fault dip to be 5° steeper than that calculated from the three-point approach.



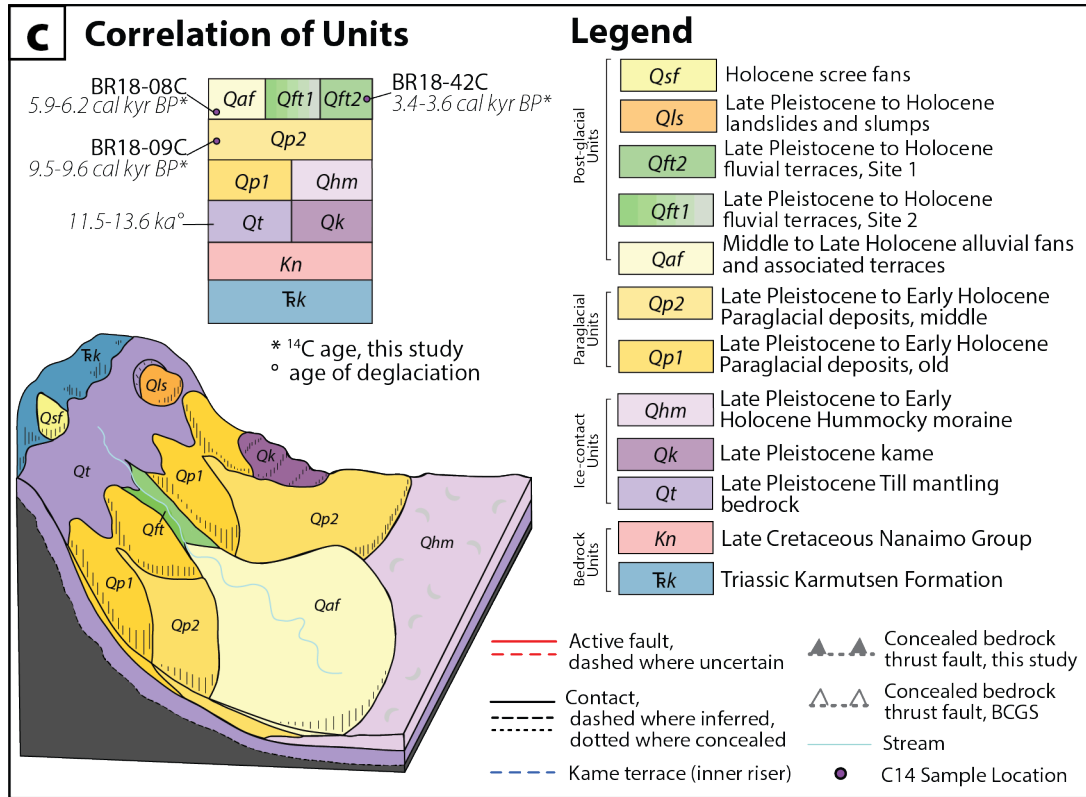


Figure 2.4: Bedrock and surficial geology of portions of the BRF (See locations in Figure 2.2a). Mapping is overlain on a composite hillshaded DEM compiled from two bare-earth lidar DEMs gridded to 0.5 m and to 2 m, and from 30 m SRTM DEM. Radiocarbon ages are reported in Table 2.1. Bedrock fault locations compiled from new field mapping and existing mapping by the British Columbia Geological Survey (BCGS; Cui et al., 2017). White boxes outlining Sites A-E correspond to locations shown in Figures 2.5 and 2.6. *Panels A and B on preceding page. A:* Map of Site 2 along the northern portion of the BRF. *B:* Map of Site 1, along the southern portion of the BRF. Fault scarps (red lines) occur at the base of the Beaufort Range and along the range front up to 1000 m above the valley floor. Mapped scarps occur in both the hanging wall and footwall of the bedrock BRF. Fault scarps offset multiple ages of glacial (Qt), paraglacial (Qp1, Qp2), and modern deposits (Qls, Qft, Qaf). Terrace generations within unit Qft1 in panel A are depicted by increasing color saturation with terrace age, delineated by thin gray lines. *C:* Correlation of units and legend for geologic maps in panels A and B. Radiocarbon ages demonstrate that these deposits are ~9600-3400 cal BP in age (Table 2.1).

We combined our fault plane solutions and topographic survey data to calculate the 3D offset of each piercing line, specifically the magnitude and direction (trend and plunge) of the slip vector (Figure 2.7). Calculations were made using an R script that performed a Monte Carlo simulation to evaluate the slip vector and associated uncertainty (script available in data repository, Lynch et al., 2023). The script requires the following user-defined inputs: the strike and dip of the fault plane, the 1σ uncertainty on strike and dip, the XYZ coordinates of the topographic survey data, the location where the fault plane intersects the ground surface, and the number of survey points in the upthrown and downthrown sides of the profile used to

define the 3D geometry of the piercing line segments. For each profile, we assigned a fault strike and dip as described above, and $\pm 1\sigma$ uncertainty (5°). We manually defined the remaining parameters—fault plane intersections, and the number of survey points used to fit linear regressions through the upthrown and downthrown surveyed piercing lines—for each topographic profile. It has been well-documented that how a user defines fault and piercing line geometry (i.e., profile regression limits) can lead to multiple admissible geologic slip reconstructions (e.g., Scharer et al., 2014). To account for this uncertainty, we performed Monte Carlo simulations for each offset profile using input defined by five different users, each trained in scarp offset analysis.

Using these inputs, we used the R script to calculate 3D linear regressions through topographic survey points on the upthrown and downthrown sides of the fault scarp and then solve for the intersection points of these lines with the fault plane (Figure 2.7). These two intersection points were then used to calculate the

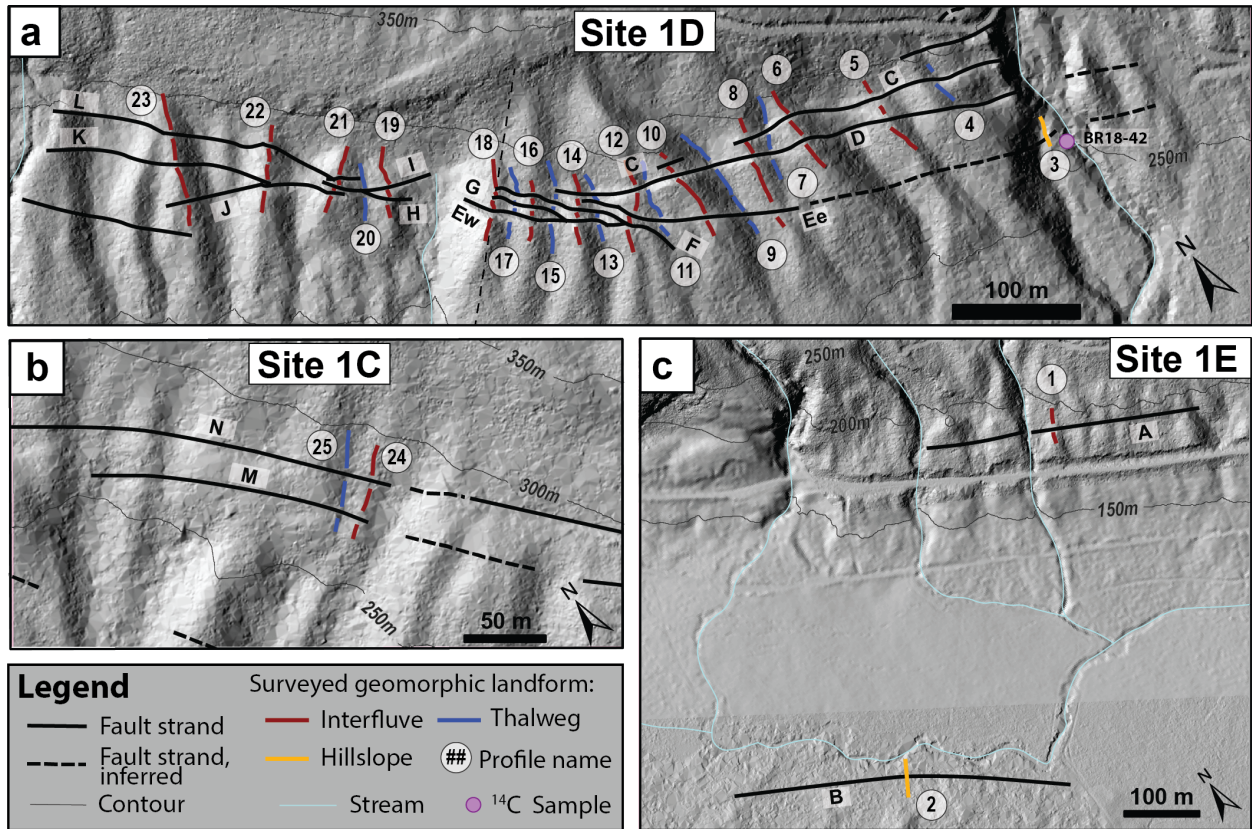
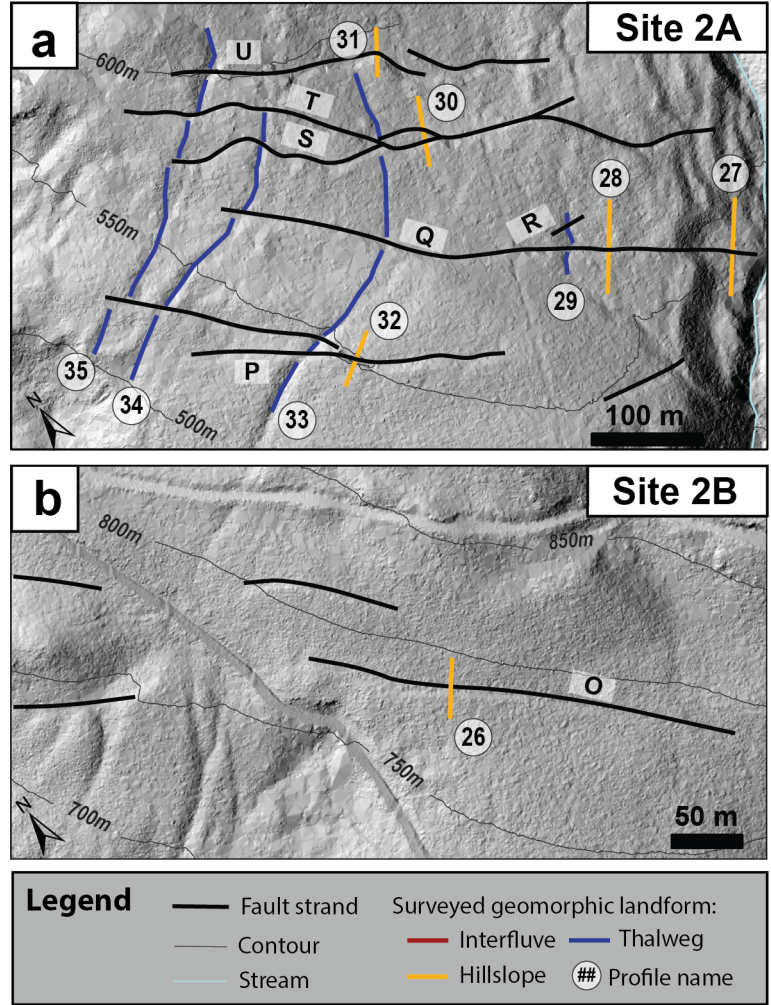


Figure 2.5: Hillshaded lidar DEMs of Site 1 showing mapped faults (labeled from A to N) and surveyed topographic profiles (numbered from 1 to 25). See Figure 2.4 for locations and Dryad data repository for topographic profile survey data (Lynch et al., 2023). **A:** Annotated hillshaded DEM showing the locations of mapped fault strands and topographic survey profiles at Site 1D. Unannotated lidar DEM is presented in Figure 2.3a. **B** and **C:** Annotated DEMs of Sites 1C and 1E. Unannotated versions of all DEMs are in Figure A.4.

Figure 2.6: Hillshaded DEMs of Site 2 showing mapped faults (labeled from O to U) and surveyed topographic profiles (numbered from 26 to 35). See Figure 2.4 for locations and Dryad data repository for topographic profile survey data (Lynch et al., 2023). **A:** Annotated hillshaded DEM showing the locations of mapped fault strands and topographic survey profiles at Site 2A. Unannotated lidar DEM is presented in Figure 2.3d. **B:** Annotated DEM of Site 2B showing mapped faults and surveyed profile. Unannotated versions of all DEMs are in Supporting Information Figure A.4.



magnitudes of strike slip (SS), dip slip (DS), and oblique slip (OS) for each piercing line, as well as the trend and plunge of the slip vector (Figure 2.7). The Monte Carlo simulation was repeated 100 times for each of the five user-defined profile selections, yielding a total of 500 simulations of fault slip for each displaced piercing line. We report the outputs as the mean $\pm$ one standard deviation of the 500 values calculated for that profile.

2.3.2.3 Inversion for fault kinematics

We use the slip vector data to invert for the kinematics of the BRF using the FaultKin 7.6 program (Marrett & Allmendinger, 1990; Allmendinger et al., 2012). Data inputs included the trend and plunge of the best-fit slip vector determined from the Monte Carlo simulations, and the corresponding fault plane strike and dip determined from the modified three-point fault plane regressions. Inversions were performed using data from each of the 55 fault scarp surveys with vertically and laterally offset piercing lines. We grouped data for kinematic slip inversions in two ways. First, we grouped data collected at each mapping

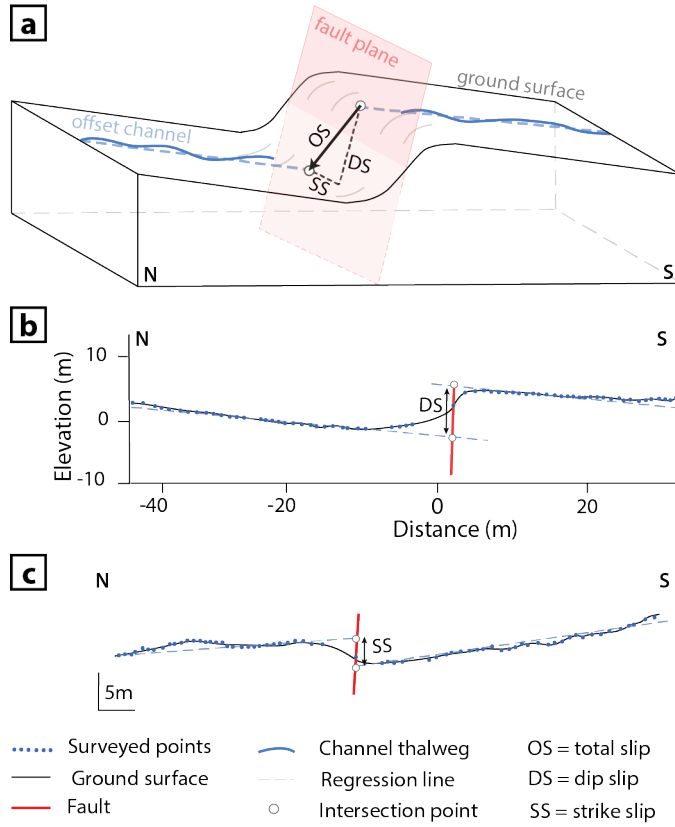


Figure 2.7: Schematic diagrams showing how surveyed geomorphic piercing lines were used to reconstruct 3D fault slip. **A:** Block diagram showing an oblique normal right lateral offset channel thalweg. Fault slip components (OS, DS, and SS) are calculated from the 3D positions of the intersections of the fault plane with the linear projections of the upthrown and downthrown channel segments. **B:** Example of a surveyed geomorphic piercing line profile in cross-section. **C:** Example of a surveyed geomorphic piercing line profile in plan view. In each survey, points were collected every ~ 0.5 -1 m at least 10-20 m beyond the fault scarp.

sub-site (A-E in Figure 2.4), to produce kinematic inversions representative of slip observed at each site location. Then, we grouped all data for the entire BRF to determine a kinematic inversion best fit to all observed data. Kinematic inversions were performed by calculating P- and T-axes from each calculated slip vector and fault plane pair, and then generating Bingham fault plane solutions from the set of P- and T-axes at each site (Marrett & Allmendinger, 1990; Allmendinger et al., 2012). These inversions assume slip occurs in the direction of maximum resolved shear stress on the fault plane, and produces mean P- and T-axes, pseudo focal mechanisms, and predicted slip vectors for each nodal plane (Marrett & Allmendinger, 1990; Allmendinger et al., 2012). These kinematic inversions and P- and T-axes provide information about paleo strain fields, and may, under certain assumptions, be used to approximate local stress axis orientations at the time of deformation (e.g., Angelier & Mechler, 1977; Riller et al., 2017).

2.3.3 Radiocarbon dating of Quaternary deposits

We collected charcoal samples from natural and manmade exposures of mapped Quaternary deposits to determine the chronologic ages of units offset by mapped faults. We focused our sampling on detrital charcoal as charcoal is present in many deposits on Vancouver Island, has previously been used to evaluate late Pleistocene to Holocene unit ages (e.g., Clague et al., 1980; Morell et al., 2018; Harriehausen et al.,

2021), and because luminescence techniques have not yielded reliable ages for late Pleistocene to Holocene deposits due to insufficient dose rate (e.g., Graham, 2017; Morell et al., 2018). We collected samples of macroscopic (macro) charcoal (>0.5 cm) where fragments were visible in outcrops of Quaternary deposits. If no macro charcoal was readily visible in an outcrop, we collected 1-2 L of bulk sediment and sieved the samples to extract any dateable macro charcoal present. For all sample sites, we completed detailed unit descriptions and noted the sample’s stratigraphic position within the deposit (Figure A.3). We collected three macro charcoal samples and five bulk sediment samples from Site 1 (see Figure 2.4b for locations). Our sampling was focused on units mapped at Site 1 (Figure 2.4b), where we identified multiple generations of Quaternary deposits (see Section 2.4.2). We were unable to date any mapped deposits at Site 2 due to a lack of exposure.

Charcoal samples were cleaned and processed at Paleotec Services, Ottawa, Ontario, Canada. Macroscopic charcoal pieces were extracted from bulk sediment samples by flotation and wet sieving in warm tap water using nested sieves of 0.85 mm and 0.425 mm. All material greater than 0.425 mm was examined using a binocular microscope, and any isolated charcoal pieces were shaved of any adhering sediment. The largest shaved fragment from each sample was further sliced into smaller fragments to look for the presence of fine modern rootlet penetration and/or fungal contamination, including mycorrhizae, and rejected if contaminants were present.

Three Quaternary units yielded dateable charcoal fragments that were processed for radiocarbon analysis (Table 2.1, Figure 2.4b). These included macro charcoal samples extracted from one outcrop (BR18-06C, -07C, and -08C), and two samples extracted from sieved bulk sediment from two additional outcrops (BR18-42C and BR18-09C). Sample BR18-08C was selected as the highest-quality sample of the three charcoal fragments extracted from the outcrop exposure. Bulk sediment sample BR18-09C included three mm-sized charcoal pieces that were combined to ensure adequate sample mass for AMS after acid-base-acid (ABA) treatment (Table 2.1, Figure 2.4b, Figure A.3). Unfortunately, the three remaining bulk sediment samples (BR18-10C, -11C, and -12C) were barren of charcoal. Samples were analyzed at the Keck Carbon Cycle AMS Laboratory at UC Irvine. Radiocarbon ages (reported following Stuiver & Polach, 1977) were calibrated using the INTCAL20 calibration curve (Reimer et al., 2020) and OxCal v4.4 (Bronk Ramsey, 1995, 2021). We report radiocarbon ages as the two-sigma (2σ) range of calendar years before present (1950).

2.3.4 Estimates of fault slip

We estimate slip rates at Site 1 using the cumulative oblique displacement measurements of three different ages of offset landforms, as well as radiocarbon dates from detrital charcoal that provide estimates of unit

ages. We use two approaches to estimate slip rates, following the methods of DuRoss et al. (2020). The first is an “open-ended” approach that uses the cumulative slip of the oldest offset unit and that unit’s estimated age. The second is a “closed interval” approach that uses the difference in slip that has occurred during a known time interval that encompasses one or more complete recurrence periods. We report both slip rate calculations and discuss the relative applicability of each.

2.4 Results

Our mapping provides several lines of evidence that the BRF is Quaternary-active, and has experienced multiple slip events since the late Pleistocene. Field mapping of the morphology and spatial distribution of fault scarps (Figures 2.3 and 2.4) indicates that the mapped scarps are of tectonic origin, produced during one or more surface-rupturing earthquakes, and are not the product of glacial, gravitational, or anthropogenic processes. An active BRF is further supported by the presence of numerous right-laterally and vertically offset abandoned stream channels incised into Late Pleistocene to Holocene till and paraglacial deposits. Below we discuss the morphology of the fault scarps, the ages of offset deposits, the kinematics of fault slip derived from measured offsets of channel networks, and our interpretations of the number and relative timing of events that have occurred along the BRF since the last glacial maximum.

2.4.1 Quaternary fault scarps

Our mapping shows that the Quaternary-active BRF is defined by a series of sub-parallel, discontinuous fault scarps ($n=153$) that offset multiple ages of Quaternary deposits preserved on the southwestern flank of the Beaufort Range (Figure 2.2a, Figure A.1). The spatial distribution of preserved scarps shows they are part of an ~ 500 m-wide fault zone, where slip is distributed across multiple ($\sim 1-6$) sub-parallel, steeply-dipping fault strands (Figure 2.8, Figure A.1). Individual fault scarps locally exhibit strike lengths of $\sim 100-1500$ m, exhibit scarp heights of $\sim 0.5-6$ m, and occur in en echelon or parallel sets with intra-fault spacings of 5-100 m. Scarp facing directions can vary locally over short distances but about two-thirds of the scarps ($n=101$) face NE. Most ($\sim 70\%$) of the mapped faults have asymmetric cross-sectional morphologies with steep uphill-facing scarps, while a smaller fraction are preserved as flat, degraded topographic features embedded in the high-gradient hillslopes (Sites A-E; Figures 2.3, A.2, A.4). Our mapping demonstrates that active fault strands generally strike NW, parallel to the range (average $\sim 287^\circ$, with variation of up to $20^\circ-38^\circ$), and our topographic field surveys (see Section 2.3.2.1) indicate that near the surface, most strands dip steeply NE ($\sim 60^\circ-88^\circ$), with a few dipping steeply SW ($\sim 70^\circ$).

The steepness and morphology of the scarp faces vary both along strike and between strands. At Site 1,

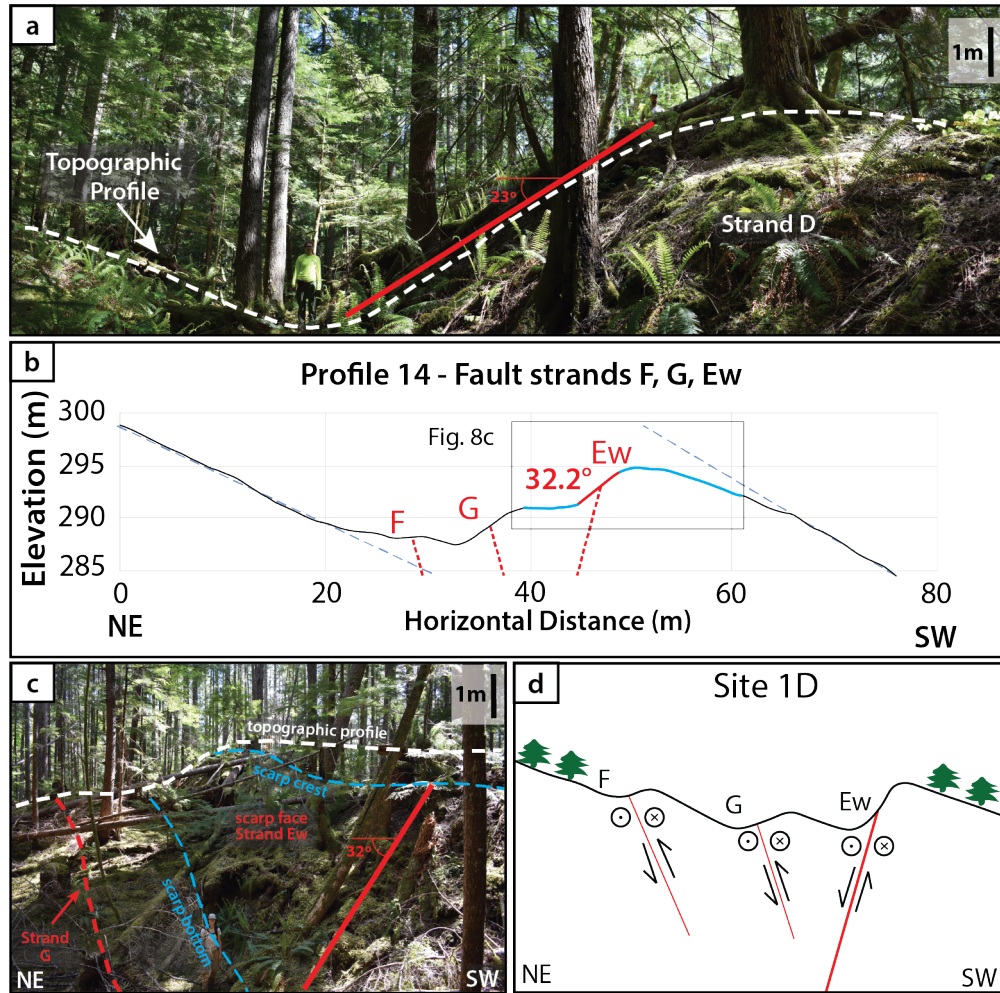


Figure 2.8: Examples of fault scarps identified along the Beaufort Range fault. **A:** Tall, uphill-facing, moderately steep ($\sim 23^\circ$) fault scarp along strand D at Site 1D (Figures 2.4b, 2.5a). **B:** Topographic profile across three scarps at Site 1D associated with fault strands F, G, and Ew extracted from bare-earth lidar DEM (Profile 14, Figure 2.5). Dashed dark blue lines show the projection of the background hillslope toward the scarps. **C:** Photo of the tall, steep preserved face of Strand Ew shown in the topographic profile in panel b. The uphill-facing fault scarp along strand Ew is $\sim 32^\circ$, nearly angle of repose, much steeper than the scarp face along strand D (panel a). **D:** Cartoon cross-section showing the schematic relationships between sets of sub-parallel and en echelon fault strands, based on observations at Site 1D. These strands are interpreted to merge at depth in a flower structure consistent with strike-slip faulting.

the steepest and tallest scarps are 4-6 m tall and have scarp faces near the angle of repose (32° strand Ew at Site 1D; Figures 2.5, 2.8). Many of the scarps at Site 1 exhibit steep, well-preserved free faces, such as strand Ew at Site 1D (Figure 2.8b-c). Other scarps at Site 1 exhibit a more moderate, 24° dipping scarp face (Figure 2.8a), such as Strand D at Site 1D. At Site 2, the scarps are 1-3 m tall and have faces near the angle of repose ($\sim 45^\circ$ strand U at Site 2A; Figures 2.6 and A.5), and some are large and steep enough to have effectively ponded large boulders sourced from uphill (Figure A.5). Several of the individual scarps

at Site 1 are part of a larger, multi-fault scarp that includes multiple emergent fault strands (Figure 2.8b), whereas individual scarps at Site 2 appear to occur as separate parallel or anastomosing fault sets (Figure 2.6).

Our mapping shows that the majority of active fault scarps are not directly co-located with known bedrock fault planes at the surface (Figures 2.4 and A.1). At Site 2, the primary bedrock thrust fault, which places Karmutsen Fm basalts over Nanaimo Gp sediments (Figure 2.4a), is exposed as a 200+ m-wide damage zone that juxtaposes hanging wall basalts against an upright, open, footwall syncline of Nanaimo Gp sandstones. Mapped Quaternary fault scarps do not appear co-located with the exposed bedrock fault plane at the surface, but instead occur in sub-parallel networks spanning up to 500 m away, in both the hanging wall and footwall. Similarly, at Site 1 where the bedrock thrust branches into two strands, mapped Quaternary faults occupy a zone that is ~500 m wide, and occur up to 500 m away from mapped bedrock thrust faults (Figure 2.4b).

Quaternary fault scarps also have different slip senses and attitudes than observed along the bedrock thrust faults. Slickenlines and Riedel shear geometries (Figure A.7) on Eocene bedrock thrust faults indicate apparent NE-side-up, dominantly dip-slip displacement, whereas the active faults exhibit southwest-side-up and right-lateral displacements (Figure A.7). Outcrop exposures of the bedrock thrust fault at both Site 1 and Site 2 exhibit strike orientations that differ from Quaternary scarps by $\sim 15^\circ$. Our field surveys indicate that the active fault BRF strands have dips of 70° - 90° NE, whereas exposures at several sites along the range suggest the inherited thrust fault has a dip of $<40^\circ$ NE to sub-vertical. These observations indicate that mapped active fault scarps are not produced by slip along inherited structures in the near subsurface, but instead occupy a zone that is generally sub-parallel to the inherited structure.

2.4.2 Quaternary mapping and stratigraphy

Quaternary fault scarps along the BRF displace a series of nine units that were deposited during the late Pleistocene to Holocene deglaciation and subsequent transition to a post-glacial environment. We develop a local Quaternary stratigraphy (Figure 2.4 and Table A.1) that groups these deposits into three categories: ice-contact glacial units deposited during the most recent glaciation, paraglacial units deposited during ice retreat and slope readjustment, and post-glacial units deposited after ice retreat.

2.4.2.1 Ice-contact glacial deposits and landforms

The ice-contact glacial units are the oldest and stratigraphically lowest Quaternary units mapped in the study area and include subglacial till (Qt), kame terraces (Qk), and hummocky moraine (Qhm) (Figure

2.4; Table A.1). The subglacial till (Qt) is a very indurated, matrix-supported diamict containing both locally-derived and exotic clasts and is up to 40 m thick. Qt mantles bedrock along the southwestern flanks of the Beaufort Range mountain front at elevations >150-400 m. Kame terraces (Qk) occur as a series of five evenly-spaced, flat-topped terrace treads with steep risers, at 150-300 masl (<150 m above the valley floor), underlain by indurated, poorly to moderately sorted, stratified sands and gravels. Hummocky moraine (Qhm) is present on the valley floor at elevations of <150 masl at Site 1.

2.4.2.2 Paraglacial deposits and landforms

Glacial deposits are overlain by two generations of paraglacial deposits, Qp1 and Qp2 (Figure 2.4c; Table A.1). Qp1 consists of indurated, clast-supported, poorly-sorted, stratified sands and gravels. Qp1 deposits occur as cone-shaped landforms whose heads merge into Qt and whose toes are buried by Qp2 at the foot of the range. Qp2 has a similar composition to Qp1 and consists of thinly-bedded, clast-supported, stratified sands and gravels with occasional coarse sand lenses. Qp2 is distinguishable from Qp1 based on inset and burial relationships and its position at lower elevations on the range front.

Qp1 is incised by a series of abandoned channels. These channels are disconnected from active streams but merge into the heads of Qp2 deposits, suggesting that they were active at the time of deposition of Qp2. Abandoned channels at Site 1 are typically ~1-4 m deep, have V-shaped cross-sectional morphologies and are separated by adjacent interfluvies with linear ridges and steep flanks, or are incised into till and colluvium-mantled hillslopes (Figures 2.4 and 2.5). We interpret these abandoned channels to have formed as the result of fluvial and debris flow scouring and filling associated with the deposition of Qp2. At Site 2, offset abandoned channels have broad cross-sectional morphologies and are moderately incised into hummocky, till-mantled hillslopes (Figures 2.4 and 2.6). These channels do not clearly merge into other mapped deposits but appear to be cross-cut by younger landslides at the foot of the range.

2.4.2.3 Post-glacial units and landforms

The youngest units include post-glacial landslides (Qls), scree fans (Qsf), alluvial fans (Qaf), and fluvial terraces (Qft1 and Qft2) that either bury or are inset into the glacial and paraglacial deposits (Figure 2.4, Table A.1). Mapped landslides (Qls) are hummocky deposits associated with curvilinear headscarps and oversteepened toes and have widths of 50-600 m. Scree fans (Qsf) are small (30-250 m across), fan-shaped deposits with rough surfaces that contain cobble to boulder-sized bedrock clasts. Qsf occurs at the bases of mapped bedrock exposures at elevations of ~750 masl. Alluvial fan deposits (Qaf) are defined as a series of broad, convex, gently-sloping fans headed in active or recently-active channels (Figure 2.4). The fans consist of poorly to moderately sorted, clast-supported, stratified alluvial and fluvial deposits containing silt, sand,

pebbles, and boulders, with occasional clast imbrication and cross-bedding (Table A.1). Qaf deposits are mapped at the base of the range front and bury portions of Qp2, Qt, and Qhm.

At two locations in Site 2, and one at Site 1, Qaf fan heads merge into deeply incised (by $\sim 1\text{--}15$ m) streams that are flanked by a series of up to five fluvial terraces (Qft1 and Qft2). Fluvial terrace treads are 20-130 m wide, slope gently downstream, and have risers up to 5-10 m tall. The deposits that underlie these terraces are moderately to well-sorted, clast-supported sediments, with sub-horizontally stratified interbeds of rounded cobbles, boulders, and pebbles. We subdivide these fluvial terraces into two generations (Qft1 and Qft2) based on the inset relationships observed at Sites 1 and 2. At Site 2, Qft1 terraces are inset into till-mantled bedrock and are, in turn, incised by channels feeding Qaf alluvial fans (Figure 2.4b). This observation shows that at Site 2, Qft1 terraces are older than Qaf. In contrast, at Site 1, Qft2 appears to grade into the channels that feed Qaf, indicating that Qft2 terraces are younger than at Site 1 and are instead correlative to upper portions of Qaf or the channels inset into Qaf (Figure 2.4a).

2.4.2.4 Radiocarbon results and inferred unit ages

We use radiocarbon ages from detrital charcoal extracted from Quaternary deposits to place brackets on the possible ages of mapped units offset by BRF scarps. We note that the interpretation of detrital charcoal radiocarbon dates can be challenging due to vertical mixing during bioturbation or soil creep, recycling of older charcoal into younger deposits, and bias from younger carbon (e.g., roots) included in older charcoal. However, the radiocarbon ages that we obtained from Quaternary units in the map area are in broad agreement with our local relative Quaternary stratigraphy (Figure 2.4) and with regional constraints on the timing of deglaciation and post-glacial processes (e.g., Halsted, 1968; Alley & Chatwin, 1979; Blaise et al., 1990; Clague, 1994). We use these data, therefore, to make the following interpretations of unit ages.

The three ice-contact glacial deposits, Qt, Qk, and Qhm, were barren of charcoal and could not be directly dated (Table 2.1). This absence is consistent with other studies on Vancouver Island that have found ice-contact deposits to be devoid of charcoal (Morell et al., 2018; Harrichhausen et al., 2021). We interpret Qt, Qk, and Qhm to be associated with the last deglaciation, which has been regionally dated to $\sim 11.5\text{--}13.6$ ka (Halsted, 1968; Alley & Chatwin, 1979; Blaise et al., 1990; Clague, 1994), although we recognize the possibility that deposits associated with prior glacial periods may be present in the study area.

We attempted to radiocarbon date both Qp1 and Qp2 debris-cone fan deposits, but only Qp2 yielded dateable charcoal. The charcoal sample was collected from a stratified fan deposit ~ 30 cm below the surface of Qp2 (BR18-09C), in a roadcut exposure located ~ 250 m SW of the fault scarps at Site 1 (Figure 2.4, Figure 2.3). This sample yielded an age of ~ 9.5 cal ka BP (Table 2.1), consistent with the older estimated age of the Late Pleistocene glacial deposits (Qt, Qk, and Qhm) of $\sim 11.5\text{--}13.6$ ka, and younger radiocarbon

Table 2.1: Radiocarbon sample data at Site 1D (see Figure 2.4 for locations).

Unit ¹	Coordinates ²	Sample Name	Sample Type	Dated Material	UCIAMS ID ³	Fraction Modern	$\delta^{14}\text{C}$ (‰)	Radiocarbon Age ⁴ (years BP, 1 σ)	Calibrated Age ⁵ (cal BP)
Qft2	364944 E, 5466432 N	BR18-42C	Bulk sediment	Charcoal (single piece)	215248	0.6658 ± 0.0013	-334.2 ± 1.3	3265 ± 20	3560-3400
Qaf	364561 E, 5465854 N	BR18-06C	Macro charcoal	Not dated	–	–	–	–	–
		BR18-07C	Macro charcoal	Not dated	–	–	–	–	–
		BR18-08C	Macro charcoal	Charcoal (single piece)	215248	0.5185 ± 0.0010	-481.5 ± 1.0	5275 ± 20	6180-5940
Qp2	364400 E, 5466402 N	BR18-09C	Bulk sediment	Charcoal (composite of three pieces)	215249	0.3428 ± 0.0008	-657.2 ± 0.8	8600 ± 20	9600-9520
		0364683E, 5466207N	Bulk sediment	barren	–	–	–	–	–
		0365209E, 5466352N	Bulk sediment	barren	–	–	–	–	–
Qp1	0364515E, 5466653N	BR18-11C	Bulk sediment	barren	–	–	–	–	–
		0364515E, 5466653N	Bulk sediment	barren	–	–	–	–	–
		BR18-12C	Sediment	barren	–	–	–	–	–

¹See Figure 2.4 and Table A.1.

²NAD83 UTM Zone 10.

³Samples were prepared at PaleoTek Services. Sample preparation backgrounds have been subtracted, based on measurements of ^{14}C -free wood. These samples were treated with acid-base-acid (1N HCl and 1N NaOH, 75°C) prior to combustion. Samples were processed at the UC Irvine Keck AMS facility.

⁴All results have been corrected for isotopic fractionation according to the conventions of Stuiver and Polach (1977), with $\delta^{13}\text{C}$ values measured on prepared graphite using the AMS spectrometer. These can differ from $\delta^{13}\text{C}$ of the original material, and are not shown.

⁵Radiocarbon ages calibrated using INTCAL20 (Reimer et al., 2020) and OxCal v. 4.4 (Bronk Ramsey, 2021). Range reported represents unmodeled 95% confidence interval as calculated by OxCal.

ages of samples from Qaf and Qft2 (see below). The ~ 9.5 cal ka BP age is also broadly consistent with the timescales of paraglacial debris cone formation documented in recently deglaciated terrains that suggest these types of deposits form in the first 100s-1000s of years following deglaciation (Ryder, 1971; Ballantyne & Benn, 1996; Ballantyne, 2002).

Post-glacial units Qaf and Qft2 also yielded dateable macro-charcoal fragments. Qaf yielded one macro-charcoal sample (BR18-08C). This sample was collected from a stratified, clast-supported sand lens within interbedded sands and gravels ~ 0.75 m below the top of the deposit located ~ 500 m SW of fault scarps at Site 1 (Figure 2.4 and 3). This sample yielded a radiocarbon age of ~ 6 cal ka BP (Table 2.1). Qft2 yielded a charcoal sample (BR18-42C) sieved from bulk sediment collected from a stream cut exposure of stratified pebbles and cobbles, located < 10 m downhill from mapped fault strand Ee (Figure 2.4b, Figure 2.3). This sample yielded a radiocarbon age of ~ 3.5 cal ka BP (Table 2.1). Both ages are younger than the ages determined from a radiocarbon sample from paraglacial deposit Qp2 (~ 9.5 cal ka), and agree with our stratigraphic interpretation that Qaf is older than Qft2.

If we assume that these samples reflect deposit ages, and are not significantly altered by recycling, bioturbation, or inclusion of younger carbon, these data suggest the following as possible brackets on the ages of mapped deposits. Qt, Qk, and Qhm are likely ~ 11 - 14 ka, paraglacial deposits Qp1 and Qp2 are likely ~ 6 to ~ 11 ka, Qaf units are likely ~ 3 to ~ 9 ka, and Qft2 deposits are likely < 4 ka. Given the uncertainties inherent with this method and with the small number of samples available for dating, we treat these as age approximations.

2.4.3 Fault offset measurements

Results of our field mapping and topographic surveys show that the BRF has accommodated several meters of vertical and right-lateral displacement, distributed over a network of one to six fault strands that offset the mapped late Pleistocene to Holocene deposits (Figure 2.3). At Site 1 (Figure 2.5), scarp heights on individual fault strands range from 0.5 to 6 m, and channels appear in the field to be right-laterally offset by ~ 0.5 - 2 m. These observations suggest cumulative displacements of several meters across multiple fault strands. Similarly, at Site 2, scarp heights range from 1 to 3 m, and a series of three stream channels visible in lidar appear to be systematically right laterally sheared by several meters across three to five fault strands (Figure 2.6). Our field observations and survey data also show that scarp heights in older deposits and landforms, including the interfluvies developed in Qp1 at Site 1 and the till-mantled hillslopes at Site 2, have larger vertical displacements than the younger channels incised into these deposits, suggesting the potential for multiple events.

2.4.3.1 Slip vectors and fault kinematics

Estimates of slip based on our topographic survey data confirm our field observations that the BRF exhibits consistent right-lateral and dip-slip offset of the ground surface. Oblique-slip magnitudes across individual fault strands range from ~ 2 to 7 m at Site 1 and from ~ 2 to 5 m at Site 2 (Table A.2). Average dip-slip magnitudes for single faults range from ~ 1 to 5 m, with the largest dip-slip magnitudes of up to ~ 9 m observed at Site 1D (Table A.2). Average right-lateral strike-slip magnitudes recorded in offset channels and interfluves at Sites 1 and 2 range from ~ 1 to 5 m. Displacements of piercing lines across individual strands yield a $\sim 0.3:1$ to $1.5:1$ ratio of strike slip to dip slip, similar to those yielded by the cumulative displacements (Table A.2). These data suggest that, while the fault system as a whole accommodates approximately equal magnitudes of strike slip and dip slip, some individual fault strands are dominated by dip slip, while others are dominated by strike slip.

kinematic slip inversions of geomorphic piercing lines offset by the BRF produce pseudo focal mechanisms that similarly indicate right-lateral transtension along a steeply NE-dipping fault (Table A.3). Inversions performed for Sites 2A, 1C, 1D, and 1E (Figure 2.9 a-d) show small variations in the average strike and dip of the primary slip plane of $292\text{--}321^\circ$ and $66\text{--}78^\circ$, and in the average trend ($095\text{--}153^\circ$) and plunge ($10\text{--}26^\circ$) of the model slip vectors. These site-specific inversion data are similar to full fault inversions, and indicate an approximate slip trend and plunge of $\sim 110/45$ along an $\sim 80^\circ$ NE dipping fault plane. These full-fault pseudo focal mechanisms yield local P- and T-axes with trends and plunges of $170/37$ and $058/26$ respectively.

2.4.3.2 Cumulative displacements

At Site 1, available exposures allowed us to calculate cumulative displacement across one to three strands for 14 interfluves developed in Qp1 and 9 channels incised into Qp1 (Figure 2.5). These data show that cumulative oblique slip at Site 1 measured in offset interfluves and channels ranges from ~ 4 to 21 m (Figure 2.6). At Site 2, cumulative displacement of channels incised into Qt was summed across two to four mapped strands showing cumulative slip magnitudes of ~ 4 to 13 m (Figure 2.6). We note that cumulative oblique-slip magnitudes at Site 2 are likely underestimated, given that it was only possible to determine cumulative displacement across a portion of the mapped strands due to limited exposure and preservation.

Our calculated vertical and oblique displacement magnitudes show that older deposits typically record greater amounts of displacement than younger deposits. Examples of this relationship can be observed in the comparison of vertical separation along adjacent profiles at Sites 1 and 2 (Figure 2.10). At Site 1 (strand D, Figures 2.5, 2.10a) there is 5.8 m of vertical separation across an offset interfluve developed in Qp1, the oldest offset deposit at the site, whereas the adjacent, younger abandoned channel shows only 4.7

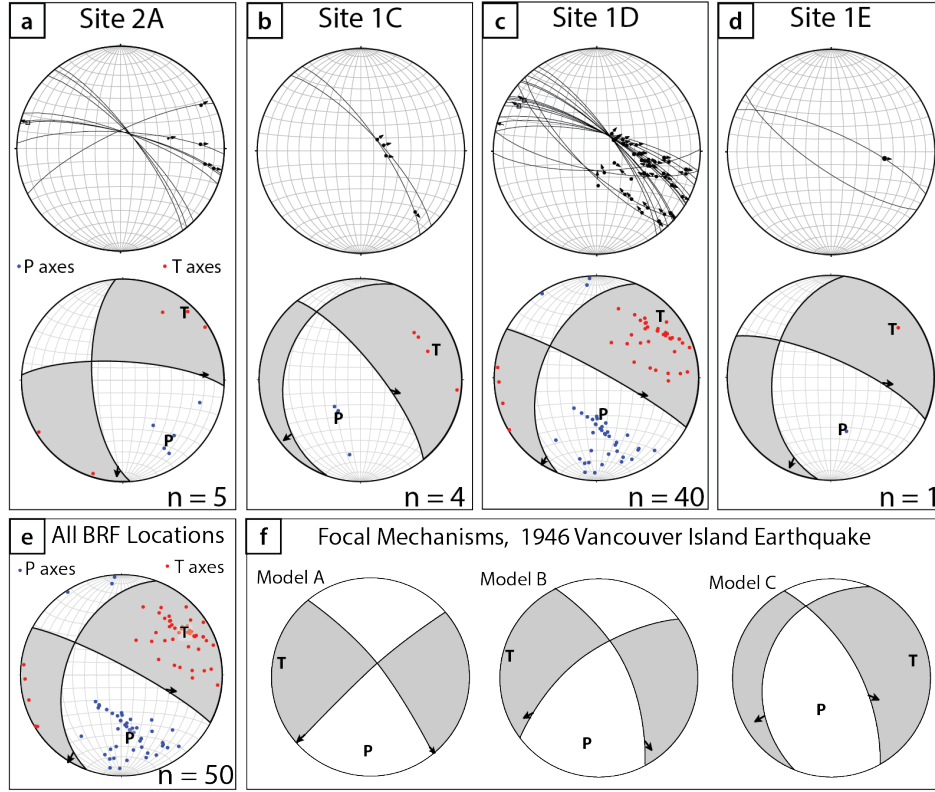


Figure 2.9: Right-lateral transtension along the Beaufort Range fault demonstrated by oblique slip of linear piercing points formed by paraglacial interfluves and channel thalwegs. Oblique slip vectors and pseudo focal mechanisms shown here are from kinematic inversions of detailed total station surveys of fault scarps and these offset geomorphic piercing lines. **A-D:** Kinematic data at four sub-sites along the BRF (see Figure 2.4 for locations). Upper panels: Lower hemisphere equal area projections showing fault planes, slip vectors, and hanging-wall motions. Lower panels: P- and T-axes and linked Bingham fault plane solutions (lower) for faults at locations 2A, 1C, 1D, and 1E. These slip vectors and kinematic inversions are consistent with right-lateral oblique motion on NE-dipping planes. **E:** Composite kinematic inversion for all surveyed sites along the BRF. Lower hemisphere equal area projection showing P- and T-axes and linked Bingham fault plane solutions. **F:** Focal mechanism solutions for the 1946 M 7.3 Vancouver Island earthquake (Rogers & Hasegawa, 1978, see Figure 2.2a for epicentral location). Model A is Rogers and Hasegawa's preferred model. Note the similarity in orientations of nodal planes and P- and T-axes for BRF fault kinematics.

m of vertical separation. A younger Qft2 fluvial terrace, which crosses adjacent fault strand Ee, has even less vertical separation (2.3 m). Similarly, at Site 2, we find that the till-mantled hillslope typically has larger vertical separation than channels incised into till. For example, profile 28 at Site 2 in Qt shows 4.1 m of vertical separation across strand Q, whereas profile 33 along a younger channel incised into Qt shows only 2.9 m of vertical separation (Figures 2.6, 2.10b). Finally, we were able to expand this assessment of cumulative displacement to a set of 23 interfluves and channels at Site 1 for which we are able to reconstruct 3D displacement. These data show that older interfluves developed in Qp1 consistently have ~4 to 10 m

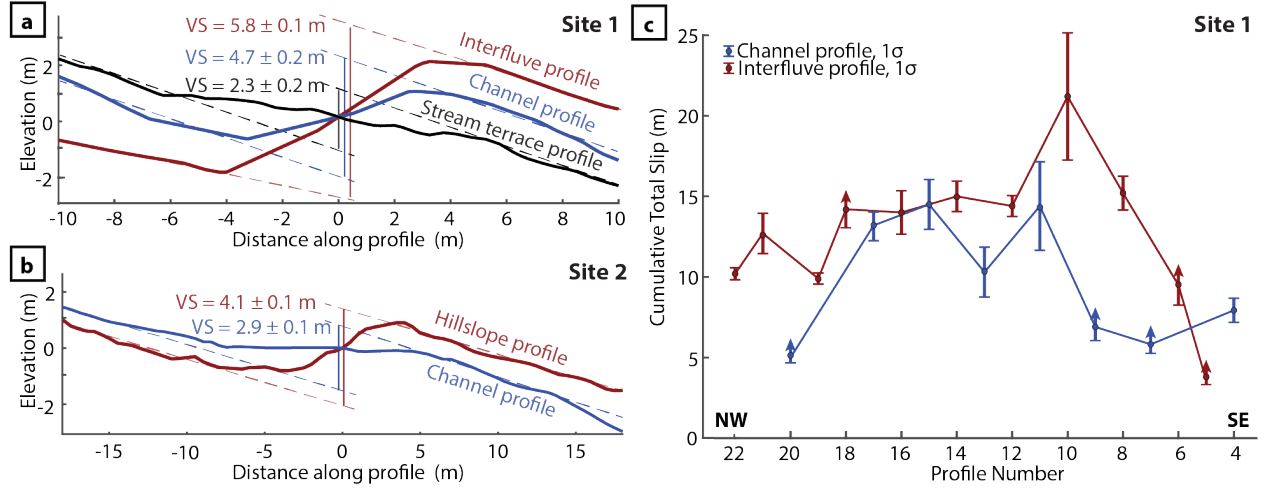


Figure 2.10: Topographic survey data showing differential magnitudes of fault offset of Quaternary landforms of different ages. **A:** Example from Site 1D where there is the largest magnitude of vertical separation across an interfluve developed in Qp1 (profile 10, strand D), intermediate magnitudes across a channel incised into Qp1 (profile 11, strand D, channel age correlative to deposit Qp2), and minimum magnitudes across a Holocene stream terrace (profile 3, strand Ee). **B:** Example from Site 2A where there is greater vertical separation across the till mantled hillslope (profile 28, Qt₁), and smaller separation across a channel incised into till (profile 33, strand Q). **C:** Cumulative slip estimates from Site 1D profiles: cumulative slip across interfluves (mean = 12.7 ± 4.4 m) is greater than for thalwegs (mean = 9.8 ± 3.9 m), suggesting interfluves have experienced at least one more event than thalwegs. Arrows indicate minimum slip estimates in locations where displacements across one or more strands could not be reconstructed. Site 1D profiles extend ~ 800 m along strike (see Figure 2.5a).

more cumulative oblique displacement as compared to young channels incised into Qp1 (Figure 2.10c).

2.5 Discussion

2.5.1 Characteristics of the Quaternary-active BRF

The field data and observations provided in this paper provide unequivocal evidence that the scarps we identify along the southwestern flank of the Beaufort Range are tectonic in origin and are associated with an active Beaufort Range fault. Mapped scarps form en echelon steps, and parallel arrays exhibit geometries common in strike-slip fault systems and pull-apart basins (e.g., Hatem et al., 2017; van Wijk et al., 2017), and occur along several tens of kilometers of strike length. The magnitudes of displacement and total fault lengths are consistent with observed displacement-length scaling relationships for active faults in Cascadia (Styron & Sherrod, 2021), and globally (D. L. Wells & Coppersmith, 1994; Wesnousky, 2008).

The scarps are inconsistent with formation processes associated with gravitational failure, glacial, or anthropogenic processes, for several reasons. First, scarps are predominantly uphill-facing, and are associated

with steep NE-dipping fault planes that produce “valley-side up” displacement. This sense of displacement is opposite to that predicted for landslide-related failures. Second, the scarps are quasi-linear and extend for several km along strike, whereas headscarps associated with landslides tend to produce curvilinear and discontinuous scarps with limited strike lengths. Third, the mapped scarps are inconsistent with formation by sackungen (McCalpin, 1990), which typically form sets of parallel scarps at range crests, rather than the en echelon scarps we observe near the base of the range (e.g., Figure 2.2b as compared to Figure 2.3c). Finally, our field observations also confirm that these scarps are not associated with roads, logging tracks, or other anthropogenic disturbances, nor are they associated with glacial scouring or glacially-streamlined deposits (see Section A.2).

Our data indicate that the BRF consists of a set of high-angle faults, with an average 60-88° NE dip, with local fault strand strikes ranging from $\sim 270^\circ$ to 320° . While individual fault strands extend for several hundred m to several km, these strands collectively define a discontinuous network of scarps that we interpret to be the surficial expression of a single fault zone at depth. Such discontinuous fault scarp networks are common in strike-slip systems, especially in immature faults with little cumulative offset (e.g., Hatem et al., 2017), and have been observed along other forearc faults in northern Cascadia (e.g., Morell et al., 2017). The mapped network of BRF fault scarps identified in this study extends for ~ 40 km from Port Alberni to Comox Lake (Figure 2.2b and 1). Additional potential scarps visible in lidar DEMs occur along strike of the active BRF outside the map area, suggesting that the BRF may have a cumulative length that is >40 km. If all of the mapped scarps in this study are associated with a continuous subsurface fault network, then the BRF is one of the longest strike-length faults identified in northern Cascadia to date.

2.5.2 Kinematics of the BRF and relationship to inherited structures

Our field mapping and topographic survey data demonstrate that the active BRF is a transtensional structure that accommodates right-lateral oblique slip along a steeply NE-dipping fault zone (Figure 2.9). Three lines of evidence support this interpretation: 1) Field observations show consistent right-lateral offset of abandoned channels and interfluvies and net NE-side-down vertical displacement. 2) Slip vectors resolved by reconstructing piercing lines similarly indicate NE-side-down hanging wall motion, consistent with right-lateral transtensional slip on a steeply NE dipping fault (Figure 2.9). These kinematics are consistent with mapped fault scarp geometries that suggest formation during right-lateral transtension. For example, at Site 1 (Figures 2.5, 2.8), there is an en echelon array of faults with opposing dips that is consistent with the map patterns expected for a right-lateral transtensional negative flower structure. 3) Pseudo focal mechanism inversion of slip vectors indicate that the BRF accommodates right-lateral transtension along a steeply NE

dipping fault plane.

The NE-side-down slip sense we determine for the Quaternary-active BRF produces a “range-side down” sense of motion. This result suggests that the high elevations and steep topography associated with the southwestern flank of the Beaufort Range were not formed by transtensional slip along the active BRF. Instead, the steep range front may be the product of differential erosion of the softer Cretaceous Nanaimo Gp sediments that underlie the Alberni Valley, relative to the more resistant Karmutsen Fm basalts that underlie the range crest (Muller & Carson, 1969). Or, this may imply that the net range-side-down (NE-side down) motion integrated over 100s kyr to Myr across the active BRF may be small relative to the amount of Eocene NE-side-up thrust fault displacement. The small cumulative magnitude of NE-side-down motion could indicate that transtension across the BRF is a relatively young phenomenon, and has not accrued a large magnitude of vertical displacement.

Finally, our data suggest that active BRF strands do not appear to directly reoccupy inherited thrust fault planes. The presence of active BRF scarps in both the hanging wall and footwall of inherited thrust faults (Figure 2.4) suggest that there is not a strong inherited lithologic or mechanical control on the position of the active BRF at the surface. Furthermore, there is an apparent difference between the near-surface dip of the Quaternary-active BRF (70-90°) and that of the inherited Eocene Beaufort Range thrust fault (~45-70°). There are two possible explanations for this apparent dip discrepancy. First, these observations could imply that the subsurface projections of the active and Eocene faults may diverge at depth. Similar discrepancies between active and inherited fault geometries have been observed in the northern Cascadia forearc along the Leech River and North Olympic faults (Morell et al., 2017; G. Li et al., 2018; Nelson et al., 2017; Schermer et al., 2021). These data suggest that it is possible that the active BRF may reflect the formation of a new fault, more optimally oriented in the forearc stress field, rather than slip on an inherited bedrock structure. The second possibility is that the dip of the BRF is steep near the surface, but has a more gentle dip at depth, such that the active transtensional fault follows the Eocene thrust fault at depth. The geometry, kinematics, and slip history of the active BRF therefore provide critical insight into the neotectonic stress and strain fields in the northern Cascadia forearc.

2.5.3 Evidence for multiple surface-rupturing late Pleistocene to Holocene earthquakes

Our tectonogeomorphic mapping, topographic surveys of offset abandoned channels and interfluvies, and field observations of fault scarp morphology support the hypothesis that the BRF has hosted multiple earthquakes since the deglaciation of the Alberni Valley (~14-11 ka). The strongest evidence for multiple events

comes from the differential scarp heights and cumulative slip magnitudes calculated for offset landforms of different ages at Sites 1 and 2 (Figure 2.10). At Site 1, interfluvial crests developed in the older paraglacial unit Qp1 have greater vertical separation (~ 1 m) and greater cumulative oblique slip (~ 1 -3 m) than abandoned channel thalwegs incised into that same unit. These abandoned channels in turn have greater vertical separation (~ 2.4 m) than the displacement surveyed across a younger Qft2 fluvial terrace. The differential offset between interfluvies, channels, and fluvial terraces indicates the occurrence of at least three events since the deposition of Qp1 at Site 1. At Site 2, differential scarp heights of ~ 2 m between those developed in till-mantled hillslopes and younger channels incised into the hillslopes indicate at least two surface-rupturing events have occurred at this site following the deposition of Qt. Furthermore, if we make the simplifying assumption that a single event produces ~ 1 -3 m of oblique slip, based on the average difference in cumulative oblique displacement between interfluvies and channels, these data suggest the BRF may have hosted more than three events since ~ 11 -14 ka.

At Site 1, we can place broad constraints on the relative timing of slip events by combining our estimates of deposit ages (Table 2.1; Figure 2.4c) with offset magnitudes (Table A.2; Figure 2.6). The timing of the first event is constrained by the observation that older interfluvies developed in Qp1 have more cumulative oblique offset than channels developed in Qp1. This observation indicates that at least one event must have occurred more recently than the deposition of Qp1, which occurred after deglaciation (~ 11 -14 ka), but before the abandonment of channels incised into Qp1. The timing of channel abandonment is not directly dated, but our correlation of channel incision to the deposition of Qp2 suggests channel abandonment occurred after the deposition of Qp2 (radiocarbon dated to ~ 6 to 11 ka) and before the deposition of Qaf (radiocarbon dated to ~ 3 -6 ka). Therefore, the first event(s) likely occurred after ~ 11 -14 ka, but before ~ 3 -6 ka. The timing of the second event is constrained by the difference in offset between channels and inset Qft2 terraces. This difference requires one or more events to have occurred after channel abandonment (which we infer occurred after 6-11 ka), but before the formation of the Qft2 terrace (radiocarbon dated to $< \sim 4$ ka). The occurrence of a third event is supported by the ~ 1.5 m of vertical offset of the Qft2 terrace. Therefore, the most recent event must have occurred after the deposition of the Qft2 terrace (since ~ 4 ka).

These data suggest that the BRF has experienced at least three events over the late Pleistocene to late Holocene. The persistence of right-lateral transtensional deformation along the BRF for several thousand years after the retreat of glaciers from the Alberni Valley indicates that deformation cannot be attributed solely to changes in crustal loads and stresses due to glacial unloading and viscoelastic relaxation of the crust and mantle (e.g., Anderson et al., 1989; Craig et al., 2016; Davenport et al., 1989; Lagerbäck, 1990; Mörner, 1991; Muir-Wood, 2000; Jarman & Ballantyne, 2002; van Loon et al., 2016). Such “glacially-induced” earthquakes typically occur during or within a few kyr of glacial retreat, when changes in ice

loads and crustal stresses from the viscoelastic rebound are greatest (Steffen et al., 2014). While our data do not constrain the precise timing of the first event(s), and cannot rule out that early events on the BRF were impacted by glacial unloading, they do indicate earthquakes have occurred in the middle to late Holocene, well after the largest stress changes due to glacial loading would have occurred. Additionally, we note that glacial unloading typically reduces vertical stress, and given that our slip data indicate the BRF accommodates transtension, such unloading would reduce the deviatoric stress making failure less likely. Overall, the persistence of right-lateral transtensional events throughout the Holocene suggests that tectonic forces are the principal drivers of deformation, and any glacial impacts are secondary.

Our estimates of displacement per event, and measurements of the total length of the active BRF from mapped or inferred fault scarps, allow us to estimate the magnitude of paleo-earthquakes at these sites using displacement scaling relationships. An estimated $\sim 1\text{-}3$ m of displacement per event suggests that the BRF could have hosted M_W 6.9 to 7.2 events (D. L. Wells & Coppersmith, 1994). These magnitudes are similar to those determined based on our total mapped fault length of 35 to >40 km, which suggests M_W 6.8 to >7.0 events (D. L. Wells & Coppersmith, 1994; Wesnousky, 2008). These earthquake magnitudes are similar in scale to the M 7.3 magnitude calculated for the 1946 Vancouver Island earthquake, which caused significant damage, including to telephone wires, underwater telegraph cables, and the hospital in Port Alberni, BC (Hodgson, 1946). An earthquake of a similar magnitude today would pose significant hazard not only to Port Alberni, but also to the nearby communities of Nanaimo, Parksville, Qualicum Beach, and Courtenay (Figure 2.2a). Failure of dams on Comox Lake and Elsie Lake could lead to flooding of communities downstream (Figure 2.2b), as well as impacts on power availability, as nearby power stations supply 11% of the electricity generated on Vancouver Island (BC Hydro, 2023).

2.5.4 Estimate of Late Pleistocene to Holocene slip rates

Estimation of fault slip rates yields important information relevant to understanding how strain is partitioned among faults, and they represent primary data used in seismic hazard analyses (Morell et al., 2020). Ideally, slip rates are calculated when at least two precise earthquake ages and the displacement associated with the bracketed event are known (e.g., a closed-interval slip rate, DuRoss et al., 2020) and estimated over a time period spanning more than 5 earthquakes (Styron, 2019). In the absence of precise earthquake ages, a geomorphic slip rate can be estimated that affords an estimate of the slip rate that has accrued since the development of the geomorphic feature, using estimates of the date of the landform and displacement data recorded in the geomorphic feature. However, such an open-interval slip rate can be biased because the age of the geomorphic feature can differ from the age of the earthquake that deformed the feature.

On the BRF, we currently do not have precise earthquake ages, and we do not know the displacement between earthquakes with precision. However, we nonetheless make broad estimates of slip rate based on the known ages and displacements between known events. The cumulative displacement of different deposits, their depositional ages, and our estimates of event timing, place bounds on fault slip rates for the active BRF. Both open-ended and closed interval approaches at Site 1 yield slip rates for the BRF that range from ~ 0.5 to ~ 2 mm/yr. The open-ended approach yields a rate of ~ 0.7 - 1.3 mm/yr, based on the ~ 10 to 15 m of cumulative oblique slip across all mapped fault strands and an estimated deglaciation age of ~ 11.5 - 13.6 ka. There is only one reliable closed interval calculation that can be made given the available offset data and uncertainties in event timing. This interval calculation uses the ~ 8 - 9 m cumulative displacement of the channels at Site 1, and the difference in age between the Qft2 terrace and the interfluvies at Site 1, which could range from ~ 3.5 to 13.6 kyr. This closed interval spans at least two events, and yields a slip rate estimate of 0.6 - 2.6 mm/yr. Given the uncertainties in the ages of events and displacement magnitudes in this closed interval calculation, we prefer the more conservative open-ended rate of 0.7 - 1.3 mm/yr.

These data demonstrate that, even at the lower bound of uncertainty, the Beaufort Range fault is one of the fastest-slipping Quaternary faults in the northern Cascadia forearc. Our slip rate estimates of ~ 0.5 to ~ 2 mm/yr indicate that the BRF has a higher slip rate than the nearby LRF (0.2 - 0.3 mm/yr; Morell et al., 2018, 2017) and Darrington-Devils Mountain fault zone DMF (0.14 ± 0.1 mm/yr; Personius et al., 2014), and a similar slip rate to the NOFZ (1.3 - 2.3 mm/yr, 3-5 post-glacial earthquakes; Schermer et al., 2021). These data suggest that the BRF is a major crustal structure that accommodates permanent deformation in the northern Cascadia forearc.

2.5.5 Comparison of the 1946 Vancouver Island earthquake kinematics with slip on the BRF

The field data reported here cannot constrain the timing of the most recent event along the active BRF beyond that it occurred after ~ 3.5 ka, and therefore cannot directly test whether the 1946 event ruptured along the BRF within our field area. However, our data from the Holocene BRF share striking similarities to the kinematics, spatial distribution, and fault plane solutions for the 1946 event.

The pseudo-focal mechanism solutions for the BRF determined from kinematic slip inversions of geomorphic piercing lines offset by the BRF have similar slip planes, slip vectors, and P- and T-axes as those determined for the 1946 Vancouver Island earthquake (Figure 2.9). Focal mechanism solutions for the 1946 earthquake (Rogers & Hasegawa, 1978) have NW-SE striking nodal planes with strikes of 319 - 332° and dips of 66 - 79° , and SW-NE striking nodal planes with strikes of 222 - 233° and dips of 36 - 85° (Figure 2.9f, Table

A.3). These nodal plane attitudes are strikingly similar to the nodal planes of the pseudo-focal mechanisms solutions derived from fault slip vectors along the active BRF (Figure 2.9). The NW-SE nodal plane of the focal mechanism preferred by Rogers and Hasegawa (1978) of 319/79 NE is sub-parallel to our calculated attitude of the active BRF of $\sim 270\text{-}320/\sim 75$, and the predicted slip vectors associated with the NW-SE striking nodal planes for the 1946 earthquake have trends ranging from 114 to 143° , and plunges ranging from 05° to 55° —similar slip vector orientations to those determined from offset piercing lines along the active BRF. Finally, the stress axes determined for the 1946 focal mechanism solutions have moderately plunging, southerly trending P-axes and sub-horizontal T-axes with trends similar to those determined for the active BRF (Figure 2.9).

The fault slip vectors and transtensional pseudo-focal mechanisms that we determine for the BRF are also similar to fault plane inversions based on geodetic motions associated with the 1946 earthquake (Slawson & Savage, 1979). Repeat surveys of topographic benchmarks across the Beaufort Range at the latitude of the earthquake epicenter ($\sim 49.45^\circ$ N) suggest right-lateral oblique slip on a steeply (70°) NE-dipping fault plane that extended for 60 km along strike (Slawson & Savage, 1979). Our mapped Sites 1 and 2 along the BRF therefore lie within the modeled event rupture area, and the fault plane dip is similar to the $60\text{-}88^\circ$ NE dip we determine for the BRF. In addition, slip inversions for the 1946 event fault planes indicate the crustal displacements are best reproduced by ~ 1 m of right-lateral and ~ 2 m dip slip, along 60 km fault length parallel to the BRF. Therefore, both the relative ratio of strike slip to dip slip ($\sim 0.5:1$) and the estimated slip per event ($\sim 1\text{-}2$ m) modeled for the 1946 event are similar to our slip ratios of $0.3\text{-}1.5:1$ (strike slip to dip slip) and estimates of $\sim 1\text{-}3$ m of oblique slip per BRF event.

These data collectively indicate that the Holocene slip observed along the active BRF is kinematically and spatially compatible with the slip inferred for the 1946 Vancouver Island earthquake. These correlations suggest that, if the 1946 event failed along a NW-SE striking, steeply NE-dipping plane, as suggested by Rogers and Hasegawa (1978) and Slawson and Savage (1979), the BRF is a likely candidate for hosting this event. Our estimated age of the most recent event of $<3\text{-}4$ ka allows for this possibility. Furthermore, our field offset data provide evidence that the active BRF has hosted at least 3 earthquakes since the late Pleistocene, each with slip that is compatible with that modeled for the NW-SE striking nodal plane for the 1946 event. These observations suggest then that the BRF hosted multiple 1946-like events over the late Pleistocene to Holocene.

2.5.6 Implications for forearc strain accommodation

These results have several key implications for the long-term permanent strain accumulation in the northern Cascadia forearc. First, our field data and kinematic slip inversions of geomorphic piercing lines offset by the BRF indicate that this portion of the Cascadia forearc has experienced right-lateral transtension over the past ~ 11 -14 ka. In addition, the similarity between the fault kinematics integrated over multiple paleoseismic events spanning the late Pleistocene to Holocene and the 1946 Vancouver Island earthquake suggests that these transtensional kinematics are representative of the local upper plate deformation field over decadal to millennial time scales. If true, these time scales would span multiple upper plate fault seismic cycles, which likely have recurrence intervals of 1000s of years (e.g., Morell et al., 2018; Schermer et al., 2021), and multiple subduction interface megathrust seismic cycles, which have recurrence periods of ~ 390 -540 years (e.g., Walton et al., 2021).

Second, the kinematic slip inversions for the active BRF suggest that the P- and T-axes inverted for long-term (late-Pleistocene to present) slip are consistent with regional stress patterns derived from historical seismicity (Figure 2.2a; Balfour et al., 2011). We find that the trends and plunges of P- and T-axes determined for the BRF are within ~ 20 -40° of the P- and T-axes determined for the 1946 Vancouver Island earthquake and from inversions determined from nearby upper plate seismicity (Balfour et al., 2011). Several studies have suggested that variations in trench-perpendicular tractions during the megathrust seismic cycle can cause inversions in principal stress orientations in the overriding plate, such that forearc fault slip sense can vary as a function of the megathrust seismic cycle (e.g., Wang et al., 1995; Loveless et al., 2010; Regalla et al., 2017). However, the consistency of P- and T-axes determined from historical earthquakes and from paleoseismic slip suggests that BRF kinematics have not changed drastically over Holocene time scales. These data suggest that there is some level of consistency in the deformation field associated with short-term (decadal) upper plate seismicity and long-term (kyr-scale) fault slip, and a potentially similar temporal consistency in the upper plate stress field.

Third, a comparison between these BRF fault kinematics to those determined for other active faults in the northern Cascadia forearc, suggests that there may be a spatial transition from a forearc deformation field promoting right-lateral transpression near the Olympic Mountains and on southernmost Vancouver Island, to one promoting right-lateral transtension in the northern Cascadia forearc on central Vancouver Island. Specifically, the North Olympic fault zone, the Darrington-Devils Mountain fault, the Southern Whidbey Island fault, the Leech River fault, and the XEOLXELEK-Elk lake fault appear to be accommodating right-lateral slip and compression, as determined by earthquake focal mechanisms and paleoseismic data (Sherrod et al., 2008; Personius et al., 2014; Schermer et al., 2021; Morell et al., 2018; Harrichhausen et al., 2021,

2023). In contrast, at the latitude of central Vancouver Island, the BRF appears to be accommodating right-lateral transtension, as determined by kinematic slip inversions of geomorphic piercing lines offset by the BRF.

This observation suggests that there may be a change in the upper plate strain field from one favoring transpression on faults near the Olympic Mountains to one favoring transtension on faults in northern Cascadia, around the latitude of 48.5-49° N. This change in strain field may be related to spatial variations in principal stress orientations and magnitudes in the upper plate that locally promote transtension along the BRF. While the data presented here are not sufficient to determine the causes of this potential change in the upper plate deformation field, there are several possibilities, including oroclinal bending (Johnston & Acton, 2003; Finley et al., 2019; Harriehausen et al., 2021), spatial changes in plate tractions, convergence rate, or obliquity (R. E. Wells et al., 1998; Wang, 2000; McCaffrey et al., 2013; S. Li et al., 2018), or the ‘escape’ of forearc crustal blocks related to north-directed shear from southern Cascadia (Nelson et al., 2017). The consistency of fault kinematics and P- and T-axes calculated near the BRF from both seismic and paleoseismic data suggest that a deformation field favoring local right-lateral transtension has persisted over both decadal and millennial timescales.

2.6 Conclusions

We provide the first geologic field evidence that the Beaufort Range fault is a seismogenic fault, and demonstrate that it actively accommodates right-lateral transtension within the northern Cascadia forearc of central Vancouver Island. Field mapping and topographic surveys document >35-40 km of northwest-striking, primarily northeast-dipping fault strands along the southwestern flank of the Beaufort Range. These scarps occur in discontinuous, en echelon and parallel sets and offset late Pleistocene to Holocene glacial, paraglacial, and post-glacial deposits. We observe an increase in scarp height and total offset with increasing unit age that provides evidence for at least three surface-rupturing earthquakes on the BRF since ~13.6-11 ka, the most recent of which occurred in the past ~3-4 kyr. Slip magnitudes reconstructed from offset piercing lines, total fault length, and the ages of offset deposits suggest that the BRF is capable of hosting earthquakes of M_W 6.5-7.5, and has a late Pleistocene to Holocene slip rate of 0.5 to 2 mm/yr. Thus the BRF is a major forearc fault accommodating deformation in the northern Cascadia subduction zone, and poses significant hazard to communities and infrastructure on Vancouver Island.

Notably, kinematic slip inversions of geomorphic piercing lines offset by the BRF yield transtensional pseudo-focal mechanisms, fault geometries, slip vectors, and P- and T-axes that are remarkably similar to those determined for the 1946 Vancouver Island earthquake. These data suggest that the BRF is a candidate

structure to have hosted this event. The consistency of right-lateral transtensional slip kinematics between the 1946 earthquake and late Pleistocene to Holocene slip on the BRF suggests that this portion of the northern Cascadia forearc has accommodated regional transtension over decadal to millennial time scales, spanning multiple earthquake cycles.

Open Research

New data produced in this study:

Data collected and analyzed in this manuscript are available in a Dryad Data Repository (Lynch et al., 2023) at <https://datadryad.org/stash/share/Ui8KejoZgz41xsl0ZUxfCQV2Nea3JsVwQoTz9DZ1iho>.

The repository contains the following data:

1. Text files containing raw field data (x,y,elevation) of surveyed offset landforms
2. Text files containing raw field data (x,y,elevation) of fault midpoint locations
3. R script for calculating 3D fault plane geometry
4. R script for calculating 3D offsets of linear piercing lines across a dipping fault

The contents of this repository are also included in Appendix B.

Previously published data and programs used in this study:

1. Lidar is available through LidarBC at <https://www2.gov.bc.ca/gov/content/data/geographic-data-services/lidarbc>
2. The USGS Quaternary faults and folds database used for is available at <https://www.usgs.gov/programs/earthquake-hazards/faults>.
3. The BC Geological Survey (BCGS) bedrock geology map used for Figures 2.1, 2.2, and 2.4 is available at <https://www2.gov.bc.ca/gov/content/industry/mineral-exploration-mining/british-columbia-geological-survey/geology/bcdigitalgeology>.
4. The OxCal program v. 4.4 by C. Bronk Ramsey used for radiocarbon calibration is available at <https://c14.arch.ox.ac.uk/oxcal/OxCal.html>.
5. The R. Allmendinger FaultKin 7.6 program used for plotting and analyzing fault plane and slip vector data in Figure 2.9 is available at <http://www.geo.cornell.edu/geology/faculty/RWA/programs/faultkin.html>.
6. The OSX Stereonet 9.9.4 program used for plotting bedrock fault planes and slickenlines in Figure 2.7 is available at <http://www.geo.cornell.edu/geology/faculty/RWA/programs/stereonet.html>.

Acknowledgments

This work was conducted on the traditional territory of the Hupačasath, K'ómoks, and čišaaʔath people, and we acknowledge their historical relationships with the land that continue to this day. Funding for this project was supported by NSF EAR Tectonics 1756834/ 2004684 to Regalla and 1756943 to Morell, by the Boston University Department of Earth and Environmental Sciences, and the Northern Arizona University Duebendorfer-Barnes Structure Endowment. We would like to thank Todd Thompson, TimberWest, and Island Timberlands for land access, and Mosaic/Island Timberlands/TimberWest for providing lidar ground return data, as well as Scott Bennett for illuminating discussions in the office and field and Edwin Nissen and the University of Victoria School of Earth and Ocean Sciences for discussion of local tectonics and help with field logistics. We would also like to acknowledge the late Vic Levson for help with understanding glacial systems of Vancouver Island, and Johanna Fisch and Tatum McLeod for assistance in the field.

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Chapter 3

Five late Pleistocene to Holocene ruptures reflect non-steady slip rates on the Beaufort Range fault: implications for seismic hazard in the northern Cascadia forearc

Abstract

Determining the timing and magnitude of past earthquakes on forearc faults is a critical step in quantifying strain accommodation in subduction zone forearcs and developing robust seismic hazard assessments. No historical earthquakes in the northern forearc of the Cascadia Subduction Zone (CSZ) have been attributed to a specific fault. Paleoseismic trenching is therefore an important tool for unearthing the rupture histories of forearc faults prior to written records. Tectono-geomorphic mapping identified the Beaufort Range fault (BRF) on central Vancouver Island, BC, Canada, as an active right-lateral oblique fault that has hosted multiple earthquakes in the Quaternary. Here we discuss the results of a paleoseismic trench study (Woodlot Trench) on the southern BRF. We find that post-glacial slope adjustment was punctuated by earthquakes that created a fault scarp and blocked the channel flow. Five M_W 6.7-7.5 surface-rupturing earthquakes (Event 1 10.9-10.3 ka, Event 2 5.2-4.9 ka, Event 3 4.9-4.2 ka, Event 4 4.2-3.0 ka, and Event 5 2.8-0.7 ka) occurred on the BRF; however, recurrence intervals and slip rates are non-steady. A long ($\sim$ 5-6 kyr) quiet period occurred in the late Pleistocene to early Holocene, whereas three earthquakes occurred between $\sim$ 5-3 ka, yielding closed-interval slip rates of $\sim$ 0.2 mm/yr to $\geq$ 2 mm/yr. The Events 3-4 and 4-5 interval slip rates are consistently higher than the early to mid-Holocene rates. Radiocarbon data suggest the most recent event in the Woodlot Trench is not the 1946 M_S 7.3 Vancouver Island earthquake, but is instead an event $\sim$ 2-2.5 kyr older.

3.1 Introduction

Forearc faults pose significant hazard to nearby communities (e.g., Johnson et al., 2004; Morell et al., 2018; Santibáñez et al., 2018; Escuder-Viruete & Pérez, 2020), and form a component of strain accommodation in subduction zones (e.g., Fitch, 1972; Wang, 2000; Symithe et al., 2015); however, to accurately assess

their hazard and quantify strain accumulation, the magnitude and timing of past ruptures on these faults must be characterized. Globally, damaging earthquakes have occurred on previously unrecognized faults in subduction zone forearcs, such as the 2010 Darfield, NZ, 2011 Christchurch, NZ, and 2022 Pasaman, West Sumatra earthquakes (Beavan et al., 2011; Supendi et al., 2023). The northern Cascadia Subduction Zone (CSZ) forearc has hosted several large and damaging earthquakes, and the Holocene rupture histories of some forearc faults have been documented (e.g., Leech River fault, North Olympic fault zone, Darrington-Devils Mountain fault zone, Canyon River fault; Morell et al., 2018; Schermer et al., 2021; Personius et al., 2014; Bennett et al., 2017), but no historic earthquakes have been linked to a particular forearc source fault. The Beaufort Range fault (BRF) on central Vancouver Island (Figure 3.1) has recently been identified as an important active forearc fault (Chapter 2). The BRF is a candidate source fault for the 1946 M_S 7.3 Vancouver Island earthquake (Rogers & Hasegawa, 1978), which was the largest historical event in Canada and occurred at a depth of <30 km near the northern end of the mapped BRF (Figure 3.1a). Previous work has determined that the BRF has ruptured in the Holocene, but no paleoseismic study that could constrain the timing of individual events and magnitude of past earthquakes, and the test if the 1946 earthquake occurred on the BRF has yet been conducted. This work is necessary to estimate recurrence intervals for use in seismic hazard assessment, and slip rates need to be constrained to understand how much strain is accommodated on forearc faults.

Our recent work (Chapter 2) has identified the BRF as one of the longest forearc faults in northern Cascadia, with right-lateral oblique offsets resulting from Quaternary ruptures. We showed that the BRF experienced right-lateral oblique displacement from at least three Holocene earthquakes. Late Pleistocene to Holocene slip rates on the BRF are ~ 0.5 -2 mm/yr, as fast as or faster than any other fault in the northern Cascadia forearc, indicating that upper plate deformation occurs farther north than previously documented. However, the number and timing of ruptures along the BRF is unconstrained.

In this study, we excavate a paleoseismic trench across the BRF to expose deposits offset by Holocene earthquakes, use radiocarbon dating to constrain the timing of individual earthquakes, and estimate closed-interval slip rates using per-event displacements determined from the trench. We find that the BRF has hosted at least five M_W 6.6-7.2 surface-rupturing earthquakes since ~ 11 ka, and that recurrence intervals vary significantly. Closed-interval slip rates on the BRF range from ~ 0.2 mm/yr in the late Pleistocene to mid-Holocene to >2 mm/yr in the middle and late Holocene, indicating that there may be temporally variable strain release over millennial timescales.

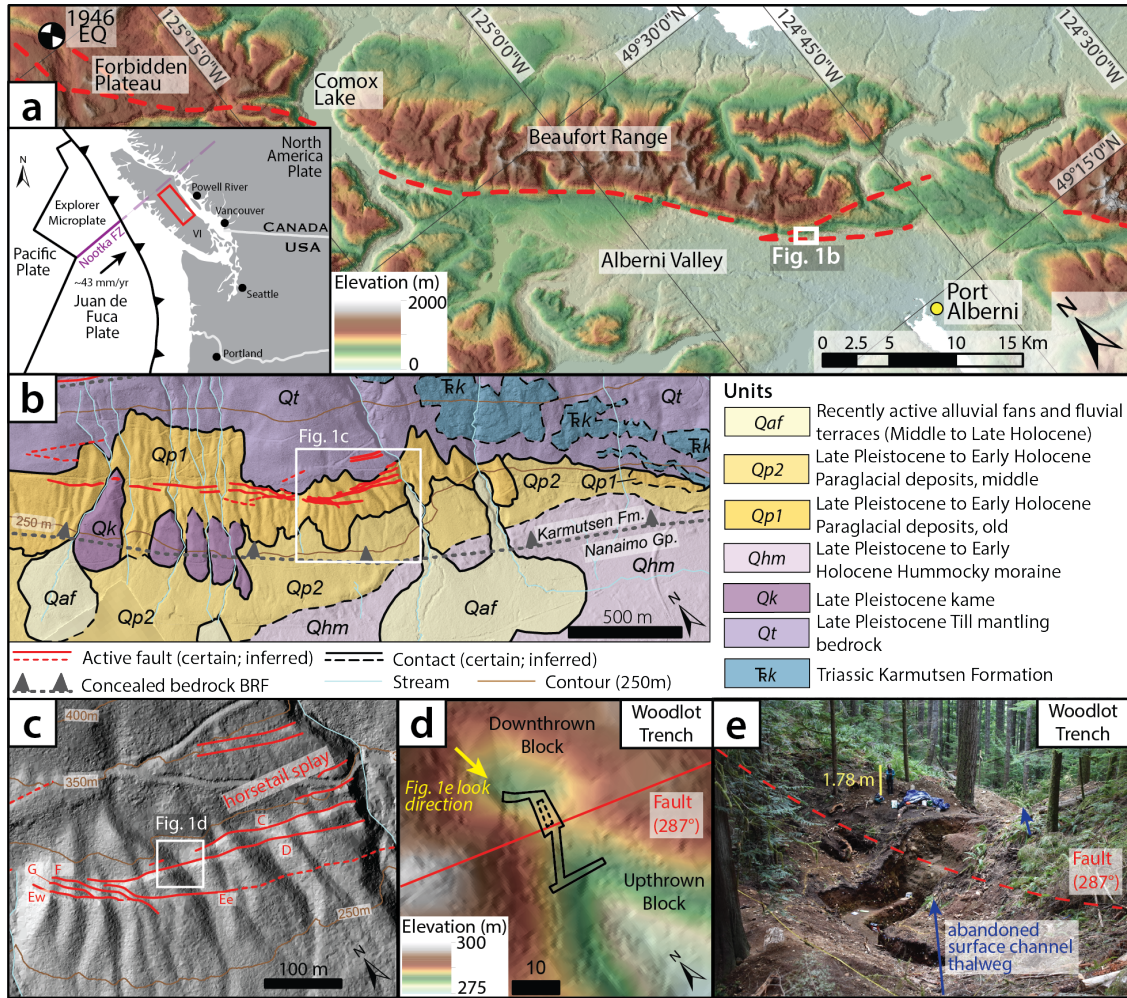


Figure 3.1: Location and setting of the Woodlot paleoseismic trench along the southern Beaufort Range Fault (BRF). **A.** Hillshaded SRTM DEM of the Alberni Valley and Beaufort Range, with inset showing location within the Cascadia Subduction Zone. Red box on the inset figure shows the location of panel a. Active fault scarps mapped in red are simplified from Chapter 2. Relocated 1946 Vancouver Island earthquake focal mechanism after Rogers and Hasegawa (1978). White box notes the location of panel b. VI - Vancouver Island. **B.** Geologic mapping of the area surrounding the Woodlot Trench (after Chapter 2). **C.** Hillshaded lidar DEM of the trench showing mapped fault strands. White box notes the location of panel d. **D.** Footprint of the Woodlot Trench overlain on a hillshaded lidar DEM, with red line showing strand D (mapped in panel c). **E.** Photo of the Woodlot Trench, with the block north of the scarp in the foreground (view looking south). Blue arrows mark the location of the modern-day expression of the surface channel thalweg on the block north of the scarp in the foreground, and the block south of the scarp in the background. Blue arrows point downslope and are offset right-laterally ~2 m.

3.2 Geologic Setting

The BRF is located in the northern Cascadia forearc in central Vancouver Island, where the Juan de Fuca plate subducts under North America at a rate of ~ 43 mm/yr (Kreemer et al., 2014). The BRF is one of a network of active faults in the Cascadia forearc (e.g., Morell et al., 2017; Harrichhausen et al., 2021; Delano et al., 2017; Nelson et al., 2014; Personius et al., 2014), and it strikes NW-SE along the SW side of the Beaufort Range (Chapter 2), sub-parallel to the bedrock thrust fault that formed in the Eocene (e.g., England & Calon, 1991; England, 1990; Cui et al., 2017, ; Figure 3.1). The BRF is the northernmost known active fault in the upper plate of the Cascadia subduction zone, located near the city of Port Alberni, BC, ~ 60 km south of the Nootka fault zone that marks the northern limit of the Juan de Fuca slab (Figure 3.1a). The M_S 7.3 1946 Vancouver Island earthquake occurred near the northern end of the mapped BRF (Figure 3.1a). The relocated focal mechanism of Rogers and Hasegawa (1978) is along strike from the BRF, and one nodal plane is parallel to the BRF (Figure 3.1a). The source fault for the 1946 event remains uncertain due to the lack of both surface ruptures and aftershocks (Rogers & Hasegawa, 1978; Slawson & Savage, 1979), but scaling relationships using a 35-60-km-long potential surface rupture, and estimates of per-event displacement, suggest that the BRF is capable of hosting M_W 6.5-7.5 earthquakes like the 1946 event (Chapter 2). However, no paleoseismic study has evaluated if this earthquake occurred on the BRF.

The active BRF, as mapped in Chapter 2, consists of >60 km of sub-parallel fault scarps, benches, and linear sags in a ~ 500 m-wide zone. At the southern end of the BRF near Port Alberni, the fault zone consists of several steeply ($>70^\circ$) NE- and SW-dipping strands (Figure 3.1b). These strands displace a series of abandoned channels and interfluvies both vertically (NE-side-down) and right-laterally. Detailed surveys of these offset landforms indicate right-lateral transtension. The landforms at this site exhibit increasing displacement with age, indicating that there have been at least three earthquakes since ~ 13 ka (Chapter 2).

We selected a location for a paleoseismic site (Woodlot Trench) where the fault crosses a disrupted channel and generates a large (~ 6 m-tall) uphill-facing fault scarp that blocks downhill flow within the channel (Figure 3.1c). This disrupted channel is incised into a late Pleistocene to early Holocene paraglacial debris cone, and is associated with deposition of late Pleistocene to mid-Holocene paraglacial alluvial fans (Chapter 2). This portion of the BRF consists of multiple fault strands (Figure 3.1c). However, at the trench site, most of the slip appears to be concentrated on one fault strand (strand D) that connects a series of en echelon fault steps with a horsetail splay (Figure 3.1c).

The Woodlot Trench was excavated in this blocked channel because of the high potential to preserve deposits that post-date formation of the scarp, either from scarp-derived colluvial wedges or up-slope derived, fine-grained channel deposits that banked up against the scarp and may contain organic material suitable for

radiocarbon dating. This site also had the potential to preserve buried paleochannel deposits that pre-date formation of the scarp that could be used as piercing lines to reconstruct a 3D slip vector. Prior mapping indicates that the stratigraphy at the trench site consists of late Pleistocene till overlain by late Pleistocene to Holocene paraglacial debris cones and channels feeding alluvial fans downslope, overlain or incised by Holocene alluvial fans and fluvial deposits (Chapter 2).

3.3 Methods

3.3.1 Trench design and excavation

The Woodlot Trench was designed with a “Z”-shape footprint, including a fault-perpendicular main leg and two fault-parallel wings (Figure 3.1d), to capture both vertical and right-lateral offset of faulted channel deposits. The length, depth, and geometry of the trench were based on existing topographic surveys of the offset abandoned channel at this site, which records ~ 7 m of displacement (Chapter 2). The main leg of the trench was excavated perpendicular to the fault and was ~ 21 m long. South of the scarp, the fault-perpendicular leg was 1-1.5 m wide and 1.5-2 m deep. North of the scarp, the trench was benched, with a ~ 4 m-wide, 2 m-deep upper level and a ~ 1 m-wide, ~ 1.5 m-deep lower level (Figure 3.2). Each parallel wall of the fault-perpendicular leg was termed Slice 1-4, increasing from east to west (Figure 3.2). The fault-parallel wing south of the scarp was ~ 13 m long, and the fault-parallel wing north of the scarp was ~ 9 m long. Both wings were ~ 1 -1.5 m wide and ~ 1 -2 m deep. The nine individual trench walls were cleaned and gridded with a 1 x 1 x 1 m local grid system (X - fault-parallel; Y - fault-perpendicular; Z - vertical; Figure 3.2), using a line level and measuring tape. Each trench wall was photographed and processed into a 3D model using Agisoft MetaShape (e.g., Reitman et al., 2015) from which we created orthorectified photomosaics. Trench walls were logged at 1:10-scale on printed photomosaics (shown at 1:35-scale in Appendix C). The footprint of the trench and the coordinates of key contacts were surveyed using Nikon XS and Spectra Precision Focus 6 total stations.

3.3.2 Trench logging

In the Woodlot Trench, we mapped faults, depositional and deformational fabrics, deposits and their contacts, modern and buried soils, and roots. The strike and dip of fault planes were measured directly in the trench by connecting correlative fault strands with strings across the open trench (Figure 3.3). We measured the strike and dip of stratification, fissility, and clast orientations where possible (Figure 3.3a) to determine whether depositional fabrics were folded or otherwise disturbed proximal to the fault zone. Clast orientations are recorded as either the strike and dip of ab-planes or the trend and plunge of long axes.

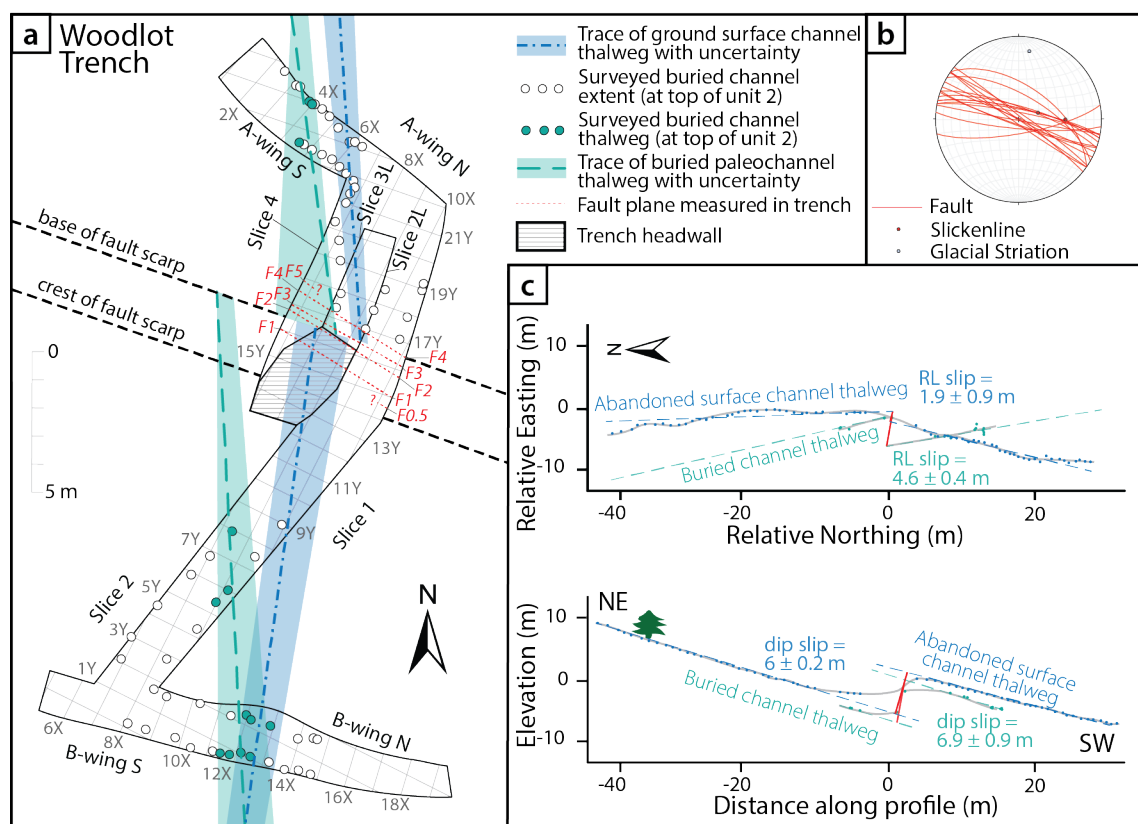


Figure 3.2: Trench footprint, faults, and channels surveyed within the Woodlot Trench. **A.** Diagram of the plan view trench footprint, wall names, and local grid coordinates (X and Y), showing the locations of mapped faults, and offset buried channel and abandoned surface channel thalwegs. **B.** Lower-hemisphere, equal area stereonet showing the orientations of structures documented in the trench (measurements reported in Table 3.2.) **C.** Map view (top) and cross-section (bottom) of the surveyed buried channel and abandoned surface channel thalwegs.

Trench units were mapped based on lithology, texture, stratification, soil development, and color (Table C.1). For each deposit, the proportion of clasts versus matrix, matrix texture and color, the distinctness and geometry of lower contacts, and unit thicknesses and shapes were all described. The average and range of clast sizes, sorting, rounding, and the presence of any soil development, detrital organic matter, and roots were also described. Soil horizons were identified by changes in texture and color and the presence of additional clays. For each soil horizon, the thickness, distinctness, matrix texture and color, and presence of roots were described following Schoeneberger et al. (2012). Both modern and buried soils were described to identify marker horizons on both sides of the fault that can be used to estimate displacement.

A significant component of our logging and unit description was aimed toward the identification of colluvial wedges. Colluvial wedges are wedge-shaped deposits of scarp-derived material deposited on the downthrown block of a fault following the rapid erosion of the fault scarp (e.g., J. P. McCalpin et al., 2009) and can be used as a proxy to estimate per-event vertical offset (e.g., Bennett et al., 2018). A deposit

was identified as a colluvial wedge if it was wedge-shaped, poorly to moderately sorted, had an abrupt lower contact that buried a soil, and had a matrix color distinct from the surrounding units. Additionally, colluvial wedges were identified as scarp-derived if they had a distinct texture (e.g., stratification) sub-parallel to the north-facing scarp face rather than the south-facing background hillslope, and if they pinched out laterally away from the fault zone. For each wedge, in addition to these unit characteristics, the presence of fault terminations and any back-rotation of the wedge were described (Table C.2).

3.3.3 Sample collection and selection for radiocarbon dating

Charcoal and non-carbonized plant material were collected from the deposits exposed in the walls of the Woodlot Trench to constrain the ages of mapped units and to bracket the timing of earthquakes. We used radiocarbon dating due to the abundance of organic materials in the temperate rainforest environment of Vancouver Island and the lack of suitable materials for luminescence dating. The basalt exposed at the top of the Beaufort Range contains minimal quartz and feldspar, and likely experiences insufficient dose rates for luminescence dating (Shannon Mahan, USGS Luminescence Dating Laboratory, pers. Comm., 2017). Two types of charcoal samples were collected in the Woodlot Trench: visible macro charcoal extracted directly from the trench walls, and macro charcoal sieved from bulk sediment samples. Bulk sediment samples were collected from key deposits and then sieved in the lab to concentrate additional dateable macro charcoal fragments. We also collected non-carbonized plant material samples preserved under boulders lying at the ground surface (Figure 3.3). This sampling strategy yielded a total of 101 charcoal samples, 9 non-carbonized plant samples, and 13 bulk samples from the Woodlot Trench.

We chose a subset of these macro charcoal samples for radiocarbon dating, prioritizing those that would best constrain deposit ages and bracket earthquake timing. Our first criterion was to select samples that were well-preserved and that lacked visible signs of contamination from the modern rootlets ($\sim 0.5\text{-}2$ mm) common throughout the Woodlot Trench. In general, we chose samples that were collected from stratified and/or clay-rich sediments with no major roots or burrows (krotovina) disrupting the unit fabric and possibly introducing charcoal that is unrelated to the depositional age of the unit. Our second criterion was to select samples for radiocarbon dating that would either help constrain earthquake timing or constrain the age of offset units. To constrain the minimum age of each event, we selected the best 1-2 samples that were near the stratigraphic base of each colluvial wedge (i.e., in the debris facies, near the base of the wedge, and near the fault zone). To constrain the maximum age of each event, we selected the best 1-3 samples that were directly below the colluvial wedge, either at the top of the underlying scarp-derived colluvial wedge or in a ponded deposit interpreted to be derived from the upslope portion of the channel. To constrain long-term

slip rates, we selected an additional 3-5 samples from the oldest and youngest colluvial wedges and dated 11 samples from a soil developed below a paleochannel deposit (discussed below in Section 3.4.2).

3.3.4 Sample preparation and analysis

Macro charcoal and non-carbonized plant samples were cleaned and prepared in the Paleoecology Lab at Northern Arizona University, in Flagstaff, Arizona (NAU). In the lab, we manually cleaned any residual sediment from macro charcoal samples using brushes, tweezers, and dissecting needles, soaked them in sodium hexametaphosphate for three days, agitated them daily to disaggregate clays from the sample, and then rinsed them in a 250- μm sieve. These steps were repeated until samples were free from adhering sediments. Cleaned samples were then examined under a binocular microscope to select charcoal or plant pieces that would be suitable for radiocarbon dating without visible contamination by modern rootlets. Most samples comprised a single large (~ 5 mm) piece, but some included multiple smaller fragments (Table 3.1). Bulk sediment samples were wet sieved first with an 850- μm sieve and then a 250- μm sieve to isolate charcoal. Any charcoal isolated from the bulk sediment was then cleaned and prepared in the same manner as the macro charcoal samples.

Samples were prepared for accelerated mass spectrometer (AMS) ^{14}C analysis at the UC Irvine Keck Carbon Cycle AMS Laboratory in Irvine, California (UCI) and the Arizona Climate & Ecosystems Isotope Lab at NAU (Table 3.1). In both labs, samples underwent acid/base/acid (ABA) treatment following the protocols of Santos and KCCAMS Prep-Laboratory Personnel (2011a). Samples that were too small for combustion and graphitization following ABA treatment were excluded from further analysis. The 43 chosen samples were then combusted and graphitized following the protocols of Santos and KCCAMS Prep-Laboratory Personnel (2011b); Santos and Personnel (2011). Sample mass is reported in Table 3.1. Using a National Electrostatics Corporation 0.5MV 1.5SDH-2 compact AMS, the ^{12}C , ^{13}C , and ^{14}C of each sample were measured at UCI following the protocols of Santos and KCCAMS Prep-Laboratory Personnel (2009).

AMS analyses yielded the radiocarbon age. Laboratory-reported radiocarbon ages are listed as ^{14}C yr BP (e.g., Stuiver & Polach, 1977), except for modern samples, for which only the fraction modern and D^{14}C is reported (Table 3.1). We calibrated radiocarbon ages using OxCal 4.4, IntCal20, and the post-bomb Northern Hemisphere Zone 1 curve for modern ages (Bronk Ramsey, 2021, 1995; P. J. Reimer et al., 2020; R. W. Reimer & Reimer, 2023; Hua et al., 2013). Calibrated ages are reported as 95% confidence age intervals based on the IntCal20 curve (yrs BP; Table 3.1; P. J. Reimer et al., 2020).

3.3.5 Earthquake age modeling

Radiocarbon ages in paleoseismic trenches are frequently out of stratigraphic order. Organic material can spend considerable time on the forest floor, e.g., between plant growth, charcoal production (i.e., forest fire), and burial in a sedimentary deposit. This can lead to in-built ages of 10s-100s of years for individual samples (e.g., Frueh & Lancaster, 2014; Bird, 2013). Gavin (2001) documented inbuilt ages of up to 670 yrs on Vancouver Island. Thus, the radiocarbon age of the sample would be considerably older than the age of deposition. Samples can also be anomalously old due to reworking of deposits, e.g., a sample in a deposit on the upthrown block of a fault being incorporated into a colluvial wedge (e.g., Bennett et al., 2018). Anomalously young samples can result from a number of processes, including downward transport through plant or animal bioturbation, roots burning in situ, and contamination by fungal mycorrhizae (e.g., Bird, 2013; Grant & Sieh, 1994). The standard practice in paleoseismology for addressing issues of stratigraphic vs. chronologic order is to combine the calibrated age PDFs and stratigraphic constraints into a Bayesian statistical model in OxCal, omitting as few outliers as necessary (e.g., Lienkaemper & Bronk Ramsey, 2009; Bennett et al., 2018; Personius et al., 2014; DuRoss et al., 2018, 2011; Schermer et al., 2021; Klinger et al., 2015; Morell et al., 2018, 2020; J. McCalpin et al., 2023). This may require omitting as many as 30% or even >50% of the samples (e.g. Schermer et al., 2021; Barnett et al., 2015; Bennett et al., 2018).

We modeled the ages of individual earthquakes using OxCal 4.4 (e.g., Lienkaemper & Bronk Ramsey, 2009). The maximum possible age for each model was 13.6 ± 1 ka (maximum regional deglaciation age; Halsted, 1968; Alley & Chatwin, 1979; Clague, 1994; Blaise et al., 1990), and the minimum possible age for each model was 1946.5 CE, when the most recent earthquake occurred on Vancouver Island (the June 23, 1946 Vancouver Island earthquake; Hodgson, 1946). We generated a series of OxCal event age model probability density functions (PDFs), testing different assumptions about the minimum and maximum possible ages of deposits, and different assumptions about which samples were outliers and unlikely to represent the age of deposition. The details of these models are explained in Section 3.4.2.2. All the models generated are geologically plausible. Therefore, we also calculated a weighted average of the earthquake event age PDFs following the methods of DuRoss et al. (2018) using normalized earthquake PDFs. Earthquake event ages from the OxCal models are reported as the 95% confidence interval, rounded to the nearest century in thousands of calendar years before 1950 (ka BP).

3.3.6 Slip rate calculations

Slip rates in the Woodlot Trench were calculated using two methods. The first is an “open-ended” slip rate using the cumulative oblique displacement from all events in the trench and the age of deglaciation.

The second is a “closed-interval” slip rate, which use a known time interval that encompasses one or more complete recurrence periods and the displacement that occurred during that time interval (e.g., Working Group on California Earthquake Probabilities, 1990). We report the open-ended slip rate, closed-interval slip rate for each recurrence period, and a closed-interval slip rate for all events in the Woodlot Trench.

3.4 Results

New paleoseismic trenching of the BRF confirms the existence of an active fault that has produced multiple surface rupturing earthquakes in the late Pleistocene to Holocene. Multiple fault planes offset hillslope and channel deposits at the Woodlot Trench site. We identify five colluvial wedge deposits that we interpret to have been deposited following five surface rupturing earthquakes. Below, we discuss the evidence for tectonic faulting, the composition of the faulted deposits, and the radiocarbon-based deposit ages. Then we use our trench logs and surveys of buried, offset paleochannel deposits to determine the magnitude of displacement per event, and the kinematics of fault slip.

3.4.1 Trench stratigraphy and structure

3.4.1.1 Unit descriptions

Units 1 and 2 - Paraglacial Deposits

Unit 1 is the stratigraphically lowest unit in the Woodlot Trench, and is an indurated, >4 m-thick, laterally continuous unit exposed in the deepest exposures of the trench. Unit 1 is a poorly sorted, matrix-supported deposit consisting of weakly stratified, evenly distributed sub-rounded to rounded basalt and exotic granodiorite pebbles to cobbles with rare boulders in a loamy sand matrix (Table C.1). We interpret unit 1 to be a colluvial deposit formed in a paraglacial environment during and immediately after glacial retreat (unit Qp1 on Figure 3.1b; Chapter 2). There is a decrease in induration at the top of unit 1, which we interpret as a weathering profile defining a gradational contact between units 1 and 2. Unit 1 is offset by the BRF and is observed both north and south of the scarp. It forms a slight synform south of the scarp (~9-15Y; Figure C.1, C.4 Slice 1 and 2 logs) that we interpret to be a result of folding caused by slip in the fault zone.

Unit 2 overlies unit 1 throughout the trench and is ~85 cm thick. Unit 2 is nearly tabular, with a planar lower contact and an upper boundary that forms a broad channel shape. Unit 2 is a poorly sorted, clast-supported deposit consisting of evenly distributed sub-angular to sub-rounded basalt pebbles and cobbles in a loamy sand matrix (Table C.1). We interpret unit 2 as a hillslope colluvium developed through weathering and downhill transport of material sourced from unit 1. The top ~10-30 cm of unit 2 are marked by a

horizon of strong red-orange coloration and higher proportion of clays that we interpret to be a paleo Bt soil horizon developed on unit 2. This red-orange Bt soil horizon is gently inclined, sub-parallel to both unit 2 and the modern ground surface and generally follows the broad channel shape at the top of unit 2 (Figure 3.3, Figures C.1-C.8). Unit 2 and related Bt soil horizon are offset by the BRF.

Units 3, 3a, 5a, 5b, 7, 9 - Channel Deposits

Unit 3 is a discontinuous and lens-shaped deposit, up to 100 cm thick, that buries unit 2 at the margins of the buried channel both north and south of the scarp (Figures C.1-C.8). Unit 3 is a moderately sorted clast-supported deposit with normal grading. The unit contains sub-angular to rounded basalt and minor exotic granodiorite clasts within pebble and cobble interbeds with little to no matrix (Table C.1). We interpret unit 3 to be a fluvial deposit formed in the paleochannel feeding a relatively younger paraglacial alluvial fan mapped downslope of the trench (mapped as Qp2 in Figure 3.1b; Chapter 2). Unit 3 is displaced by the BRF, but due to its discontinuous nature, unit 3 is not in direct fault contact with the BRF in the Woodlot Trench.

Unit 3a is a lens-shaped deposit exposed only on the A-wing north wall, where it is up to 56 cm thick and fills a ~1.5 m-wide paleochannel that is locally incised into unit 2, adjacent to the buried channel thalweg marked by the strong red-orange horizon developed on unit 2 (e.g., 8X, Figure C.5). Unit 3a is a moderately sorted matrix-supported deposit with distinct reverse grading. The clasts are subangular basalt and minor exotic granodiorite pebbles and cobbles in a moderately sorted sandy loam matrix (Table C.1). We interpret unit 3 to be a muddy debris flow facies of the same channel deposits as unit 3, both of which fed the paraglacial alluvial fan deposits mapped downslope of the trench (unit Qp2 on Figure 3.1b; Chapter 2).

Unit 5a is a lens-shaped deposit, up to 66 cm thick, located north of the scarp, where it appears buttressed against a local, uphill-facing paleotopographic high formed in unit 2 (Figure 2.9b). Unit 5a is a moderately sorted clast-supported deposit with normal grading. The clasts are primarily sub-angular to sub-rounded basalt pebbles interbedded with lithic sands, whereas minor matrix exists as clay at the base of the deposit (Table C.1). We interpret unit 5a to be a ponded channel deposit that was deposited after slip on the BRF formed an uphill-facing scarp that blocked downstream channel flow.

Unit 5b is a 10-43 cm-thick lens-shaped deposit overlying unit 5a north of the scarp. Unit 5b is a moderately sorted matrix-supported deposit with sub-angular to sub-rounded basalt pebbles and few rounded cobbles and boulders in a silty clay loam matrix (Table C.1). The clasts in unit 5b are coarser toward the southwest (near the fault zone, missing clasts on Figure C.4). We interpret unit 5b to be a time-transgressive channel deposit that was deposited against an uphill-facing fault scarp that blocked channel flow, causing ponding north of the scarp, similar to unit 5a slightly upstream. Unit 5b is truncated by fault F5 in Slice 4

(Figure 3.3b), and appears to be drag-folded into a syncline, as shown by the strong curvature of the lower contact in Slices 2L and 3L (Figures 3.3b, C.2, and C.3). The top ~5 cm of unit 5b at the corner between Slice 4 and the A-wing south wall is a very dark brown to black organic-rich horizon with abundant charcoal. We interpret this to be a charcoal-rich soil A horizon associated with a forest fire.

Unit 7 is an 8-17 cm-thick, ~1 m-wide lens-shaped deposit exposed north of the scarp (Slice 1; Figure 3.3a). Unit 7 is a poorly sorted matrix-supported deposit with sub-angular to sub-rounded basalt pebbles in a loamy matrix (Table C.1). We correlate unit 7 with a later stage of deposition of the ponded channel deposit unit 5b; unit 5b and unit 7 are separated stratigraphically by unit 6, a colluvial wedge unit described below. Unit 7 is truncated by fault F4 on Slice 1.

The stratigraphically highest channel deposit, unit 9, is a lens-shaped deposit that is up to 10 cm thick and interfingers with unit 5b north of the scarp (Slice 4; Figure 3.3b). Unit 9 is a well sorted matrix-supported deposit with sub-angular to sub-rounded basalt pebbles in a loamy matrix (Table C.1). We correlate with the uppermost and youngest portion of ponded channel deposit unit 5b. Unit 9 is not in contact with the fault zone, but rather pinches out toward the fault zone. The thickness and average grain size of ponded sediment deposits decrease up section. Such a pattern may reflect a decrease in stream discharge and sediment flux over the period of deposition of units 3 to 9.

Units 4, 6, 8, 10, 11, and 5c - Scarp-derived Colluvium

Unit 4 is a wedge-shaped deposit, up to 40 cm thick, located north of the scarp, and is exposed in Slices 1, 2L, 3L, and 4 (Figures 3.3, C.1-C.4). Unit 4 is a poorly sorted, clast-supported deposit containing angular to sub-rounded basalt and minor exotic granodiorite pebbles to small boulders within a sandy loam matrix (Table C.1). The clasts in unit 4 are coarser toward the fault and form a line at the base deposit in the distal portion of the wedge. Unit 4 thins and pinches out ~1.8 m away from the BRF (Figures 3.3b and C.1-C.4). We interpret unit 4 to be a scarp-derived colluvial wedge, truncated by faults F4 and F5.

Unit 6 is a wedge-shaped deposit, up to 60 cm thick, located north of the scarp, and is exposed in Slices 1 and 4, where it buries the lower portions of unit 5b and interfingers with channel deposits unit 5b, 5d, and unit 9 (Figures 3.3, C.1, C.4). Unit 6 is a poorly sorted, clast-supported deposit of angular to rounded pebbles and sub-angular to sub-angular basalt cobbles in a sandy clay loam matrix (Table C.1). Unit 6 has cobbles along its base and at its toe. Unit 6 thins and pinches out ~2.5 m away from the BRF (Figures 3.3b and C.1 and C.4). We interpret unit 6 to be a scarp-derived colluvial wedge, truncated by faults F4 and F5.

Unit 8 is a wedge-shaped deposit, up to 57 cm thick, located north of the scarp, where it buries unit 7 and both overlies and interfingers with hillslope colluvium deposit unit 5d (Slice 2L; Figures 3.3a, C.1). Unit 8 also overlies unit 6 (Slice 4; Figures 3.3b, C.4). Unit 8 is a poorly sorted clast-supported deposit with angular to rounded pebbles and sub-angular to sub-rounded basalt and minor exotic granodiorite cobbles

and small boulders in a loamy sand matrix (Table C.1). Unit 8 thins and pinches out ~ 2.0 m away from the BRF (Figures 3.3b and C.1, C.4). We interpret unit 8 to be a scarp-derived colluvial wedge, truncated by faults F4 and F5.

Unit 10 is a wedge-shaped deposit, up to 57 cm thick, located north of the scarp, where it buries units 8 and 9 (Slices 1 and 4; Figures 3.3, C.1, C.4). Unit 10 also interfingers with channel deposits units 5b and 5d, and overlies sheared sediment and fissure fill within the fault zone. Unit 10 is a poorly sorted clast-supported deposit with angular to sub-rounded pebbles and sub-rounded basalt cobbles in a sandy matrix (Table C.1). Unit 10 thins and pinches out (or interfingers with channel deposits) ~ 1.9 m away from the BRF (Figures 3.3, C.1, C.4). We interpret unit 10 to be a scarp-derived colluvial wedge. Unit 10 buries fault F5, but is truncated by faults F3 and F4.

The youngest and stratigraphically highest unit observed in the Woodlot Trench is unit 11, a lens-shaped deposit, up to 50 cm thick, draped across the scarp face (Figures 3.3, C.1, C.4). There are two facies in unit 11. First, the debris facies is poorly sorted and clast-supported, and consists of sub-angular to sub-rounded pebbles and sub-rounded to rounded cobbles and boulders in a loamy sand matrix (Table C.1). The clasts in the debris facies consist of both basalt and clasts of till, and are primarily located in one clump above fault F1. The second facies is the sheet wash facies, which is moderately sorted and clast-supported, with angular to rounded basalt pebbles and cobbles in a sandy loam matrix (Table C.1). The sheet wash facies of unit 11 includes slope-parallel stratification of pebbles. Unit 11 thins and pinches out in both directions, up to ~ 2.0 m north and south of the BRF (Figures 3.3, C.1, C.4). We interpret unit 11 to be a scarp-derived colluvial wedge. Unit 11 is the youngest generation of scarp-derived colluvium, is the only colluvial wedge deposit observed on both sides of the BRF, and buries the entire fault zone.

Unit 5c is a deposit, up to 35 cm thick, with both lens-shaped and tabular components that buries unit 5b. Unit 5c is best exposed in the A wing and at the corner between Slice 4 and the A wing (Figures 3.3, C.4-C.6). Unit 5c is a moderately to well-sorted clast-supported deposit with angular to sub-rounded sand and pebbles, and angular to sub-angular basalt cobbles in a silty matrix (Table C.1). The cobbles are primarily located at the base and lateral terminations of the deposit. Unit 5c is mapped on the north-facing slope of the scarp face, uphill of the modern surface channel thalweg ($\sim 2-4X$ on A-wing S, Figure 3.2). We therefore interpret it as scarp-derived colluvium that may correlate to unit 10 or 11. However, unit 5c may be hillslope derived, as it is discontinuous with the colluvial wedge exposures on Slices 1 and 4 (Figure 3.3).

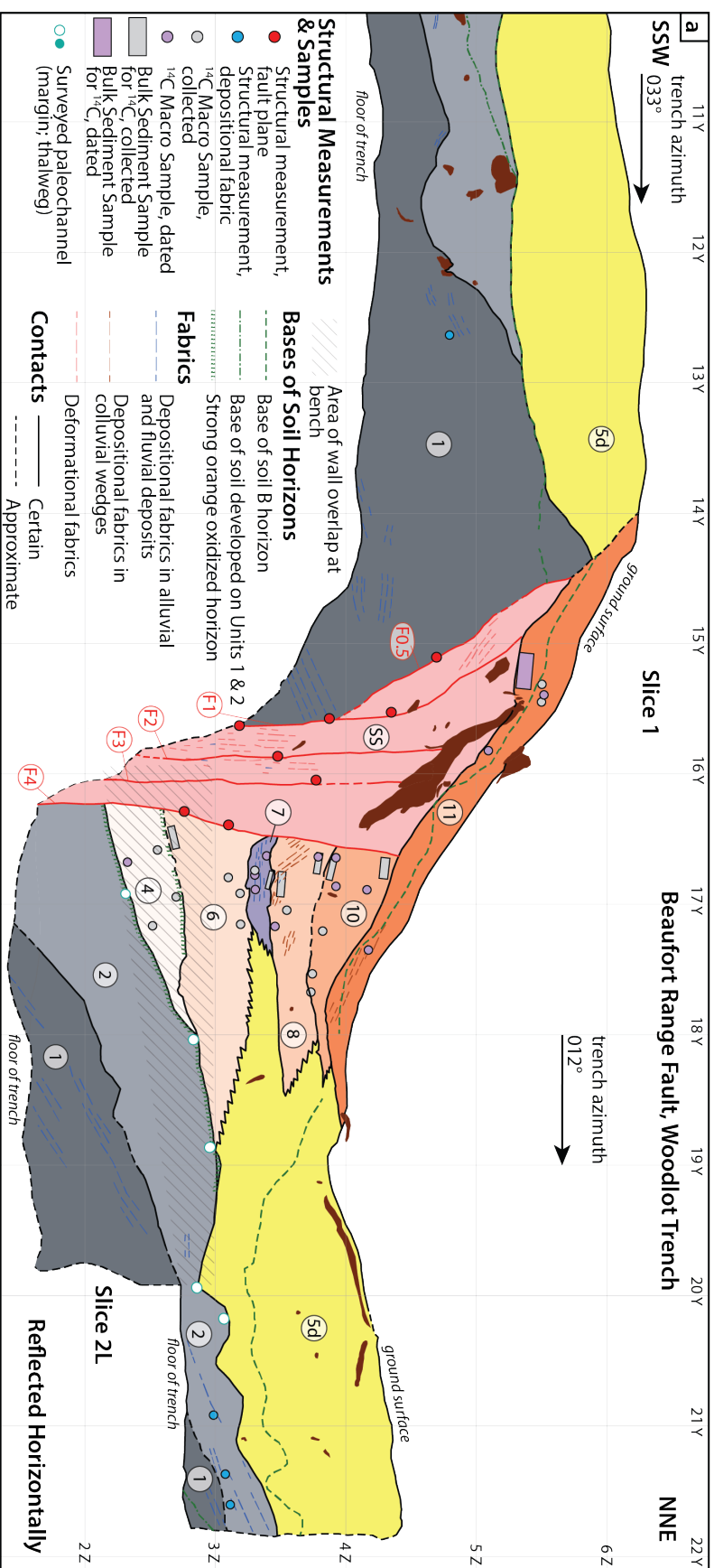
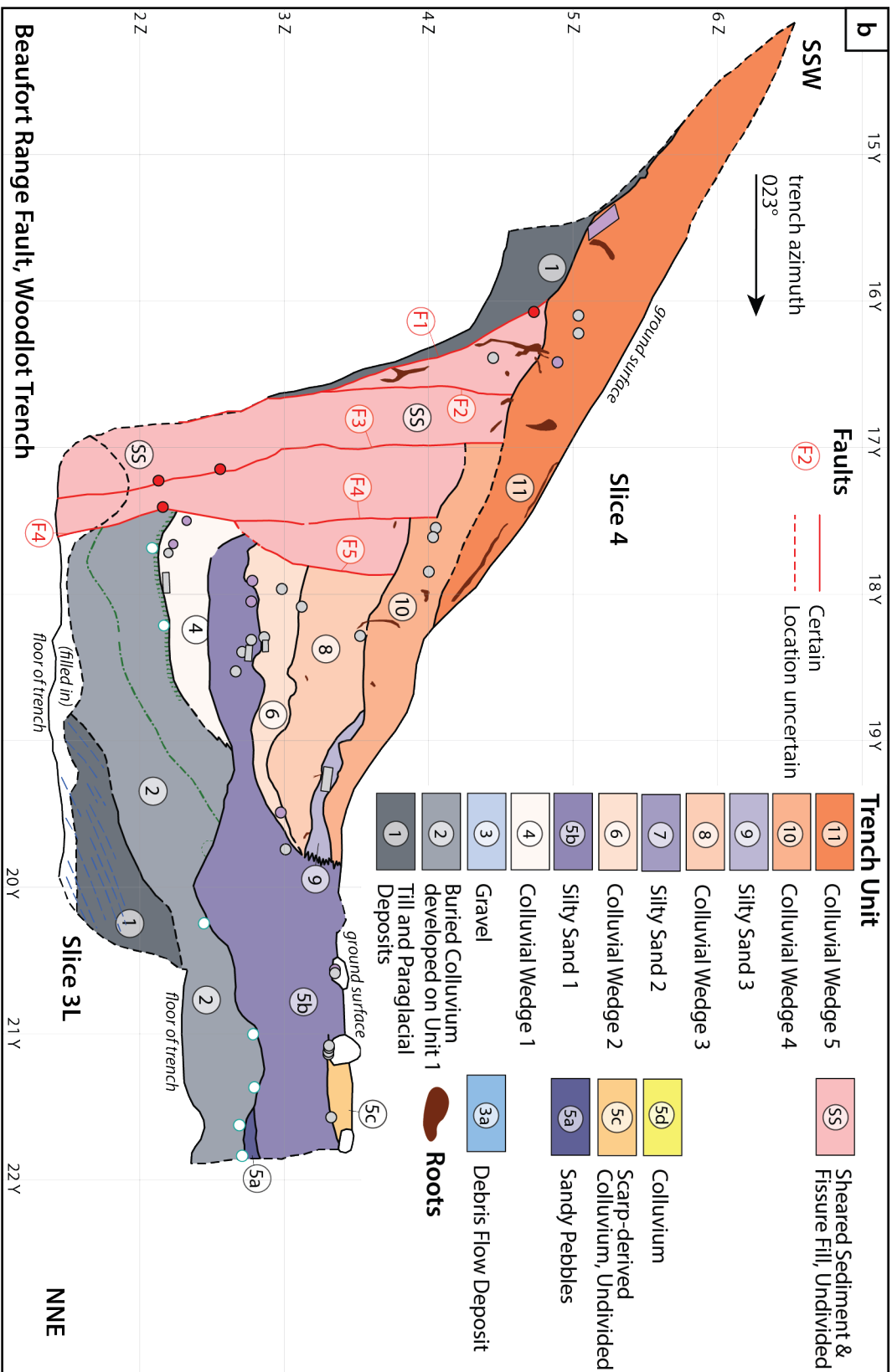


Figure 3.3: Simplified composite trench logs of fault zones in the Woodlot Trench. See Appendix C for complete trench logs of all nine walls. **A.** Simplified trench log of Slices 1 and 2L (see Figure 3.2 for locations), reflected horizontally. **B.** Simplified trench log of Slices 3L and 4 (see Figure 3.2 for locations). Unit numbers refer to stratigraphic correlations, whereas unit colors refer to lithologic correlations. The locations of all collected radiocarbon samples are shown. See Section 3.3.3 for details on which samples were chosen for analysis.



Unit 5d – Hillslope colluvial deposit

Unit 5d is an unconsolidated 25-113 cm-thick tabular deposit that overlies units 1, 2, 3, 3a, and 5b in the channel flanks on both sides of the scarp (Figures 3.3, C.1-C.8). Unit 5d is poorly sorted and matrix-supported, with angular to sub-rounded pebbles to boulders in a loamy sand matrix (Table C.1). The clasts in unit 5d exhibit chaotic distribution and orientations. We interpret unit 5d to be a time-transgressive hillslope colluvium, whose base grades laterally into unit 6 and whose top grades laterally into unit 10. Unit 5d interfingers with colluvial wedge units 6, 8, and 10, and channel deposit unit 7. It is unclear whether unit 5d developed continuously across the fault and was then displaced, or if it developed discontinuously on both sides of the fault. However, its time-transgressive nature, presence on the channel flank (Slice 1), and absence in the modern channel thalweg (Slice 3L) all suggest that both are likely true: unit 5d formed by similar hillslope processes occurring on both sides of the scarp and its time-transgressive nature requires that it also offset by the BRF. The interfingering colluvial wedges and ponding deposits indicate that the Quaternary alluvial and fluvial deposition was punctuated by periodic earthquakes creating and subsequently growing a fault scarp blocking the channel.

3.4.1.2 Faults

The Woodlot Trench exposes a WNW-striking, sub-vertical, >1 m-wide fault zone consisting of six distinct fault strands bounding a zone of sheared sediment (Figure 3.3). The main shear zone is bounded by fault strands F1 and F4. Fault strands F2 and F3 are discrete planes within the exposed zone of sheared sediment, and fault strands F0.5 and F5 splay off of strands F1 and F4, respectively. The fault strands in the Woodlot Trench strike between 260° and 309°, subparallel to the topographic fault scarp mapped at the surface (strike of 289°). The fault zone widens from ~0.5 m wide at the floor of the trench to ~2 m wide near the ground surface, with 57°-77° dips near the surface, and 66°-90° dips at the bottom of the trench. The long axes of cobbles within the sheared sediment are aligned parallel to the faults (e.g., Slice 1, Figure C.1). There were two slickenlines within the sheared sediment that were consistent with slip along measured fault planes (Figure 3.2b). There was also a striation on a clast plucked from the trench headwall in unit 1. This striation had a trend and plunge of 009°/18°, inconsistent with slip on the sub-vertical F1 that forms the trench headwall. We interpret this striation to be a result of glacial processes, rather than fault slip.

3.4.1.3 Evidence for five surface-rupturing earthquakes

Two lines of evidence support a history of multiple surface-rupturing earthquakes in the Woodlot Trench. First, the development of a colluvial wedge reflects an instantaneous rupture and subsequent stability, during which the crest of the fault scarp on the upthrown block degrades and eroded material is deposited on the

downthrown block, and in some cases develops a soil (J. P. McCalpin et al., 2009). Five such scarp-derived colluvial wedges were identified in the Woodlot Trench: units 4, 6, 8, 10, and 11, indicating that there have been five surface-rupturing earthquakes. These units are wedge-shaped deposits that formed against the scarp face and pinch out over short distances ($\sim 1\text{--}2.5$ m) from the fault. For all but the youngest colluvial wedges, the wedge-shape geometry has been accentuated by fault truncation during later ruptures. These colluvial wedges have distinct lower contacts, stratification parallel to the local north-facing scarp due to clasts and matrix being transported down the scarp face from the upthrown block to the downthrown block, texture and color distinct from neighboring units, and any organic material is dispersed throughout the deposit from being transported downslope in a chaotic jumble, as opposed to forming in a distinct soil horizon. All of the colluvial wedges, with the exception of the wedge resulting from the most recent event (MRE; unit 11), also have post-depositional deformation in the form of back-rotation and local folding, where the wedge is tilted in subsequent events. Details for each wedge and each trench wall (Slice) are in Table C.2. In the case of the Woodlot Trench, channel flow was blocked by the formation of the scarp, resulting in channel deposits being ponded on the downthrown block and interfingering with the scarp-derived colluvial wedge deposits. These ponded deposits are a further indication of surface stability. The opposing direction of unit thickening (colluvial wedges thickening to the south, towards the fault, and channel deposits thinning to the south, as they onlap the north-facing colluvial wedge paleo ground surface) provide very clear contrast between the interfingering channel deposits and colluvial wedges.

Second, upward fault terminations are used to constrain which strands slipped in which events (e.g., Sieh, 1978; Nelson, 2000). Faults F1, F2, and F3 (and F4 in Slice 1) terminate at the base of unit 11 (MRE wedge), so they intersected the ground surface immediately prior to deposition of the MRE wedge. Therefore, all of these strands slipped in the MRE. Fault F4 in Slice 4 terminates within unit 10 (penultimate event, or PE wedge), which also indicates that it slipped during the MRE. On the other hand, fault F5 (Figure C.4) terminates at the base of unit 10, indicating that this strand slipped in the PE, but not in the MRE.

3.4.2 Geochronology

A total of 43 samples from a collection of 123 were dated to constrain the ages of 11 of the 15 mapped trench deposits and to bracket earthquake timing (see Section 3.3.3 for sample selection criteria). Ten of these units were dated directly with radiocarbon samples (units 2, 3, 4, 5b, 5d, 6, 7, 8, 10, and 11). The age of unit 1 is constrained through stratigraphic relationships and constraints on deglaciation as it was barren of organic matter. Three of the mapped trench deposits lack direct age constraints from radiocarbon samples (units 3a, 5a, and 9). Calibrated radiocarbon ages from the Woodlot Trench ranged from 13.4 cal

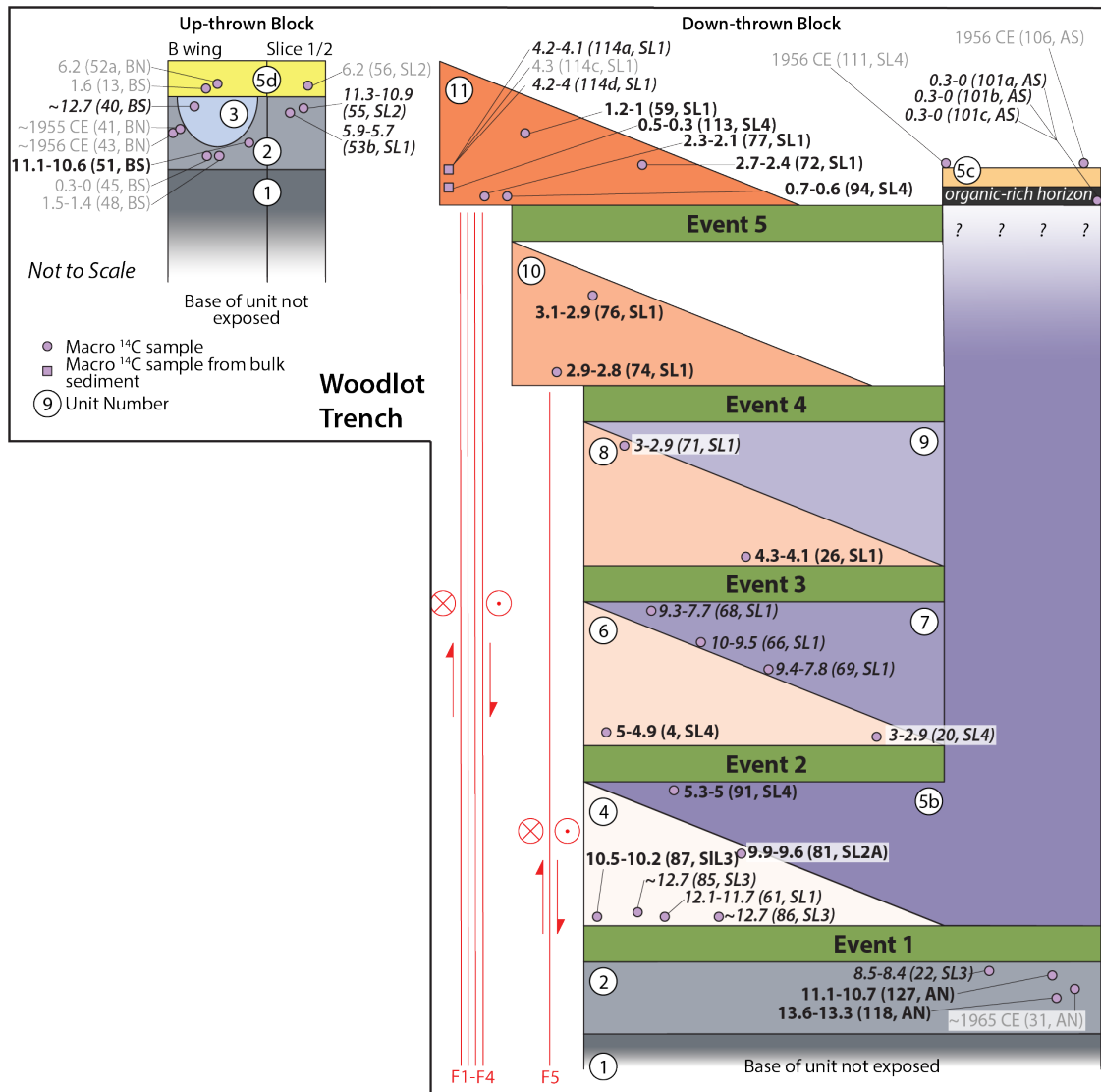


Figure 3.4: Schematic diagram showing stratigraphic relationships between trench units and locations of radiocarbon dated detrital charcoal and non-carbonized plant samples. Unit colors follow the legend in Figure 3.3. Sample numbers and the wall (SL - Slice) from which the sample was collected are shown in parentheses (see Figure 3.2 for wall locations). Radiocarbon samples are shown as either purple circles (macro ¹⁴C samples) or purple squares (macro ¹⁴C from bulk sediment samples). Ages in bold are included in our preferred OxCal model, shown in Figure 3.5a. Italicized ages are included in some models, but not our preferred model (model A.1). Ages in gray are excluded from all models.

ka BP to modern.

3.4.2.1 Unit ages

Units 1 and 2 - Paraglacial deposits

Unit 1 did not contain dateable charcoal, but its age can be estimated based on understandings of

paraglacial slope adjustment elsewhere. Regional ages constrain the timing of deglaciation on central Vancouver Island to ~ 13.6 -11 ka (Mosher & Hewitt, 2004; Easterbrook, 1992; Clague & James, 2002). Paraglacial debris cone (colluvial) deposits, like unit 1, would have begun to form in the Alberni Valley in the first 10s to 100s of years following the retreat of the Cordilleran Ice Sheet and subsequent valley glacier (e.g., Ballantyne, 2002; Ballantyne & Benn, 1994, 1996). If deglaciation continued through 11 ka, paraglacial debris cones formation may have continued until ~ 10.5 ka. Therefore, we interpret the age of unit 1 to be ~ 13.6 -10.5 ka.

11 charcoal samples were dated from unit 2 that yielded calibrated ages ranging from 13.5 cal ka BP to 1965 CE (Figure 3.4, Table C.3). Five of these samples, BR19-31C, 41C, 43C, 45C, and 48C, are likely outliers, as they yielded calibrated ages ranging from ~ 1.5 cal ka BP to modern, significantly younger than samples in all stratigraphically higher deposits. The remaining 6 samples from unit 2 yielded ages from ~ 13.5 to ~ 5.8 cal ka BP (Figure 3.4, Table C.3). We interpret the maximum age of unit 2 to be ~ 13.5 cal ka BP, which is consistent with previous estimates of deglaciation age (see above). Which samples we interpret to be the minimum age of unit 2 in different earthquake age models (ranging from ~ 10.7 -5.8 cal ka BP) are discussed below in Section 3.4.2.2.

Units 3, 5b, 7 - Channel deposits

One charcoal sample from channel deposit unit 3, BR19-40C, yielded an age of 12.5 cal ka BP. Sample BR19-40C is likely recycled, as it is older than ten of the eleven samples dated from unit 2. The paleochannel marked by the red-orange horizon developed on unit 2 is likely older than ~ 10.7 cal ka BP, the oldest minimum age of unit 2. Because unit 3 was deposited in this paleochannel (Section 3.4.1.1), we interpret unit 3 to be younger than ~ 10.7 cal ka BP.

One sample from channel deposit unit 5b, BR19-91C, yielded a calibrated age of ~ 5.2 cal ka BP (Figure 3.4, Table C.3). We interpret this sample to be representative of the age of unit 5b because it is in stratigraphic order and was collected from undisturbed clays directly below colluvial wedge unit 6. Three samples from the organic-rich horizon at the top of unit 5b, BR19-101C-a, 101C-b, and 101C-c, all located ~ 10 cm below the modern ground surface (Figure 3.3b), yielded ages of ~ 100 -120 cal yr BP. The Woodlot Trench site was clear-cut and then burned in the early 1900s CE (Todd Thompson, pers. comm., 2018), within the 95% confidence intervals of these samples. We interpret this organic-rich horizon to be a buried soil A horizon associated with a forest fire (Section 3.4.1.1). These age results thus support a time-transgressive interpretation for unit 5b, which is likely as old as ~ 5.2 ka, was aggrading for 100s to 1000s of years coeval with many adjacent and interfingering scarp-derived colluvial wedge deposits and was experiencing near-surface soil development into the early 1900s CE.

Three samples from channel deposit unit 7, BR19-66C, 68C, and 69C, yielded ages of ~ 11 -8.4 cal ka BP (Figure 3.4, Table C.3). These samples are all older than the stratigraphically lower sample BR19-91C in

unit 5b, which was collected from undisturbed clays. Samples BR19-66C, 68C, and 69C are therefore likely outliers. We interpret the age of unit 7 to be ~ 4.9 - 4.1 cal ka BP, as constrained by the ages of units 6 and 8 (see below).

Units 4, 6, 8, 10, 11, and 5c - Scarp-derived colluvium

Five samples from colluvial wedge unit 4 (Event 1 wedge) yielded calibrated ages ranging from ~ 12.7 to 9.6 cal ka BP (Figure 3.4, Table C.3). Three of these samples, BR19-61C, 85C, and 86C, yielded ages of ~ 12.8 - 11.9 cal ka BP. These samples are older than half of the samples in the stratigraphically lower unit 2 and are likely outliers. These samples are excluded from all but one of our age models (Section 3.4.2.2). We interpret the remaining samples, BR19-81C and 87C, to be representative of the deposit age. These samples yielded calibrated ages of 10.5 - 9.6 cal ka BP.

Two samples from the base and thin distal portion of colluvial wedge unit 6 (Event 2 wedge), BR19-04C and 20C, yielded calibrated ages of 4.9 - 2.8 cal ka BP (Figure 3.4, Table C.3). BR19-20C yielded a calibrated age of ~ 2.8 cal ka BP, significantly younger than ages from stratigraphically higher units, indicating that it is unlikely to be representative of the deposit age. We interpret the remaining sample, BR19-04C, to be representative of the deposit age. This sample yielded a calibrated age of ~ 4.9 cal ka BP.

Two samples from colluvial wedge unit 8 (Event 3 wedge) were dated, one from the base of the wedge, and one from the top of the wedge. These samples, BR19-26C and 71C, yielded calibrated ages of 4.3 - 2.9 cal ka BP (Figure 3.4, Table C.3). These samples were in stratigraphic order, and we interpret them to be representative of the deposit age.

Two samples from colluvial wedge unit 10 (PE/Event 4 wedge) were dated, one from the base of the wedge, and one from the top of the wedge. These samples, BR19-74C and 76C, yielded calibrated ages of ~ 3.1 - 2.8 cal ka BP (Figure 3.4, Table C.3). These samples were in stratigraphic order, and we interpret them to be representative of the deposit age.

Eight samples from colluvial wedge unit 11 (MRE/Event 5 wedge) yielded calibrated ages of ~ 4.3 - 0.5 cal ka BP (Figure 3.4, Table C.3). Three of these samples appear to be out of stratigraphic order (BR19-114C-a, 114C-c, and 114C-d, which yielded calibrated ages of ~ 4.3 - 4.1 cal ka BP) and are likely outliers. We interpret the remaining five samples, BR19-59C, 72C, 77C, 94C, and 113C, to be representative of the deposit age. These samples yielded calibrated ages of ~ 2.5 - 0.5 cal ka BP.

We dated two samples from unit 5c (scarp-derived colluvium, undifferentiated), BR19-106C and 111C, that yielded calibrated ages of 1956 CE. These samples were non-carbonized plant material collected from beneath boulders on the ground surface (Figures 3.3b, 3.4, Table C.3). We therefore interpret these samples to constrain the minimum age of unit 5c. Unit 5c buries unit 5b, and therefore must be younger than unit 5b (~ 5.2 cal ka BP). We interpret the age of unit 5c to be ~ 5.2 cal ka BP to modern. Though unit 5c is

discontinuous with the scarp-derived colluvial wedges, its base on Slice 4 projects toward the base of colluvial wedge 10. Unit 5c likely correlates with scarp-derived colluvial wedge units 10 or 11, $<\sim 3.1$ ka B.

Unit 5d - Other colluvial units

We dated three samples from unit 5d (hillslope colluvium), BR19-13C, 52C-a, and 56C, that yielded calibrated ages of ~ 6.2 - 1.6 cal ka BP (Figure 3.4, Table C.3). This large age range is consistent with the interpretation that unit 5d represents continued hillslope transport over several thousands of years. We therefore do not use the radiocarbon ages from unit 5d to constrain earthquake timing.

3.4.2.2 Modeled earthquake ages

We modeled earthquake ages in OxCal using five different combinations of radiocarbon sample ages. Notably, some of these models required disregarding our previous assessments of anomalously old or young samples to fit other assumptions, yielding age models that are less likely. This is reflected in the weighting of our mean model, discussed below. The models use radiocarbon ages from directly above the lower contacts of colluvial wedges to constrain minimum earthquake ages and radiocarbon ages from directly below the lower contacts of colluvial wedges (e.g., from the top of the unit buried by the colluvial wedge) to constrain the maximum earthquake ages (Figure 3.4). The maximum allowed age for each model was 13.6 ± 1 ka (maximum regional deglaciation age; Halsted, 1968; Alley and Chatwin, 1979; Ryder and Clague, 1989; Blaise et al., 1990; Clague, 1994), and the minimum allowed age for each model was 1946.5 CE, when the most recent earthquake occurred on Vancouver Island (the June 1946 Vancouver Island earthquake). Based on these initial constraints, we exclude five modern sample ages from all models.

The first model iteration, model A, included as many sample ages as possible while maintaining the stratigraphic order of the units (e.g., Lienkaemper & Bronk Ramsey, 2009). model A included 17 samples and excluded 21 samples (Table C.3). model A yielded ages of 10.9-10.3 ka for Event 1, 5.2-4.9 ka for Event 2, 4.9-4.2 ka for Event 3, 3.0-2.9 ka for Event 4, and 2.8-2.4 ka for Event 5 (Table C.3).

Table 3.1: Calibrated and uncalibrated radiocarbon ages. Sample locations are in Figures C.1-C.8. Sample numbers in italics underwent acid-base-acid (ABA) treatment, combustion, and graphitization at UC Irvine; all others underwent ABA treatment, combustion, and graphitization at Northern Arizona University before undergoing AMS ^{14}C dating at UC Irvine.

Field sample # (BR19-)	Lab ID # (AMS and NAU) ¹	IGSN # ²	Wall ³	Unit ⁴	Fraction modern $\pm 1\sigma$	$\text{D}^{14}\text{C} \pm 1\sigma$ (‰)	^{14}C age $\pm 1\sigma$ (yrs BP)	Calibrated age range ⁵ (cal yrs BP ⁶)	Median calibrated age (cal yrs BP ⁶)	Sample type ⁷
04C	239896, 2256	0003	4	6	0.5822 ± 0.0013	-422.7 ± 1.3	4345 ± 20	4964 – 4853	4905	S, charcoal (0.39 mg C)
13C	239897, 2257	000C	BS	11	0.8121 ± 0.0017	-194.7 ± 1.7	1670 ± 20	1688 – 1525	1556	S, charcoal (0.37 mg C)
<i>20C</i>	230832, –	000J	4	6	0.7114 ± 0.0011	-288.6 ± 1.1	2735 ± 15	2863 – 2776	2814	S, charcoal (~ 0.7 mg C)
<i>22C</i>	230835, –	000L	3L	2	0.3857 ± 0.0008	-614.3 ± 0.8	7655 ± 20	8519 – 8387	8427	S, charcoal (~ 0.7 mg C)
<i>26C</i>	230830, –	000P	1	8	0.6227 ± 0.0010	-377.3 ± 1.0	3805 ± 15	4243 – 4098	4192	S, charcoal (~ 0.7 mg C)
31C	239898, 2258	000U	AN	2	1.1885 ± 0.0020	178.5 ± 2.0	Modern	1958 – 1988 CE	1965 CE	S, burned wood (0.27 mg C)
40C	239899, 2259	0013	BS	3	0.2706 ± 0.0013	-731.7 ± 1.3	10500 ± 40	12672 – 12198	12547	S, charcoal (0.41 mg C)
41C	239900, 2260	0014	BN	2	1.0106 ± 0.0075	2.0 ± 7.5	Modern	1694 – 1957 CE	1955 CE	M, charcoal (0.042 mg C)
43C	239901, 2261	0016	BN	2	1.0246 ± 0.0069	16.0 ± 6.9	Modern	1955 – 1957 CE	1956 CE	M, charcoal (0.050 mg C)
45C	229902, 2262	0018	BS	2	0.9799 ± 0.0018	-28.3 ± 1.8	165 ± 15	285 – 0	187	S, unburned wood (0.19 mg C)
48C	239903, 2263	001B	BS	2	0.8205 ± 0.0021	-186.4 ± 2.1	1590 ± 25	1530 – 1405	1467	M, charcoal (0.11 mg C)

(continued).

Field sample # (BR19-)	Lab ID # (AMS and NAU) ¹	IGSN # ²	Wall ³	Unit ⁴	Fraction modern $\pm 2\sigma$	D ¹⁴ C $\pm 2\sigma$ (‰)	¹⁴ C age $\pm 1\sigma$ (yrs BP)	Calibrated age range ⁵ (cal yrs BP ⁶)	Median calibrated age (cal yrs BP ⁶)	Sample type ⁷
51C	239904, 2264	001E	BS	2	0.3066 $\pm$ 0.0014	-695.9 $\pm$ 1.4	9495 $\pm$ 40	11070 – 10578	10758	S, charcoal (0.46 mg C)
52C-a	239905, 2265	001F	BN	5d	0.5075 $\pm$ 0.0015	-496.8 $\pm$ 1.5	5450 $\pm$ 25	6300 – 6200	6237	S, charcoal (0.20 mg C)
53C-b	239906, 2266	001G	1	2	0.5325 $\pm$ 0.0014	-472.0 $\pm$ 1.4	5060 $\pm$ 25	5901 – 5741	5820	S, charcoal (0.46 mg C)
55C	239907, 2267	001I	2	2	0.2975 $\pm$ 0.0014	-705.0 $\pm$ 1.4	9740 $\pm$ 40	11245 – 10897	11188	S, charcoal (0.64 mg C)
56C	239908, 2268	001J	2	5d	0.5087 $\pm$ 0.0013	-495.6 $\pm$ 1.3	5430 $\pm$ 20	6290 – 6199	6234	S, charcoal (0.55 mg C)
59C	239910, 2269	001M	1	11	0.8670 $\pm$ 0.0016	-140.3 $\pm$ 1.6	1145 $\pm$ 15	1177 – 973	1021	S, charcoal (0.63 mg C)
61C	239911, 2270	001O	1	4	0.2802 $\pm$ 0.0014	-722.2 $\pm$ 1.4	10220 $\pm$ 45	12417 – 11739	11893	M, charcoal (0.33 mg C)
66C	239912, 2271	001T	1	7	0.3013 $\pm$ 0.0107	-701.3 $\pm$ 10.7	9640 $\pm$ 290	11969 – 10226	11011	S, charcoal (0.034 mg C)
68C	230831, –	001V	1	7	0.3910 $\pm$ 0.0182	-609.0 $\pm$ 18.2	7540 $\pm$ 380	9400 – 7673	8412	S, charcoal (0.013 mg C)
69C	239913, 2272	001W	1	7	0.3793 $\pm$ 0.0014	-623.9 $\pm$ 1.4	7790 $\pm$ 30	8636 – 8458	8565	M, charcoal (0.78 mg C)
71C	230829, –	001Y	1	9	0.7041 $\pm$ 0.0011	-295.9 $\pm$ 1.1	2820 $\pm$ 15	2964 – 2865	2918	S, charcoal (~0.7 mg C)
72C	239914, 2273	001Z	1	11	0.7382 $\pm$ 0.0014	-268.1 $\pm$ 1.4	2440 $\pm$ 20	2697 – 2360	2474	S, charcoal (0.87 mg C)
74C	230824, –	0021	1	10	0.7145 $\pm$ 0.0010	-285.5 $\pm$ 1.0	2700 $\pm$ 15	2849 – 2757	2977	S, charcoal (~0.7 mg C)

(continued).

Field sample # (BR19-)	Lab ID # (AMS and NAU) ¹	IGSN # ²	Wall ³	Unit ⁴	Fraction modern $\pm 2\sigma$	D ¹⁴ C $\pm 2\sigma$ (‰)	¹⁴ C age $\pm 1\sigma$ (yrs BP)	Calibrated age range ⁵ (cal yrs BP ⁶)	Median calibrated age (cal yrs BP ⁶)	Sample type ⁷
76C	230823, –	0023	1	10	0.7004 $\pm$ 0.0015	-299.6 $\pm$ 1.5	2860 $\pm$ 20	3064 – 2881	2977	S, charcoal (~0.7 mg C)
77C	230817, –	0024	1	11	0.7618 $\pm$ 0.0013	-238.2 $\pm$ 1.7	2185 $\pm$ 15	2304 – 2121	2250	S, charcoal (~0.7 mg C)
81C	239915, 2274	0028	2L	4	0.3363 $\pm$ 0.0014	-666.5 $\pm$ 1.4	8755 $\pm$ 35	9897 – 9557	9739	S, charcoal (0.44 mg C)
85C	230834, –	002C	3L	4	0.2582 $\pm$ 0.0007	-741.8 $\pm$ 0.7	10875 $\pm$ 25	12832 – 12747	12789	S, charcoal (~0.7 mg C)
86C	239916, 2275	002D	3L	4	0.2571 $\pm$ 0.0014	-745.1 $\pm$ 1.4	10910 $\pm$ 45	12907 – 12746	12811	M, charcoal (0.41 mg C)
87C	239917, 2276	002E	3L	4	0.3178 $\pm$ 0.0017	-684.8 $\pm$ 1.7	9205 $\pm$ 45	10499 – 10246	10363	M, charcoal (0.19 mg C)
91C	230833, –	002I	4	5b	0.5724 $\pm$ 0.0011	-427.6 $\pm$ 1.1	4480 $\pm$ 15	5285 – 5042	5184	S, charcoal (~0.7 mg C)
94C	239919, 2277	002L	4	11	0.9205 $\pm$ 0.0017	-87.3 $\pm$ 1.7	665 $\pm$ 15	668 – 563	590	S, charred wood (0.38 mg C)
101C-a ⁸⁹	230820, –	002S	AS	5b	0.9837 $\pm$ 0.0018	-16.3 $\pm$ 1.8	130 $\pm$ 15	268 – 14	102	S, charcoal (~0.7 mg C)
101C-b ⁸⁹	230821, –	002S	AS	5b	0.9846 $\pm$ 0.0023	-11.7 $\pm$ 1.6	95 $\pm$ 15	256 – 33	120	S, charcoal (~0.7 mg C)
101C-c ⁸⁹	230822, –	002S	AS	5b	0.9837 $\pm$ 0.0018	-15.4 $\pm$ 2.3	125 $\pm$ 20	268 – 14	106	S, charcoal (~0.7 mg C)
106C	230818, –	002X	AS	11	1.0379 $\pm$ 0.0019	37.9 $\pm$ 1.9	Modern	1956 – 2009 CE	1956 CE	S, fern (brown; ~0.7 mg C)
111C	230819, –	0032	4	11	1.0279 $\pm$ 0.0018	27.9 $\pm$ 1.8	Modern	1955 – 1957 CE	1956 CE	S, fern (green; ~0.7 mg C)

(continued).

Field sample # (BR19-)	Lab ID # (AMS and NAU) ¹	IGSN # ²	Wall ³	Unit ⁴	Fraction modern $\pm 2\sigma$	D ¹⁴ C $\pm 2\sigma$ (‰)	¹⁴ C age $\pm 1\sigma$ (yrs BP)	Calibrated age range ⁵ (cal yrs BP ⁶)	Median calibrated age (cal yrs BP ⁶)	Sample type ⁷
113C	239920, 2278	0034	4	11	0.9544 $\pm$ 0.0018	-53.7 $\pm$ 1.8	375 $\pm$ 20	498 – 325	451	M, charcoal (Sv) ¹⁰ (0.69 mg C)
114C-a ⁹	239921, 2279	0035	1	11	0.6235 $\pm$ 0.0014	-381.8 $\pm$ 1.4	3795 $\pm$ 20	4242 – 4092	4181	S, charcoal (Sv) (1.03 mg C)
114C-c ⁹	239922, 2280	0035	1	11	0.6157 $\pm$ 0.0014	-389.5 $\pm$ 1.4	3895 $\pm$ 20	4413 – 4248	4339	S, charcoal (Sv) (0.75 mg C)
114C-d ⁹	239923, 2281	0035	1	11	0.6254 $\pm$ 0.0014	-379.9 $\pm$ 1.4	3770 $\pm$ 20	4235 – 4013	4129	S, charcoal (Sv) (0.49 mg C)
118C	239924, 2282	0039	AN	2	0.2359 $\pm$ 0.0014	-766.1 $\pm$ 1.4	11600 $\pm$ 50	13585 – 13339	13462	S, charcoal (0.77 mg C)
127C	239925, 2283	003I	AN	2	0.3054 $\pm$ 0.0014	-697.1 $\pm$ 1.4	9525 $\pm$ 40	11078 – 10663	10859	S, charcoal (0.68 mg C)

¹Six-digit AMS lab ID numbers listed first are from the UC Irvine Keck Carbon Cycle AMS Lab (UCIAMS#), and four-digit NAU Lab ID numbers listed second are from the Arizona Climate & Ecosystems Isotope Lab at NAU where samples underwent ABA treatment, combustion, and graphitization (NAU#).

²Four-digit IGSN numbers refer to the International Geoscience Sample Number (10.58052/IEBRF#).

³1 - Slice 1; 2 - Slice 2 (upper); 2L - Slice 2L (lower); 3L - Slice 3L (lower); 4 - Slice 4; AN - A-Wing North; AS - A-Wing South; BN - B-Wing North; BS - B-Wing South. See Figure 3.2 for trench footprint.

⁴Units numbered and described in Table C.1. SS refers to sheared sediment within the fault zone.

⁵Radiocarbon ages calibrated using INTCAL20 (P. J. Reimer et al., 2020) and OxCal v. 4.4 (Bronk Ramsey, 2021). Range reported represents unmodeled 95% confidence interval as calculated by OxCal.

⁶Calibrated age is reported in calibrated years Before Present (1950) except where modern.

⁷S – single piece; M – multiple pieces; Sv - sieved from bulk sediment.

⁸Samples BR19-101C-a, BR19-101C-c, and BR19-101C-b are from an organic-rich horizon at the top of unit 5b (see Fig 3.4).

⁹Lower-case letters indicate that samples were selected from the same location on the trench walls.

However, we recognize two limitations of model A. First, it over-constrains the timing of Event 4 (to within a 100-year interval), beyond reasonable geologic precision in coarse colluvial deposits. To address this point, we modified model A to exclude a sample from the top of unit 8 (Event 3 colluvial wedge). The results of this OxCal model are presented in model A.1 (Figure 3.5a). Model A.1 yielded the same earthquake ages as model A for Events 1-3, 4.2-3.0 ka for Event 4, and 2.8-2.4 ka for Event 5 (Figure 3.5, Table C.3). Model A.1 is our preferred model. Second, model A also appears to contradict the geomorphic observation of a preserved free face (surface exposed by faulting, e.g., Wallace, 1977) at the modern ground surface along portions of the BRF (e.g., Figure 2.8 c). Free faces are typically preserved for years to centuries following earthquakes, but not millennia (McCalpin, 2009). This suggests that the MRE may be younger than the model A.1 age of 2.8-2.4 ka, particularly if the samples from unit 11 had in-built ages of 100s of years (e.g., Gavin, 2001). We allow for this possibility in model A.2, in which we assume that all unit 11 (MRE wedge) samples that are older than ~0.3-0.7 ka are recycled (Figure 3.5, Table C.3). Model A.2 included 14 samples and excluded 24 samples. Model A.2 yielded the same ages as model A.1 for Events 1-4, and 2.8-0.7 ka for Event 5.

We recognize that the preferred age model above (model A.1) is only one plausible interpretation of the ages of each event. To better evaluate the potential range of permissible ages for each event, we ran two additional OxCal models, models B and C, using permissible combinations of sample ages. In model B, we preferentially selected the oldest sample ages to maximize the age of each colluvial wedge, therefore providing estimates on the oldest possible age for each earthquake. Model B included 17 samples and excluded 21 samples; however, this model included several samples earlier assessed to be outliers due to anomalously old ages. Model B yielded ages of 13.5-12.8 ka for Event 1, 10.1-9.6 ka for Event 2, 8.4-4.1 ka for Event 3, 4.3-4.1 ka for Event 4, and 4.2-4.1 ka for Event 5 (Figure C.11, Table C.3). In model C, we preferentially selected the youngest sample ages in each colluvial wedge to minimize the age of each colluvial wedge, therefore providing estimates on the youngest possible age for each earthquake. Model C included 12 samples and excluded 26 samples; however, this model included several samples earlier assessed to be outliers due to anomalously young ages and potential for sediment mixing. Model C yielded ages of 5.9-5.2 ka for Event 1, 5.2-4.9 ka for Event 2, ~2.9 ka for Event 3, 2.9-2.8 ka for Event 4, and 2.7-0.5 ka for Event 5 (Figure C.11, Table C.3). Despite omitting more than two-thirds of the samples, model C over-constrains the timing of Events 3 and 4 beyond reasonable geologic precision.

Because all four models we present here are geologically plausible, we combined all four of these models into a weighted mean (e.g., DuRoss et al., 2018, Section 3.3.5) to assess the overlap in model ages and develop a most likely age PDF for each event. We considered model A.1 to be the most likely of the four models, followed by model A.2. We gave model A.1 a weight of 3, model A.2 a weight of 2, and models B

and C weights of 1. This weighted mean, presented in Figure C.12, yielded earthquake age 95% confidence intervals of 13.7-5.4 ka for Event 1, 9.8-4.8 ka for Event 2, 6.7-2.8 ka for Event 3, 4.5-2.6 ka for Event 4, and 4.9-0 ka for Event 5.

3.4.3 Reconstructions of fault slip

3.4.3.1 Cumulative oblique displacement

We use the vertical and lateral offset of a buried paleochannel developed on unit 2 (paleo hillslope colluvium) to reconstruct cumulative oblique slip across the BRF since the late Pleistocene. This paleochannel underlies, and therefore predates, the oldest colluvial wedge, and has been displaced by all five

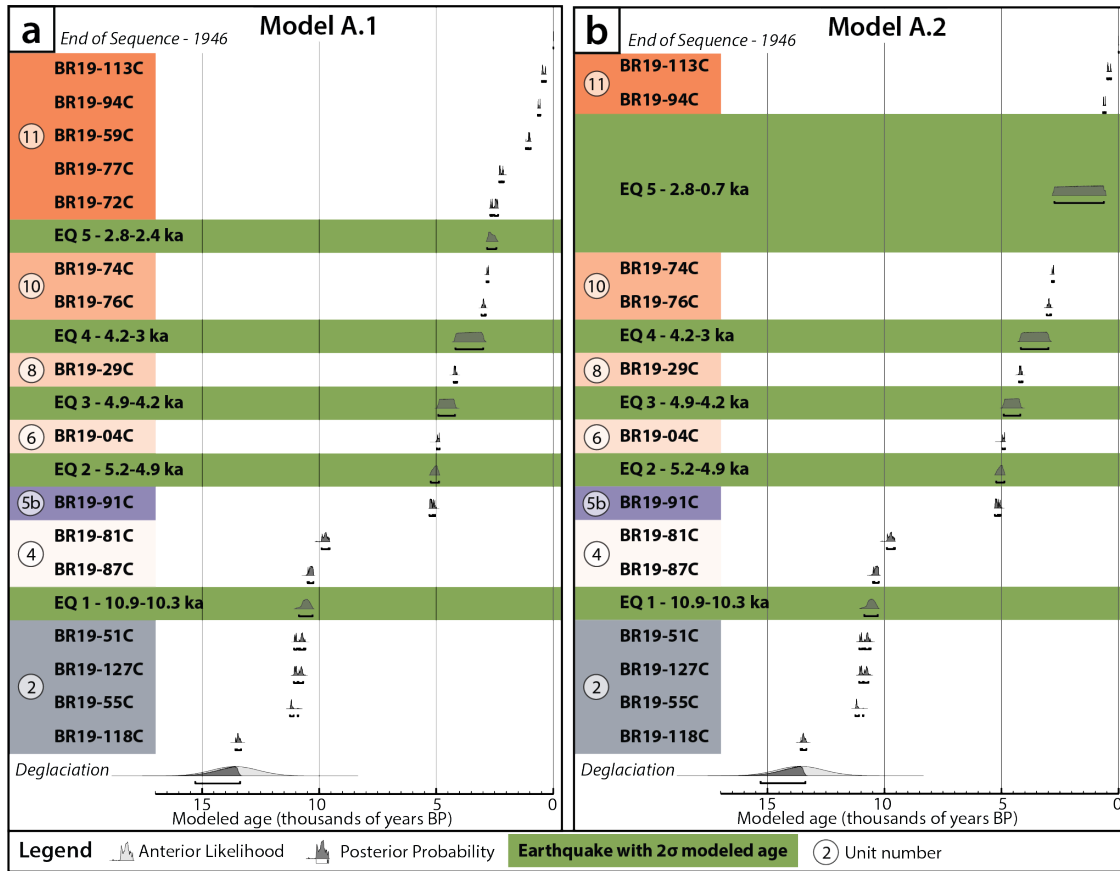


Figure 3.5: Synthesis figure showing results of OxCal models A.1 and A.2. The calibrated (anterior likelihood) and modeled (posterior probability) age PDFs are shown for each radiocarbon sample. See Section 3.3.5 for OxCal methods. **A:** Preferred OxCal model A.1 with modeled earthquake age PDFs and 2σ earthquake age ranges highlighted in green. Model A.1 includes 16 samples and excludes 22 samples. **B:** OxCal model A.2 with modeled earthquake age PDFs and 2σ earthquake age ranges highlighted in green. Model A.2 includes 14 samples and excludes 24 samples. Results of OxCal models B and C are in Figure C.11a-b.

paleoearthquake events. Field surveys of the XYZ coordinates of the paleochannel thalweg and margins at its intersection with several trench walls (Figure 3.2; Appendix C) indicate that the paleochannel is displaced both vertically and laterally by the BRF. However, only the eastern margin of the buried channel, not the thalweg, was exposed in the lower part of the trench just north of the scarp. We constrain the buried channel thalweg north of the scarp with the lowest point in Slice 3L (Figure 3.3b). The buried paleochannel thalweg may project further west (shown by the shaded region in Figure 3.2). The best fit paleochannel thalweg projection had a trend and plunge of $\sim 167^\circ/05^\circ$ and $\sim 171^\circ/18^\circ$ in the blocks north and south of the scarp, respectively, across an average fault plane with a strike and dip of $297^\circ/80^\circ$ NE (Table C.3). Reconstruction of fault slip of this buried paleochannel thalweg yielded a cumulative right-lateral oblique displacement of 7.6 ± 0.3 m (6.0 ± 0.2 m dip slip, 4.6 ± 0.4 m strike slip; Table 3.2). These data suggest the BRF at this site has accommodated an $\sim 0.8:1$ ratio of strike slip to dip slip since late Pleistocene time.

3.4.3.2 Reconstructions of slip per event

We use a combination of different methods to reconstruct the oblique slip that occurred in each earthquake, using three methods to quantify the dip-slip component and two methods to quantify the strike-slip component of slip per event. First, we reconstruct the dip slip per event using direct offset of trench units along mapped faults by reconstructing the offset between the correlative hanging wall and footwall cutoffs. However, correlative cutoffs on both sides of the fault are only present for two units (units 2 and 5d; Figure 3.3). Therefore, in the event where one of the cutoffs is not preserved (typically the hanging wall cutoff), we assume that the length of faults in direct contact with the offset deposits is an estimate of the minimum dip slip that occurred in that event. Where multiple faults are in contact with the offset deposits, we sum the length of all involved faults, and where displacement can be measured on both walls, we report displacements that average the offsets from both walls. This method yields per-event dip slip estimates of 0.3-0.9 m (Table 3.2). However, this second method does not account for slip along faults within the shear zone and is, therefore, a minimum estimate.

Third, we reconstruct dip slip for individual events using the thickness of the colluvial wedges deposited following each event (Figure 3.3). This method assumes that the maximum exposed thickness of the colluvial wedge, proximal to the fault zone, measured on a trench log represents approximately half of the scarp height, and therefore half of the total vertical component of slip per event (e.g., J. P. McCalpin et al., 2009; Bennett et al., 2018). The colluvial wedge thicknesses we measure of ~ 0.3 - 0.6 m yield per-event dip slip estimates of ~ 0.6 - 1.1 m (Table 3.2). These reconstructions likely also underestimate dip slip because the thickest parts of wedges, adjacent to the fault zone, may have been truncated and incorporated into the sheared sediment within the fault zone in subsequent events, such that the full wedge thickness is no longer preserved. In

addition, the MRE wedge thickness may be an underestimate as it may not yet have built up to half of the scarp height. We prefer estimates of per-event dip slip determined from colluvial wedge thickness for two reasons. First, this method does not overlook the slip on the faults in the sheared sediments, as is the case for the second method using slip along discrete fault planes. Second, estimates of per-event dip slip from colluvial wedge thickness are consistent with cumulative displacements estimated for Events 1-5 by matching unit 2 cutoffs and for Events 2-5 by matching unit 5d cutoffs, shown on retrodeformed logs (see Figures C.9-C.10).

Fourth, we estimate per-event strike slip using two approaches. The first approach assumes that the cumulative oblique displacement of the buried paleochannel of 7.6 ± 0.3 m is the result of five events, each with ~ 1.5 m of oblique slip (~ 0.9 m right-lateral slip, ~ 1.2 m dip slip). This approach assumes similar magnitude of slip in each event, which is supported by the similarity in dip slip per event. The second approach uses our estimates of per-event dip slip from colluvial wedge thickness to back calculate per-event oblique slip, assuming a constant ratio of $\sim 0.8:1$ strike slip to dip slip. This approach yields per-event strike-slip magnitudes of 0.5-1 m (Table C.5).

We combine each approximation of dip slip and strike slip to calculate oblique slip per event, along a fault plane with a strike and dip of $297^\circ/80^\circ$ NE. These oblique slip estimates range from 0.75 to 1.54 m per event. These per-event oblique displacements correspond to cumulative oblique slip of 5.7-7.5 m in five events. This is comparable to our estimates of oblique slip from the buried paleochannel of ~ 7.6 m (Section 3.4.3.1). The slight discrepancy between these methods may be the result of distributed deformation affecting the buried paleochannel away from the fault zone and/or the possible underestimates of dip slip discussed above.

3.4.3.3 Earthquake magnitudes

We estimated earthquake magnitude directly from per-event displacements (D. L. Wells & Coppersmith, 1994), and indirectly through estimates of surface rupture length from per-event displacements (25-150 km; Wesnousky, 2008). We calculated surface rupture length and earthquake magnitudes using relationships for both normal and strike-slip faults to account for the oblique kinematics of the BRF (Table C.5). The above estimates of per-event displacements and scaling relationships from D. L. Wells and Coppersmith (1994) and Wesnousky (2008) yield estimated moment magnitudes of 6.7-7.5 (Table C.5).

3.4.4 Recurrence intervals and slip rates

modeled earthquake ages for the five paleoearthquakes documented in the Woodlot Trench show that the earthquake recurrence interval on the BRF varies significantly, regardless of which OxCal model is used.

Our preferred OxCal model (model A.1) and our mean OxCal model (Figure C.12; Section 3.3.5) include one earthquake ($\sim 10.9\text{-}10.3$ ka) that occurred during or a few kyr after deglaciation. Approximately 5.1-6 kyr passed before the second event occurred $\sim 5.2\text{-}4.9$ cal ka BP. There is no evidence in the Woodlot Trench for any earthquakes in the $\sim 10.3\text{-}5.2$ ka time period; however, we cannot rule out the possibility that any such events ruptured other strands of the BRF north and south of the strand examined at the Woodlot Trench (e.g., those mapped in Figure 3.1b). In model A.1, Event 2 occurred $\sim 5.2\text{-}4.9$ ka, Event 3 occurred $\sim 4.9\text{-}4.2$ ka, and Event 4 occurred $\sim 4.2\text{-}3$ ka. Units 6, 8, and 10 (the colluvial wedges associated with Events 2, 3, and 4, respectively) are well-developed wedges with distinct lower boundaries and soil development, separated by ponded sediments in units 7 and 9, consistent with three distinct events. However, the overlap in event ages makes it more difficult to determine recurrence intervals between these events. Following the ages from model A.1, ≤ 1 kyr passed between Events 2 and 3, and ≤ 1.9 kyr passed between Events 3 and 4. However, each of these events has an associated colluvial wedge of similar thickness and soil development (Table C.2), suggesting these events were spaced far apart in time as to allow deposition and stabilization of a separate colluvial wedge. We can estimate an average recurrence interval of 0.5-1.1 kyr for these three events from the 1-2.2 kyr that passed between Events 2 and 4. This pattern of non-steady recurrence exists in all our OxCal models, though the precise timing varies (Table C.6). In models A.1 and A.2, there is a long ($\sim 5\text{-}6$ kyr) gap between Events 1 and 2, while Events 2, 3, and 4 are closely spaced in time (Table C.6). In model B, approximately $\sim 2\text{-}3$ kyr pass between Events 1 and 2, and Events 2 and 3, while Events 4 and 5 occur in quick succession (Figure C.11; Table C.6). In model C, there are no events in the late Pleistocene to early Holocene. Instead, Events 1 and 2 occur in a ~ 1 kyr time span, followed by a ~ 2 kyr quiet period, and then Events 3, 4, and 5 within $\sim 0.2\text{-}2$ kyrs (Figure C.11; Table C.6).

These non-steady recurrence intervals correspond with non-steady slip rates on the BRF. The open-ended slip rate from the timing of deglaciation to the present is $\sim 0.5\text{-}0.6$ mm/yr (Figure 3.6; Table 3.3). However, this may be an underestimate as it includes additional time both before Event 1 and since Event 5. Using closed intervals (following Working Group on California Earthquake Probabilities, 1990) and the earthquake ages from model A.1, the Holocene slip rate between Events 1 and 5 is $\sim 0.6\text{-}0.7$ mm/yr. The slip rates for individual intervals are ~ 0.2 mm/yr between Events 1 and 2, >1.4 mm/yr between Events 2 and 3 and Events 3 and 4, and $0.8\text{-}7.7$ mm/yr between Events 4 and 5. These closed-interval slip rates vary somewhat between earthquake age models (Table C.6), but the Events 3-4 and 4-5 interval slip rates are consistently higher than the early to mid-Holocene rates in each model. The slip rate estimated from surficial mapping and ground surface surveys along ~ 6 km of the southern BRF is $\sim 0.5\text{-}2$ mm/yr (Chapter 2), which is comparable to our mid-Holocene slip rate from the Woodlot Trench.

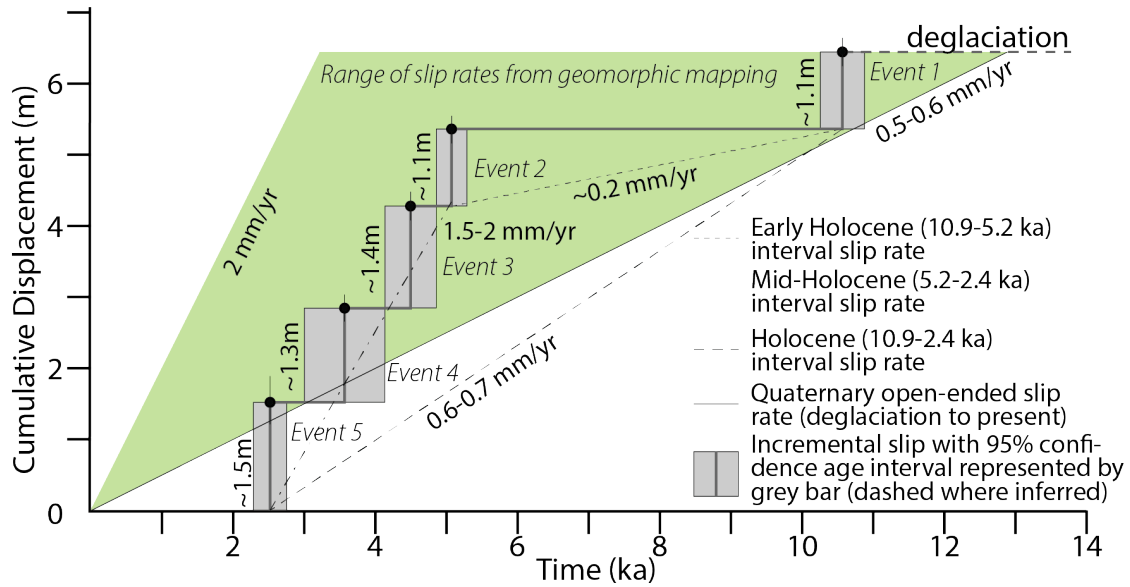


Figure 3.6: Slip and slip rates calculated for the Beaufort Range fault from the Woodlot Trench. Cumulative displacement estimates combine individual vertical displacements from colluvial wedge heights and average right-lateral displacement from the buried paleochannel (see Section 3.4.3.2; Table 3.2). Timing of individual events shown here follows our preferred OxCal model A.1 (see Section 3.4.2.2 and Figure 3.5 for more details). Closed-interval slip rate estimates shown by dashed and dotted lines use the range of recurrence intervals. Open-ended slip rate from deglaciation to the present is shown by the solid line. Green shaded area shows the range of open-ended slip rates estimated from geomorphic mapping and topographic surveys (0.5-2 mm/yr; Chapter 2).

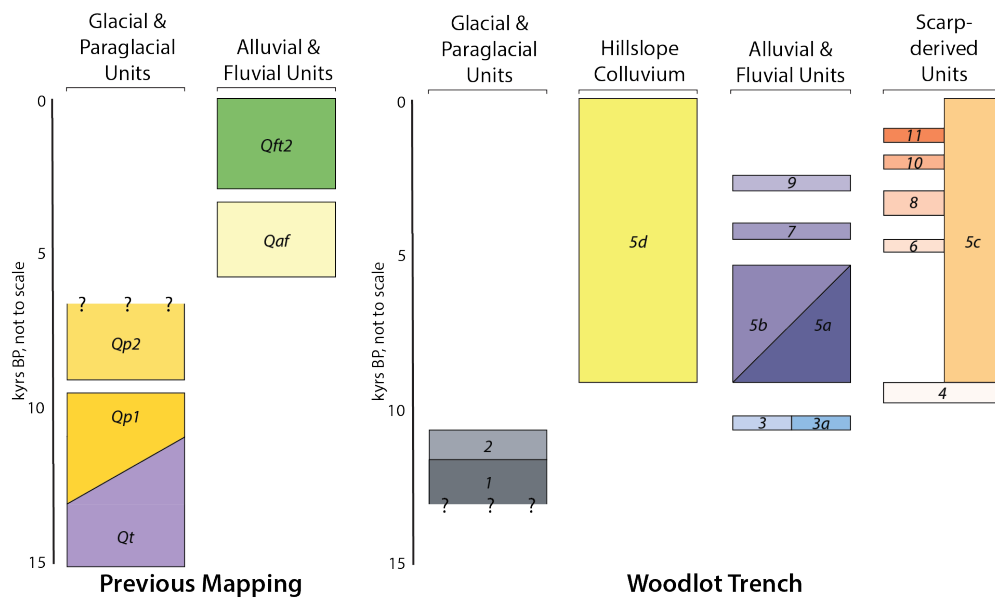


Figure 3.7: Stratigraphic correlation between previously mapped units along the southern BRF (left) and the units described in this paper from the Woodlot Trench paleoseismic site (right).

Table 3.2: Earthquake timing and displacements in the Woodlot Trench.

Event	Earthquake Age (ka) ¹	Measured from trench logs			Reconstructed from channel displacement			Range of Oblique Slip estimates (m)
		Sum of dip slip on faults (m) ²	Vertical displacement from wedge height (m) ³	Vertical Separation (m) ⁴	<i>Average Oblique Slip (m)</i>	<i>Average Strike Slip (m)</i>	<i>Average Dip Slip (m)</i>	
Event 5	2.8-2.4	0.47 (0.46-0.48)	1.14 (0.88-1.44)	~3.9	~1.5	~0.9	~1.2	1.45-1.54
Event 4	4.2-3	0.53 (0.46-0.61)	0.96 (0.86-1.04)		~1.5	~0.9	~1.2	1.29-1.5
Event 3	4.9-4.2	0.65 (0.4-0.89)	1.08 (0.92-1.26)		~1.5	~0.9	~1.2	1.41-1.5
Event 2	5.2-4.9	0.49 (0.27-0.65)	0.61 (0.43-0.81)		~1.5	~0.9	~1.2	0.82-1.5
Event 1	10.9-10.3	–	0.56 (0.44-0.73)	–	~1.5	~0.9	~1.2	0.75-1.5
Sum of all events measured from trench logs								
		>2.1 (1.6-2.6)	4.4 (3.5-5.3)	4.9 (3.7-5.1)				5.45-6.46
Sum of all events since formation of buried channel⁵								
					7.57±0.33	4.57±0.43	6.02±0.15	
Sum of all events since formation of abandoned surface channel⁶								
		–	–	–	6.93±0.89	1.9± 0.92	6.62±0.74	–

¹Earthquake ages reported as 2σ ranges from our preferred OxCal model.²Mean dip slip (minimum-maximum in parentheses) calculated as the sum of fault-parallel offset on all faults, averaged from both trench walls where applicable.³Determined assuming that vertical displacements are double the maximum colluvial wedge(s) thickness.⁴Cumulative vertical separation calculated by projecting the top of unit 2 for Events 2-5, and the base of unit 5d for Events 1-5.⁵Average oblique, strike slip, and dip slip calculated assuming the cumulative displacement of the paleochannel thalweg represents five events with equal displacements.⁶Calculated from 3D regression of the channel thalweg on the modern ground surface from topographic profile TH5.

Table 3.3: Slip rates in the Woodlot Trench

Interval	Recurrence period (kyr) ¹	Slip Rate (mm/yr) ²		
		Between events ³	All events ³	Since deglaciation ⁴
Event 4 to 5	1.8-0.2	0.8-7.5	0.6-0.7	0.5-0.6
Event 3 to 4	<1.9	>1.4		
Event 2 to 3	<1	>1.4		
Event 1 to 2	6-5.1	~0.2		

¹Recurrence periods are from OxCal model A.1.

²Slip rates are calculated using the total displacement estimates reported in Table 3.2.

³Closed interval

⁴Open-ended

3.5 Discussion

3.5.1 Five surface-rupturing earthquakes since c. 15 ka

The five scarp-derived colluvial wedges in the Woodlot Trench record five late Quaternary earthquakes. All units exposed in the Woodlot Trench, and therefore all events, post-date deglaciation, which occurred $<\sim 13.6$ ka (Mosher & Hewitt, 2004; Easterbrook, 1992; Clague & James, 2002). The scarp-derived colluvial wedges, units 4, 6, 8, 10, and 11, each record a surface-rupturing earthquake (Section 3.4.1.3), and the older 4 of these wedges interfinger with the channel deposits of units 5a, 5b, 7, and 9, indicating that post-glacial slope adjustment was punctuated by earthquakes that created a fault scarp and blocked the channel flow. These five late Quaternary earthquakes in the Woodlot Trench are consistent with the three or more earthquakes on the southern BRF documented in prior work, which included one or more post-glacial event, one or more mid-Holocene events, and one or more event since $<\sim 4$ ka (Chapter 2). All earthquake age models, with the exception of model C (youngest possible events), include one earthquake soon after deglaciation, a several kyr period of quiescence, followed by several mid-Holocene events. The MRE likely occurred within the past <2.7 kyrs. These repeated surface ruptures throughout the Holocene indicate that there is a persistent source of strain accumulation over millennial timescales, rather than the cluster of post-glacial earthquakes that have occurred in response to glacial isostatic adjustment in other regions (e.g., northern Fennoscandia, Lagerbäck, 1990).

3.5.2 Did the 1946 Vancouver Island earthquake rupture the Woodlot Trench site?

Previous work has suggested that the BRF may have been the source fault for the 1946 M_S 7.3 Vancouver Island earthquake, based on proximity of relocated focal mechanisms to the BRF, geodetic surveys, and

kinematic analysis (e.g., Rogers & Hasegawa, 1978; Slawson & Savage, 1979, Chapter 2). The Woodlot Trench site occurs within the rupture area modeled by Slawson and Savage (1979), and the 1.00 ± 0.25 m of right-lateral slip and 2.50 ± 0.65 m of normal slip they modeled is broadly similar to our estimates of per-event displacement. The steeply NE-dipping fault planes we measured in the Woodlot Trench are similar to one of the nodal planes determined by Rogers and Hasegawa (1978, Figure 3.1), and to the fault plane used by Slawson and Savage (1979) to recreate their observed geodetic deformation. While numerous ground failures occurred as a result of the 1946 earthquake, including slides, slumps, and sand blows, no surface ruptures were identified at the time (Hodgson, 1946). Free faces mapped elsewhere along the southern BRF (e.g., Chapter 2; Figure C.14) are consistent with a recent (<100s yrs) MRE, however the Woodlot Trench site was clear-cut and burned at the turn of the 20th century. One would therefore expect to find charcoal from that time incorporated into the colluvial wedge from a 1946 CE earthquake. However, none of the radiocarbon samples we collected from unit 11 (the colluvial wedge associated with the MRE) were younger than 0.3 ka (Figure 3.4). A 1946 CE MRE would require discarding all eight dated samples from unit 11. Slawson and Savage (1979) modeled a 60 km surface rupture continuing south from an epicenter in the Forbidden Plateau, and the Woodlot Trench is located 59 km south of the relocated epicenter of Rogers and Hasegawa (1978) (Figure 3.1a). Therefore, a 1946 rupture on the BRF may not have included surface ruptures as far south as the Woodlot Trench, an interpretation corroborated by the lack of evidence for a 1946 rupture in the Woodlot Trench. Additionally, the strand excavated in the Woodlot Trench is one of many strands in a horsetail splay. Strands to the west are steeper and better preserved, including free faces (Chapter 2). This suggests that these horsetail splay strands may have ruptured in a recent event that did not rupture the Woodlot Trench site, such as the 1946 Vancouver Island earthquake. While we find no evidence for a 1946 CE rupture in the Woodlot Trench, we cannot rule out that the 1946 Vancouver Island earthquake ruptured the BRF at another site or on another strand.

3.5.3 Fault slip, kinematics, & earthquake magnitude

The magnitude (~ 7.5 m) and kinematics (right-lateral oblique) of fault slip we document in the Woodlot Trench are consistent with the ~ 7 m of right-lateral oblique slip determined from surficial mapping of late Pleistocene to Holocene deposits (Chapter 2), indicating that the right-lateral transtension is a persistent signal through the Holocene. The dip-slip magnitude determined from the buried paleochannel (~ 6 m; Table 3.2) is also consistent with the estimates of per-event dip slip determined from slip along faults and colluvial wedge heights measured from the trench logs, and with the abandoned surface channel (Table 3.2). The ~ 2 m cumulative right-lateral slip estimated from the modern-day expression of the channel at the ground

surface is significantly less than the ~ 4.6 m estimated from the buried paleochannel (Figure 3.2c; Table 3.2). This discrepancy in the right-lateral component (Figure 3.2a) is consistent with progressive displacement in multiple events and subsequent channel adjustment (e.g., Sieh, 1978; Sieh & Jahns, 1984; Lienkaemper & Sturm, 1989; Liu-Zeng et al., 2006). The displacement of the buried paleochannel demonstrates right-lateral oblique slip with an 0.8-1.6:1 ratio of strike to dip slip over the late Pleistocene to Present, which is consistent with kinematic inversions for the BRF both near the Woodlot Trench and further north near Comox Lake (described in Chapter 2; see Figure 2.9), as well as the focal mechanism for the 1946 Vancouver Island earthquake (Figure 3.1). ca

Scaling relationships based on average displacements of ~ 0.75 -1.5 m per event yield magnitude estimates of M_W 6.7-7.5 (Table C.5), which would pose considerable hazard to the surrounding communities and infrastructure. These magnitude estimates are consistent with the M_W 6.8 to >7.0 determined from average displacements estimated from surficial mapping and topographic surveys (Chapter 2). The surface rupture lengths of 25-150 km estimated from average displacement encompass the 35-60 km total length of mapped scarps along the BRF and corresponding magnitudes (M_W 6.5-7.5; Chapter 2). The M_S 7.3 1946 Vancouver Island earthquake caused damage to buildings in communities both on Vancouver Island and in Powell River on the mainland (Hodgson, 1946). In particular, damage affected the hospital in Port Alberni where those injured in the earthquake were treated and the telephone wires crossing the Beaufort Range connecting Port Alberni with the more populous communities along the northeastern coast of Vancouver Island (Hodgson, 1946). Today, Highway 4 remains the only major road crossing the Beaufort Range and providing access to the west coast communities of Tofino and Ucluelet. The Comox, Alberni-Clayoquot, and Nanaimo Regional Districts have a combined population of over 250,000 people who would be at risk from a M_W 6.7-7.5 earthquake on the BRF.

3.5.4 Non-steady slip rates on the BRF & implications for regional seismotectonics

The varying recurrence intervals in the Woodlot Trench (Section 3.4.4), and the consistent slip per event (Section 3.4.3.2) result in changing slip rates in the Holocene, which may reflect a change in strain accumulation rate in the mid-Holocene. Closed-interval slip rates on the southern BRF range from ~ 0.2 mm/yr during the late Pleistocene to early Holocene to as high as 7.7 mm/yr during the middle and late Holocene (Section 3.6). The high slip rates indicate that the BRF accommodates a significant portion of strain accumulation in the overriding plate. These rates may still be underestimates, as they do not account for any off-fault deformation that may have occurred outside of the trench exposure, or ruptures on other

strands (e.g., those in Figure 3.1c). The mid-Holocene acceleration in slip rates is a robust pattern, as it exists in all models (Table C.6). All models have a relatively slow slip rate in the early Holocene, between Events 1 and 2. In contrast, one or more mid-Holocene interval slip rates are 1-2 orders of magnitude faster than the early Holocene interval slip rate for that same model. The varying slip rates on the BRF may reflect a significant acceleration in strain accumulation in the mid-Holocene. Such a change may have occurred as a result of changing stresses in the lithosphere due to glacial isostatic adjustment (e.g., Steffen & Steffen, 2021), “incomplete” strain release (e.g., Weldon et al., 2004), or interactions with other fault systems (e.g., Dolan et al., 2007). Further work is necessary to determine if this is a regional pattern, and if so, to assess the possibility of these or other drivers.

The varying recurrence and slip rates on the BRF affect how we assess the hazard faced by nearby communities. Slip-predictable models of earthquake recurrence (e.g., Shimazaki & Nakata, 1980) use the time since the previous event to predict how large the next earthquake will be. However, the wide range in recurrence intervals on the BRF (<1-6 kyr) but similar magnitudes of offset (~ 0.75 -1.5 m, Table 3.2) would suggest that using the time since the previous event is not the most robust approach to estimate the expected slip and magnitude of future events. Active faults in other tectonic settings have documented similar increases in late Holocene slip rates, and related shorter recurrence intervals, all with no significant change in the magnitude of slip per event (e.g., Bennett et al., 2018), which suggests that this dilemma is not unique to forearc faults systems. Other models use the slip rate of a fault to assess hazard (e.g., Youngs & Coppersmith, 1985; Petersen et al., 2014), but the varying slip rates on the BRF make this approach problematic as well. A conservative approach would use the Quaternary open-ended slip rate of ~ 0.5 -0.6 mm/yr; however, this may be an underestimate, and therefore result in inadequate hazard preparation. On the other hand, using the fastest interval slip rate of 0.8-7.7 mm/yr between Events 4 and 5 may significantly overestimate the current slip rate. This wide range of slip rates makes it difficult to accurately assess the hazard posed by the BRF.

The five Quaternary earthquakes and fast slip rates on the southern BRF may indicate higher strain rates in the northernmost Cascadia forearc than farther south, or that strain is more localized in the northernmost forearc, concentrated on the BRF, rather than distributed amongst a network of faults as in the central and southern Cascadia forearc. On Vancouver Island, three earthquakes have been documented on the Leech River fault (LRF; Figure 1.1; Morell et al., 2018), and one earthquake has been documented on the ~~XEOLXELEK~~-Elk Lake fault (XELF; Figure 1.1; Harrichhausen et al., n.d., *in review*). Mid-Holocene slip rates on the southern BRF (>1.4 mm/yr, Section 3.4.4) are also higher than the ~ 0.1 -0.3 mm/yr on the LRF, XELF, and Darrington-Devils Mountain fault (DDMF; Morell et al., 2017, 2018; Personius et al., 2014, Harrichhausen et al., *in review*). Slip rates on the BRF are more comparable to the 1.3-2.3

mm/yr of the North Olympic fault zone (Schermer et al., 2021). The BRF is located ~ 100 km northwest of the LRF, the closest documented active forearc fault. The XELF is within ~ 15 km of the LRF, which projects across the Salish Sea towards the DDMF and Southern Whidbey Island fault zone (Figure 1.1). If forearc strain is localized on the BRF in the northernmost Cascadia forearc, it may be comparable to the cumulative strain accommodated on the multiple faults distributed across southern Vancouver Island and Puget Lowlands. However, if there are additional active faults near the BRF that have yet to be identified, the strain accumulation in the northernmost Cascadia forearc may be even higher. This would be contrary to models that cast Vancouver Island north of the LRF as minimally- or non-deforming (e.g., R. E. Wells et al., 1998; McCaffrey et al., 2007, 2013) and to the model of locking on the subduction interface as the primary driver of upper plate deformation (Wang, 2000). Models of interseismic coupling show a decrease in slip deficit north of the Olympic Peninsula (e.g., McCaffrey et al., 2013; Li et al., 2018). A higher strain rate in the northernmost Cascadia forearc would therefore require some additional source of strain accumulation, such as oroclinal bending (e.g., Johnston & Acton, 2003; Finley et al., 2019; Harrichhausen et al., 2021) or impacts from the Nootka fault zone and Explorer microplate (e.g., Braunmiller & Nábělek, 2002; Audet et al., 2008; Hutchinson et al., 2023). More focused geomorphic mapping and paleoseismologic studies are necessary to determine if there are additional active faults in the northernmost Cascadia forearc and to further quantify strain accumulation.

3.6 Conclusions

We find that the Beaufort Range fault has hosted at least five surface-rupturing earthquakes since deglaciation (~ 13.6 -11 ka), along a steeply WNW-striking ($\sim 300^\circ$), NE-dipping ($>70^\circ$ NE) fault. Earthquakes created a fault scarp and blocked channel flow, punctuating post-glacial slope adjustment. The fault geometry and right-lateral oblique kinematics observed in the Woodlot Trench are consistent with evidence for right-lateral oblique transtension documented by recent surficial mapping (Chapter 2). Each paleoearthquake resulted in ~ 0.75 -1.5 m of oblique displacement, consistent with M_W 6.7-7.5 earthquakes. Future earthquakes of this magnitude would pose significant hazard to the central Vancouver Island region. The first earthquake represented in the Woodlot Trench occurred ~ 0.9 -10.3 cal ka BP, shortly after deglaciation, followed by a ~ 5 -6 kyr period of quiescence, before four additional earthquakes (~ 5.2 -4.9, ~ 4.9 -4.2, ~ 4.2 -3, and ~ 2.8 -0.7 cal ka BP). Recurrence intervals on the BRF range from <1 kyr to 5-6 kyr; however, the consistent slip in each event suggests that slip on the BRF is not time-predictable. Slip rates on the BRF range from ~ 0.2 mm/yr during the late Pleistocene to early Holocene to 0.8-7.7 mm/yr during the middle and late Holocene. This mid-Holocene increase in slip rate may reflect an increase in strain accumulation in

the northernmost Cascadia forearc.

Acknowledgments

We thank Todd Thompson for land access and tree felling, and the BC Ministry of Forests for assistance with permitting. Lidar ground returns were provided by TimberWest. Walter Langer excavated the trench, and Maximilian Lauch and Megan Caston assisted with trench logging and sample collection. We thank Colin Amos, Liz Schermer, and John Clague for productive discussions in the trench, and Darrell Kaufman, R. Scott Anderson, Jordon Bright, and Chris Ebert for assistance and support with radiocarbon sample processing. Funding for this project was supported by NSF EAR Tectonics 1756834/ 2004684 to C. Regalla and 1756943 to K. Morell, by the Boston University Department of Earth and Environmental Sciences, and the Northern Arizona University Duebendorfer-Barnes Structure Endowment. S. Bennett was supported by funding from the USGS Earthquake Hazards Program and the USGS National Cooperative Geologic Mapping Program. Any use of trade, firm, or product names is for descriptive purposes only and does not imply endorsement by the U.S. Government.

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Chapter 4

Non-steady recurrence and slip rates on Cascadia Subduction Zone forearc faults

Abstract

Earthquake recurrence and slip rates are primary components of probabilistic seismic hazard assessment (PSHA). Most PSHA models use average recurrence rates derived from seismic catalogs and geologic records. Seismically quiet areas like the Cascadia Subduction Zone (CSZ) forearc lack historical seismicity (10s-100s yrs) and require the use of paleoseismic records to estimate recurrence. Previous studies on CSZ forearc faults have suggested that recurrence and slip rates are non-steady. Non-steady recurrence and slip rate may not be well-represented by the average values used in PSHA. In this study, I compare the timing and recurrence of earthquakes on thirteen CSZ forearc faults and calculate slip rates for five faults to determine if they are steady or non-steady. I find that recurrence and slip rates are non-steady on most faults, and that faults tend to have long recurrence intervals in the late Pleistocene to early Holocene, and short recurrence intervals in the mid-late Holocene. Furthermore, recurrence and displacement on these forearc faults are inconsistent with periodic, time-predictable, or slip-predictable models of earthquake recurrence, and suggest that regional processes such as glacial isostatic adjustment or Coulomb stress transfer may help promote non-steady strain release.

4.1 Introduction

Earthquake frequency is one of three primary components of probabilistic seismic hazard assessment (PSHA), along with fault or area sources and spectral amplitude attenuation curves (Morell et al., 2020; Thenhaus & Campbell, 2002). Earthquake frequency is described through either magnitude-frequency (time-independent) models based on historical seismicity and/or geologic slip rate, or recurrence-frequency (time-dependent) models based on historic and/or paleoseismic recurrence intervals (e.g., Thenhaus & Campbell, 2002). Time-independent models assume a constant slip rate through time (e.g., Thenhaus & Campbell, 2002; Salditch et al., 2020), inconsistent with irregular slip rates documented in numerous tectonic settings

(e.g., Weldon et al., 2004; Dolan et al., 2016; Zinke et al., 2017, 2019, 2021; Hatem et al., 2020; DuRoss et al., 2020). Time-dependent models can be periodic or quasi-periodic, or they can be non-periodic. Periodic or quasi-periodic models assume that the recurrence interval does not vary significantly and can be described by the mean recurrence determined from historical seismicity and/or paleoseismic records (e.g., Matthews, 2002; Morell et al., 2020; Schwartz & Coppersmith, 1984; Sornette & Knopoff, 1997; Thenhaus & Campbell, 2002; S.-C. Wu et al., 1995; Youngs & Coppersmith, 1985). However, the assumption of steady recurrence in periodic or quasi-periodic models is inconsistent with irregular recurrence intervals documented in numerous tectonic settings (e.g., Weldon et al., 2004; Goldfinger et al., 2013; Mouslopoulou et al., 2014; Dolan et al., 2016; Koukouvelas et al., 2017; Zinke et al., 2017, 2021, 2019; Philibosian & Meltzner, 2020; Hatem et al., 2020; Salditch et al., 2020; DuRoss et al., 2020; Bennett et al., 2018). Faults with irregular recurrence intervals may be described better through non-periodic recurrence-frequency models. The primary non-periodic recurrence-frequency models are the time-predictable model (Shimazaki & Nakata, 1980), the slip-predictable model (Shimazaki & Nakata, 1980), and the long-term fault memory model (e.g., Salditch et al., 2020).

The forearc of the Cascadia Subduction Zone (CSZ, Figure 4.1) has long been a focus of probabilistic seismic hazard assessments, primarily using time-independent recurrence models (e.g., Atkinson & Charlwood, 1983; Harmsen, 2001; Frankel et al., 2015; Petersen et al., 2020; Hobbs et al., 2023). However, historical records include fewer than 15 $M > 5$ on CSZ forearc faults since 1850, which means that earthquake frequency for these faults must be estimated from paleoseismic records. Some paleoseismic records suggest that CSZ forearc faults exhibit non-steady recurrence and slip rates. For example, a cluster of earthquakes occurred on eight different forearc faults in the Puget Lowland (Figure D.1) around 1000 cal yrs BP that has been dubbed the millennial earthquake series (e.g., Nelson et al., 2014) and the Puget Sound regional cluster (e.g., Sherrod & Gombert, 2014). Recent studies have documented similar patterns of non-steady recurrence and slip rates in the Holocene between three separate CSZ forearc faults spanning nearly 300 km. The Beaufort Range fault (BRF) and Leech River fault (LRF) on Vancouver Island both exhibit changing recurrence intervals, with one earthquake in the late Pleistocene to early Holocene, a ~ 5 -7 kyr period of quiescence, and then multiple mid-late Holocene earthquakes (Morell et al., 2018, ; Chapter 3). Nelson et al. (2014) observed a similar phenomenon of varying recurrence on the Seattle fault, noting a ~ 6 -8 kyr quiet period before a series of late Holocene earthquakes. These observations raise the question if these are isolated examples or if non-steady recurrence is typical of all CSZ forearc faults. In this study, I use paleoseismic earthquake records to address the following questions (1) Are recurrence and slip rates on CSZ forearc faults steady or non-steady? (2) If non-steady, can their behavior be described by the time- or slip-predictable models typically used to describe faults with non-steady recurrence? (3) What external tectonic or climatic factors

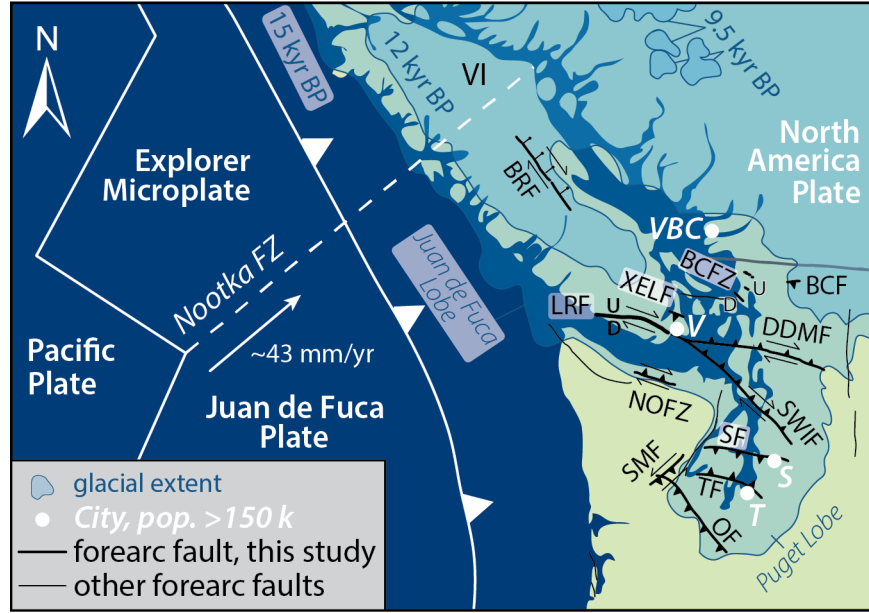


Figure 4.1: Regional tectonic setting showing the locations of active forearc faults in the northern Cascadia forearc of Canada and the USA. Also shown is the extent of Cordilleran Ice Sheet coverage during retreat, at 15 kyr BP, 12 kyr BP, and 9.5 kyr BP, after Clague et al. (1989). Juan de Fuca-North America convergence vector after Kreemer et al. (2014). Active faults in the US after U.S. Geological Survey (n.d.), Leech River fault after Morell et al. (2017), Beaufort Range fault after Chapter 2. BCF - Boulder/Canyon Creek fault; BRF - Beaufort Range fault; DDMF - Darrington-Devils Mountain fault; LRF - Leech River fault; NOFZ - North Olympic fault zone; OF - Olympia fault; S - Seattle, WA; SF - Seattle fault; SMF - Saddle Mountain fault; SPF - Sandy Point fault; SWIF - Southern Whidbey Island fault; T - Tacoma, WA; TF - Tacoma fault; V - Victoria, BC; VBC - Vancouver, BC; VI - Vancouver Island; XELF - ~~XEOLXELEK~~-Elk Lake fault.

could be affecting the recurrence on these forearc faults?

I use new and existing compilations of earthquake timing and displacements on thirteen forearc faults and the CSZ subduction zone to examine earthquake recurrence and interval slip rates. Paleoseismic records on forearc faults in the northern CSZ extend back to ~15 ka, when most of the region was under the Cordilleran Ice Sheet (e.g., Clague & James, 2002; Ryder et al., 2007; Booth et al., 2003). Recurrence and slip rate data exhibit non-steady patterns that are inconsistent with periodic, time-predictable, and slip-predictable recurrence models. The apparently systematic variability in recurrence and slip rate that we observe on numerous forearc faults suggests there may be a regional change in strain accumulation caused by some external tectonic or climatic factors affecting loading on forearc faults, such as interactions between faults, impacts from the CSZ, and glacial isostatic adjustment.

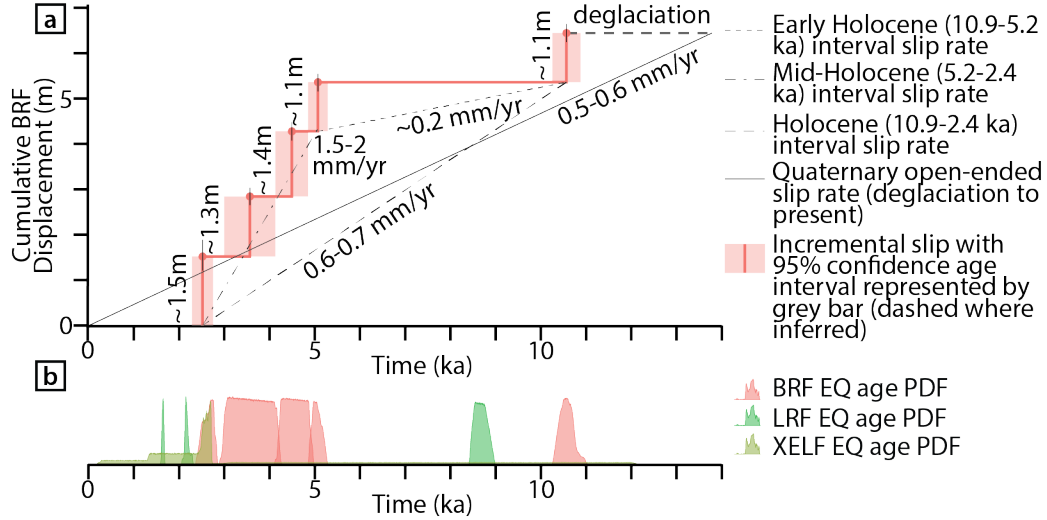


Figure 4.2: Non-steady slip and recurrence on Vancouver Island faults in the Holocene. **A:** Cumulative displacement on the BRF from five discrete events. The solid line shows the 0.5-0.6 mm/yr Quaternary open-ended slip rate, while the dashed and dotted lines show several closed-interval slip rates. The early Holocene slip rate of ~ 0.2 mm/yr is an order of magnitude slower than the mid-Holocene slip rate of 1.5-2 mm/yr. **B:** Timing of individual events on Vancouver Island Quaternary-active faults. Probability density functions (PDFs) for each event were created using OxCal modeling (Chapter 3; Morell et al., 2018; Har-richhausen et al., n.d., *in review*). The BRF and LRF both have one event in the late Pleistocene to early Holocene, followed by a long quiet period (~ 5 -6 kyr on BRF, ~ 6 -7 kyr on LRF) and then several events in the mid-late Holocene. The XELF has one event with a wide range of possible ages, with a strong peak at ~ 2.8 cal ka.

4.2 Background

4.2.1 Models for earthquake recurrence

There are two main categories of earthquake recurrence models used in probabilistic seismic hazard analysis (PSHA)—magnitude-frequency or time-independent models, and recurrence-frequency or time-dependent models—that rely on different input parameters and assumptions (Figure 4.2.1; Thenhaus & Campbell, 2002). In this study, I consider whether these models describe the behavior of CSZ forearc faults and whether their assumptions are valid in this region.

The first main category, time-independent models, assumes that the timing of future earthquakes is unrelated to the timing of any past earthquakes (Thenhaus & Campbell, 2002; Morell et al., 2020; Salditch et al., 2020). There are four main time-independent models that each use the assumption that earthquake frequency does not vary with time (Figure 4.2.1). (1) The most commonly used model in PSHA is the Poisson distribution, which is based on the mean and variance of an annual instrumental seismicity rate (Thenhaus & Campbell, 2002; Salditch et al., 2020). (2) The exponential model, developed by Richter

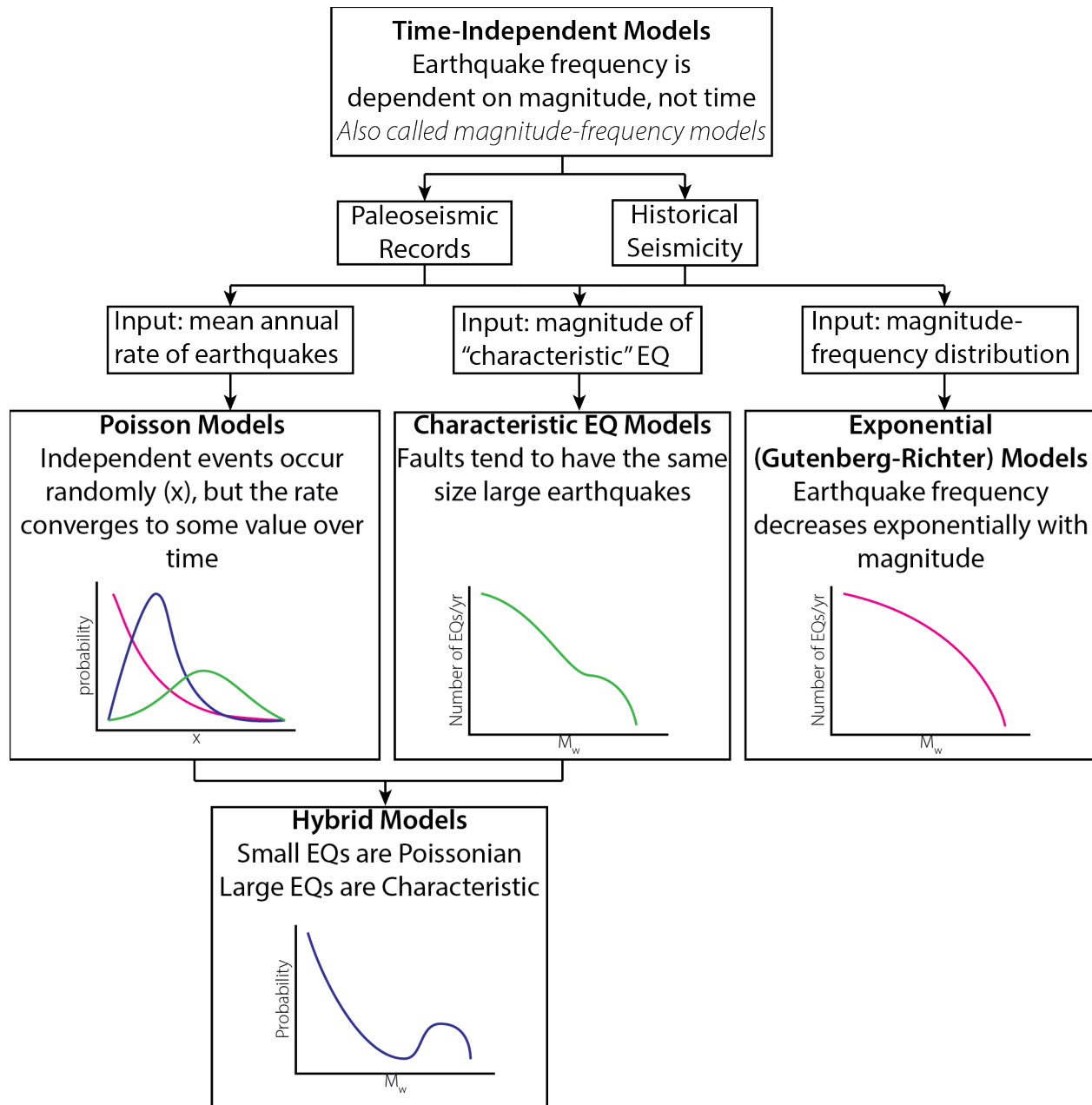
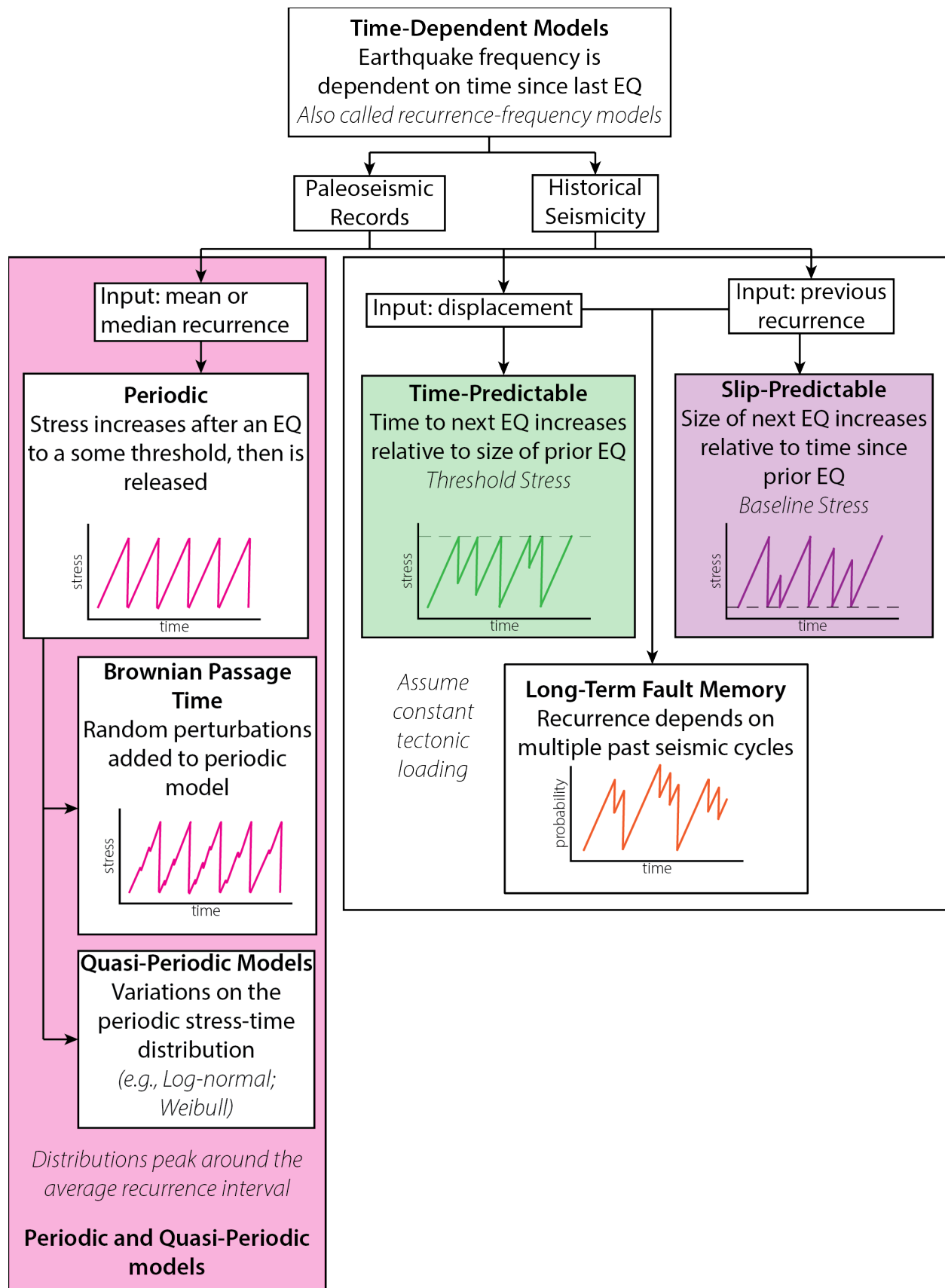


Figure 4.3: Types of earthquake recurrence models, their inputs, and assumptions. **A:** Time-independent models. These models are not evaluated in this study. Cartoon frequency plots are modified from Youngs and Coppersmith (1985) and S.-C. Wu et al. (1995). **B:** Time-dependent models. Only the models in colored boxes are evaluated in this study: periodic models, the time-predictable model, and the slip-predictable model. Cartoon plots are modified from Shimazaki and Nakata (1980); Matthews (2002); Salditch et al. (2020).

and Gutenberg (1954) and updated by Cornell and Vanmarcke (1969), assumed the probability of rupture decreases exponentially relative to rupture magnitude and is used primarily to describe the behavior of regions



with many faults rather than individual faults (Schwartz & Coppersmith, 1984; Wesnousky, 1994; Thenhaus & Campbell, 2002). (3) The characteristic earthquake model, which is typically applied to individual faults. This model posits that faults tend to have earthquakes of similar (or characteristic) magnitudes (Schwartz & Coppersmith, 1984; Youngs & Coppersmith, 1985). (4) Finally, hybrid models, such as those developed by S.-C. Wu et al. (1995) assume that smaller earthquakes on a fault follow a Poissonian distribution, while the larger characteristic earthquakes follow a separate distribution. For the purposes of this study, I group these four models together into a category called “time-independent models” (Figure 4.2.1).

The second main category is time-dependent or renewal models, in which the probability of rupture is related to the time since the most recent event (Thenhaus & Campbell, 2002). The most basic of these renewal models is periodic recurrence, where stress increases following an earthquake until it hits some threshold where the fault ruptures and releases that accrued stress, following the elastic rebound theory developed by Reid (1910). A modification of the periodic recurrence model is the Brownian Passage Time model of Matthews (2002), which adds random perturbations to the stress accrual. Other periodic and quasi-periodic models using mean and/or median recurrence intervals and variance have been developed (e.g., Uniform, Semi-Gaussian, lognormal, Weibull, power-law, Gamma; Sornette & Knopoff, 1997; Thenhaus & Campbell, 2002; Morell et al., 2020). Importantly, these recurrence models consider only the time since the most recent event and a representative recurrence interval and are independent of slip or earlier events on that fault (Thenhaus & Campbell, 2002). For the purposes of this study, I consider the many periodic and quasi-periodic models together as a single category (Figure 4.2.1).

Within the category of time-dependent models, there are also non-periodic models (Figure 4.2.1). The three main models in this category include the time-predictable model, the slip-predictable model, and the long-term fault memory model. These non-periodic models incorporate additional parameters beyond the mean recurrence interval. Shimazaki and Nakata (1980) proposed two recurrence models: a time-predictable model in which faults rupture at some threshold stress such that the length of a recurrence period increases proportionally to the slip in the prior earthquake and a slip-predictable model in which rupture returns faults to some stress level such that the size of an earthquake increases proportionally to the length of time since the prior earthquake. Time- and slip-predictable models can have varying recurrence interval lengths, but the assumption of a constant stress accumulation rate means that these models result in short-term slip rates that match long-term slip rates (Shimazaki & Nakata, 1980). Weldon et al. (2004) proposed another model, which they termed strain-predictable, in which the length of a recurrence period increases proportionally to the “relative strain level” after the prior earthquake. They used cumulative slip relative to the long-term slip rate as a proxy for strain. Similar behavior has been termed “supercycles” or a long-term fault memory (LTFM) model by subsequent work (e.g., Goldfinger et al., 2013; Salditch et al., 2020). In these models,

earthquakes do not necessarily release all of the accumulated strain (Salditch et al., 2020). As a result, the probability of an earthquake does not necessarily return to zero after rupture (Salditch et al., 2020). The strain-predictable, or LTFM model, assumes that the strain accumulation rate, or background tectonic loading, is constant through time but that the short-term slip rate can vary considerably (Weldon et al., 2004; Salditch et al., 2020). Unfortunately, we don't know the background tectonic loading rates of CSZ forearc faults. If differential GPS rates for individual faults could be isolated from the signal of CSZ interseismic coupling similar to faults in southern California (e.g., Loveless & Meade, 2011; Titus et al., 2011; Evans et al., 2015), we could then assess if CSZ forearc faults are consistent with the LTFM model. For the purposes of this study, I consider only the time- and slip-predictable non-periodic models (Figure 4.2.1).

4.2.2 Earthquake clustering in paleoseismic records

Earthquake clustering occurs frequently in paleoseismic records. This has been observed in numerous tectonic settings, including rifts (e.g., Mouslopoulou et al., 2009) and strike-slip settings (e.g., Grant & Sieh, 1994; Weldon et al., 2004; Marco et al., 1996; Dolan et al., 2016; Oskin et al., 2007). At convergent boundaries, earthquake clustering has been documented both on the megathrust (e.g., Goldfinger et al., 2013; Sieh et al., 2008), and in the forearc (e.g., Schlagenhauf et al., 2011; Zinke et al., 2017, 2019; Hatem et al., 2020; Zinke et al., 2021; Mouslopoulou et al., 2014). The limited time lengths of paleoseismic records can make it difficult to determine whether paleo-earthquakes are clustered in time or random. Williams et al. (2019) considered 31 earthquake chronologies of five or more earthquakes and found that the fewer earthquakes on record for a fault, the less likely it was to be described by periodic or quasi-periodic recurrence. Non-steady recurrence has been proposed to be caused by external tectonic or climatic factors. Tectonic factors may include changes in the rate of subduction (e.g., S. Li et al., 2020), Coulomb stress transfer from neighboring faults (e.g., King et al., 1994; Stein, 2003, 1999), changes in interseismic strain accumulation on the subduction interface (e.g., Wang, 2000; Loveless et al., 2010; Symithe et al., 2015; Delano et al., 2017), or co-seismic triggering from subduction megathrust events (e.g., Farías et al., 2011; Toda, Stein, & Lin, 2011; Imanishi et al., 2012; McCloskey et al., 2005). A climatic factors that has affected the timing of crustal fault rupture elsewhere is glacial isostatic adjustment (e.g., Brandes et al., 2011; Olesen, 1988; Lagerbäck, 1990; Lund & Näslund, 2009; Hetzel & Hampel, 2005; Hampel et al., 2007; Hampel, 2017). Below, I discuss the major tectonic and climatic events in the northern CSZ forearc that could potentially impact recurrence on forearc faults.

4.2.3 Evolution and rupture history of the Cascadia Subduction Zone interface

The Cascadia Subduction Zone (CSZ) is the plate boundary where the Juan de Fuca plate subducts obliquely under North America at ~ 43 mm/yr (Figure 4.1; Kreemer et al., 2014). The obliquity of subduction affects how strain is partitioned between the plate interface and the upper plate, and therefore where forearc faults form (e.g., Fitch, 1972; McCaffrey, 1992; ten Brink & Lin, 2004). The Juan de Fuca plate has been subducting underneath North America since ~ 28 Ma (McCroory & Wilson, 2013), and the northern terminus of this boundary has migrated slowly northward since 15 Ma (Madsen et al., 2006). Around 4 Ma, the buoyant northern portion of the Juan de Fuca plate began to resist subduction, resulting in the formation of the Explorer microplate, separated from the Juan de Fuca plate by the left-lateral Nootka fault zone (Figure 4.1; Audet et al., 2008). The Explorer microplate may be partially coupled with the Pacific plate, but it is still partially subducting under North America (Figure 4.1; Riddihough, 1977; Madsen et al., 2006). There have been no significant changes in plate configuration or geometry since ~ 4 Ma, such that convergence rate and obliquity have likely been steady over the past ~ 15 ka.

The CSZ is presently locked, accumulating both strain on the plate interface that will be recovered in megathrust earthquakes and strain in the upper plate that will be accommodated through permanent deformation on forearc faults (e.g., Miller et al., 2001; McCaffrey et al., 2007; Wang, 2007; Delano et al., 2017; McKenzie et al., 2022). The CSZ has ruptured approximately 42 times in the past 10 kyrs (Goldfinger et al., 2013; Walton et al., 2021). Of the 42 ruptures documented by Goldfinger et al. (2012), only 19-20 have ruptured the full margin, as far north as Vancouver Island (Table 4.1). These full-margin ruptures were $M_W \sim 8.5-9.1$ and estimates of per-event slip range from 4.3 to 43.5 m. The oldest documented full-margin rupture occurred ~ 9.8 ka, while the most recent CSZ megathrust rupture occurred in 1700 CE (Goldfinger et al., 2012; Yamaguchi et al., 1997). The evidence for these events primarily comes from the marine turbidite record, with additional on-shore data from tsunami deposits, microfossil-based estimates of coseismic subsidence, liquefaction studies, and lacustrine turbidite deposits (Goldfinger et al., 2012; Walton et al., 2021).

4.2.4 Glacial history of the CSZ forearc

Glacial loading and subsequent unloading affect the stress state in the lithosphere (e.g., Hetzel & Hampel, 2005; Hampel & Hetzel, 2006; Lund & Näslund, 2009; Hampel et al., 2009, 2010; Steffen et al., 2014), and can suppress or promote slip on crustal faults (e.g., Lagerbäck, 1990; Muir-Wood, 2000; Sauber et al., 2000; Brandes et al., 2011; Hampel, 2017), and therefore modify the length of earthquake recurrence. The Cordilleran Ice Sheet (CIS) covered Western Canada and parts of Washington state during the last glacial

period (Cordilleran Fraser glaciation, e.g., Booth et al., 2003). The Puget and Juan de Fuca lobes (see Figure 4.1) reached their maximum extent after 16 cal ka BP (Haugerud, 2021). Glacial extents at 15, 12, and 9.5 ka BP are shown in Figure 4.1 (after Clague et al., 1989). The ice surface of the Puget and Juan de Fuca lobes at glacial maximum was >2,000 m above sea level (asl) at their highest (Booth, 1986), 1035 m asl above Seattle, 770 m asl above Tacoma, and 465 m asl above Olympia (Figure 4.1; Thorson, 1980). Paleogeographic reconstructions suggest sea level in the eastern Strait of Juan de Fuca was ~ 75 m higher at the last glacial maximum (LGM) than today, falling to a low stand ~ 20 -60 m lower than today around ~ 9 -10 ka BP (Mosher & Hewitt, 2004; James et al., 2009). Since ~ 5.5 ka BP, crustal uplift has more or less matched eustatic sea level rise (Mosher & Hewitt, 2004). Postglacial rebound models suggest the deglaciation-related uplift has decreased from as fast as >100-200 mm/yr immediately post-glacial to <10 mm/yr at 10 cal ka BP, and is <0.1-0.8 mm/yr today (James et al., 2000, 2005; Clague & James, 2002). Postglacial rebound, or glacial isostatic adjustment (GIA) can suppress or promote slip on crustal faults due to changes in vertical loading stress and horizontal flexure (e.g., Hetzel & Hampel, 2005; Hampel et al., 2007; Steffen et al., 2014; Hampel, 2017).

4.2.5 Crustal faults of the Cascadia forearc

Over the past ~ 50 years, more than 20 active faults have been identified in the northern CSZ forearc. Of these active faults, 13 have paleoseismic data that can be used to evaluate recurrence and slip rates. These faults are located in the northern Cascadia forearc, from the Olympia fault in the southern Puget Lowland (Figures 4.1, D.1) to the Beaufort Range fault on Vancouver Island, the farthest north documented-active CSZ forearc fault. There are age estimates for earthquakes on all of these faults, and the majority have documented recurrence and displacement estimates.

The **Beaufort Range fault (BRF)** is located on central Vancouver Island, along the southwestern flank of the Beaufort Range (Figure 4.1; Chapter 2). It is the northernmost active fault in the CSZ forearc, ~ 60 km Southwest of the northern terminus of the Juan de Luca slab within the accreted Wrangellia terrane (Chapter 2). The BRF is >60 km long and strikes $\sim 260^\circ$ - 309° , with individual fault strands 100-1500 m long (Chapter 2; Chapter 3). Both steeply ($>70^\circ$) NE- and SW-dipping strands make up a ~ 500 m-wide zone that accommodates right-lateral transtension, as exhibited by vertical (NE-down) and right-lateral displacement of glacial till, and paraglacial and post-glacial deposits and landforms across ~ 1 -6 m-tall scarps (Chapter 2; Chapter 3). Kinematic inversions show a strike slip to dip slip ratio of 0.9:1 (Chapter 2; Chapter 3). One trench has been completed on the southern BRF, which revealed evidence that the BRF has hosted at least five earthquakes since the late Pleistocene (Chapter 3), the ages of which are included in Table 4.1. The

BRF may have hosted the 1946 Ms 7.3 Vancouver Island earthquake (Rogers & Hasegawa, 1978; Slawson & Savage, 1979), however, this was not resolved in the trench, which recorded a most recent event that most likely occurred ~ 2.8 -2.4 cal kyr BP (Table 4.1; Chapter 3). Cumulative oblique displacements on the BRF range from 4-21 m measured from surface displacements (Chapter 2), and 7.6 ± 0.3 m measured from displacements in the trench (Chapter 3). Relative magnitudes of dip slip in each event (~ 0.6 -1.1 m) and right-lateral slip in each event (~ 0.5 -1 m) are consistent with the ratio of strike slip to dip slip estimated from surface displacements and result in oblique displacements of 0.75-1.54 m per event (Chapter 3). Displacement estimates for individual events are listed in Table 4.2. Slip rates on the BRF range from ~ 0.5 -0.7 mm/yr since deglaciation (Chapter 2; Chapter 3) to >2 mm/yr in the mid-late Holocene (Chapter 3).

The **Leech River fault (LRF)** is located on southern Vancouver Island, approximately 100 km south of the BRF (Figure 4.1). The LRF is >60 km long and strikes $\sim$ E-W in the western part of the island, and then turns to be more NW-SE ($\sim 130^\circ$ - 150°) west of Victoria (Figure 4.1; Morell et al., 2017; Harrichhausen et al., 2021). Individual fault strands are 100s of m to >2.5 km long, and form <5 m-tall N- and S-facing scarps in a zone ~ 1 km wide (Morell et al., 2017). Fault exposures are steeply N-dipping, and vertical separation of the hillslope suggests dip slip on the LRF (Morell et al., 2018). However, relocated microseismicity is consistent with right-lateral kinematics (G. Li et al., 2018), which is further supported by sub-horizontal slickenlines in a trench exposure, and the presence of pressure ridges and tension fractures, despite the lack of clear lateral offset of geomorphic features (Morell et al., 2018, 2017; Harrichhausen et al., 2021). Harrichhausen et al. (2021) interpreted the LRF as an oblique right-lateral fault consistent with bending of the Olympic orocline or extrusion of the Olympic peninsula (e.g., Nelson et al., 2017; Finley et al., 2019). Four trenches have been completed on the LRF, exposing evidence of at least three earthquakes since the late Pleistocene (Morell et al., 2018; Harrichhausen et al., 2021). One site experienced three earthquakes since ~ 9 ka, whereas ~ 7 km east, another site experienced only one earthquake 9.4 ± 3.4 ka (Morell et al., 2018; Harrichhausen et al., 2021). The ages of these earthquakes are reported in Table 4.1. Cumulative displacements from vertical separation of a <15 ka colluvial surface are ~ 4 -6 m, slightly more than the ~ 1.1 -3 m apparent dip slip measured in trench exposures, a discrepancy that may be the result of off-fault deformation and folding (Morell et al., 2018; Harrichhausen et al., 2021). Estimates of per-event vertical displacement range from $>0.6 \pm 0.4$ m to 1.2 ± 0.5 m (Table 4.2; Morell et al., 2018; Harrichhausen et al., 2021). The slip rate on the LRF is ≥ 0.2 -0.3 mm/yr since ~ 9 ka (Morell et al., 2018). It is important to note that per-event displacements, and therefore slip rates, on the LRF are likely minima. These estimates are based only on dip slip and do not include any off-fault deformation or right-lateral displacement.

The third documented active fault on Vancouver Island is the ~~**XEOLXELEK**~~-**Elk Lake fault (XELF)** on the Saanich Peninsula, ~ 10 km north of the Victoria city center (Figure 4.1; Harrichhausen et al., n.d.,

in review). The XELF is <15 km long and strikes WNW ($\sim 127^\circ$), with several NE-facing scarps offsetting Quaternary sediments and landforms (Harrichhausen et al., n.d., *in review*). The straight trace of the fault suggests a sub-vertical dip, however a three-point problem and ERT profiles suggest the XELF dips moderately, $\sim 50^\circ \pm 5^\circ$ toward the SW (Harrichhausen et al., n.d., *in review*). The XELF accommodates compression through reverse faulting and a fault-propagation fold (Harrichhausen et al., n.d., *in review*). One trench has been completed on the XELF, which revealed evidence of one earthquake in the late Pleistocene to Holocene (Harrichhausen et al., n.d., *in review*). There are multiple plausible earthquake ages, so we use the age determined from the mean PDF of multiple geologically plausible age models (Harrichhausen et al., n.d., *in review*). Because of the long tail of the mean PDF and mid-late Holocene mode (Figure 4.2b), we report only the 6% age range (Table 4.1). Trishear modeling of the fault-propagation fold exposed in the trench shows ~ 3.2 m of dip slip in a single earthquake (Harrichhausen et al., n.d., *in review*). A maximum slip rate for the XELF is ~ 0.34 mm/yr (Harrichhausen et al., n.d., *in review*).

Table 4.1: Late Pleistocene and Holocene earthquakes on northern Cascadia forearc faults and the CSZ megathrust.

Main Fault ¹	Strand/Site Name	EQ name ²	Maximum Age (yrs BP)	Minimum Age (yrs BP)	Source	Comments
Forearc events						
BRF	Woodlot Trench	Event 1	10900	10300	Chapter 3	
BRF	Woodlot Trench	Event 2	5200	4910	Chapter 3	
BRF	Woodlot Trench	Event 3	4900	4210	Chapter 3	
BRF	Woodlot Trench	Event 4	4200	3000	Chapter 3	
BRF	Woodlot Trench	Event 5	2800	2400	Chapter 3	
LRF	Good Hope Trench	GH	12800	6000	Harrichhausen et al., 2021	
LRF		Event 3	9000	8400	Morell et al., 2018	
LRF		Event 2	2300	2100	Morell et al., 2018	
LRF		Event 1	1800	1600	Morell et al., 2018	
XELF		XELF	4700	2300	Harrichhausen et al., <i>in review</i>	68% interval due to very long tail
BCFZ	Sandy Point fault	SPF-O	15000 ³	2100	Kelsey et al., 2012	
BCFZ	Sandy Point fault	SPF-M	2100	100 ⁴	Kelsey et al., 2012	

Table 4.1: (continued)

Main Fault ¹	Strand/Site Name	EQ name ²	Maximum Age (yrs BP)	Minimum Age (yrs BP)	Source	Comments
BCFZ	Sandy Point fault	SPF-Y	2100	100 ⁴	Kelsey et al., 2012	
BCFZ	Birch Bay fault	BB	1280	1070	Kelsey et al., 2012	
BCF	Boulder Creek fault/ Hornet trench	EQ A	8050	7250	Sherrod et al., 2013	
BCF	Boulder Creek fault/ Hornet trench	EQ B	3190	2980	Sherrod et al., 2013	
BFC	Boulder Creek fault/ Hornet trench	EQ C	910	740	Sherrod et al., 2013	
DDMF		CE2	8200	8000	Personius et al., 2014	
DDMF		CE1	2300	2100	Personius et al., 2014	Using charcoal ages
DDMF	Utsalady Point fault	Duffers 1	2200	1200	Johnson et al., 2004	
DDMF	Utsalady Point fault	Duffers 2	500	100 ⁴	Johnson et al., 2004	
SWIF		FO		16400	Sherrod et al., 2008	Occurred before last glacial advance, no maximum age
SWIF		BB1	16400	11670	Sherrod et al., 2008	Occurred before BB2
SWIF		BB/MB	11670	2850	Sherrod et al., 2008	
SWIF		CH	3800	3200	Kelsey et al., 2004	
SWIF		BB3	2700	100§	Sherrod et al., 2008	
SF	Vasa Park	Vasa	12880	12310	Sherrod, 2002; Sherrod and Gomberg, 2014	
SF	Restoration Point	EQ A	7200	6900	Sherrod et al., 2000	
SF		EQ B	2650	1940	Nelson et al., 2014	
SF		EQ C	1350	1170	Nelson et al., 2014	

Table 4.1: (continued)

Main Fault ¹	Strand/Site Name	EQ name ²	Maximum Age (yrs BP)	Minimum Age (yrs BP)	Source	Comments
SF		EQ D	1040	910	Nelson et al., 2014	
SF		EQ E	940	380	Nelson et al., 2014	
SF/TJF	Toe Jam Hill fault/ Crane Lake	Event A	6000	2500	Nelson et al., 2003	
SF/TJF	Toe Jam Hill fault/ Crane Lake	EQ B	2500	1900	Nelson et al., 2003	
SF/TJF	Toe Jam Hill fault/ Crane Lake	EQ C	1700	1200	Nelson et al., 2003	Correlated with SF EQ C by Nelson et al., 2014
SF/TJF	Toe Jam Hill fault/ Mossy Lane	EQ D	1200	100 ⁴	Nelson et al., 2003	Correlated with SF EQ D by Nelson et al., 2014
TF		EQ D	1050	980	Nelson et al., 2014	
OF	Little Skookum	OLY	1150	1010	Sherrod, 2001]	
NOFZ	Sadie Creek fault	GB1	12200	10000	Schermer et al., 2021	
NOFZ	Sadie Creek fault	GB2	8700	4100	Schermer et al., 2021	
NOFZ	Sadie Creek fault	GB3	3600	3100	Schermer et al., 2021	Age from four-event model
NOFZ	Sadie Creek fault	GB4	2500	300	Schermer et al., 2021	
NOFZ	Sadie Creek fault	BL1	12200	10300	Schermer et al., 2021	
NOFZ	Sadie Creek fault	BL2	8900	8100	Schermer et al., 2021	
NOFZ	Sadie Creek fault	BL3	7000	6400	Schermer et al., 2021	
NOFZ	Sadie Creek fault	BL4	6200	2300	Schermer et al., 2021	
NOFZ	Sadie Creek fault	BL5	1200	900	Schermer et al., 2021	
NOFZ	Lake Creek-Boundary Creek fault	EQ A	12000	8000	Nelson et al., 2017	
NOFZ	Lake Creek-Boundary Creek fault	EQ B	8000	5000	Nelson et al., 2017	

Table 4.1: (continued)

Main Fault ¹	Strand/Site Name	EQ name ²	Maximum Age (yrs BP)	Minimum Age (yrs BP)	Source	Comments
NOFZ	Lake Creek-Boundary Creek fault	EQ C	3500	2300	Nelson et al., 2017	
NOFZ	Lake Creek-Boundary Creek fault	EQ D	2100	500	Nelson et al., 2017	
SMF	Saddle Mountain West	E1	15000 ³	7500	Witter et al., 2008; Barnett et al., 2015	
SMF	Sund Creek fault	E2	3700	1000	Barnett et al., 2015	
SMF	All strands	E3	1170	970	Barnett et al., 2015	
Megathrust events						
CSZ		T18	9979	9563	Goldfinger et al., 2012	
CSZ		T17a	9429	8989	Goldfinger et al., 2012	
CSZ		T17	9390	8810	Goldfinger et al., 2012	
CSZ		T16	9066	8761	Goldfinger et al., 2012	
CSZ		T15	8356	8038	Goldfinger et al., 2012	
CSZ		T14	7763	7487	Goldfinger et al., 2012	
CSZ		T13	7304	7062	Goldfinger et al., 2012	
CSZ		T12	6612	6333	Goldfinger et al., 2012	
CSZ		T11	6100	5824	Goldfinger et al., 2012	
CSZ		T10	4940	4579	Goldfinger et al., 2012	
CSZ		T9	4278	3918	Goldfinger et al., 2012	
CSZ		T8	3596	3287	Goldfinger et al., 2012	
CSZ		T7	3162	2865	Goldfinger et al., 2012	
CSZ		T6	2673	2389	Goldfinger et al., 2012	
CSZ		T5	1731	1384	Goldfinger et al., 2012	
CSZ		T4	1348	1119	Goldfinger et al., 2012	
CSZ		T3	905	679	Goldfinger et al., 2012	

Table 4.1: (continued)

Main Fault ¹	Strand/Site Name	EQ name ²	Maximum Age (yrs BP)	Minimum Age (yrs BP)	Source	Comments
CSZ		T2	573	384	Goldfinger et al., 2012	
CSZ		T1	371	139	Goldfinger et al., 2012	

¹See Figure 4.1 for fault abbreviations and locations

²Event names included in original publications or earlier compilations were kept where possible.

³Events with maximum ages of 15,000 yrs BP had no published maximum age. I have imposed a maximum age of 15,000 yrs BP, corresponding with the maximum age of deglaciation (e.g., Haugerud, 2021)

⁴Events with minimum ages of 100 yrs BP had no published minimum age. I have imposed a minimum age of 100 yrs BP, corresponding with 1850 CE.

The **Bellingham Coastal fault zone (BCFZ)** is a group of several faults located between Bellingham, Washington, and the USA-Canada border (Figure 4.1). The Drayton Harbor fault is nearest to the border and consists of >8 km of S-side-up scarps and disrupted drainages (Kelsey et al., 2012). The Birch Bay and Sandy Point faults are southwest of the Drayton Harbor fault and were identified through magnetic anomalies that matched up with scarps identified on lidar (Kelsey et al., 2012). The Birch Bay fault is buried, however uplift of a beach platform is consistent with one Holocene earthquake (Kelsey et al., 2012). Radiocarbon ages from a core collected from the raised beach platform indicate this event occurred in the late Holocene (Kelsey et al., 2012). The Sandy Point fault trends toward 310°, and extends for ~12 km (Kelsey et al., 2012). Uplifted shorelines and tide flats are consistent with dip slip, however, it is unclear if the slip is reverse or normal (Kelsey et al., 2012). Three uplifted tide flats are consistent with three late Pleistocene to Holocene earthquakes on the Sandy Point fault (Kelsey et al., 2012). Radiocarbon ages from a soil pit in the tide flats are minimal, so the earthquake ages are poorly constrained. The two younger events likely occurred after 2100 cal yrs BP, while the older event occurred before 2100 cal yrs BP (Table 4.1; Kelsey et al., 2012). The ages of these BCFZ earthquakes are reported in Table 4.1. The oldest tide flat has experienced ~4 m of cumulative uplift Table 4.1;. The relative uplift of the younger tide flats indicates that there was 2-2.5 m of uplift during the oldest event, and ≤ 1 m of uplift during the younger two events Table 4.1;).

The **Boulder and Canyon Creek faults (BCF)** are located northwest of Mount Baker in the foothills of the North Cascades (Figure 4.1). Together, the faults are ~12 km long and bound the eastern edge of the Bellingham Basin (Sherrod et al., 2013). The BCF is shallowly (~30°) south-dipping, and forms ~2-4 km-long S-side-up scarps up to ~6.9 m tall (Sherrod et al., 2013). The BCF accommodates reverse and/or dextral oblique slip, with the lateral component occurring primarily on the Canyon Creek segment to the east (Sherrod et al., 2013). Two trenches have been excavated and many cores have been collected

on the BCF, which revealed evidence that the BCF has hosted three earthquakes during the Holocene (Sherrod et al., 2013), the ages of which are reported in Table 4.1. These ages are well-constrained by both numerous radiocarbon samples and the presence of Mazama ash in cores (Sherrod et al., 2013). Cumulative displacement of 3.4-6.9 m from vertical separation of scarps and terraces is consistent with ~ 0.25 -1.7 m vertical separation per event measured in the trenches (Sherrod et al., 2013). Displacement estimates for individual events are listed in Table 4.2. Slip rates on the BCF range from interval rates of 0.3 ± 0.1 to 0.5 ± 0.2 mm/yr to a long-term rate of 0.65 ± 0.1 mm/yr from terrace displacement (Sherrod et al., 2013).

The **Darrington-Devils Mountain fault zone (DDMF)** is a group of faults located offshore Victoria, BC, crossing Puget Sound, and extending into the foothills of the North Cascades (Figure 4.1; Personius et al., 2014). It is >125 km long and forms the northern boundary of the Everett basin, between Mesozoic and Upper Mesozoic accreted terranes (Personius et al., 2014). The DDMF strikes $\sim 286^\circ$, with ~ 0.5 -1 km-long fault scarps within a ~ 6 -10 km-wide zone of main and conjugate faults, including the Utsalady Point and Strawberry Point faults (Johnson, Nelson, et al., 2004; Personius et al., 2014; Barrie & Greene, 2015). The DDMF dips $>30^\circ$ N to vertical, with both N- and S-side-up scarps identified on lidar (Johnson, Nelson, et al., 2004; Personius et al., 2014; Barrie & Greene, 2015, 2018). Johnson et al. (2001) interpreted the DDMF as a transpressive left-lateral fault based on NW-striking en echelon faults, however later work has identified right-lateral offsets and conjugate faults through surficial mapping, paleoseismic trenching, and offshore seismic reflection (Personius et al., 2014; Barrie & Greene, 2015, 2018). The DDMF most likely accommodates right-lateral transpression through N-side-up reverse faulting (Personius et al., 2014). The Utsalady Point fault (UPF), however, does accommodate left-lateral motion on a sub-vertical NW-striking fault ($\sim 305^\circ$; Johnson, Nelson, et al., 2004). The UPF is south of the main DDMF fault, where it extends >28 km across northern Whidbey Island (Johnson, Nelson, et al., 2004). Personius et al. (2014) estimated a strike slip to dip slip ratio of $\sim 5:1$ on the main strand of the DDMF. Seven trenches and a coring transect have been completed on the DDMF, including two trenches on the UPF (Johnson, Nelson, et al., 2004; Personius et al., 2014). These studies have identified two Holocene earthquakes each on the main strand of the DDMF and the UPF (Johnson, Nelson, et al., 2004; Personius et al., 2014). The ages for these events (see Table 4.1) were constrained by detrital charcoal, wood, microfossil plant and animal remains, and Mazama ash (Johnson, Nelson, et al., 2004; Personius et al., 2014). However, extensive burrowing may have obscured evidence of additional earthquakes in the main strand trenches (Personius et al., 2014). Relative magnitudes of dip slip (~ 0.4 -0.5 m) and strike slip (~ 1.1 -2.8 m) in the main strand trenches are consistent with predominantly right-lateral transpression and result in $\sim 2.3 \pm 1.1$ m of cumulative oblique displacement on the main strand, comparable to the ~ 2.2 -2.65 m oblique displacement estimated for the UPF (Johnson, Nelson, et al., 2004; Personius et al., 2014). This corresponds to 1.15 m oblique slip per

event (Personius et al., 2014). Johnson et al. (2001) estimated a vertical slip rate of 0.05-0.31 mm/yr, which is consistent with the 0.14 ± 0.1 mm/yr estimated by Personius et al. (2014).

The **Southern Whidbey Island fault (SWIF)** extends ~ 100 -150 km from offshore Vancouver Island through Whidbey Island to E of Seattle (Figure 4.1). It may merge with the DDMF in the Strait of Juan de Luca, where it projects toward the eastern portion of the LRF (Figure 4.1; Sherrod et al., 2008). The SWIF strikes NW-SE and forms a broad (6-100 km wide) steeply NE-dipping zone with two to three strands in Puget Sound and on Whidbey Island (Johnson et al., 1996). Further SE, it broadens to a zone up to 20 km wide east of Seattle (Sherrod et al., 2008). Reverse displacement (N-side-up) was documented in Puget Sound through seismic reflection (Johnson et al., 1996). The presence of positive flower structures, along with apparent displacement of sedimentary units prompted Johnson et al. (1996) to interpret a transpressional history on the SWIF. Sherrod et al. (2008) expanded on this interpretation, identifying the northwestern portion of the SWIF as predominantly N-side-up compression, whereas the southeastern portion experiences oblique right-lateral kinematics. Similar to the BCFZ, topographic lineaments and scarps were identified parallel to aeromagnetic lineaments (Sherrod et al., 2008). Two such scarps were excavated in four trenches, in which at least four folding and faulting events were identified (Sherrod et al., 2008). The ages of these earthquakes, as well as an additional event identified from surface uplift by Kelsey et al. (2004), are included in Table 4.1. Scarps are consistent with <1 -2 m reverse slip per event, or 4 m in a single event (Sherrod et al., 2008). These displacements are consistent with the 1-2 m minimum vertical displacement documented by Kelsey et al. (2004). However, these displacement estimates are imprecise due to extensive faulting and few discrete offsets (Sherrod et al., 2008).

The **Seattle fault (SF)** extends ~ 70 km east from Hood Canal, across Puget Sound, to the Cascade foothills (Figure 4.1; Sherrod et al., 2000; Nelson, Johnson, Kelsey, Wells, et al., 2003). The SF forms the boundary between the Seattle uplift to the south and the Seattle basin to the N (Nelson, Johnson, Kelsey, Sherrod, et al., 2003; Nelson, Johnson, Kelsey, Wells, et al., 2003). It is primarily E-W striking, with some W to NNW-striking strands at its eastern extent (Sherrod et al., 2000; Sherrod, 2002). There are few surface traces of the SF, but several uplifted wave-cut platforms have been identified (Bucknam et al., 1992; Sherrod et al., 2000)). The central SF contains short en echelon backthrusts, some of which rupture the surface, however, there are multiple interpretations for how these strands relate to the primary SF structure, which accommodates reverse slip and dips $\sim 40^\circ$ - 80° S (Sherrod et al., 2000; Nelson, Johnson, Kelsey, Wells, et al., 2003; Nelson et al., 2014; Moore et al., 2022). Numerous paleoseismic studies have been completed on the SF, including cores collected from uplifted platforms, marshes, tide flats (e.g., Bucknam et al., 1992; Sherrod et al., 2000; Nelson et al., 2014), tree-ring studies (e.g., Atwater & Moore, 1992), and six paleoseismic trenches (Sherrod, 2002; Nelson, Johnson, Kelsey, Sherrod, et al., 2003; Nelson et al., 2014).

These studies have identified a total of five distinct events, four of which are correlated and well-described in Nelson et al. (2014) as earthquakes A-D. Event D occurred ~ 1000 cal kyr BP, and has been termed the “millennial earthquake series” (Nelson et al., 2014). However, the present radiocarbon ages do not resolve whether Event D was a single event, or whether there were smaller earthquakes on the SF, Tacoma fault, or Saddle Mountain fault within ~ 100 -200 yrs (Nelson et al., 2014). The fifth event is a late Pleistocene event identified by Sherrod (2002) along the eastern portion of the SF, with updated age data reported in Sherrod and Gombert (2014). The ages of these five events are included in Table 4.1. Notably, Nelson et al. (2014) suggested that late Holocene earthquakes A-E on the SF occurred following a ~ 6 -8 kyr period of quiescence, similar to the BRF.

The **Toe Jam Hill fault (TJF)** is located on Bainbridge Island, just north of the Seattle fault (Figure 4.1). The TJF is a N-dipping backthrust to the SF, merging with a moderately S-dipping ($\sim 40^\circ$ - 70°) SF < 10 km depth (Nelson, Johnson, Kelsey, Wells, et al., 2003). A S-facing scarp cuts across N-S trending streamlined glacial landforms (Nelson, Johnson, Kelsey, Wells, et al., 2003). The TJF accommodates compression through thrust and reverse faulting, as well as folding (Nelson, Johnson, Kelsey, Wells, et al., 2003). Five trenches have been completed on the TJF, in which 3-4 Holocene earthquakes were identified (Nelson, Johnson, Kelsey, Wells, et al., 2003). The ages of these events, termed Event A and Earthquakes B-D, are listed in Table 4.1. There is a range of displacement estimates for each event due to the range of fault dips from sub-horizontal to near-vertical in trench exposures (Nelson, Johnson, Kelsey, Wells, et al., 2003). Cumulative displacement of 5.2 m determined from surface displacement is consistent with ~ 0.6 -1.9 m of slip per event (Nelson, Johnson, Kelsey, Wells, et al., 2003). Displacement estimates for individual events are listed in Table 4.2. Slip rates on the TJF range from 0.2-0.3 mm/yr since 16 ka to 0.8-3.4 mm/yr since 3 ka (Nelson, Johnson, Kelsey, Wells, et al., 2003). The TJF may have ruptured both independently from and in conjunction with the SF (Nelson, Johnson, Kelsey, Wells, et al., 2003; Nelson et al., 2014). Nelson et al. (2014) correlated events in two of the TJF trenches with the Event B-E chronology on the SF. However, because the events of the TJF are well-correlated between events and include displacement estimates for individual events, I consider them separately from the SF for this paper.

The **Tacoma fault (TF)** extends ~ 50 km from Hood Canal through the city of Tacoma (Figure 4.1; Sherrod et al., 2004). The TF is the southern boundary of the Seattle uplift, > 20 km south of the surface projection of the SF, to which it may or may not be antithetic (Brocher et al., 2001; Johnson, Blakely, et al., 2004; Sherrod et al., 2004; Nelson et al., 2014). The TF strikes ~ 330 - 350° and dips steeply towards the north (Nelson et al., 2014). South-facing scarps are up to 4-6 m tall and displace glacial lineations and ice-contact deposits N-side-up, consistent with reverse slip (Sherrod et al., 2004; Nelson et al., 2014). Six trenches have been excavated across the TF and nearby scarps, however, one trench did not yield radiocarbon ages

(Sherrod et al., 2004; Nelson et al., 2014). Radiocarbon ages from the five later trenches and additional cores correlated slip on the TF with the millennial earthquake series identified on the SF (Earthquake D; Nelson et al., 2014). Plant fossils associated with tidal flat uplift also correlate coastal deformation with the millennial earthquake series (Sherrod et al., 2004; Nelson et al., 2014). This deformation may have occurred simultaneously with rupture on the SF, or within ~ 200 yrs of the SF event (Nelson et al., 2014).

The **Olympia fault (OF)** is ~ 80 km long, extending NW from Olympia, the Washington State capital (Odum et al., 2016). It forms the northern boundary of the Olympia basin, however, its location is poorly constrained due to the lack of clear scarps (Brocher et al., 2001; Sherrod, 2001). Recent magnetic and gravity anomaly models have refined its location and extended the OF to the northwest, where it projects toward the Saddle Mountain fault on the Olympic Peninsula (Figure 4.1; Odum et al., 2016). The OF trends $\sim 315^\circ$ and consists of a ~ 2 km-wide zone of deformed and faulted reflectors (Sherrod, 2001; Odum et al., 2016). Coastal deformation indicates N-side-up deformation (Sherrod, 2001; Odum et al., 2016). One event has been identified ~ 1150 - 1010 cal yr BP from relative sea level histories, however Sherrod (2001) did not directly attribute this abrupt submergence of a salt marsh to a specific fault.

The **North Olympic fault zone (NOFZ)** is located on the northern Olympic Peninsula and is composed of the Lake Creek-Boundary Creek (LCBCF) and Sadie Creek faults (Figure 4.1; Schermer et al., 2021). It is north of the Olympic Mountains and the Olympic accretionary complex (Nelson et al., 2017). The NOFZ is ~ 80 km long and strikes WNW, with ~ 0.5 - 2 km-long individual strands (Nelson et al., 2017; Schermer et al., 2021). The NOFZ is sub-vertical, with both N- and S-facing scarps offsetting postglacial landforms and re-routing drainages (Nelson et al., 2017; Schermer et al., 2021). Multiple recent studies have documented right-lateral and S-side-down displacement of glacial and postglacial deposits and landforms (Nelson et al., 2017; Schermer et al., 2021; Duckworth, 2019). Nelson et al. (2017) interpreted a small graben on the LCBCF to be part of a releasing stepover and estimated a strike slip to dip slip ratio of $\sim 10:1$. Further west on the Sadie Creek fault, Duckworth (2019) estimated a strike slip to dip slip ratio of $\sim 6:1$. Seven trenches have been completed on the NOFZ, which together with seismic reflection and cores collected from Lake Crescent, indicate the NOFZ has experienced 3-5 earthquakes since the late Pleistocene (Nelson et al., 2017; Leithold et al., 2019; Schermer et al., 2021). Nelson et al. (2017) identified 3-5 events on the LCBCF, correlated between their five trenches. Schermer et al. (2021) identified 4-5 events in two trenches (GB and BL) on the Sadie Creek fault and created multiple geologically plausible models for earthquake ages. I use the conservative models in this paper, except for event GB3, for which I use their model that allows for four events in the GB trench. Schermer et al. (2021) completed a preliminary correlation between the Sadie Creek fault and LCBCF faults, however as they did not publish modeled age ranges for that correlation, I consider the Sadie Creek and LCBCF earthquake histories separately for this paper. The ages

for all of these events are presented in Table 4.1. Cumulative displacements from vertical separation of the ground surface are 2.6 ± 0.2 m and 5.4 ± 0.6 m on the Sadie Creek fault (Schermer et al., 2021), more than the ~ 1 -2 m vertical separation estimated across scarps on the LCBCF (Nelson et al., 2017). Neither study was able to constrain right-lateral displacement in the trench, therefore oblique displacement must be estimated using the ~ 23 -28 m cumulative right-lateral displacement of channels and terrace risers (Nelson et al., 2017; Schermer et al., 2021). These cumulative displacements are consistent with 4 ± 1 m right-lateral displacement, similar to the 3.6-6.6 m estimated using the ratio of strike slip to dip slip (Schermer et al., 2021). Together with estimates of per-event vertical separation that range from 0.6-1.2 m (Schermer et al., 2021), per-event oblique displacement on the Sadie Creek fault is consistent with the 2.5-7.8 m estimated by Nelson et al. (2017). Displacement events for individual events are listed in Table 4.2. The dextral slip rate on the NOFZ is ~ 1.3 -2.3 mm/yr (Nelson et al., 2017; Duckworth, 2019; Schermer et al., 2021).

The **Saddle Mountain fault zone (SMF)** is located on the SE slope of the Olympic Mountains on the Olympic Peninsula, where it forms the boundary between the uplifting Olympic Mountains and the Puget Lowland (Figure 4.1; Barnett et al., 2015). The SMF trends NE and forms a 45 km-long system with the Canyon River, Frigid Creek, and Sund Creek faults (Barnett et al., 2015). The Saddle Mountain East strand strikes 022° and dips steeply (75°) toward the SE, and the Sund Creek strand strikes 240° and dips steeply (75° - 80°) toward the NW (Barnett et al., 2015). Barnett et al. (2015) documented 15 km of scarps up to 7.5 m tall, accommodating reverse (E-side-up) motion. Slickenlines in trenches suggest some component of lateral motion, however, there is no clear lateral displacement in the landscape (Barnett et al., 2015). Five trenches have been completed on the SMF, including a re-excavation of the initial trench documented by (Wilson et al., 1979). These trenches, together with a coring transect, reveal evidence for at least three earthquakes on faults within the SMF system since glacial retreat ~ 17 ka (Witter et al., 2008; Nelson et al., 2014; Barnett et al., 2015). Only the most recent event (E3) ruptured multiple strands of the SMF—earlier events ruptured either the Sund Creek fault or the Saddle Mountain West fault (Witter et al., 2008; Barnett et al., 2015). Nelson et al. (2014) correlated E3 with the millennial earthquake series on the SF and TF, but could not resolve whether SMF ruptured simultaneously with the other faults, or within 10s to 100s of years. The ages of these earthquakes are reported in Table 4.1. Displacement ranges from 1-1.5 m on the Saddle Mountain West fault to >3.2 m on the Saddle Mountain East fault (Barnett et al., 2015). An estimated vertical slip rate is 0.2 mm/yr since deglaciation (Barnett et al., 2015).

Table 4.2: Interseismic periods and displacements for forearc fault earthquakes.

Main Fault ¹	Strand/Site Name	EQ name ²	Time since previous event (yrs)	Time to next event (yrs)	Displacement (m)	Source
BRF	Woodlot Trench	Event 1	–	5100-5990	0.8-1.1	Chapter 3
BRF	Woodlot Trench	Event 2	5100-5990	10-990	0.7-1.1	Chapter 3
BRF	Woodlot Trench	Event 3	10-990	10-1900	0.9-1.5	Chapter 3
BRF	Woodlot Trench	Event 4	10-1900	200-1800	0.7-1.3	Chapter 3
BRF	Woodlot Trench	Event 5	200-1800	–	0.6-1.5	Chapter 3
LRF		Event 3	–	6100-6900	0.2-1.7	Morell et al., 2018
LRF		Event 2	6100-6900	300-700	0.4-1.4	Morell et al., 2018
LRF		Event 1	300-700	–	0.2-1.3	Morell et al., 2018
BCFZ	Sandy Point fault	SPF-O	–	<14900	2-2.5	Kelsey et al., 2012
BCFZ	Sandy Point fault	SPF-M	<14900	<2000	<1	Kelsey et al., 2012
BCFZ	Sandy Point fault	SPF-Y	–	–	<1	Kelsey et al., 2012
BCF	Boulder Creek fault/ Hornet trench	EQ A	–	4060-5270	0.8-1.4	Sherrod et al., 2013
BCF	Boulder Creek fault/ Hornet trench	EQ B	4060-5270	2070-2450	0.25-1.7	Sherrod et al., 2013
BFC	Boulder Creek fault/ Hornet trench	EQ C	2070-2450	–	0.4-0.7	Sherrod et al., 2013
DDMF		CE2	–	5700-6100	0.6-1.7	Personius et al., 2014
DDMF		CE1	5700-6100	–	0.6-1.7	Personius et al., 2014
DDMF	Utsalady Point fault	Duffers 1	–	700-2100	1-1.2	Johnson et al., 2004
DDMF	Utsalady Point fault	Duffers 2	700-2100	–	0.9-1.1	Johnson et al., 2004
SWIF	Beef Barley trench	BB3	9270-16400	–	0.3-1	Sherrod et al., 2008
SF/TJF	Toe Jam Hill fault/ Crane Lake	Event A	–	<4100	1-1.5	Nelson et al., 2003

Table 4.2: (continued)

Main Fault ¹	Strand/Site Name	EQ name ²	Time since previous event (yrs)	Time to next event (yrs)	Displacement (m)	Source
SF/TJF	Toe Jam Hill fault/ Crane Lake	EQ B	<4100	200-1300	1.6-3.3	Nelson et al., 2003
SF/TJF	Toe Jam Hill fault/ Crane Lake	EQ C	200-1300	<700	1.5-2.1	Nelson et al., 2003
SF/TJF	Toe Jam Hill fault/ Crane Lake	EQ D	<700	–	1.8-2.5	Nelson et al., 2003
SF/TJF	Toe Jam Hill fault/ Mossy Lane	EQ C	–	<400	2.3-3.3	Nelson et al., 2003
SF/TJF	Toe Jam Hill fault/ Mossy Lane	EQ D	<400	–	5.7-9.2	Nelson et al., 2003
NOFZ	Sadie Creek fault	GB1	–	1300-8100	3-5	Schermer et al., 2021
NOFZ	Sadie Creek fault	GB2	1300-8100	500-5600	3-5	Schermer et al., 2021
NOFZ	Sadie Creek fault	GB3	500-5600	600-3300	3-5	Schermer et al., 2021
NOFZ	Sadie Creek fault	GB4	600-3300	–	3-5	Schermer et al., 2021
NOFZ	Sadie Creek fault	BL1	–	1400-4100	3-5	Schermer et al., 2021
NOFZ	Sadie Creek fault	BL2	1400-4100	1100-2500	3-5	Schermer et al., 2021
NOFZ	Sadie Creek fault	BL3	1100-2500	200-4700	3-5	Schermer et al., 2021
NOFZ	Sadie Creek fault	BL4	200-4700	1100-5300	3-5	Schermer et al., 2021
NOFZ	Sadie Creek fault	BL5	1100-5300	–	3-5	Schermer et al., 2021
SMF	Sund Creek fault	E2	–	2430-2530	1.4-3.6	Barnett et al., 2015
SMF	Sund Creek fault	E3	2430-2530	–	1.9-3.2	Barnett et al., 2015
SMF	Saddle Mountain West	Cargill-A	–	6700-15300	1.9-2.4	Witter et al., 2008
SMF	Saddle Mountain West	Cargill-B	6700-15300	–	1.3-2.7	Witter et al., 2008

¹See Figure 4.1 for fault abbreviations and locations.²Event names included in original publications or earlier compilations were kept where possible.

4.3 Methods

4.3.1 Compilation of paleoearthquake chronologies

I compiled earthquake age data for thirteen forearc fault systems to examine spatial and temporal patterns of rupture and recurrence in the northern Cascadia forearc. The existing compilations of Sherrod and Gombert (2014) and Styron and Sherrod (2021) focus on forearc faults in western Washington. To these compilations, I added recent work on Vancouver Island faults (Morell et al., 2018, 2017; Harriehausen et al., 2021, *in review*; Chapter 2; Chapter 3) and the North Olympic fault zone (Nelson et al., 2017; Duckworth, 2019; Schermer et al., 2021). To compare the timing of earthquakes on different faults, I plotted the 95% confidence interval age range for each forearc earthquake, as determined from published paleoseismic studies. If no maximum bounding age was reported, I used a maximum of 15,000 cal yr BP, the average age of deglaciation in the region (Haugerud, 2021), as most of the trenches included only post-glacial deposits. If no minimum bounding age was reported, I used a minimum of 100 cal yr BP (1850 CE), the start of written historical earthquake records in the region (Seattle Municipal Archives, 2023; City of Victoria, 2023). The ages and sources of all events in this paper are in Table 4.1, and the recurrence periods are in Table 4.2.

I compiled displacement data associated with the forearc events in Table 4.1 to calculate slip rates and assess earthquake recurrence models. Paleoseismic records in Cascadia are few and temporally limited, due to rapid soil development and dense forests obscuring geomorphic evidence of faulting, landscape modification from glacial and anthropogenic processes, and slow strain rates. Of the thirteen fault systems described above, nine had displacement estimates for two or more consecutive events (Table 4.2). I used the range of possible displacement values for each specific event as identified by the authors of the original studies. Per-event displacement data were unavailable for the NOFZ, so I used per-event displacements estimated from geomorphic mapping (Schermer et al., 2021, Schermer, pers. comm., 2023).

4.3.2 Slip rate calculations

I used earthquake age and displacement data to calculate closed-interval slip rates for five of the forearc faults described in Section 4.2.5 and evaluate whether slip rates were steady in the Holocene. This required at least two consecutive intervals, or complete seismic cycles, for each fault. Only five faults (BRF, BCF, LRF, NOFZ, and TJF) had paired age and displacement data available for two or more intervals (Table 4.2). For each closed interval, I then used an R script that performed a Monte Carlo simulation to calculate the slip rate and associated uncertainty (script available in Appendix D). The script requires a user-defined input of a .csv file with the following data for each event on a fault: the minimum and maximum values of

the 2σ earthquake age range, the minimum and maximum values for event displacement, and the minimum and maximum values for cumulative displacement. The script then calculates a slip rate from earthquake ages and displacements chosen from a normal distribution defined by the parameters in the input file and repeats this calculation 10,000 times. I report the outputs as the mean $\pm$ one standard deviation of the 10,000 runs.

4.3.3 Testing models for earthquake recurrence

I assessed whether faults were consistent with periodic, time-predictable, or slip-predictable models of earthquake recurrence by plotting the frequency of recurrence intervals and the relationship between displacement and recurrence (Table 4.2). If the forearc faults are consistent with a periodic model, they should have approximately equal recurrence intervals between each event. In order to assess whether faults were consistent with the time- and slip-predictable models, both the timing and displacement of individual events on a fault must be well-constrained. However, earthquake displacement paired with the prior or subsequent recurrence interval was not available for all faults, so I assessed these models on a subset of faults (Table 4.2). If the forearc faults are consistent with the time-predictable model, they should have a strong positive linear relationship between the displacement in an earthquake and the length of time until the next earthquake. I plotted the length of the previous interseismic period on the X-axis, and earthquake displacement on the Y-axis. The displacement associated with the oldest known event on any forearc fault is not included in this analysis as the length of the previous interseismic period is unknown. If the forearc faults are consistent with the slip-predictable model, they should have a strong positive linear relationship between the length of time since the previous earthquake and the displacement in the subsequent earthquake. I plotted the earthquake displacement on the X-axis and the length of the subsequent interseismic period on the Y-axis. The displacement associated with the most recent event on any forearc fault is not included in this analysis as the length of the subsequent interseismic period is unknown.

4.4 Results

The timing and recurrence of earthquakes on CSZ forearc faults (Figures 4.4 and 4.5), shows two common patterns. First, many faults have an early-mid Holocene quiet period. At least six faults have one earthquake in the late Pleistocene to early Holocene ($>\sim 7$ ka), followed by a long gap of over 4 kyrs before multiple mid-late Holocene earthquakes each separated by less than ~ 2 kyrs (Figure 4.4). This pattern is evident on the BRF, BCF, LRF, DDMF, SF, and SMF. The BCFZ may also be included if the SPF-O event occurred in the earlier part of its possible range, as well as the SWIF if the BB/MB event on the SWIF occurred in

the later part of its range, for example, if it correlates with the CH event at another site on the SWIF (Table 4.1; Figure 4.4). Some of these late Pleistocene to early Holocene events may have occurred within a few kyrs of deglaciation (e.g., BRF, SWIF, SF, SMF; Figure 4.4); however, the long and short intervals occur at slightly different times on different faults, indicating that there may be spatial variability in this pattern. Second, there appear to be two populations in the distribution of recurrence intervals, denoted in Figure 4.5 as Set A and Set B. Set A includes recurrence intervals less than ~ 3 kyrs, whereas Set B includes recurrence intervals of ~ 4.5 - 7.5 kyrs, with an outlier on the SMF at ~ 11 kyr (mean of a ~ 6.7 - 15.3 kyr range; Table 4.2). The longer intervals are generally the earliest intervals on a fault, whereas the shorter intervals are more recent (Table 4.2). All the faults included in Figure 4.5 have intervals in both Set A and Set B (Figure 4.5), indicating that this pattern does not reflect a dichotomy between faults with frequent earthquakes and those with infrequent earthquakes, but rather a behavior shared by all. Recurrence on these faults is therefore not consistent with a periodic or quasi-periodic model of earthquake recurrence.

The varying recurrence intervals (Figure 4.5), when paired with slip in individual events (Figure 4.6), result in closed-interval slip rates that vary by orders of magnitude on the same fault (Table 4.3; Figure 4.7). I calculated interval slip rates of ~ 0.2 - 3.3 mm/yr for the BRF, ~ 0.2 mm/yr for the BCF, ~ 0.1 - 1.5 mm/yr for the LRF, ~ 0.9 - 4.8 mm/yr for the NOFZ, and ~ 1.9 - 16.1 mm/yr for the TJF (see Table 4.3 for full details). These slip rates are generally higher in the mid-late Holocene than in the late Pleistocene to early Holocene, but the fast and slow intervals do not occur at the same time or in the same succession on different faults (Figure 4.7). The slowest slip rate interval on the BRF from ~ 10.5 to ~ 5 ka is followed by the fastest slip rate interval from ~ 5 to ~ 4.5 ka, but then the slip rate declines through the late Holocene (Table 4.3; Figure 4.7). This pattern is similar to the GB site on the NOFZ, but slip rates on the same fault at the BL site increase through the early-mid Holocene with the slowest slip rate being the most recent (Table 4.3; Figure 4.7). The LRF and TJF both increase in the mid-late Holocene (Table 4.3; Figure 4.7). Only the BCF slip rate does not appear to change (Table 4.3; Figure 4.7). Such drastic changes in slip rate are inconsistent with the assumptions of the time- and slip-predictable recurrence models.

The time- and slip-predictable earthquake recurrence models can only be assessed for the BRF and TJF, as these are the only faults with paired age-displacement data for the three or more intervals required to calculate a regression analysis. Neither the BRF nor the TJF is time-predictable as they show a negative relationship between displacement and length subsequent interseismic period ($R^2=0.1$ and 0.5 , respectively; Figure 4.8), in contrast to the strong positive relationship expected in a time-predictable model. The BRF also has a negative relationship between the length of the previous interseismic period and earthquake displacement, indicating that it is not slip-predictable ($R^2=0.8$; Figure 4.8). There is a weak positive relationship for the TJF ($R^2=0.5$); however, the regression does not go through the origin, suggesting constant slip, or creep,

inconsistent with observations of the Seattle fault zone (Figure 4.8a). Neither the BRF nor the TJF are consistent with the time- or slip-predictable models of earthquake recurrence. Unfortunately, the data do not exist to assess these models for other CSZ forearc faults.

Figure 4.4 (*following page*): Timing of events on CSZ forearc faults from north to south, and on the CSZ megathrust. Also shown is the approximate timing of deglaciation, after Haugerud (2021), Easterbrook (1992), and Clague et al. (1980). CSZ timing from Goldfinger et al. (2012); see Table 4.1 for forearc fault timing sources. There is a high frequency of earthquakes in the past ~ 2 ka, including the “millennial earthquake series” of Nelson et al. (2014). Also evident are the long gaps between late Pleistocene to early Holocene and mid-late Holocene events on several faults (BRF, BCF, LRF, DDMF, SF, SMF).

Table 4.3: Calculated slip rates for CSZ forearc faults

Forearc fault	Interval	Length of interval ¹ (yrs)	Cumulative slip ² (m)	Slip rate ³ (mm/yr)
BRF	Events 4-5	1000 $\pm$ 313	1.1 $\pm$ 0.2	1.4 $\pm$ 10.5
BRF	Events 3-4	954 $\pm$ 338	2.1 $\pm$ 0.4	1.8 $\pm$ 27.5
BRF	Events 2-3	502 $\pm$ 182	3.3 $\pm$ 0.5	3.3 $\pm$ 11.4
BRF	Events 1-2	5545 $\pm$ 164	4.1 $\pm$ 0.6	0.2 $\pm$ 0.02
BRF	Event 1	–	5.0 $\pm$ 0.7	–
BCF	EQ B-C	2260 $\pm$ 68	0.5 $\pm$ 0.1	0.2 $\pm$ 0.03
BCF	EQs A-B	4665 $\pm$ 256	1.5 $\pm$ 0.4	0.2 $\pm$ 0.08
BCF	EQ A	–	2.6 $\pm$ 0.6	–
LRF	Events 2-1	500 $\pm$ 70	0.8 $\pm$ 0.3	1.5 $\pm$ 0.6
LRF	Events 3-2	6500 $\pm$ 159	1.7 $\pm$ 0.5	0.1 $\pm$ 0.04
LRF	Event 3	–	2.6 $\pm$ 0.9	–
NOFZ	BL4-5	3193 $\pm$ 970	4 $\pm$ 0.5	1.4 $\pm$ 1.1
NOFZ	BL3-4	2478 $\pm$ 958	8 $\pm$ 1	4.8 $\pm$ 240
NOFZ	BL2-3	1799 $\pm$ 251	12 $\pm$ 1.5	2.3 $\pm$ 0.4
NOFZ	BL1-2	2748 $\pm$ 508	16 $\pm$ 2	1.5 $\pm$ 0.4
NOFZ	BL1	–	20 $\pm$ 2.5	–
NOFZ	GB3-4	1954 $\pm$ 555	4 $\pm$ 0.5	2.3 $\pm$ 1.1
NOFZ	GB2-3	3063 $\pm$ 1146	8 $\pm$ 1	2.9 $\pm$ 104
NOFZ	GB1-2	4701 $\pm$ 1279	12 $\pm$ 1.5	0.9 $\pm$ 1.1
NOFZ	GB1	–	16 $\pm$ 2	–
TJF	EQs C-D	308 $\pm$ 138	2.2 $\pm$ 0.2	16.1 $\pm$ 225.5
TJF	EQs B-C	750 $\pm$ 195	4 $\pm$ 0.3	2.6 $\pm$ 1.3
TJF	EQs A-B	2070 $\pm$ 863	6.4 $\pm$ 0.8	1.9 $\pm$ 10.7
TJF	EQ A	–	7.7 $\pm$ 0.9	–

¹Length of interval reported as mean $\pm$ one standard deviation of the time between the events. For event age range inputs, see Table 4.1.

²Cumulative slip reported as the mean $\pm$ one standard deviation. For individual event displacement inputs, see Table 4.6.

³Slip rate reported as the mean $\pm$ one standard deviation of the 10,000 modeled slip rates for that interval.

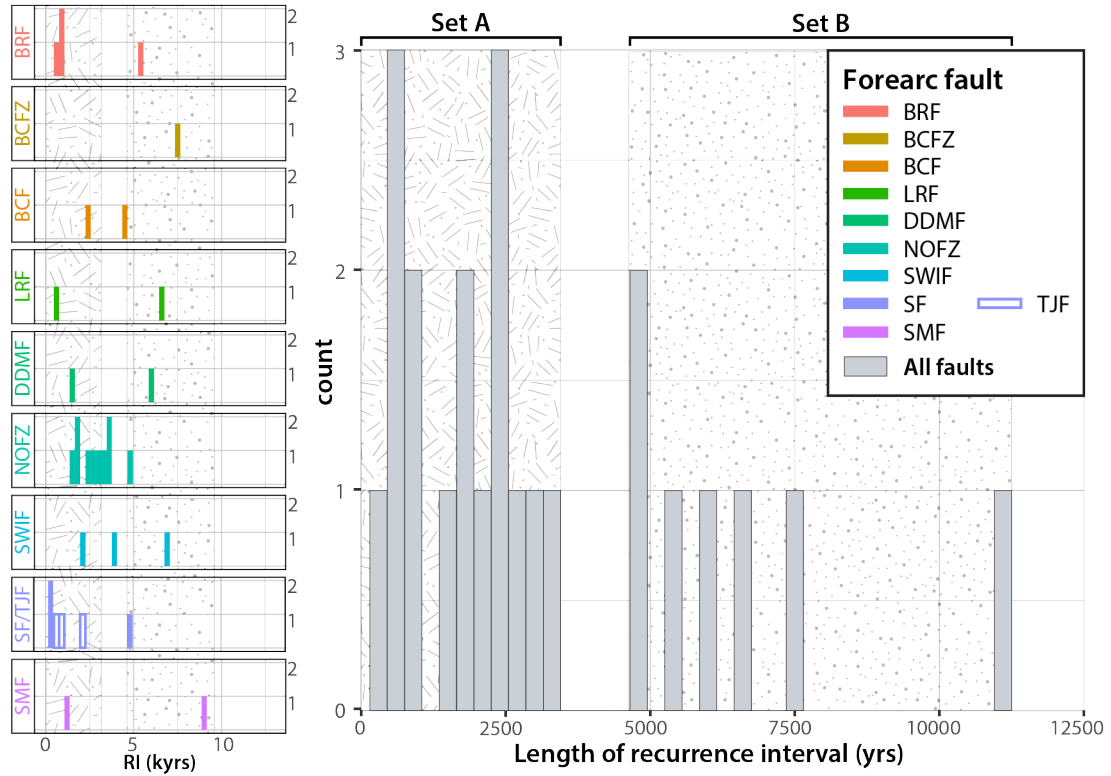


Figure 4.5: Histogram of interseismic recurrence intervals on CSZ forearc faults, (bin size = 300 yrs). The frequency of recurrence intervals of individual faults is depicted by thin bars colored by fault (left). The frequency of recurrence intervals among all faults is depicted by thick grey bars (right). Forearc fault colors and abbreviations are the same as in Figure 4.4. Note that there are two distinct populations of recurrence intervals, Set A ($< \sim 3$ kyr), and Set B ($> \sim 4.5$ kyr).

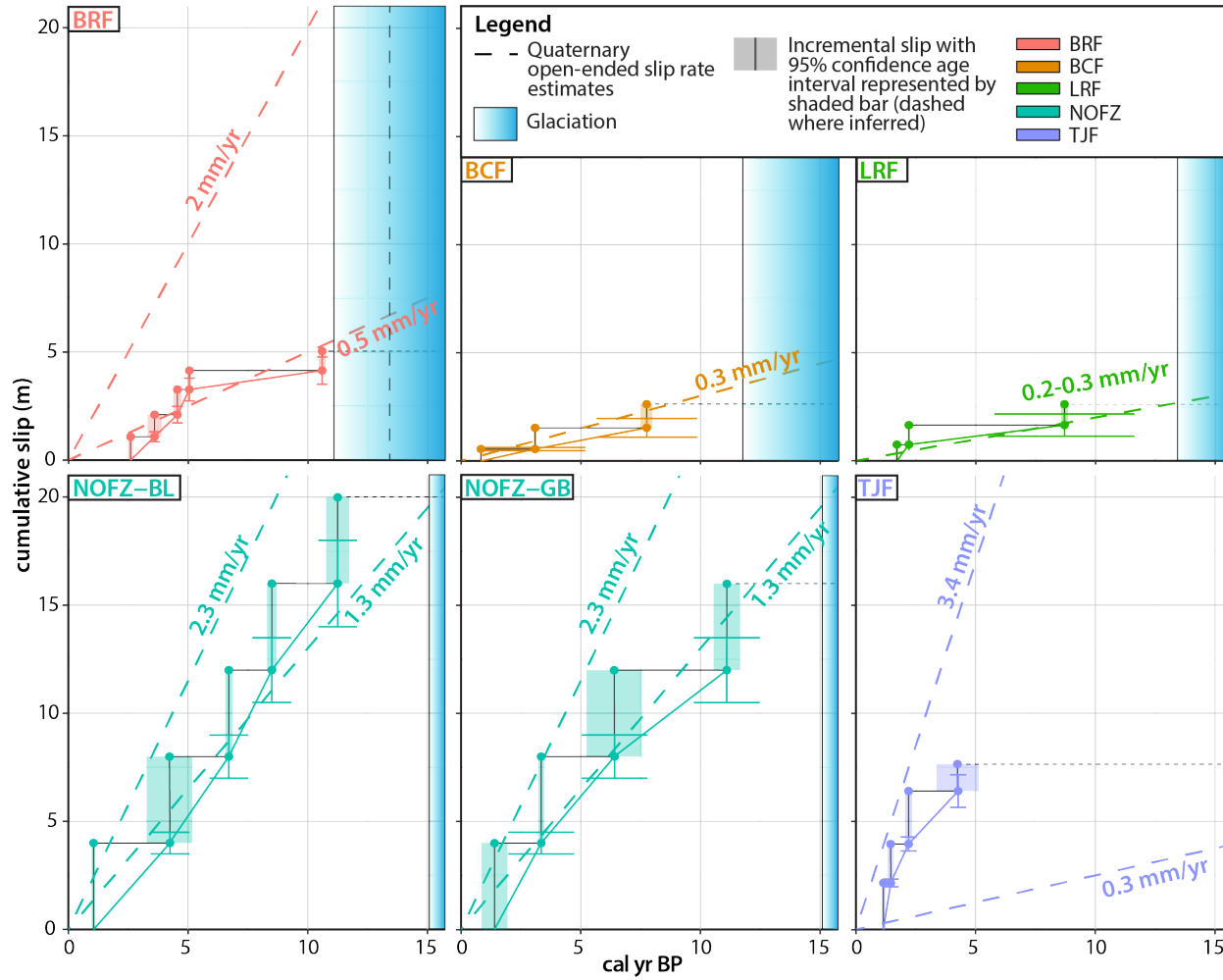


Figure 4.6: Cumulative displacements on forearc faults, separated by fault. Shaded areas represent age ranges of individual events, and mean displacement in those events. Error bars represent uncertainty on cumulative displacement. Also shown are the Quaternary open-ended slip rate estimates for each fault. BRF slip rate estimates of $\sim 0.5\text{--}2$ mm/yr from Chapter 3. BCF slip rate estimate of $\sim 0.3 \pm 0.1$ mm/yr from Sherrod et al. (2013). LRF slip rate estimate of $\sim 0.2\text{--}0.3$ mm/yr from Morell et al. (2018). NOFZ slip rate estimates of $\sim 1.3\text{--}2.3$ mm/yr from Schermer et al. (2021) TJF slip rate estimates of $0.2\text{--}0.3$ mm/yr post-glacial, and $0.8\text{--}3.4$ mm/yr since 3 ka from Nelson, Johnson, Kelsey, Wells, et al. (2003).

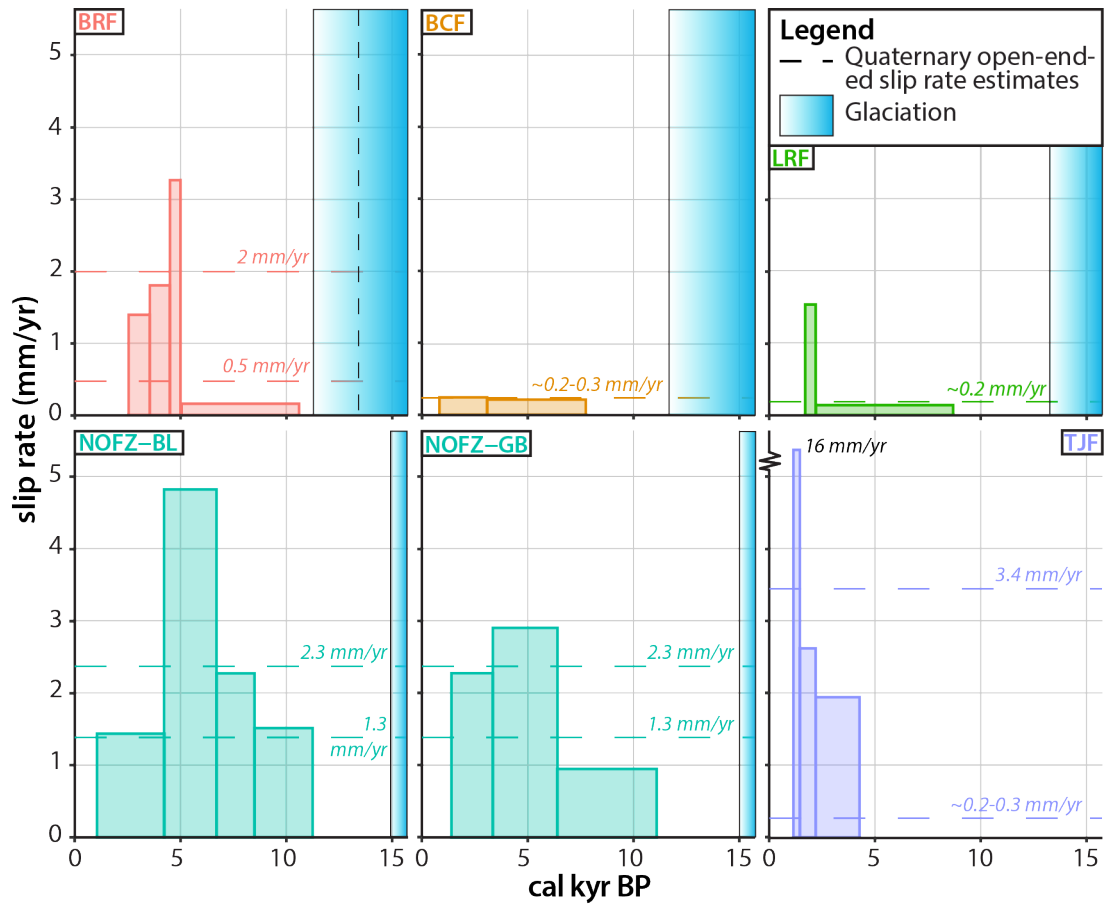


Figure 4.7: Calculated closed-interval slip rates on forearc faults. Minimum and maximum Quaternary open-ended slip rate estimates are shown by dashed lines. The wide range of closed-interval slip rates shows that the slip rates on forearc faults are not constant through time, and in fact, vary by orders of magnitude from one interval to the next.

4.5 Discussion

The above data allow us to evaluate the three motivating questions of this study: (1) Are recurrence and slip rates on CSZ forearc faults steady or non-steady? (2) If non-steady, can their behavior be described by the time- or slip-predictable models typically used to describe faults with non-steady recurrence? (3) What external tectonic or climatic factors could be affecting the recurrence on these forearc faults?

All CSZ forearc faults with a record of two or more recurrence intervals show a distinction between longer (~ 4.5 - 7.5 kyr) and shorter ($< \sim 3$ kyr) intervals (Figure 4.5). Slip rates on the BRF, LRF, NOFZ, and TJF vary by an order of magnitude on the same fault. These data directly answer the first question: these recurrence intervals and slip rates are non-steady and are inconsistent with periodic or quasi-periodic recurrence models.

In the time-predictable recurrence model, there is some threshold stress at which an earthquake occurs

(Shimazaki & Nakata, 1980). The model predicts that the length of a recurrence interval depends on how much stress was released in the prior earthquake, proportional to slip. In the slip-predictable recurrence model, there is some baseline stress that a fault returns to after an earthquake. The model predicts that the amount of stress released in an earthquake (proportional to slip; Shimazaki & Nakata, 1980) depends on the length of the preceding recurrence interval. There are two datasets we can use to answer the second question. The first dataset is recurrence interval length paired with the displacement in the preceding or following earthquake. The eight faults for which I have displacement data tend to have similar displacements in each event (Table 4.2). These similar displacements correspond to similar magnitude earthquakes, and therefore a similar stress drop in each event. This same stress drop is independent of recurrence interval length.

I ran linear regressions from this dataset of recurrence interval length paired with displacement. These regressions require data from three or more complete seismic cycles, which exists only for the BRF and TJF. The time- and slip-predictable recurrence models predict a positive linear relationship between earthquake displacement and the length of the subsequent (for time-predictable) or prior (for slip-predictable) recurrence interval. Neither the BRF nor the TJF are consistent with this prediction (Figure 4.8). The second dataset is the slip rates calculated for the BRF, LRF, TJF, and NOFZ (Figure 4.7, Table 4.3). The time- and slip-predictable recurrence models predict that the short-term slip rate should match the long-term slip rate due to the assumption that the rate of stress accumulation is steady through time (e.g., Weldon et al., 2004). The varied slip rates calculated in this study are inconsistent with this assumption.

To answer the third question, I consider four possible factors that could affect recurrence on CSZ forearc faults: plate reorganization, interseismic strain accumulation or coseismic triggering from the CSZ megathrust cycle, Coulomb stress changes from neighboring faults, and glacial isostatic adjustment. I assess whether the impact of each factor is consistent with the observed recurrence intervals and slip rates on CSZ forearc faults.

4.5.1 What factors could affect recurrence?

4.5.1.1 Changes in plate motion

The first factor that could affect how much strain is accommodated in the upper plate is a change in plate motion. Several authors have suggested that a change in the rate of subduction (e.g., S. Li et al., 2020), a change in the angle of subduction (e.g., Fitch, 1972; McCaffrey, 1992; ten Brink & Lin, 2004), or accretion of an exotic terrane (e.g., England & Calon, 1991; McCrory, 1996; Wells et al., 2014) could change where forearc faults form, their kinematics, and how much strain they accommodate. However, there is no

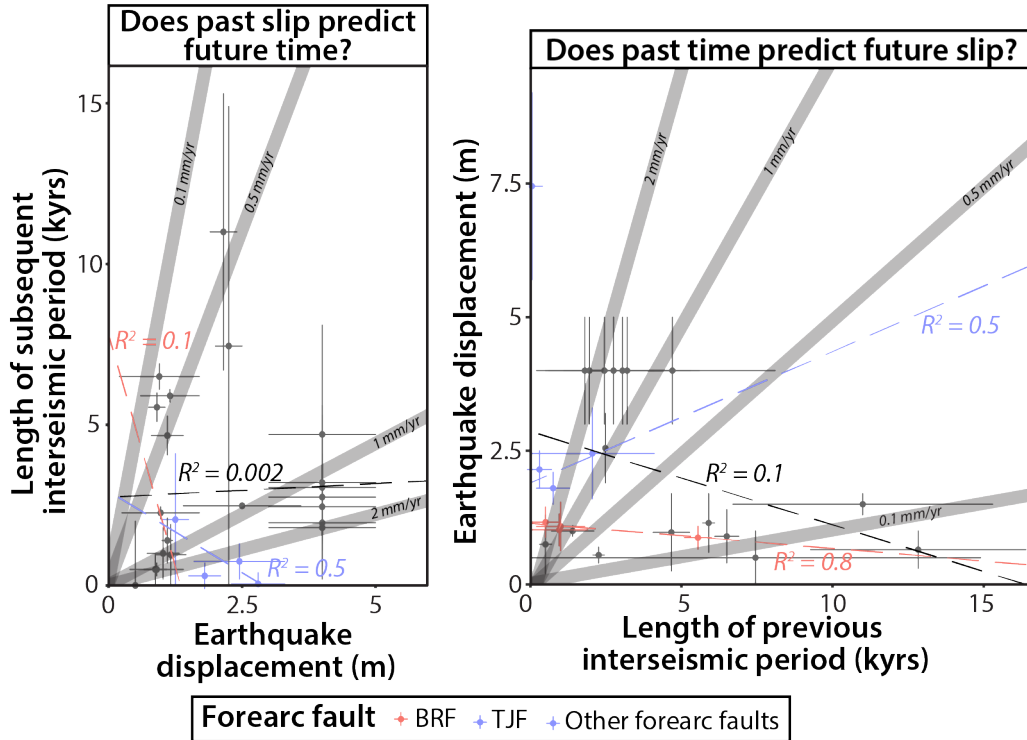


Figure 4.8: Plots showing the relationship between earthquake displacement and recurrence for CSZ forearc faults (after Weldon et al., 2004). We would expect a positive linear relationship in the plot on the left if faults followed a time-predictable model, and on the right if they followed a slip-predictable model. Grey bars show expected relationships for different slip rates. TJF - Toe Jam Hill fault (SF splay).

evidence of a change in the rate or obliquity of the CSZ on the timescales of this study, and the most recent accretionary event was that of Siletzia ~ 50 -45 Ma (Wells et al., 2014). The most recent plate reorganization in northern Cascadia was the formation of the Explorer plate and Nootka fault zone ~ 4 Ma (Audet et al., 2008), considerably earlier than the Holocene changes that are the focus of this study. Geodetic rotations are consistent with rotations from paleomagnetic data, suggesting that rates have been consistent since ~ 18 Ma (e.g., Finley et al., 2019). A change in plate motion is, therefore, not a factor that could impact Holocene changes in recurrence or slip rates.

4.5.1.2 Interactions between crustal faults

A second factor that could be impacting CSZ forearc earthquake timing is interactions between faults or groups of faults. Slip on one fault or group of faults could cause Coulomb stress transfer onto nearby faults, either promoting or inhibiting slip on those receiver faults (e.g., King et al., 1994; Charles & King, 1995; Stein, 1999, 2003). Studies in other tectonic settings have shown that Coulomb stress changes can promote slip on neighboring faults either through direct triggering or increased loading (e.g., King et al., 1994; Sarkar & Chander, 2003; Ma et al., 2005; Freed, 2005; Blasweiler et al., 2022). It is also possible for

Coulomb stress changes to reduce loading and inhibit slip on neighboring faults (e.g., Bakun, 1999; Toda & Stein, 2002; Sarkar & Chander, 2003; Freed, 2005). Models suggest that Coulomb stress changes can occur as far as 1-4 fault lengths away from a crustal source fault, with stress increasing or decreasing depending on the kinematics of the source fault (e.g., King et al., 1994; Stein, 1999). If we consider the BRF as a source fault, it could affect faults 30-240 km away (e.g., LRF, XELF, DDMF, SWIF). If Coulomb stress transfer is affecting CSZ forearc faults, we would expect neighboring faults to rupture within 10s-100s of years of the source fault if Coulomb stress increases. Conversely, if Coulomb stress decreases on neighboring faults, we would not expect them to rupture at a similar time as the source fault.

The timing compilation shows that many CSZ forearc faults do not rupture within a few 100 yrs of their neighbors (Figure 4.4). However, there are two sets of adjacent faults in this study whose earthquake timing suggests that they may be candidates for fault interaction through static stress transfer. The first group consists of the BRF, XELF, and LRF on central and southern Vancouver Island (Figure 4.1). The one earthquake recorded on the XELF (~ 4.7 - 2.3 ka) occurred during the cluster of four mid-Holocene earthquakes on the BRF (~ 5 - 2.5 ka; Figure 4.9a), with the largest peak in the XELF PDF, and therefore the most likely age of the XELF event, corresponding with the BRF MRE (Figure 4.1b). This temporal overlap would be consistent with right-lateral transtension on the BRF increasing the horizontal stress on the compressional XELF, potentially promoting slip and triggering an earthquake (Figure 4.9a). The BRF is ~ 30 - 60 km long (Chapter 2) and is ~ 130 km away from the XELF (≤ 15 km long; Harrichhausen et al., 2021). A receiver XELF is within ~ 2 - 4.5 source-fault-lengths of the BRF, within the distances at which Coulomb stress transfer can affect adjacent faults. However, a receiver BRF is >8 source-fault-lengths away from the XELF, making it unlikely that slip on the XELF would cause Coulomb stress changes on the BRF. The timing of earthquakes on the LRF is also consistent with Coulomb stress changes from the BRF. The cluster of four earthquakes on the BRF during the mid-Holocene ~ 5 - 2.5 ka and the XELF event ~ 4.7 - 2.3 ka correspond with the interseismic period between Events 2 and 3 on the LRF of Morell et al. (2018). Events 2 and 1 on the LRF occur during a late Holocene quiet period on the BRF (Figure 4.9b). This anti-correlation in earthquake timing may reflect rupture on the BRF inhibiting slip on the LRF. Right-lateral transtension on the BRF could decrease the horizontal stress on the E-W striking portion of the LRF, potentially inhibiting slip on the compressional LRF (Figure 4.9a). The BRF and LRF are both ~ 30 - 60 km long and are ~ 100 km apart (~ 1.5 - 3.5 fault lengths; Morell et al., 2017; Harrichhausen et al., 2021, Chapter 2; Figure 4.1). These length scales are within the distances at which Coulomb stress transfer can affect adjacent faults.

The second group consists of the BCF and DDMF (Figure 4.9c). The BCF (~ 12 km long; Sherrod et al., 2013) is ~ 35 km away from the DDMF (125 km long; Figure 4.1; Personius et al., 2014), certainly close enough to be affected by Coulomb stress changes from a DDMF rupture. However, only a portion of the

Table 4.4: Likelihood of forearc fault earthquakes overlapping with CSZ events.

Forearc fault	Event name	Likelihood of overlap with CSZ event		
		Individually (100 yr)	Individually (150 yr)	Together (150 yr)
T4				
BCFZ	SPF-M/Y	10%	15%	–
BCFZ	BB	17%	37%	1%
DDMF	D1	20%	29%	1%
NOFZ	BL5	4%	12%	1%
NOFZ	GB4	9%	13%	–
NOFZ	EQ D	13%	19%	–
SWIF	BB3	8%	12%	–
TJF	EQ D	0%	2%	–
TJF	EQ C	40%	57%	1%
SF	EQ C	1%	8%	–
SMF	E3	0%	0%	–
SMF	E2	10%	15%	–
OF	OLY	0%	3%	–
T7				
BRF	Event 4	17%	25%	1%
XELF	XELF	1%	1%	–
BCF	EQ B	0%	1%	–
BCFZ	SPF-O	3%	4%	–
NOFZ	EQ C	17%	25%	1%
NOFZ	BL4	5%	8%	–
NOFZ	GB3	36%	51%	1%
SWIF	BB/MB/FS	3%	4%	–
TJF	EQ A	6%	9%	–
SMF	E2	8%	11%	–

DDMF is within a few fault lengths of the BCF, the length range over which Coulomb stress transfer has the greatest impact (Figure 4.1). Earthquakes A and B on the BCF occur during a quiet period on the DDMF, while Earthquake C occurs in a quiet period after CE1 and between UPF events D1 and D2 (Figure 4.9d). The DDMF is a right-lateral thrust fault, and slip on the DDMF would decrease the horizontal stresses on the BCF, inhibiting slip (Figure 4.9c).

Outside of these two groups, the timing of CSZ forearc earthquakes does not appear to be affected by neighboring faults. Coulomb stress transfer modeling could further assess the feasibility of this factor by determining the magnitude of potential Coulomb stress changes.

4.5.1.3 Impact of the Cascadia megathrust

A third factor that could impact non-steady recurrence and slip rates on forearc faults is the megathrust earthquake cycle, either through interseismic strain accumulation or coseismic triggering. Strain accumulation in the overriding plate during the interseismic period of the megathrust earthquake cycle has been

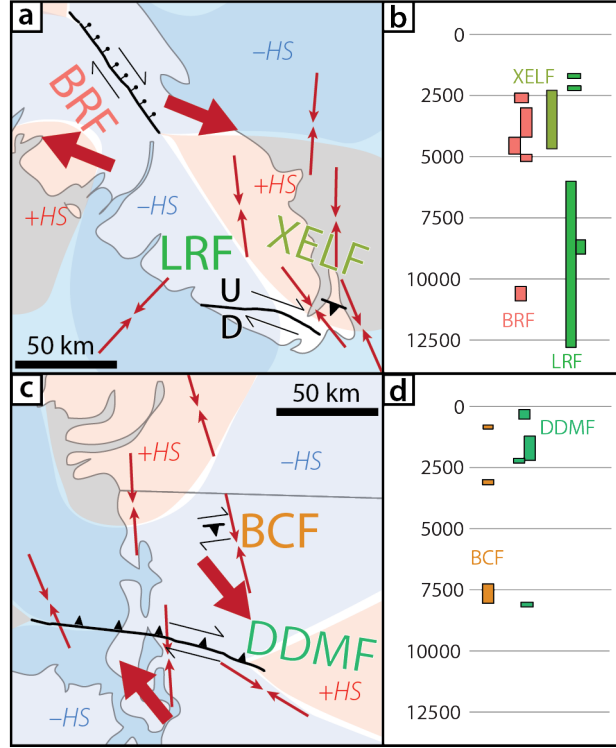


Figure 4.9: Comparison of the earthquake histories of forearc faults that may be impacted by Coulomb stress transfer. **A:** Simplified map showing locations and kinematics of faults on Vancouver Island, and the direction of maximum horizontal stress (small red arrows, after Balfour et al., 2011). Large red arrows show the direction of crustal motion resulting from right-lateral oblique slip on the BRF. Shading indicates areas of theoretical stress change from a BRF earthquake, with red for areas of increased maximum horizontal stress component (+HS), and blue for areas of reduced maximum horizontal stress component (-HS; after King et al., 1994). **B:** Comparison of earthquake event timing on the BRF and XELF, and those on the LRF. **C:** Simplified map showing locations and kinematics of the BCF and DDMF, and the direction of maximum horizontal stress (small red arrows, after Balfour et al., 2011). Large red arrows show the direction of crustal motion resulting from right-lateral oblique slip on the DDMF. Shading indicates areas of theoretical stress change from a DDMF earthquake, with red for areas of increased maximum horizontal stress component (+HS), and blue for areas of reduced maximum horizontal stress component (-HS; after King et al., 1994). **D:** Comparison of earthquake event timing between events on the BCF and those on the DDMF.

proposed to drive a component of permanent deformation in the upper plate in the form of forearc faulting (e.g., Fitch, 1972; Wang, 2000; Loveless & Meade, 2011; Symithe et al., 2015; Delano et al., 2017). I propose that longer interseismic periods, or gaps, may cause increased loading on forearc faults. This increased loading would increase the likelihood of failure, and therefore the number of earthquakes during long interseismic periods. Goldfinger et al. (2013) identified a pattern of four “superclusters” of smaller CSZ megathrust events that terminate in a large megathrust event, followed by a quiet period of ~ 1 -2 kyr.

If forearc faults were accumulating increased loading during these gaps, we would expect there to be

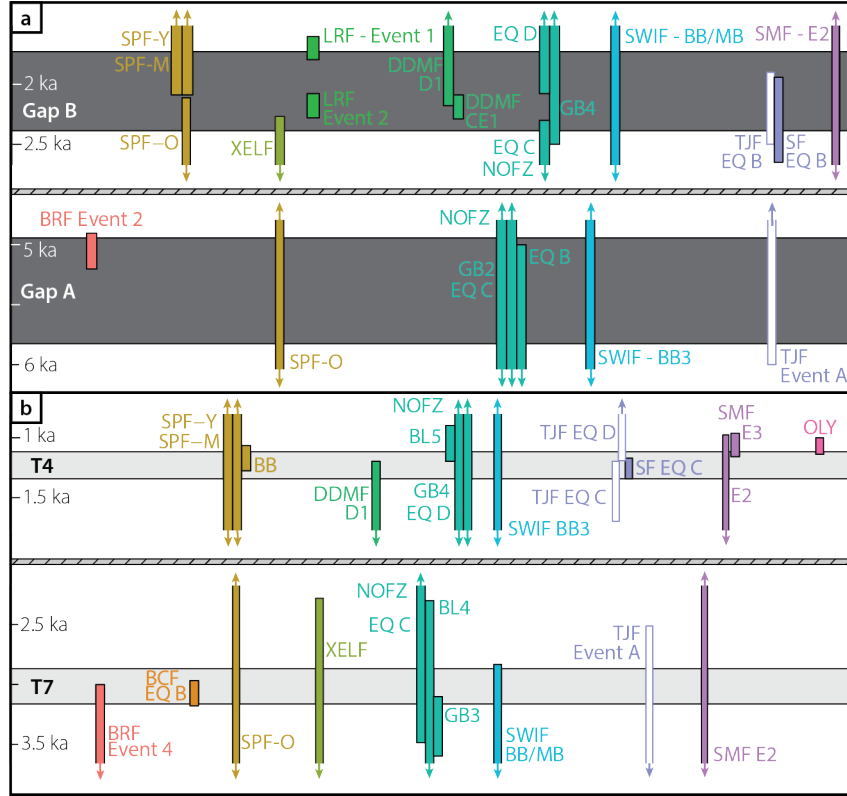


Figure 4.10: Comparison of the earthquake histories of the CSZ and forearc faults. **A:** Forearc fault events that could take place during Gaps A and B on the CSZ megathrust (see Figure 4.4). Event age ranges that extend beyond the figure are indicated by arrows (see Figure 4.4 for full extent). **B:** Forearc fault events that could overlap in time with CSZ megathrust events T4 and T7. Longer age ranges that extend beyond the figure are indicated by arrows (see Figure 4.4 for full extent).

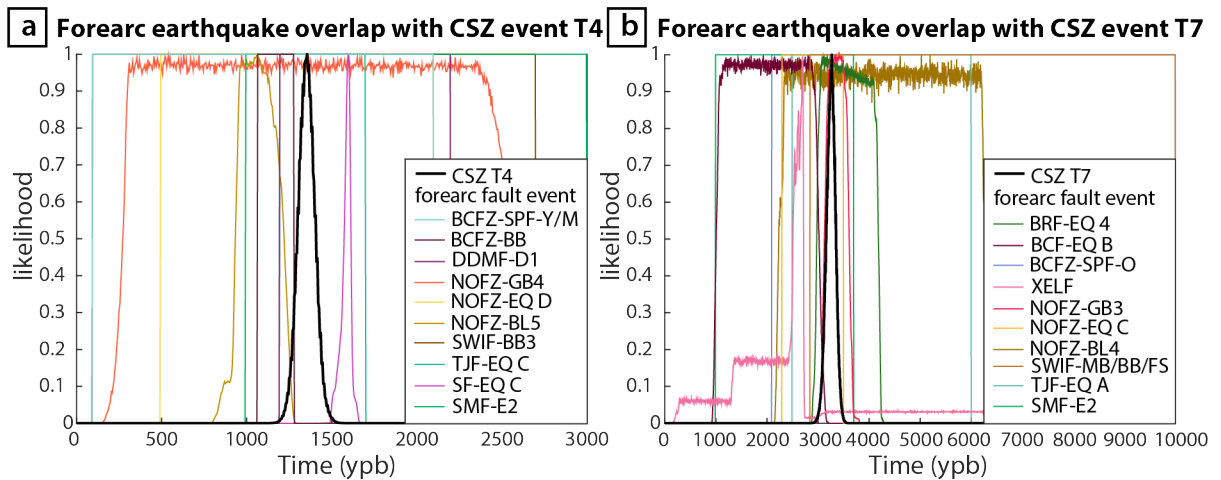


Figure 4.11: PDFs showing the timing of CSZ events and forearc earthquakes with potential overlap. The probability of overlap is listed in Table 4.4. Most forearc earthquakes have broad age ranges, and the few well-constrained forearc earthquakes have minimal overlap with the CSZ events.

more frequent earthquakes in the forearc during the gaps, releasing the excess strain that accumulated during the gap. The frequency of forearc earthquakes does not appear to change relative to the length of the CSZ interseismic period. However, I identify two quiet periods that might correspond to more frequent forearc earthquakes: Gap A, corresponding with the ~ 6 -5 ka period between clusters 3 and 2, and Gap B, corresponding with the ~ 2.5 -1.5 ka period between clusters 2 and 1 of Goldfinger et al. (2013, Figures 4.4, 4.10a). There are earthquakes on five faults that overlap in time with Gap A (~ 6 -5 ka; Figures 4.4, 4.10a). The best-constrained of these earthquakes is Event 2 on the BRF, which occurred ~ 5.2 -4.9 ka (Table 4.1). The earthquakes on the other four faults, BCFZ, NOFZ, SWIF, and TJF, are poorly constrained and may have occurred 100s-1000s of years before or after Gap A (Figures 4.4, 4.10a). There are earthquakes on eight faults that overlap in time with Gap B (~ 2.5 -1.5 ka; Figures 4.4, 4.10a). The timing of four of these earthquakes is well-constrained: Event 2 on the LRF (2.3-2.1 ka), Event 1 on the LRF (1.8-1.6 ka), CE1 on the DDMF (2.3-2.1 ka), and EQ B on the SF and TJF (2.7-1.9 ka and 2.5-1.9 ka, respectively). Earthquakes on the BCFZ, UPF strand of the DDMF, NOFZ, SWIF, and SMF may also have occurred during Gap B, but their timing is poorly constrained (Table 4.1; Figures 4.4, 4.10a). However, forearc earthquakes occur with similar frequency before, during, and after Gaps A and B (Figure 4.4). While individual faults have ruptured during these gaps, there is no pattern of multiple faults rupturing in response to higher overall strain in the forearc. Therefore, it is unlikely that forearc faults accommodate higher strain accumulation during longer interseismic periods.

Forearc earthquakes may also occur as a result of direct coseismic triggering by slip on the megathrust. Promotion of slip on forearc faults as a result of coseismic slip on the megathrust has been documented in other subduction zones (e.g., McCloskey et al., 2005; Fariás et al., 2011; Toda, Lin, & Stein, 2011; Hiratsuka & Sato, 2011; Imanishi et al., 2012). If coseismic triggering were driving regional patterns in forearc fault recurrence, we would expect to find CSZ megathrust events that overlap in time with multiple forearc earthquakes. However, the data do not support such an overlap, and any possible overlaps that exist overlap only because the timing of forearc earthquakes is poorly constrained. CSZ megathrust events overlap only with poorly constrained forearc earthquakes (Figure 4.4). There are however, two possible exceptions. There are two CSZ megathrust events that may each overlap with several well-constrained forearc earthquakes: T4, and T7 (Figures 4.4, 4.10b). T4, which occurred ~ 1.3 -1.1 ka (Table 4.1), may overlap with earthquakes on the BCFZ, DDMF, NOFZ, SWIF, SF/TJF, SMF, and/or OF (Figures 4.4, 4.10b). T7, which occurred ~ 3.2 -2.9 ka (Table 4.1), may overlap with the BRF, BCF, BCFZ, XELF, NOFZ, SWIF, TJF, and/or SMF (Figures 4.4, 4.10b). Even events that could possibly overlap within uncertainty may be statistically unlikely to occur at the same time, so I used statistical methods to compare earthquake age probability distribution functions (PDFs) for forearc faults and the CSZ megathrust and assess the likelihood that multiple forearc

earthquakes overlap with a CSZ event. I used a Monte Carlo simulation, adapted from Regalla et al. (2022) that calculates the likelihood that two earthquakes, each with their own PDF, overlap within a defined time window. I used a time window of 150 years, 30% of the average recurrence interval for the northern CSZ (Goldfinger et al., 2012) to ensure I was only considering the time of the selected CSZ megathrust and not the event before or after. The likelihood that five or more forearc earthquakes occurred within 150 years of either CSZ event is 0%. There is a 1% chance that CSZ event T4 overlapped with as many as four forearc events (BCFZ-BB, DDMF-D1, NOFZ-BL5, TJF-EQ C; Table 4.4), and a 1% chance that CSZ event T7 overlapped with as many as three forearc events (BRF-Event 4, NOFZ-EQ C, NOFZ-GB3; Table 4.4). The negligible likelihoods I calculate for overlap of multiple forearc earthquakes indicate that individual CSZ events are unlikely to trigger earthquakes throughout the forearc. This is consistent with the findings of Sherrod and Gombert (2014), who compared the timing of the Puget Sound regional cluster (see Section 4.2.5) and found that it was unlikely that the cluster was triggered by CSZ events T3, T4, or T5 (their events W, U, and S).

The only triggering scenario that is somewhat likely is that of a CSZ event coseismically triggering slip on an individual upper plate forearc fault (see Table 4.4 for specific faults). While there are some special cases where this overlap may exist, most forearc earthquakes do not appear to overlap with individual CSZ megathrust events, as their ages are more loosely constrained than the timing of CSZ events (Figure 4.4). I consider the case study of forearc events that may overlap with CSZ events T4 and T7. I calculated the likelihood that the individual events identified above overlapped with CSZ events T4 or T7 within 100 or 150 years (20-30% of the average recurrence interval for the northern CSZ; Goldfinger et al., 2012). Likelihood of overlap between CSZ event T4 and forearc earthquakes ranges from 0% to 57%, with most events ranging from ~10-20% (Table 4.4). Likelihood of overlap between CSZ event T7 and forearc earthquakes ranges from 0% to 51%, with most events ranging from 5% to 10% (Table 4.4). These models suggest that while triggering of individual forearc events is possible, it is unlikely to explain the non-steady recurrence and slip rates on all forearc faults.

4.5.1.4 Glacial Isostatic Adjustment

A fourth factor that may affect how much strain is accommodated in the upper plate is the lithospheric response to the advance and retreat of the Cordilleran ice sheet. Changes in differential stress due to glacial isostatic adjustment (GIA) have been shown to trigger slip on faults (e.g., NW Germany and Fennoscandia; Brandes et al., 2011; Olesen, 1988; Lagerbäck, 1990; Lund & Näslund, 2009), and modulate the timing of active faults (e.g., Basin and Range; Hetzel & Hampel, 2005; Hampel et al., 2007; Hampel, 2017). During glaciation, vertical stress increases due to the weight of the ice, and horizontal stress increases due to the

flexural bending of the crust (e.g., Steffen et al., 2014). During glacial retreat, vertical stress decreases quickly in response to the immediate removal of the ice load, but horizontal stress decreases more slowly due to the slow flexural response (e.g., Steffen et al., 2014). The relative magnitudes of the added vertical and horizontal stress depend on the size of the ice sheet (e.g., Johnston et al., 2002), and how these increased vertical and horizontal stresses contribute to the principal stresses depends on the orientation of those stresses (e.g., Steffen et al., 2014).

I evaluate whether the timing of CSZ forearc earthquakes could be impacted by GIA by comparing their behavior to published models of fault response to GIA by Hampel et al. (2009, 2010). Hampel et al. (2010) model cumulative slip on 30°-60° dipping thrust and normal faults before, during, and after a glacial load is imposed. These dips cover the majority of active fault dips for the 13 faults investigated here, which range from 30° to vertical. These models predict different fault slip responses depending on the relative viscosities of the upper mantle and lower crust. In Cascadia, the velocity of the upper mantle has been the subject of extensive modeling ($\sim 10^{18} - 10^{20}$ Pa s; e.g., James et al., 2000, 2009; Clague & James, 2002), but the viscosity of the lower crust is unknown. Because fault behavior in GIA models is so dependent on the relative viscosities, I consider two sets of models. The first set of models is based on a stronger lower crust than upper mantle, as suggested for tectonically active areas by Bürgmann and Dresen (2008). The second set of models is based on a weaker lower crust than upper mantle, consistent with modeled viscosities at the base of the crust for a forearc with a high sedimentation rate ($\sim 10^{15} - 10^{16}$ Pa s; Fernández-Blanco et al., 2020). The timing of glacial rebound varies with location (e.g., P. Wu & Johnston, 2000), so Hampel et al. (2009) model slip on 30°-60° dipping thrust faults above a weak lower crust, focusing on the impact of the location of a fault relative to the ice sheet.

First, I discuss the case for thrust faults and a strong lower crust. This scenario is relevant to most of the Cascadia forearc faults, which are thrust or oblique thrust (e.g., LRF, XELF, BCF, DDMF, SWIF, SF, TF, OF, SMF, NOFZ; Figure 4.1), consistent with a horizontal σ_1 . Published models for this scenario predict accelerated slip during glacial retreat followed by a cessation of slip for 10s of kyrs (Figure 4.11a; experiment 2 of Hampel et al., 2010). However, all CSZ forearc faults have continued to rupture in the ~ 15 kyrs since deglaciation (Figure 4.4), inconsistent with the predictions of this model.

Second, I consider thrust faults (e.g., LRF, XELF, BCF, DDMF, SWIF, SF, TF, OF, SMF, NOFZ; Figure 4.1) above a weak lower crust. Hampel et al. (2010) modeled thrust faults above a weak lower crust and predicted that slip would be suppressed during glaciation and then accelerate during glacial retreat before returning to background tectonic slip rates (Figure 4.11c; experiment 1 of Hampel et al., 2010). I use two datasets to evaluate this model. First, earthquakes are more frequent in the mid-late Holocene than during or shortly after glacial retreat. The only thrust faults that ruptured during or shortly after glacial retreat

are the SWIF, SF, and SMF (Figure 4.4). Second, the slip rates I calculated (Table 4.3; Figure 4.7) are generally higher in the mid-late Holocene than during or after glacial retreat, rather than the decreasing slip rates predicted by this model. The slip histories of thrust faults in the CSZ forearc are therefore inconsistent with the predictions of this model.

Third, I consider the special case of a thrust fault at the edge of the ice, such as the NOFZ (Figure 4.1). Numerical modeling suggests that a fault in this scenario would slip at approximately the same rate as its background tectonic state, with variations depending on the dip of the fault and whether the ice was in the hanging wall or the footwall (Figure 4.11e-f; Hampel et al., 2009). At ~ 15 ka, the NOFZ was located at approximately the edge of the Juan de Fuca lobe (Figure 4.1), so I compare the history of the NOFZ to models of a thrust fault at the edge of the ice (Hampel et al., 2009). However, the NOFZ is sub-vertical (e.g., Schermer et al., 2021), so I must consider both the scenario where the ice is in the hanging wall and the scenario where the ice is in the footwall. If the ice were in the hanging wall, we would expect slip on the NOFZ to be suppressed during glacial retreat and for 10s of kyrs after (Figure 4.11e; Hampel et al., 2009). The NOFZ experienced five earthquakes in the Holocene, clearly inconsistent with this model. If the ice were in the footwall, we would expect the NOFZ to approximately maintain its background tectonic slip rate (Figure 4.11f; Hampel et al., 2009). However, we do not know the background tectonic slip rate of the NOFZ, and the closed-interval slip rates I calculated are based on the same average displacement in each event, so I cannot accurately test this scenario.

Fourth, I consider normal faults above a strong lower crust. This scenario is relevant for the oblique transtensional BRF (Figure 4.1), which may have a more vertical σ_1 . Hampel et al. (2010) modeled normal faults above a strong lower crust and predicted that slip would be suppressed during glacial retreat and then accelerate for kyrs until returning to the background tectonic slip rate (Figure 4.11b; experiment 2 of Hampel et al., 2010). There is one earthquake on the BRF shortly after deglaciation followed by a ~ 5 -6 kyr quiet period and a cluster of earthquakes in the mid-Holocene (Figures 4.2, 4.4; Table 4.1; Chapter 3). If the mid-Holocene cluster represents post-deglaciation accelerated slip or “catch-up,” it could have returned to its background tectonic slip rate, but with paleoseismic data limited to the past ~ 15 kyrs, we do not know what that background slip rate is. The BRF may be consistent with this model.

Fifth, I consider normal faults like the BRF (Figure 4.1) above a weak lower crust. Hampel et al. (2010) modeled normal faults above a weak lower crust and predicted an increase in slip during glaciation, decreasing during glacial retreat, before stopping for 10s of kyrs (Figure 4.11d; experiment 1 of Hampel et al., 2010). The mid-Holocene cluster of earthquakes and higher slip rates on the BRF are inconsistent with this model.

The above interpretations demonstrate that the rupture histories of the CSZ forearc faults are inconsistent with almost all permutations of the GIA models. Only the BRF and NOFZ are possibly consistent with

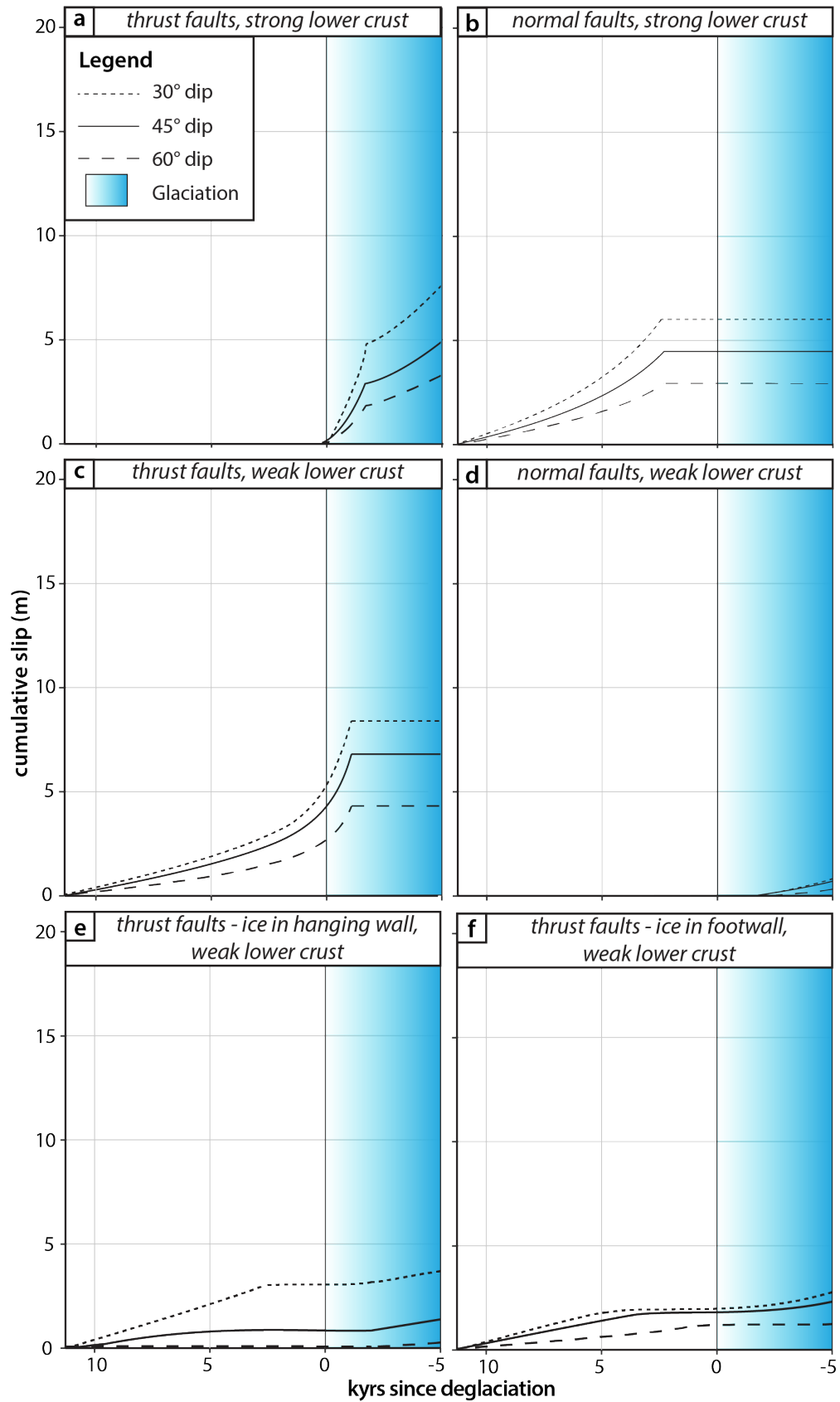
these models. It is important to note that the BRF is consistent with the model of a normal fault above a strong lower crust—realistic for the BRF as the surrounding crust consists of the accreted crystalline Wrangellia terrane (e.g., Coney et al., 1980; Clowes et al., 1987; DeBari et al., 1999; Groome et al., 2003). In contrast, the lack of slip rate change on the NOFZ may be consistent with the model of a thrust fault outside the ice above a weak lower crust. A difference in weak vs. strong crust may be possible given the bedrock geology that underlies these two faults. The NOFZ likely projects at depths into the basalt mixed and pelagic sediments of the Olympic Accretionary Complex (e.g., Delph et al., 2021), and is likely weak (e.g., Fernández-Blanco et al., 2020), whereas the BRF is located within strong crystalline bedrock.

4.5.2 Implications for regional seismotectonics

None of the four factors discussed above appear to be able to explain the variable recurrence on CSZ forearc faults. Plate reorganization hasn't occurred in the Holocene, and while there are some groups of forearc faults that are candidates for Coulomb stress transfer, this doesn't explain the patterns we see throughout the northern CSZ forearc. Individual CSZ ruptures could trigger individual forearc earthquakes, but again, this doesn't explain spatial or temporal patterns in recurrence and slip rates. GIA can only partially explain the behavior of two forearc faults.

While no single factor can explain all faults, it is possible that a combination of the above factors may be driving Holocene changes in recurrence and slip rate on CSZ forearc faults. For example, earthquakes on the BRF in response to glacial loading could then promote slip on the XELF and inhibit slip on the LRF through Coulomb stress transfer. However, no combination of the above factors can explain changes in recurrence or slip rate on 6 of the 13 faults (the SPF, SWIF, SF/TJF, OF, TF, or SMF). In particular, the slip rate of the TJF varies by 1-2 orders of magnitude with no clear cause. This behavior is inconsistent

Figure 4.12 (following page): Cumulative displacement on crustal faults during and after glaciation under different rheological conditions, modeled by Hampel et al. (2010) (panels A-D) and Hampel et al. (2009) (panels E-F). Displacement curves are shown for faults with 30°, 45°, and 60° dips. Note that the scales for both time and displacement are the same as in Figure 4.6. A-B: Thrust and normal faults, respectively, with a stronger lower crust than upper mantle, after Hampel et al. (2010), experiment 2. These models are relevant for comparison to forearc faults LRF, XELF, BCF, DDMF, SWIF, SF, TF, OF, SMF, NOFZ. C-D: Thrust and normal faults, respectively, with a weaker lower crust than upper mantle, after Hampel et al. (2010), experiment 1. These models are relevant for comparison to forearc fault BRF. E-F: Special case of thrust faults located just outside the ice sheet, with ice in the hanging wall (panel E), or the footwall (panel F), after Hampel et al. (2009). The NOFZ is located right at the maximum extent of the Cordilleran Ice Sheet, this may be the most applicable model. The NOFZ is sub-vertical, so the ice may have been in the hanging wall (panel E) or the footwall (panel F).



with the effects of GIA, and while it may be due to Coulomb stress transfer, we cannot test this without more complete rupture histories for neighboring faults and modeling. Further work is needed to explain this regional pattern of non-steady recurrence and slip rate and identify potential drivers.

4.6 Conclusions

Recurrence intervals and slip rates on CSZ forearc faults are non-steady in the Holocene. Non-steady recurrence intervals are inconsistent with periodic models of earthquake recurrence. Displacement and slip rates indicate that the time- and slip-predictable models do not describe CSZ forearc fault behavior either. Closed-interval slip rates for five CSZ forearc faults vary by an order of magnitude, indicating that there is a change in strain release in the mid-Holocene. Glacial isostatic adjustment and fault interactions may explain some of the variability for individual faults, but they require many assumptions and cannot explain why a majority of CSZ faults have long recurrence periods in the early Holocene and much shorter recurrence periods in the mid-late Holocene. The observed patterns of non-steady recurrence and slip rates cannot be described by the individual tectonic and climatic factors of changes in plate motion, interseismic strain accumulation or coseismic triggering from the CSZ megathrust, Coulomb stress transfer from neighboring faults, or glacial isostatic adjustment.

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Chapter 5

Conclusions

In this dissertation, I document the first geologic field evidence that the Beaufort Range fault is Quaternary-active. The BRF is in the northern Cascadia Subduction Zone (CSZ) forearc of central Vancouver Island and consists of >35-40 km of northwest-striking, primarily northeast-dipping fault strands occurring in discontinuous, en echelon and parallel sets and offset late Pleistocene to Holocene glacial, paraglacial, and post-glacial deposits. Through field mapping, topographic surveys, and paleoseismic trenching, I demonstrate that it actively accommodates right-lateral transtension, kinematics that are consistent over millennial timescales. Slip vectors determined from kinematic inversions of offset Pleistocene to Holocene geomorphic piercing lines and fault geometries measured in a paleoseismic trench are consistent with the 1946 Vancouver Island earthquake. This suggests that the BRF is a candidate source fault for this event.

Furthermore, I show through paleoseismic trenching that the BRF has hosted at least five surface-rupturing earthquakes since deglaciation. The oldest of these events occurred ~ 10.9 - 10.3 cal kyr BP, shortly after deglaciation. A ~ 5 - 6 kyr period of quiescence followed, before four additional earthquakes occurred in the mid-late Holocene (~ 5.2 - 4.9 , ~ 4.9 - 4.2 , ~ 4.2 - 3 , and ~ 2.8 - 0.7 cal kyr BP). Mid-late Holocene slip rates on the BRF may exceed 2 mm/yr, making the BRF the fastest-slipping fault in Canada. Slip magnitudes and total fault length suggest that the BRF is capable of hosting Mw 6.5-7.5 earthquakes, posing significant hazard to the nearby communities of Port Alberni, Parksville, and Nanaimo, BC, as well as to multiple hydroelectric power plants.

Finally, I show that multiple faults in the CSZ forearc exhibit non-steady recurrence. Slip rates on the BRF, Leech River fault, North Olympic fault zone, and Toe Jam Hill fault vary by an order of magnitude. These non-steady recurrence intervals and slip rates pose complications for probabilistic seismic hazard assessment. The behavior of Cascadia forearc faults is inconsistent with periodic, time-predictable, or slip-predictable models of earthquake recurrence, three of the most common time-dependent models in use. The mid-Holocene increase in slip rates may reflect an increase in strain accumulation in the northernmost Cascadia forearc. The behavior of individual faults may in part be a result of glacial isostatic adjustment of

Coulomb stress transfer from neighboring faults, but this cannot explain regional patterns in recurrence and slip rate. The causes of any Holocene changes in Cascadia forearc strain accommodation remain an open question.

Appendix A

Supporting Information for Chapter 2

A.1 Introduction

The Supporting Information contains (1) supporting text outlining criteria used to differentiate tectonic lineaments from those produced by glacial, gravitational, anthropogenic, or differential erosion processes, (2) supporting tables describing surficial units and radiocarbon samples, calculated displacements and slip vectors, and results of kinematic inversions, and (3) supporting figures illustrating tectonic and non-tectonic lineaments, and detailed fault mapping and structural measurements.

A.2 Lineament differentiation

We differentiated tectonic lineaments (Figure A.1) from lineaments that were produced by glacial processes such as glacial scouring, sub-glacial sedimentation (e.g., Figure A.2a), and plucking by performing three tests. First, we extracted elevation profiles both along and across scarp crests. While the elevation of crests of glacial lineaments (both erosional and depositional) typically decrease toward the glacial flow direction (down valley, toward 120° azimuth), fault related lineaments can cut across topography following the Rule of ‘V’s. Second, in the field we evaluated whether glacial and tectonic landforms were underlain by till or had glacially striated or scoured surfaces. Third, while both fault-related scarps and glacially plucked surfaces may have asymmetric profiles across the landform, glacially plucked surfaces face in the ice transport direction ($\sim 120^\circ$ azimuth), whereas fault-related lineaments primarily face uphill.

We differentiated lineaments that were produced by gravitational failure, such as landslides or sackungen, from tectonic lineaments by considering the context of the deposits around the lineaments. Landslide surfaces are often hummocky and disturbed, whereas fault scarps may offset surfaces of any preservation. Lineaments formed by landslide headscars or “toes” are typically curvilinear, whereas fault-related lineaments are typically linear. Lineaments formed by sackungen (e.g., Figure A.2b) typically occur near the top

of the range in parallel linear sets, whereas fault-related lineaments can occur at any elevation, often in splay or en echelon geometries, and produce vertical separation of the hillside.

We differentiated lineaments that were produced by anthropogenic disturbance from tectonic lineaments by comparing mapped lineaments with logging roads, timber hauling paths, and damage from heavy machinery on modern and historical air photos and road maps. Lineaments associated with logging roads or paths (e.g., Figure A.2c) typically have a flat base with an oversteepening of the lateral flanks from road construction, whereas fault-related lineaments do not.

We tested that mapped scarps were not the result of differential erosion of bedrock faults, bedding, or flow banding from tectonic lineaments using the following criteria. Scarps produced by differential erosion are not associated with vertical separation of the hillside. Vertical separation requires displacement of the ground surface. We also measured the inclination of any Nanaimo Fm. beds and Karmutsen flow tops located near scarps. Lineaments formed by differential erosion are typically co-located with steeply dipping bedding planes (e.g., within the Nanaimo Fm.), whereas fault-related lineaments can occur in beds of any dip magnitude, and do not have to be co-located with or parallel to bedding planes.

A.3 Figures

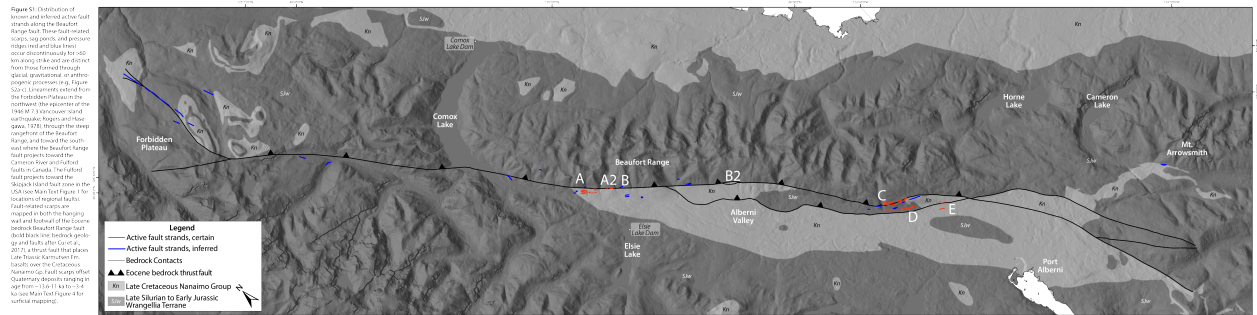


Figure A.1: Distribution of known and inferred active fault strands along the Beaufort Range fault. These fault-related scarps, sag ponds, and pressure ridges (red and blue lines) occur discontinuously for >60 km along strike and are distinct from those formed through glacial, gravitational, or anthropogenic processes (e.g., Figure A.2a-c). Lineaments extend from the Forbidden Plateau in the northwest (the epicenter of the 1946 M_S 7.3 Vancouver Island earthquake; Rogers & Hasegawa, 1978), through the steep rangefront of the Beaufort Range, and toward the southeast where the Beaufort Range fault projects toward the Cameron River and Fulford faults in Canada. The Fulford fault projects toward the Skipjack Island fault zone in the USA (see Figure 2.1 for locations of regional faults). Fault-related scarps are mapped in both the hanging wall and footwall of the Eocene bedrock Beaufort Range fault (bold barbed black line; bedrock geology and faults after Cui et al., 2017), a thrust fault that places Late Triassic Karmutsen Fm. basalts over the Cretaceous Nanaimo Gp. Fault scarps offset Quaternary deposits ranging in age from ~13.6-11 ka to ~3-4 ka (see Figure 2.4 for surficial mapping). *NB: This is a scaled-down version of Figure A.1. Please see supplemental files for the full size.*

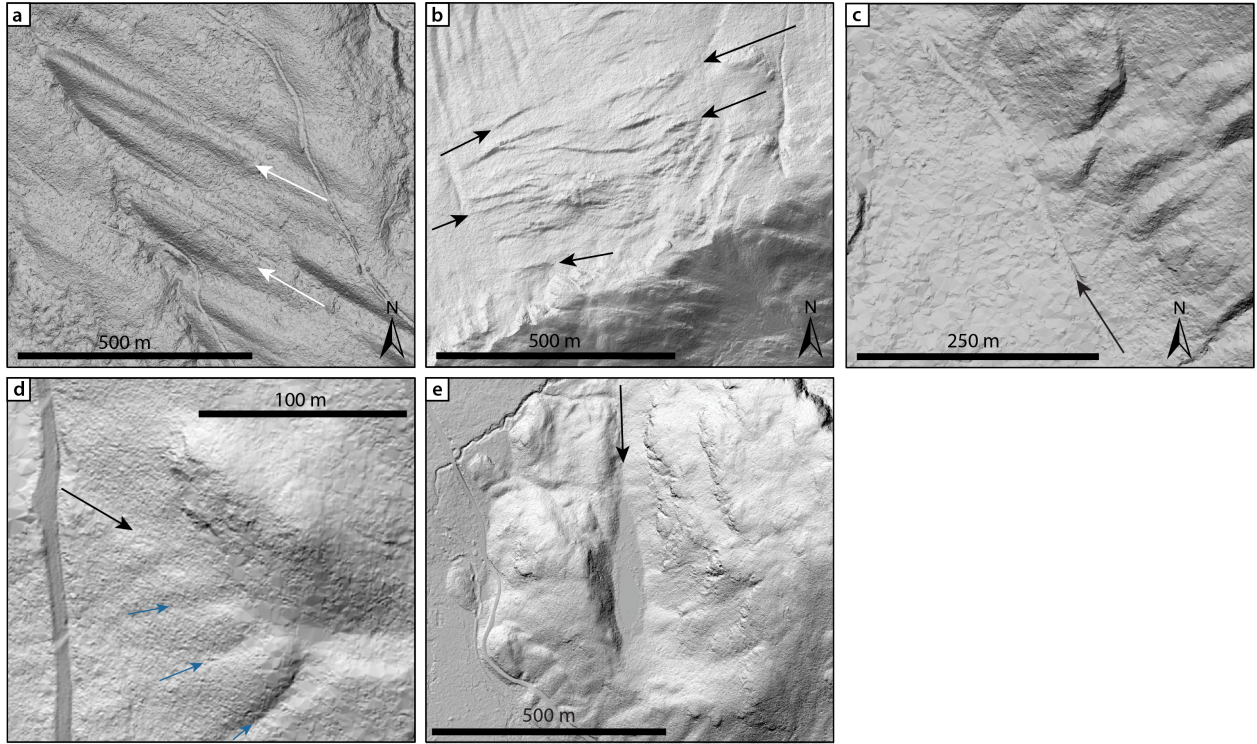
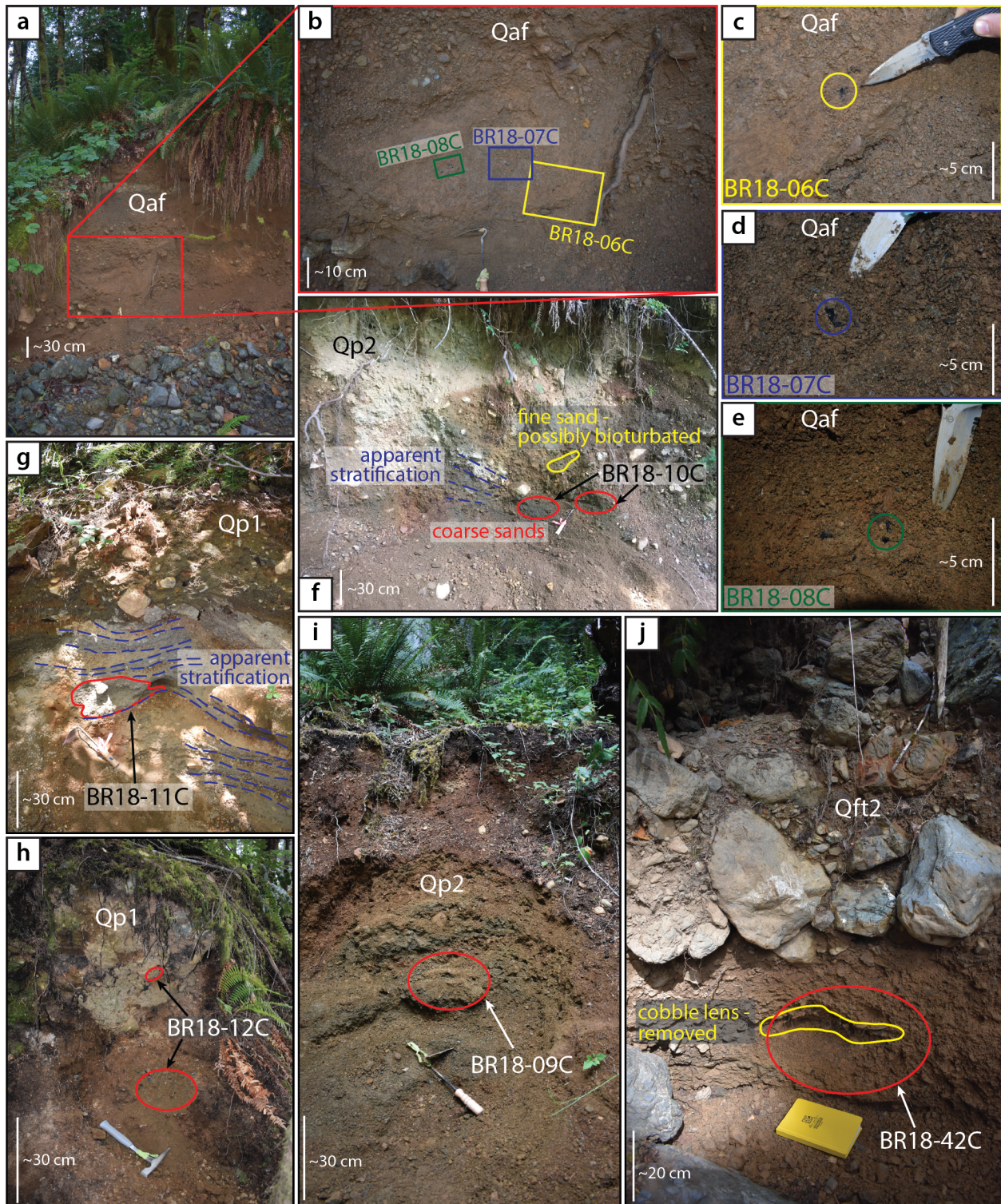


Figure A.2: Hillshaded lidar DEMs showing examples of non-tectonic (panels a-c) and tectonic (panels d and e) lineaments along the Beaufort Range. **A:** White arrows indicate streamlined glacial drumlins and lineaments. These lineaments trend down-valley toward $\sim 120^\circ$, sub-parallel to the valley glacier flow direction of Fyles (1963). **B:** Black arrows indicate sackungen, parallel linear scarps near ridge crests associated with gravitational failures after retreat of a glacial buttress (e.g., Li et al., 2010). **C:** Black arrow indicates a decommissioned logging road. The uphill side (right) is carved out, while the downhill side (left) is oversteepened due to deposition of excess material during road building. **D:** Black arrow points to a “bench,” or a flat, degraded topographic feature embedded in the high-gradient hillslope, that truncates several paleochannels (blue arrows). **E:** A linear depression developed in Karmutsen Fm. basalt that collects water (a sag pond), the trend of which is oblique to the range front.

Figure A.3 (following page): Photos of radiocarbon sample locations and surrounding deposits. **A:** Alluvial fan (Qaf) exposure in active stream channel where samples BR18-06C, -07C and -08C were collected. Deposit consists of interbedded cobbles and gravels with some finer sand beds and lenses. Beds are roughly horizontal and clast supported, with sub-angular to sub-rounded clasts. The matrix is similar throughout, red-brown with a muddy composition, likely composed of Fe oxides with clay and sands. The alluvial fan unit is 2-3 m thick and topped by a fluvial deposit, overlain by colluvium. **B:** Close-up of the location in panel A showing the individual sample locations within the Qaf deposit. Samples BR18-06C, -07C, and -08C were collected from a ~1 m wide by 30-40 cm thick lens of pebbles and coarse sands, ~1 m above the active channel floor, ~2 m below the fan surface. **C:** Sample BR18-06C in situ. **D:** Sample BR18-07C in situ. **E:** Sample BR18-08C in situ. **F:** Road cut exposure where sample BR10C was collected from an indurated coarse sand lens (outlined in red) within stratified sands and gravels (Qp2). **G:** Roadside exposure where sample BR18-11C was collected from sandy interbeds outlined in red (Qp1). Colluvium and detritus on the surface was removed, and bulk sediment sample was collected from freshly exposed sediments. **H:** Paraglacial deposit exposure (Qp1) where sample BR18-12C was collected. This deposit was very indurated, and required hammering to collect bulk sediment (area sampled outlined in red). **I:** Roadside exposure where sample BR18-09C was collected. Colluvium and detritus on the surface was removed, and bulk sediment sample was collected from freshly exposed Qp2 sediments (outlined in red). **J:** Fluvial terrace (Qft2) exposure in active stream channel where sample BR18-42C was collected from a pebble bed (outlined in red). The cobble lens indicated in yellow was plucked out prior to sampling due to an abundance of plant litter.



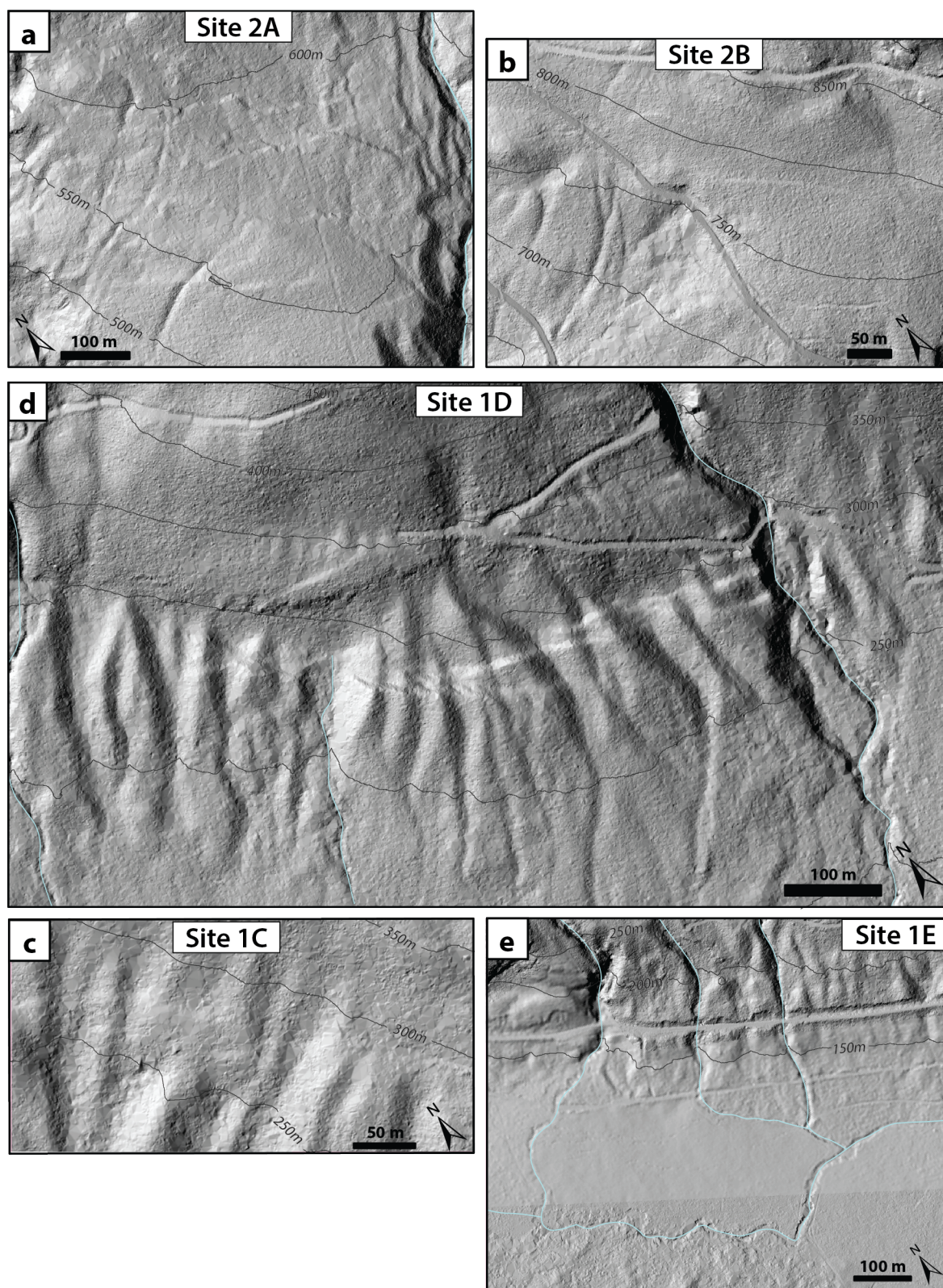


Figure A.4: Unannotated versions of bare earth lidar DEMs for Sites 2A, 2B, 1C, 1D, and 1E.

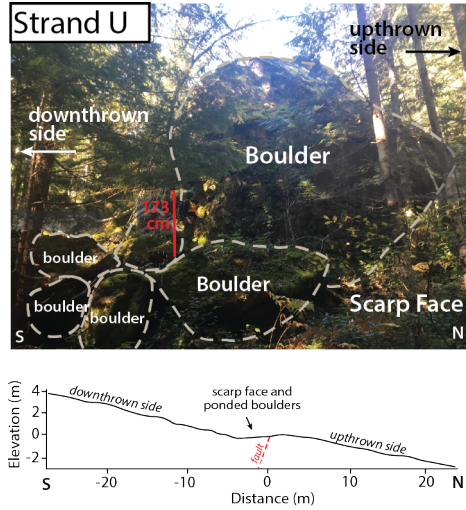


Figure A.5: Field photos of a fault scarp (Strand U) at Site 2A. **A:** Photograph showing large boulders (~ 1.5 – 8 m; outlined in dashed gray) ponded against fault scarp U at Site 2A. See Figure 2.6 for location. **B:** Topographic profile across strand U.

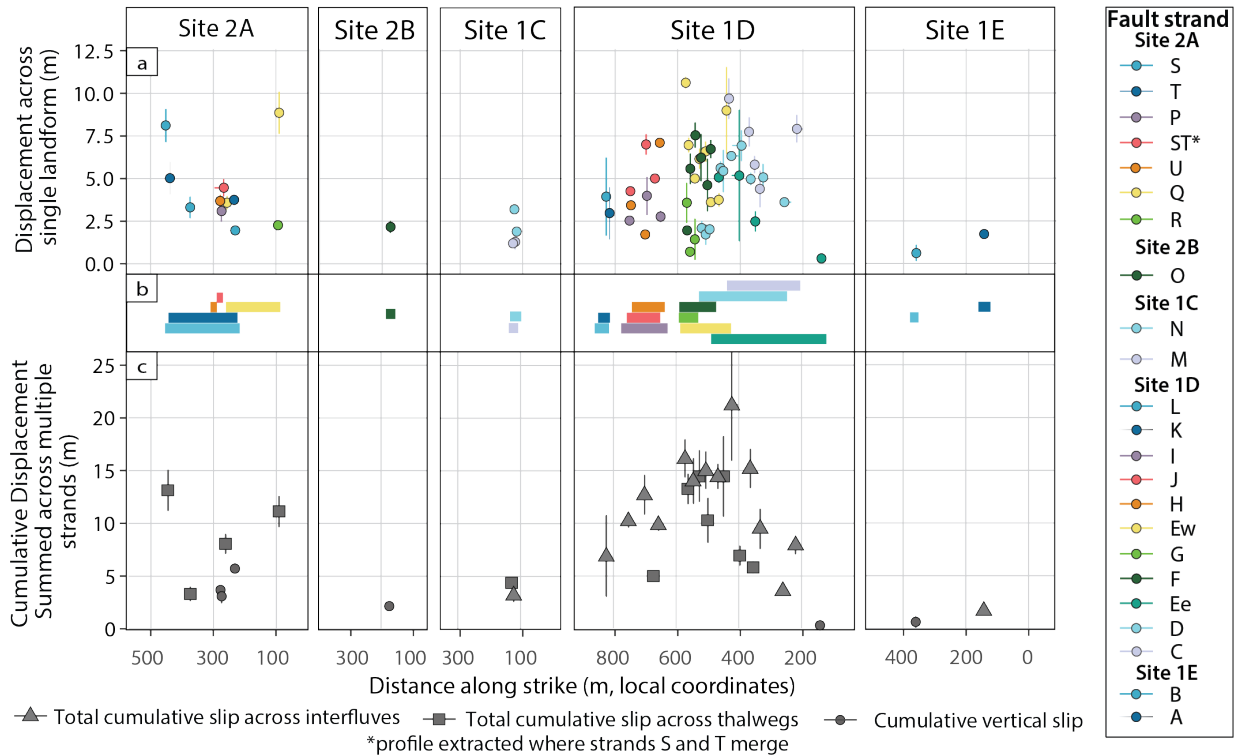
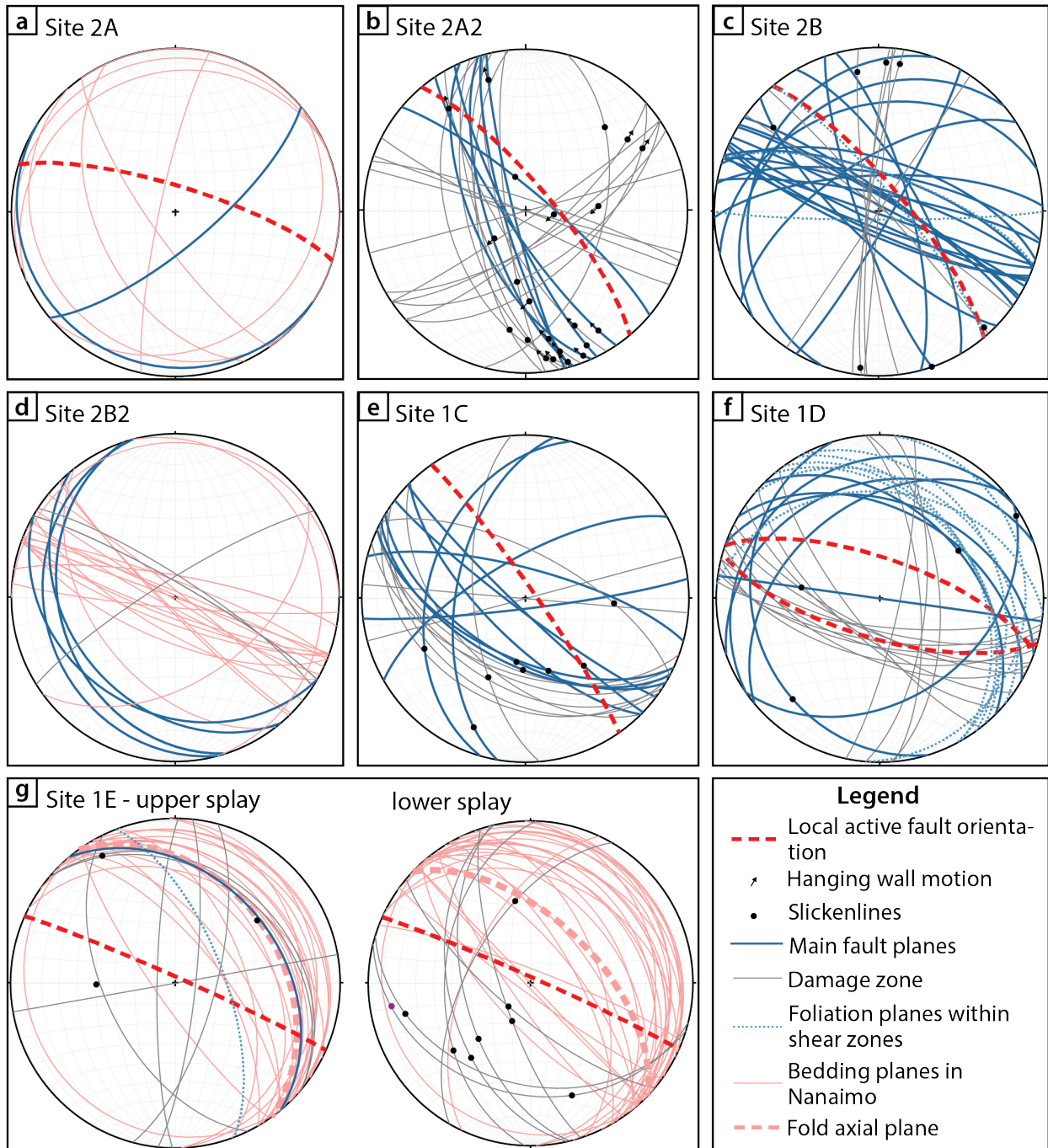


Figure A.6: Displacement measured across the Beaufort Range fault at Sites A–E. Where error bars are not visible, the magnitude of error is smaller than the diameter of the symbol. **A:** Displacement of geomorphic landforms across individual fault strands (see Figure 2.7). Y axis error bars in panel c reflect 1 standard deviation calculated via a Monte Carlo simulation (see offset data in Table A.2). **B:** Mapped extent of individual fault strands along strike (see detailed mapping in Figures 2.5 and 2.6). **C:** Cumulative displacement of geomorphic landforms summed across all surveyed fault strands. Error bars reflect the propagated sum of the errors on individual strands. Note that interfluvial crests (triangles) show more displacement than channel thalwegs (squares). Straight-line profile (circles) do not capture any right-lateral displacement.

Figure A.7 (following page): Lower hemisphere equal-area projections showing structural measurements in bedrock at Sites A-E (presented northwest to southeast; see Figure 2.4 and Figure A.1 for locations). **A:** At Site 2A, bedrock faults and bedding planes are mostly gently dipping, in contrast with the active fault that is steeply dipping. Steeply-dipping bedding at this site is part of the upturned limb of a syncline in the footwall of the Eocene bedrock thrust fault. **B:** At Site 2A2, bedrock fault planes have similar dip magnitudes to the active fault, but opposite dip directions. Slickenlines and Riedel shear orientations (light gray) at Site 2A2 primarily record right-lateral, northeast-up motion. **C:** At Site 2B, the majority of bedrock fault planes are $\sim 20^\circ$ oblique to the active fault strike. **D:** Site 2B2 is an exposure of Nanaimo bedding in a footwall syncline. **E:** At Site 1C, bedrock fault planes primarily dip southwest, and strikes are $>10^\circ$ apart from the active fault orientation. **F:** Site 1D records gently-dipping foliation and main bedrock fault planes, in contrast to steeply-dipping active fault planes. **G:** Site 1E records two footwall synclines in Nanaimo Fm., related to the two splays of the Eocene bedrock thrust fault (see Figure 2.4b).



A.4 Tables

Table A.1: Sediment and surface morphology descriptions of units mapped in Figure 2.4.

Unit	Sediment description ¹	Surface morphology	Unit Interpretation	Radiocarbon Samples
Pleistocene till (Qt)	Indurated, poorly-sorted, matrix-supported diamict. Matrix is primarily clay to silt with occasional fine sand and displays occasional fissility. Clasts are sub-rounded to rounded and consist of fine to medium pebbles (50-70%), very coarse pebbles (20-40%), cobbles (~10%), and rare boulders. Clasts derived from both local Karmutsen Fm basalt, and exotic lithologies (e.g., diorite). This deposit is at least 1-2 m thick, and up to 40 m thick in some locations.	Smooth mantle over bedrock, mirroring bedrock morphology and thinning over prominent bedrock features (such as basalt flows). Till primarily covers the top of the range and extends down to ~150 masl at the southern end of the range, and the valley floor (~400 masl) at the northern end of the range. Buried by all younger Quaternary units. Deeply incised by active or recently active channels, with minor abandoned channels preserved.	Subglacial traction till deposited along the hillside during the Fraser glaciation.	—
Pleistocene kame (Qk)	Indurated, poorly to moderately sorted stratified sands and gravels. Primarily matrix-supported with clast supported cobble interbeds. Matrix is clay-rich with sandier pockets. Clasts are sub-rounded to sub-angular and consist of pebbles (~70%), cobbles (25%), and boulders (5%). Both local Karmutsen Fm. basalt and exotic lithologies represented. Beds are 5-10 cm thick with occasional reverse grading. This deposit is 20-40 m thick.	Evenly spaced terrace risers approximately 20 m tall and 50 m wide. Terraces are fairly smooth, and trend ~120° consistent with other glacial lineation orientations. This unit is expressed as discontinuous outcrops at 150-300 masl elevation on the range front, inset by the younger paraglacial units deposited in channels incised through the kame.	Terraces formed by meltwater streams between glacier and range-front during valley glacier retreat.	—

(continued)

Unit	Sediment description ¹	Surface morphology	Unit Interpretation	Radiocarbon Samples
Pleistocene to Holocene hummocky moraine (Qhm) ²	Stratified sands and gravels containing Karmutsen basalt clasts and other lithologies. <i>This deposit minimally exposed in outcrop. Mapped primarily based on surface morphology.</i>	Chaotic surface comprised of many small depressions and mounds, 10 m tall and 10-50 m wide. Mounds are both round and oblong, are discontinuous, and have no preferred orientation. Located on the valley floor.	Glacial deposit deformed under stagnant ice during glacial retreat. Coeval with Qp1.	—
Pleistocene to Holocene paraglacial deposits, old (Qp1)	Indurated, poorly-sorted, stratified sands and gravels. Primarily clast-supported with rare matrix-supported well sorted interbeds and occasional sand lenses. Matrix throughout is primarily silt to fine sand, with minimal clay. Clasts are sub-angular to sub-rounded and consist of sand to fine pebbles (40-60%), medium to very coarse pebbles (30-35%), and cobbles (10-25%) with rare rounded small boulders. Clast lithologies include local Karmutsen Fm. basalt, exotic lithologies (including diorite, chert, and assorted volcanics), and clasts of till. Weak to moderate stratification sub-parallel to the range-front with some aligned clasts and paleo channel forms.	Wedge- and cone-shaped slope-foot accumulation at the base of the range-front (~150 - 500 masl). Cone tips are at ~300 or ~500 masl and are not co-located with active channels. This unit buries Qt, is inset into Qk, is inset by Qp2 and Qaf, and is extensively incised by younger channels associated with Qp2 and Qaf.	Paraglacial unit derived from unstable hillslope sediments and Qt during valley glacier retreat. Coeval with Qhm.	BR18-11C, BR18-12C
<i>Pleistocene to Holocene paraglacial deposits, old (Qp1), continued</i>	Moderately to strongly cemented. At least 4 m thick based on most continuous exposure, may extend up to ~8-12 m thick.			

(continued)

Unit	Sediment description ¹	Surface morphology	Unit Interpretation	Radiocarbon Samples
Pleistocene to Holocene paraglacial deposit, young (Qp2)	Thinly-bedded, clast-supported, stratified sands and gravels with occasional coarse sand lenses and interbeds. The deposit is matrix-poor, with alternating orange oxidized clay-rich and darker gray silty-sandy matrix. Clasts are mostly sub-angular, and consist of pebbles (~80%) and cobbles (~20%). Clasts are primarily local Karmutsen Fm. basalt. Stratification gently dips down-slope.	Smooth surface blanketing the bottom of the range front (120-290 masl). Deposit heads in channels incised into Qt, Qk, and Qp1 and is buried by Qaf. Minor rilling primarily below anthropogenic disturbance (i.e. roads). Deeply incised by active channels.	Paraglacial unit generated from reworked Qp1 and Qt in water-rich and later mud-rich flows.	BR18-09C, BR18-10C
<i>Pleistocene to Holocene paraglacial deposit, young (Qp2), continued</i>	Younger strata within this unit contain little to no matrix. Clasts consist of sand (20-30%), fine pebbles (30-40%), and medium pebbles (40-50%). Beds are sub-horizontal, have reverse grading and have scour surfaces at the base. This deposit is less than 15 m thick.			

(continued)

Unit	Sediment description ¹	Surface morphology	Unit Interpretation	Radiocarbon Samples
Holocene alluvial and fluvial deposits (Qaf)	Stratified alluvial deposits. Clast-supported and poorly to moderately sorted with common sand lenses. Matrix is clay-rich with silt throughout and minor sands. Clasts are angular to sub-rounded and consist of fine pebbles (~80%) and medium to coarse pebbles (20%), interbedded with cobble and boulder layers. Clasts consist of local Karmutsen Fm. basalt with minor local Nanaimo Fm. and exotic lithologies. Beds are sub-horizontal, with cross-bedding present in some sand lenses and imbricated clasts in the direction of flow. This deposit is less than 10 m thick.	Broad, convex, gently-sloping fans that are headed in active or recently active channels. Fan surfaces have preserved bar and swale morphology. This unit is located on the valley floor where channels exit the range front (~500 masl at the northern end of the range and ~300-400 masl at the southern end). Buries Qp2 and Qhm, but inset into Qt, Qp1, and Qp2. Moderate incision by active channels and buried in places by Qft2.	Alluvial fans associated with active or recently active channels.	BR18-06C, BR18-07C, BR18-08C
Holocene fluvial terraces, old (Qft1)	Moderately to well-sorted deposits with minimal matrix and large, rounded cobbles and boulders interbedded with pebbles. Beds are sub-horizontal. Total thickness up to 40 m comprised of 5-10 m terraces.	Multiple generations (up to 5) of terraces within active channels on the range front (~250-750 masl). This unit is inset into Pt and incised by Qaf. Terrace width is controlled by the width of the canyon on the range front.	Fluvial terraces associated with active or recently active channels.	—

(continued)

Unit	Sediment description ¹	Surface morphology	Unit Interpretation	Radiocarbon Samples
Holocene fluvial terraces, young (Qft2)	Stratified fluvial deposits. Clast-supported and moderately sorted. Matrix is clay-rich with silt throughout and minor sands. Clasts are angular to sub-rounded and consist of medium to coarse pebbles (~80%) and cobbles (20%), interbedded with sand and boulder layers. Beds are sub-horizontal. This deposit is less than 5 m thick.	Minor terraces in active channels. This unit is located near the base of the rangefront (~300-400 masl). Inset into Qp1 and Qp2, and grades laterally into Qaf. Moderate incision by active channels.	Fluvial deposits associated with active or recently active channels.	BR18-42C
Holocene landslides and slumps (Qls)	<i>This deposit not exposed in outcrop. Mapped based on surface morphology.</i>	Curvilinear headscarps are steeper than surrounding hillsides. Deposit surfaces are hummocky and have more gentle slopes than surrounding hillsides. Deposit toes are oversteepened. These deposits are lower on the rangefront (~400-700 masl) and bury Qt and Qaf while inset into Qt. No visible incision.	Mass wasting deposits along planar or curved surfaces on hillslopes or into fluvial channels.	—
Holocene scree fans (Qsf)	<i>This deposit not accessible in outcrop. Mapped based on surface morphology.</i>	Small, overlapping fans of scree. Rough surfaces with large clasts (cobble to boulder sized) exposed. Developed at the rangefront to hillslope transition (~750 masl). Deposited on top of Qt.	Rockfall deposits derived from exposed bedrock along the top of the range.	—

¹Clast sizes are reported following the modified Wentworth scale after Lane (1947).²Eyles et al. (1999).

Table A.2: BRF displacements and slip vectors determined from reconstructions of offset geomorphic piercing lines. *Table A.2 is available as an Excel file.*

Table A.3: Nodal planes and P- and T-axis orientations for the 1946 Vancouver Island earthquake, and for kinematic inversions of BRF displacement data.

Solution	Nodal Plane 1 (strike/dip)	Nodal Plane 2 (strike/dip)	P-axis (trend/plunge)	T-axis (trend/plunge)	Slip vector (trend/plunge)¹
1946 focal mechanism A ²	319/79 NE	228/85 N	183/12	274/05	183/05
1946 focal mechanism B ²	332/66 NE	233/70 NW	191/32	283/02	143/20
1946 focal mechanism C ²	330/67 NE	222/36 NW	198/58	080/17	114/54
Site 2A ³	276/78 N	181/64 W	141/27	046/09	096/10
Site 1C ³	321/75 NE	199/26 NW	204/54	068/27	135/26
Site 2D ³	292/66 NE	160/34 SW	165/61	040/17	107/15
Site 2E ³	294/75 NE	187/43 W	164/45	053/20	109/19
All Sites ³	296/83 NE	199/43 NW	170/37	058/26	109/48

¹Calculated slip vector along NW-SE striking plane

²Focal mechanism solutions for the 1946 Vancouver Island earthquake (Rogers & Hasegawa, 1978)

³FaultKin inversions for the Beaufort Range fault, from offset piercing lines, this study

References

- Cui, Y., Miller, D., Schiarizza, P., & Diakow, L. (2017). *British Columbia digital geology* (Tech. Rep. No. 2017-8). British Columbia Ministry of Energy, Mines and Petroleum Resources.
- Eyles, N., Boyce, J. I., & Barendregt, R. W. (1999). Hummocky moraine: Sedimentary record of stagnant Laurentide Ice Sheet lobes resting on soft beds. *Sedimentary Geology*, 123(3-4), 163–174. (Publisher: Elsevier) doi: 10.1016/S0037-0738(98)00129-8
- Fyles, J. G. (1963). Surficial geology of Horne Lake and Parksville map-areas, Vancouver Island, British Columbia. *Geological Survey of Canada Memoir*, 318, 142.
- Lane, E. (1947). Report of the subcommittee on sediment terminology. *Transactions of the American Geophysical Union*, 28(6), 936–938.
- Li, Z., Bruhn, R. L., Pavlis, T. L., Vorkink, M., & Zeng, Z. (2010). Origin of sackung uphill-facing scarps in the Saint Elias orogen, Alaska: LIDAR data visualization and stress modeling. *Bulletin of the Geological Society of America*, 122(9-10), 1585–1589. (ISBN: 0016-7606) doi: 10.1130/B30019.1
- Rogers, G. C., & Hasegawa, H. S. (1978). A second look at the British Columbia earthquake of June 23, 1946. *Bulletin of the Seismological Society of America*, 68(3), 653–676.

Appendix B

Additional Supporting Information for Chapter 2

B.1 Read Me file

The Dryad data repository (Lynch et al., 2023) contains everything needed to recreate the 3D offset calculations of Chapter 2.

In order to calculate the 3D offset of your own geomorphic piercing points, you need:

1. topographic profiles as XYZ points (can be surveyed in the field or extracted from a DEM)
2. strike and dip of your fault plane(s)
3. selections of points to include in regressions

Also included is a script to calculate the strike and dip of a fault plane from surveyed scarp midpoints: Script B1 “`fault_plane_regression.R`”

To account for the uncertainty in profile selection, I had other members of my lab select the points they would include in each regression by visually assessing whether individual points reflected the linear geomorphic feature in question, or whether they reflected a degraded scarp crest or colluvial wedge.

1. The script to share 3D plots of your profiles is Script B2 “`profile_viewer.Rmd`”
2. knit to html to share with others
3. instructions are included for others to install the required tools for viewing the 3D plots

The script to calculate 3D offset is Script B3 “`3d_regression.R`”

The raw profiles and subgroups of fault midpoints included in Chapter 2 are also included in the Dryad data repository.

References

- Lynch, E. M., Regalla, C., Morell, K. D., Harrichhausen, N., & Leonard, L. J. (2023). *Three-dimensional offsets of geomorphic piercing lines displaced by the Quaternary-active Beaufort Range fault, northern Cascadia forearc, BC, Canada*. Dryad. (Dataset) doi: <https://doi.org/10.5061/dryad.nvx0k6dvc>

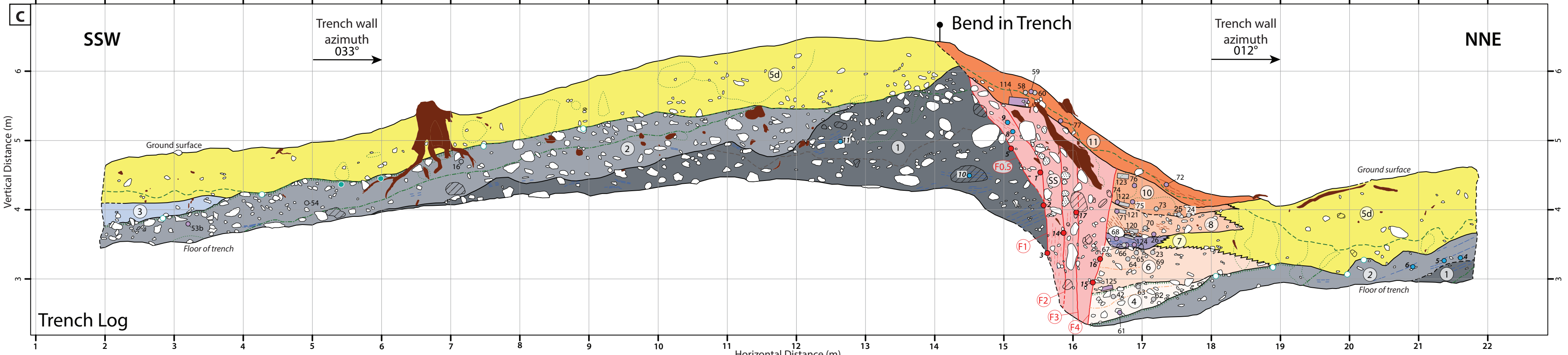
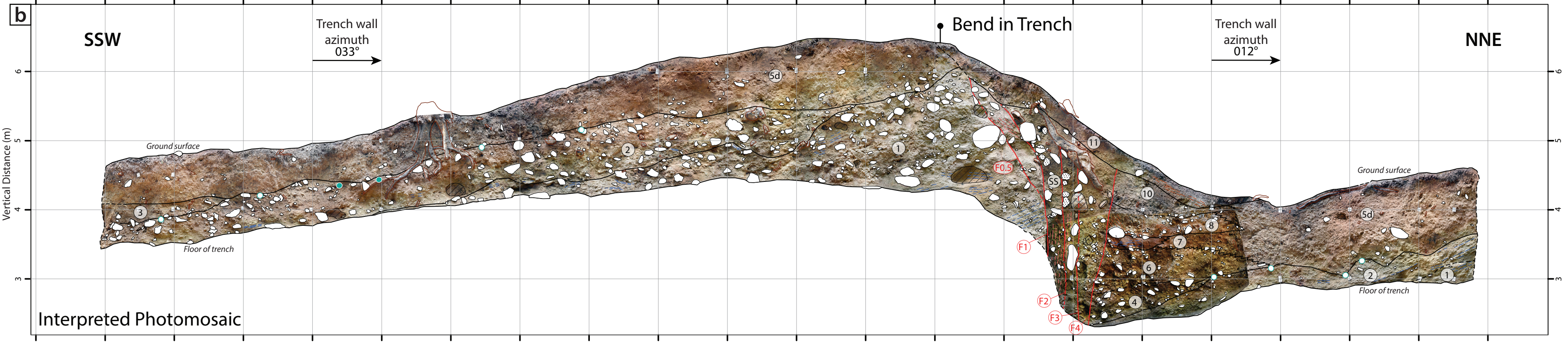
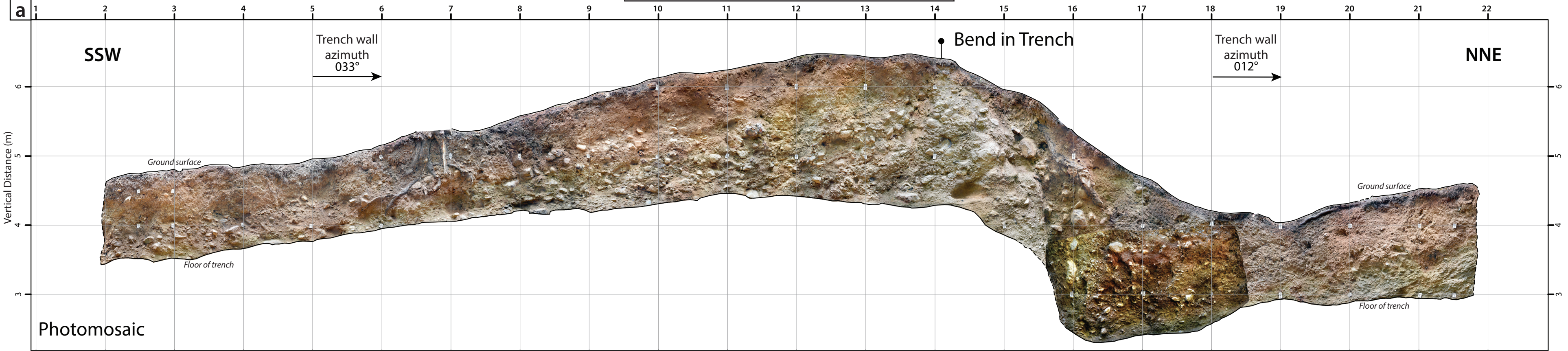
Appendix C

Supporting Information for Chapter 3

C.1 Figures

C.1.1 Supplemental trench logs

Beaufort Range Fault, Woodlot Trench
Slice 1 (reflected horizontally)



Legend

Contacts

- Contact, certain
- Contact, approximate
- Fault, certain
- Fault, location uncertain
- Base of soil B horizon
- Bioturbated soil
- Base of buried soil developed on Unit 4
- Base of buried soil developed on Unit 6
- Base of buried soil developed on Units 1 & 2
- Strong orange oxidized horizon
- Depositional fabrics in colluvial wedges
- Depositional fabrics in other units
- Deformational fabrics

Clasts & Roots

- Clasts
- Missing clasts
- Clasts of till
- Roots

Trench Unit

- 11 Colluvial Wedge 5
- 10 Colluvial Wedge 4
- 9 Silty Sand 3
- 8 Colluvial Wedge 3
- 7 Silty Sand 2
- 6 Colluvial Wedge 2
- 5b Silty Sand 1
- 4 Colluvial Wedge 1
- 3 Gravel
- 2 Buried Colluvium developed on Unit 1
- 1 Till and Paraglacial Deposits

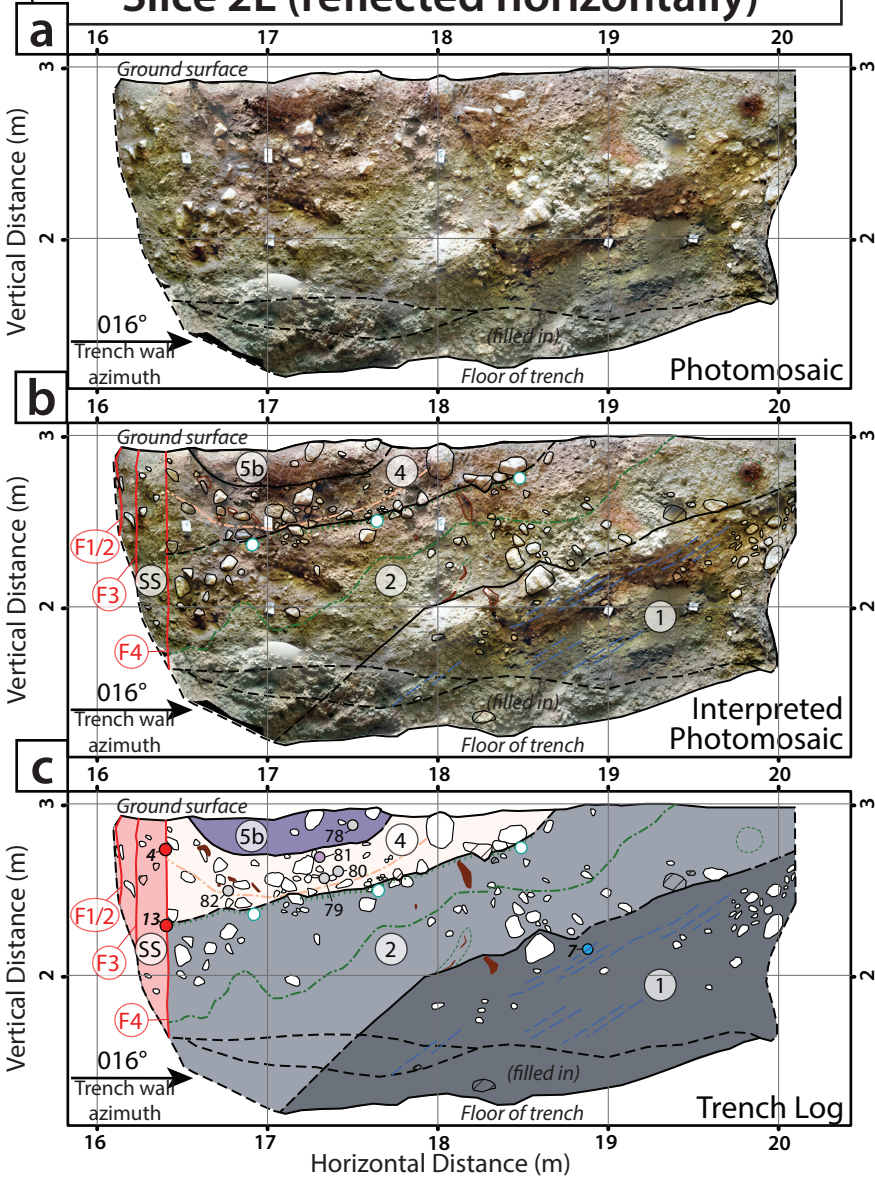
Structural Measurements & Samples

- Structural measurement, fault plane
- Structural measurement, depositional fabric
- Surveyed paleochannel (margin; thalweg)
- ¹⁴C Macro Sample, collected
- ¹⁴C Macro Sample, dated
- Bulk Sediment Sample for ¹⁴C, collected
- Bulk Sediment Sample for ¹⁴C, dated
- Sheared Sediment & Fissure Fill, Undivided
- Colluvium
- Scarp-derived Colluvium, Undivided
- Sandy Pebbles
- Debris Flow Deposit

Legend includes all mapped units. Not all units crop out in all walls.

F1 Major faults labeled F0.5-5

Beaufort Range Fault, Woodlot Trench Slice 2L (reflected horizontally)



Legend

Contacts

- Contact, certain
- - - Contact, approximate
- Fault, certain
- - - Fault, location uncertain
- - - Base of soil B horizon
- - - Bioturbated soil
- - - Base of buried soil developed on Unit 4
- - - Base of buried soil developed on Unit 6
- - - Base of buried soil developed on Units 1 & 2
- - - Strong orange oxidized horizon
- - - Depositional fabrics in colluvial wedges
- - - Depositional fabrics in other units
- - - Deformational fabrics

Clasts & Roots

- Clasts
- ▨ Missing clasts
- ▤ Clasts of till
- Roots

Structural Measurements & Samples

- Structural measurement, fault plane
- Structural measurement, depositional fabric
- ¹⁴C Macro Sample, collected
- ¹⁴C Macro Sample, dated
- Bulk Sediment Sample for ¹⁴C, collected
- Bulk Sediment Sample for ¹⁴C, dated

Trench Unit

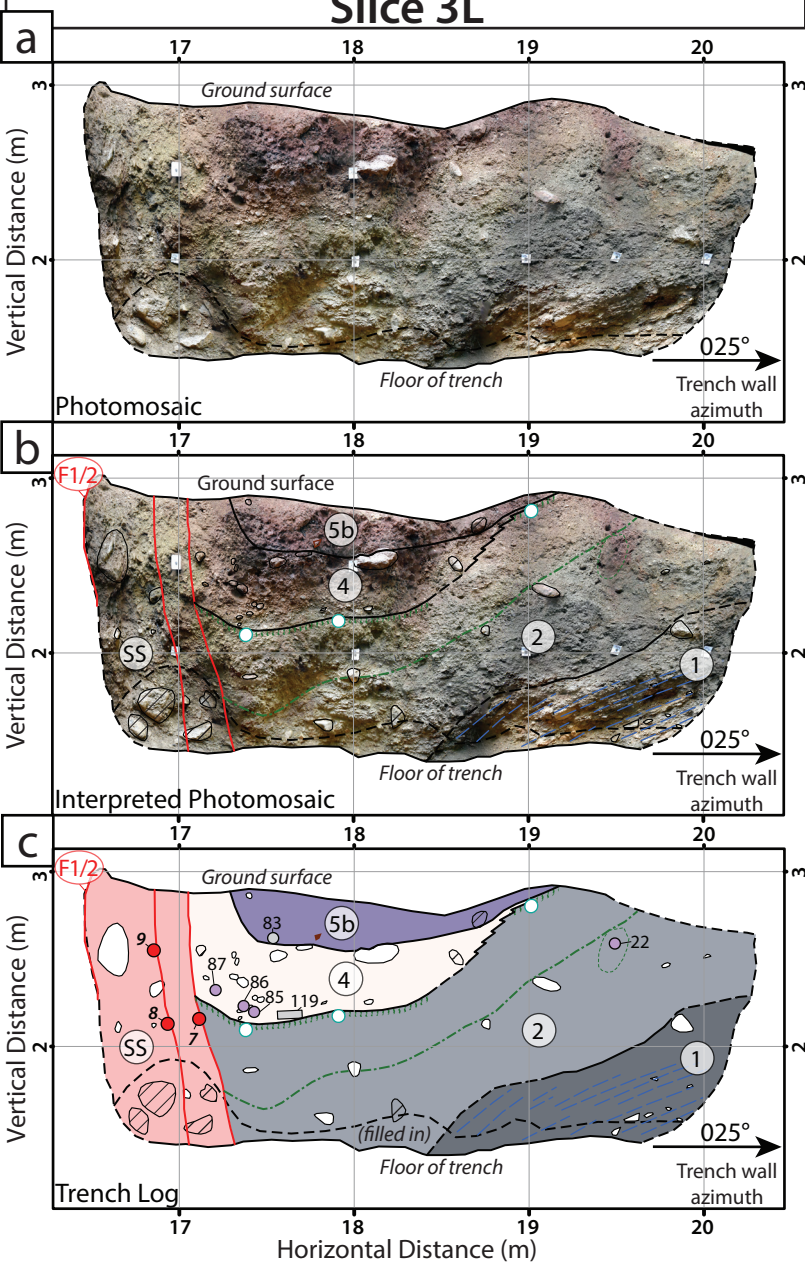
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- 10 Colluvial Wedge 4
- 9 Silty Sand 3
- 8 Colluvial Wedge 3
- 7 Silty Sand 2
- 6 Colluvial Wedge 2
- 5b Silty Sand 1
- 4 Colluvial Wedge 1
- 3 Gravel
- 2 Buried Colluvium developed on Unit 1
- 1 Till and Paraglacial Deposits
- SS Sheared Sediment & Fissure Fill, Undivided
- 5d Colluvium
- 5c Scarp-derived Colluvium, Undivided
- 5a Sandy Pebbles
- 3a Debris Flow Deposit

Legend includes all mapped units. Not all units crop out in all walls.

- Surveyed paleochannel (margin; thalweg)
- F1 Major faults labeled F0.5-5

Beaufort Range Fault, Woodlot Trench

Slice 3L



Contacts

- Contact, certain
- - - Contact, approximate
- Fault, certain
- - - Fault, location uncertain
- - - Base of soil B horizon
- Bioturbated soil
- - - Base of buried soil developed on Unit 4
- - - Base of buried soil developed on Unit 6
- - - Base of buried soil developed on Units 1 & 2
- Strong orange oxidized horizon
- - - Depositional fabrics in colluvial wedges
- - - Depositional fabrics in other units
- - - Deformational fabrics

Clasts & Roots

- Clasts
- ▨ Missing clasts
- ▤ Clasts of till
- Roots

Structural Measurements & Samples

- Structural measurement, fault plane
- Structural measurement, depositional fabric
- ¹⁴C Macro Sample, collected
- ¹⁴C Macro Sample, dated
- Bulk Sediment Sample for ¹⁴C, collected
- Bulk Sediment Sample for ¹⁴C, dated

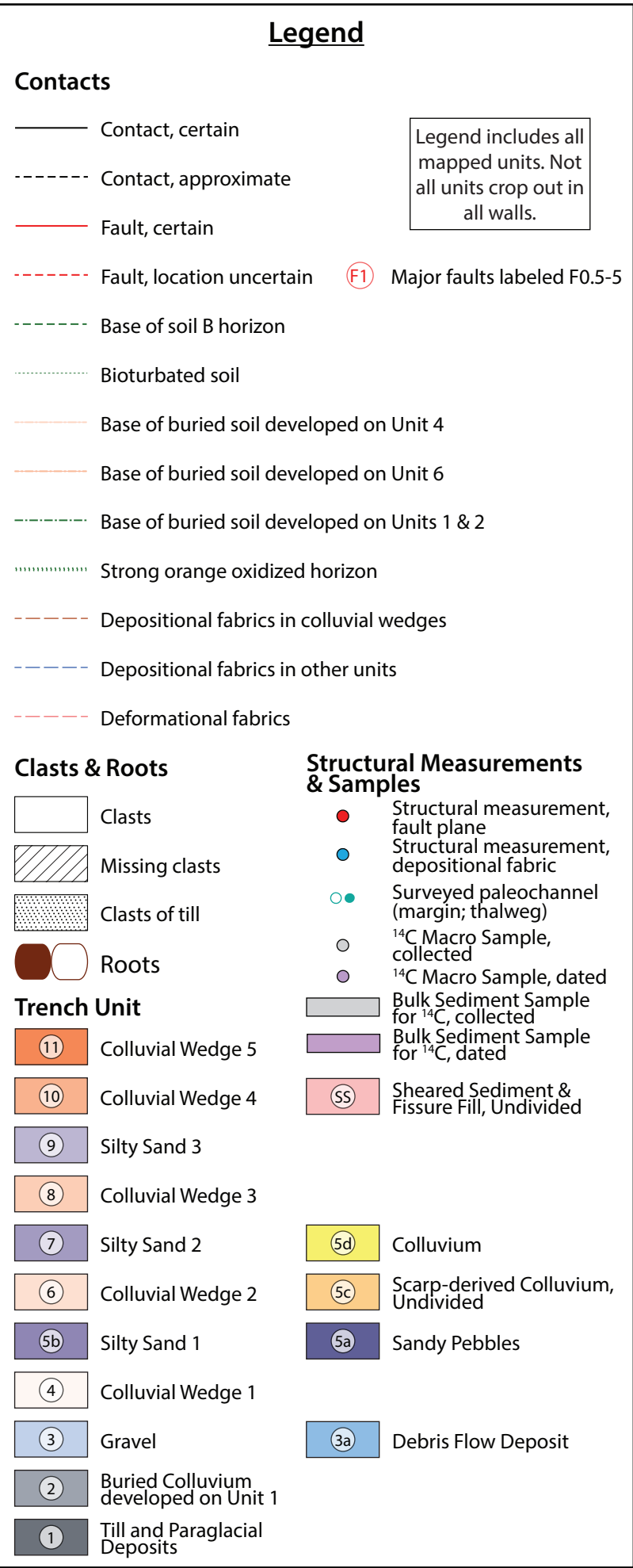
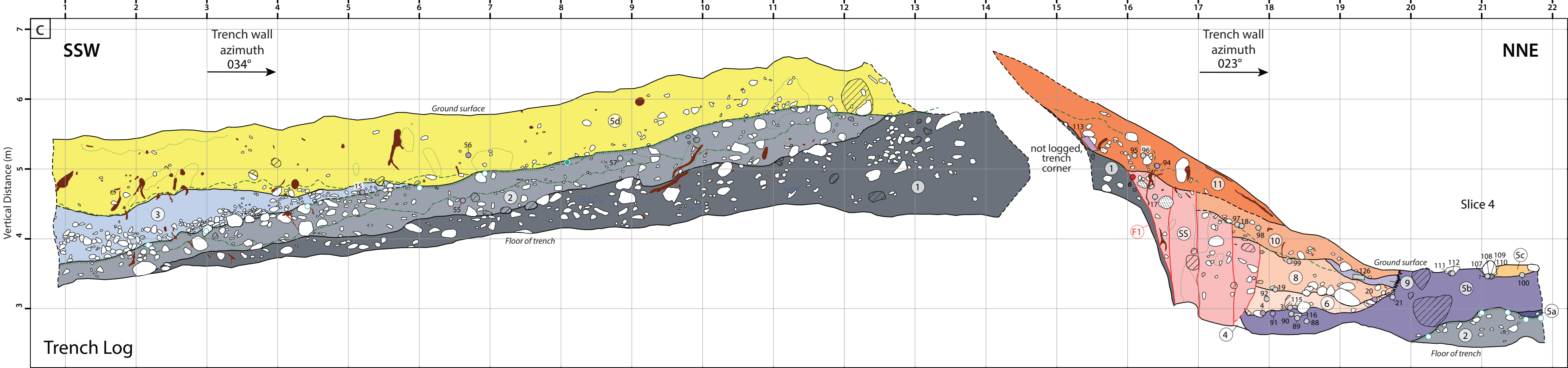
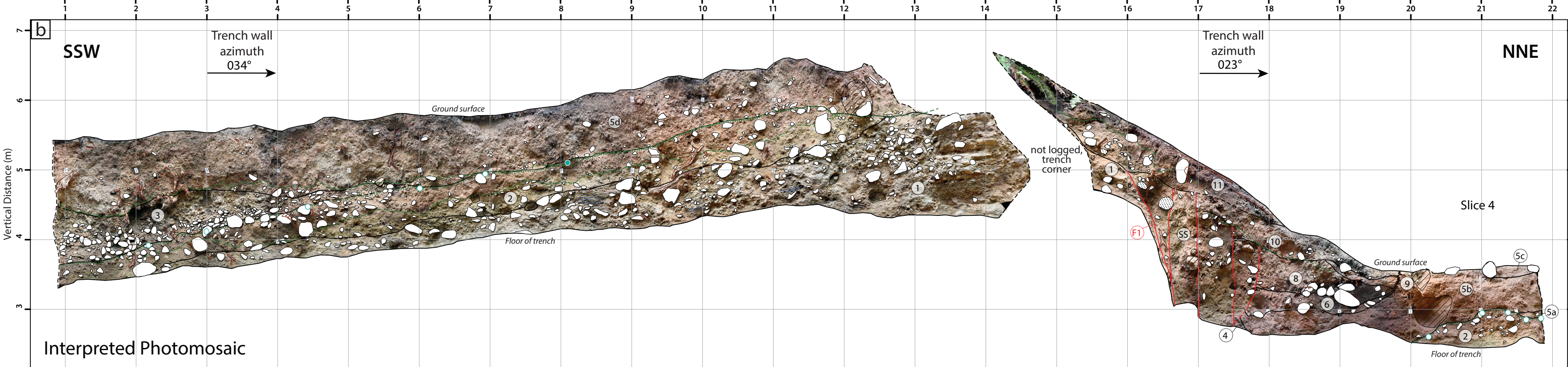
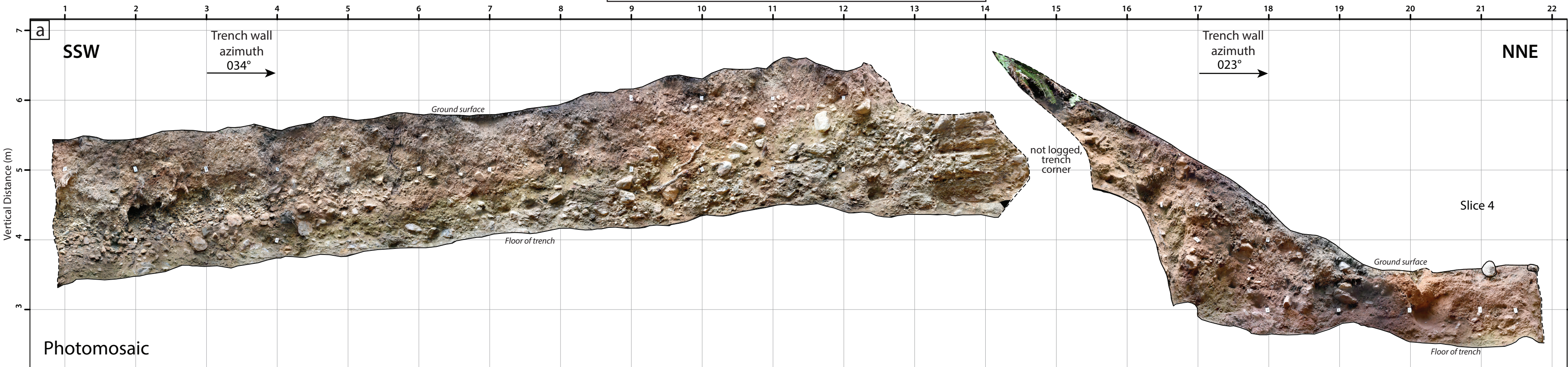
Trench Unit

- 11 Colluvial Wedge 5
- 10 Colluvial Wedge 4
- 9 Silty Sand 3
- 8 Colluvial Wedge 3
- 7 Silty Sand 2
- 6 Colluvial Wedge 2
- 5b Silty Sand 1
- 4 Colluvial Wedge 1
- 3 Gravel
- 2 Buried Colluvium developed on Unit 1
- 1 Till and Paraglacial Deposits
- SS Sheared Sediment & Fissure Fill, Undivided
- 5d Colluvium
- 5c Scarp-derived Colluvium, Undivided
- 5a Sandy Pebbles
- 3a Debris Flow Deposit

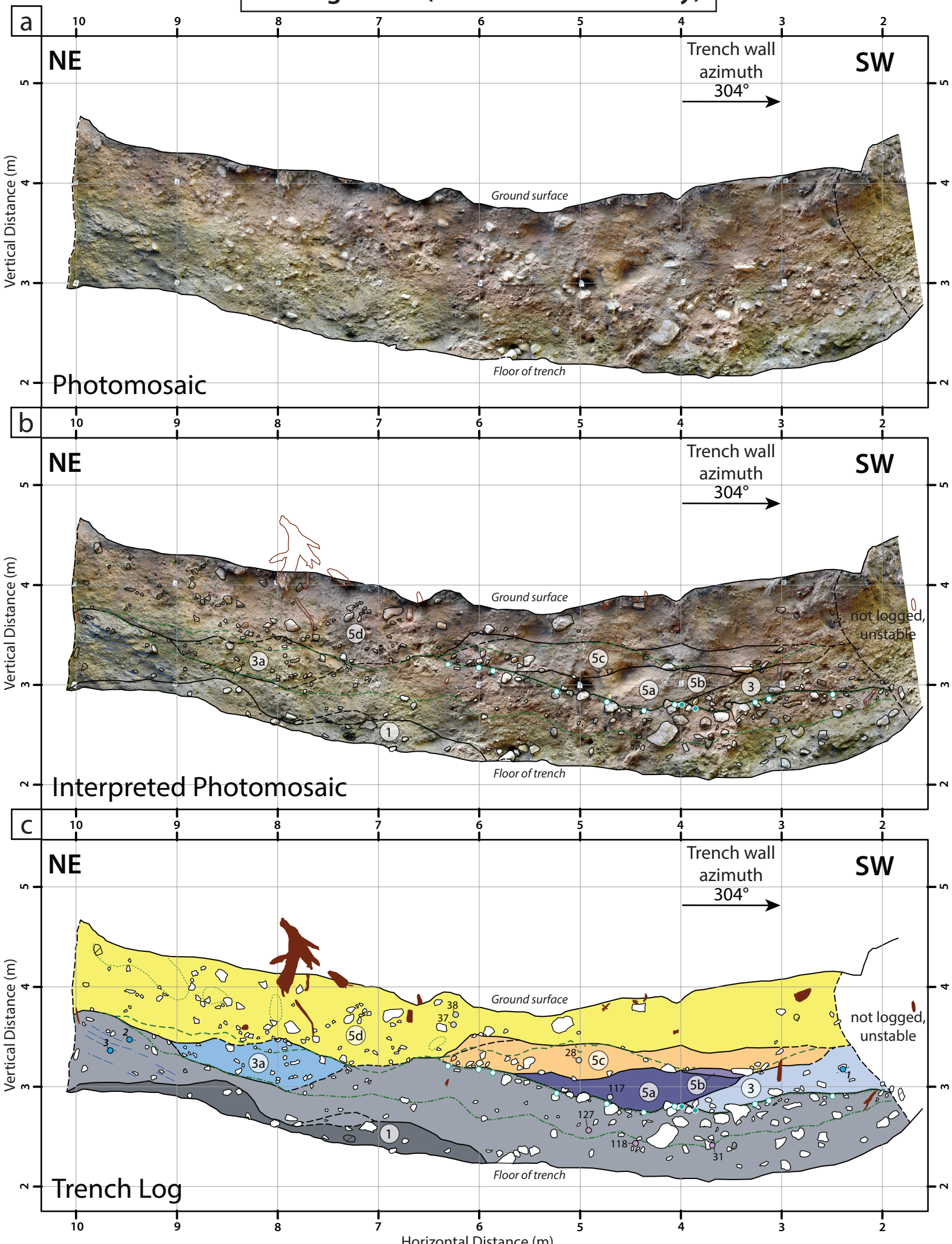
Legend includes all mapped units. Not all units crop out in all walls.

- Surveyed paleochannel (margin; thalweg)
- F1 Major faults labeled F0.5-5

Beaufort Range Fault, Woodlot Trench
Slices 2 and 4



Beaufort Range Fault, Woodlot Trench
A-Wing North (reflected horizontally)



Legend

Legend includes all mapped units. Not all units crop out in all walls.

Contacts

- Contact, certain
- Contact, approximate
- Fault, certain
- Fault, location uncertain
- Base of soil B horizon
- Bioturbated soil
- Base of buried soil developed on Unit 4
- Base of buried soil developed on Unit 6
- Base of buried soil developed on Units 1 & 2
- Strong orange oxidized horizon
- Depositional fabrics in colluvial wedges
- Depositional fabrics in other units
- Deformational fabrics

Clasts & Roots

- Clasts
- Missing clasts
- Clasts of till
- Roots

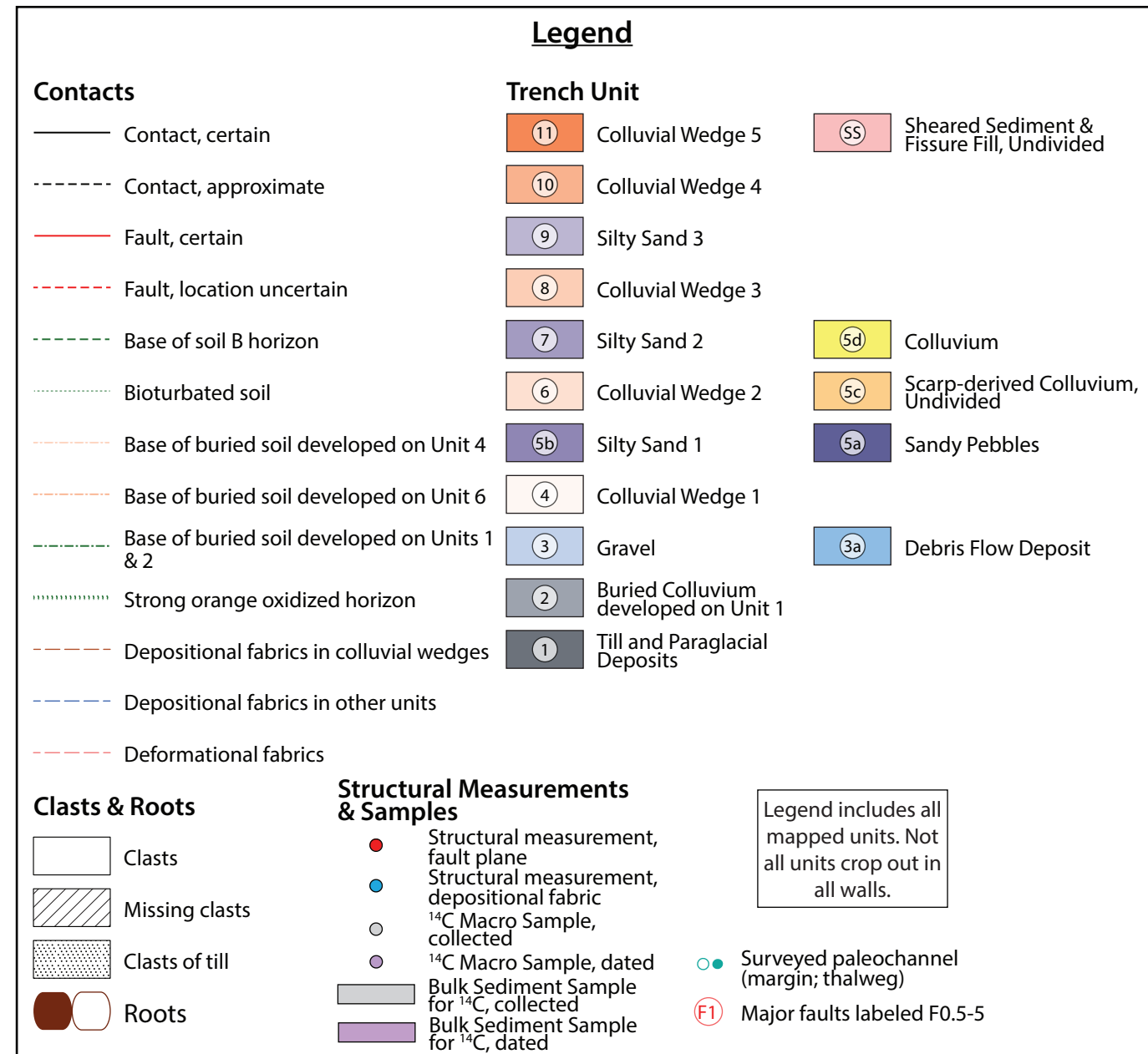
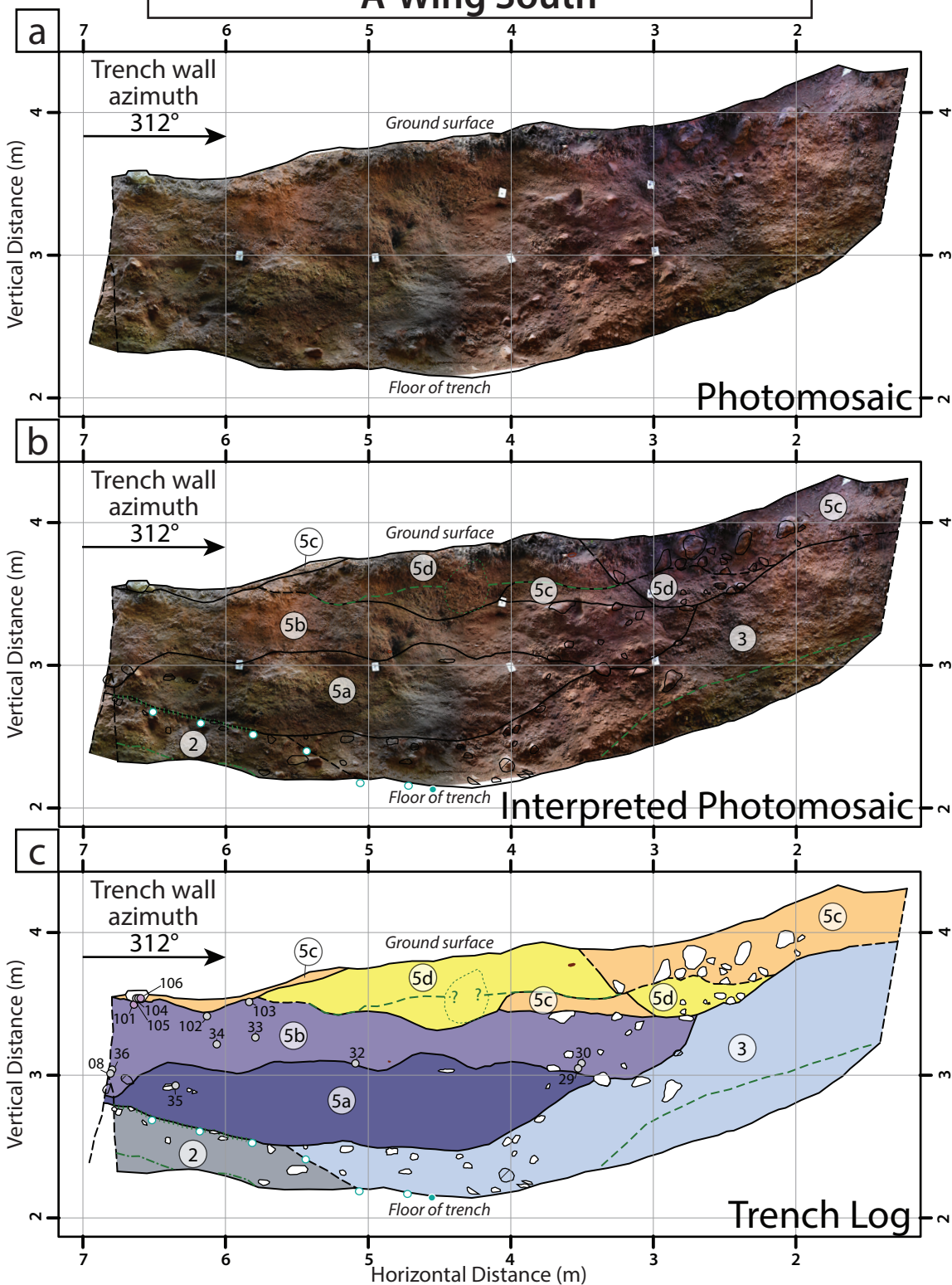
Trench Unit

- Colluvial Wedge 5
- Colluvial Wedge 4
- Silty Sand 3
- Colluvial Wedge 3
- Silty Sand 2
- Colluvial Wedge 2
- Silty Sand 1
- Colluvial Wedge 1
- Gravel
- Buried Colluvium developed on Unit 1
- Till and Paraglacial Deposits

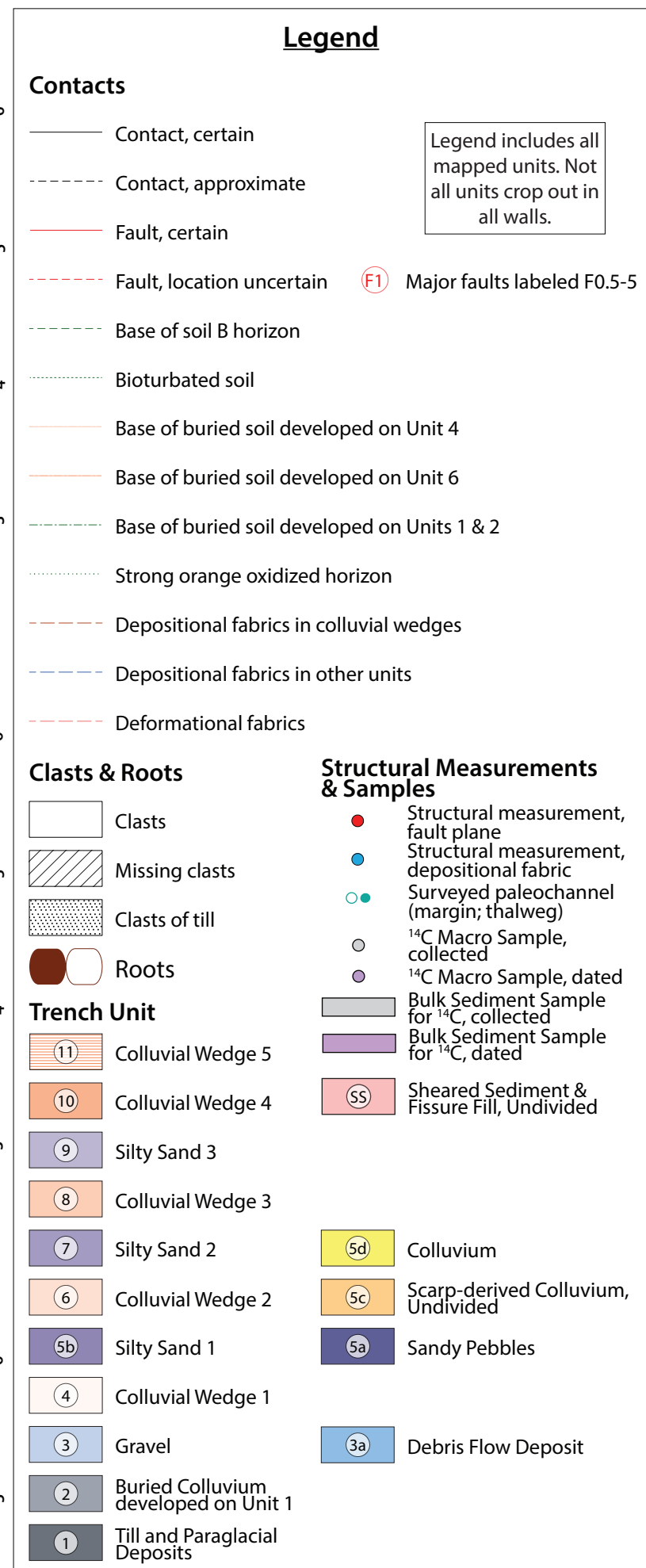
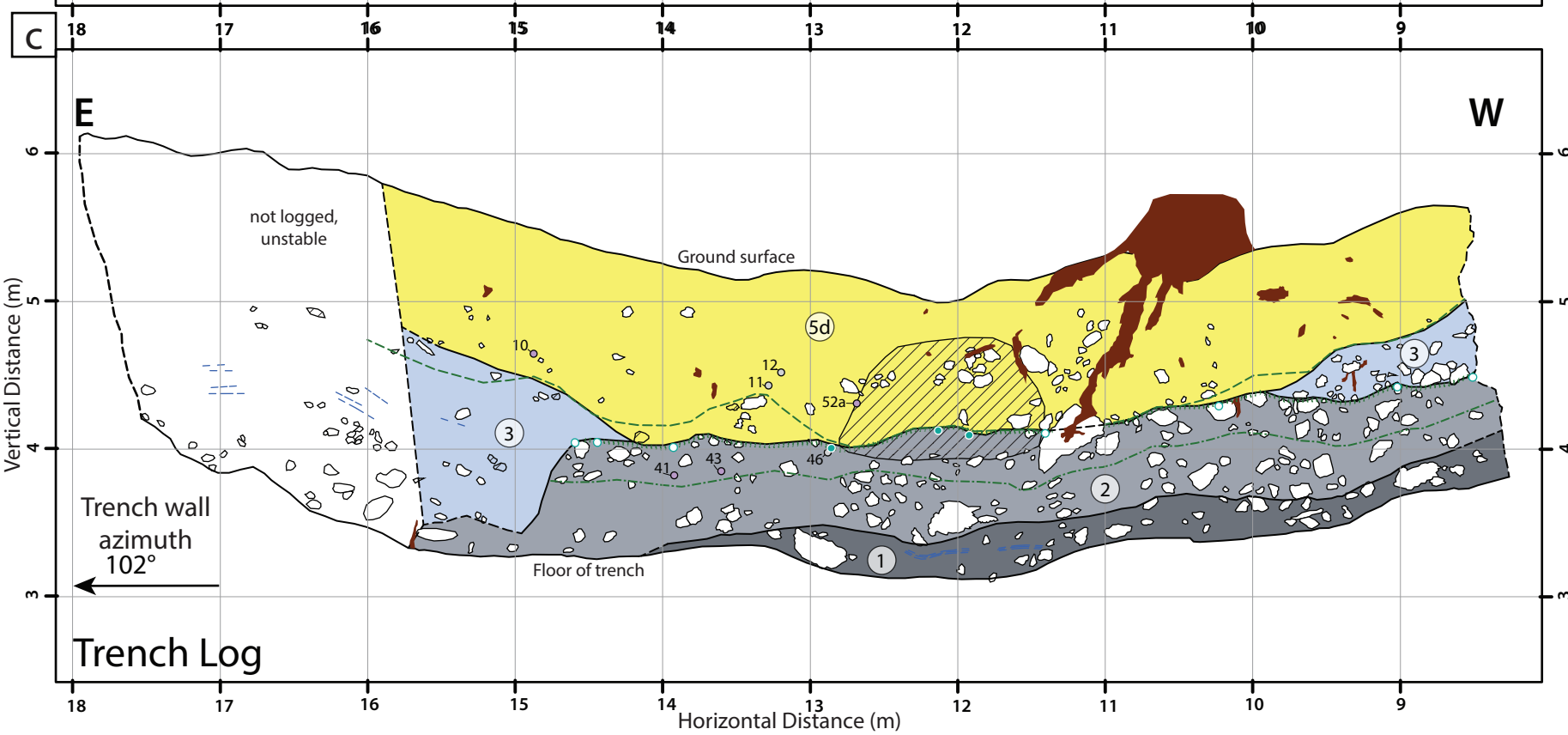
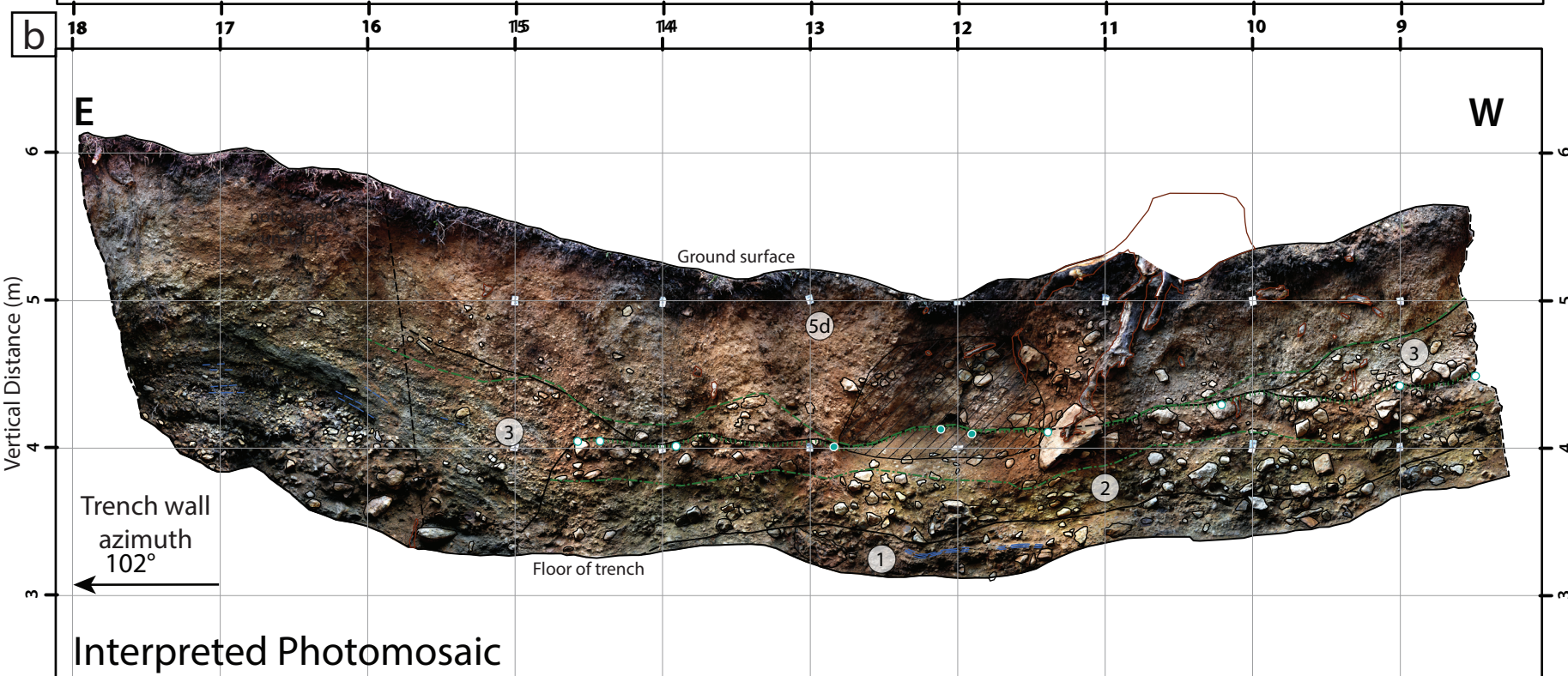
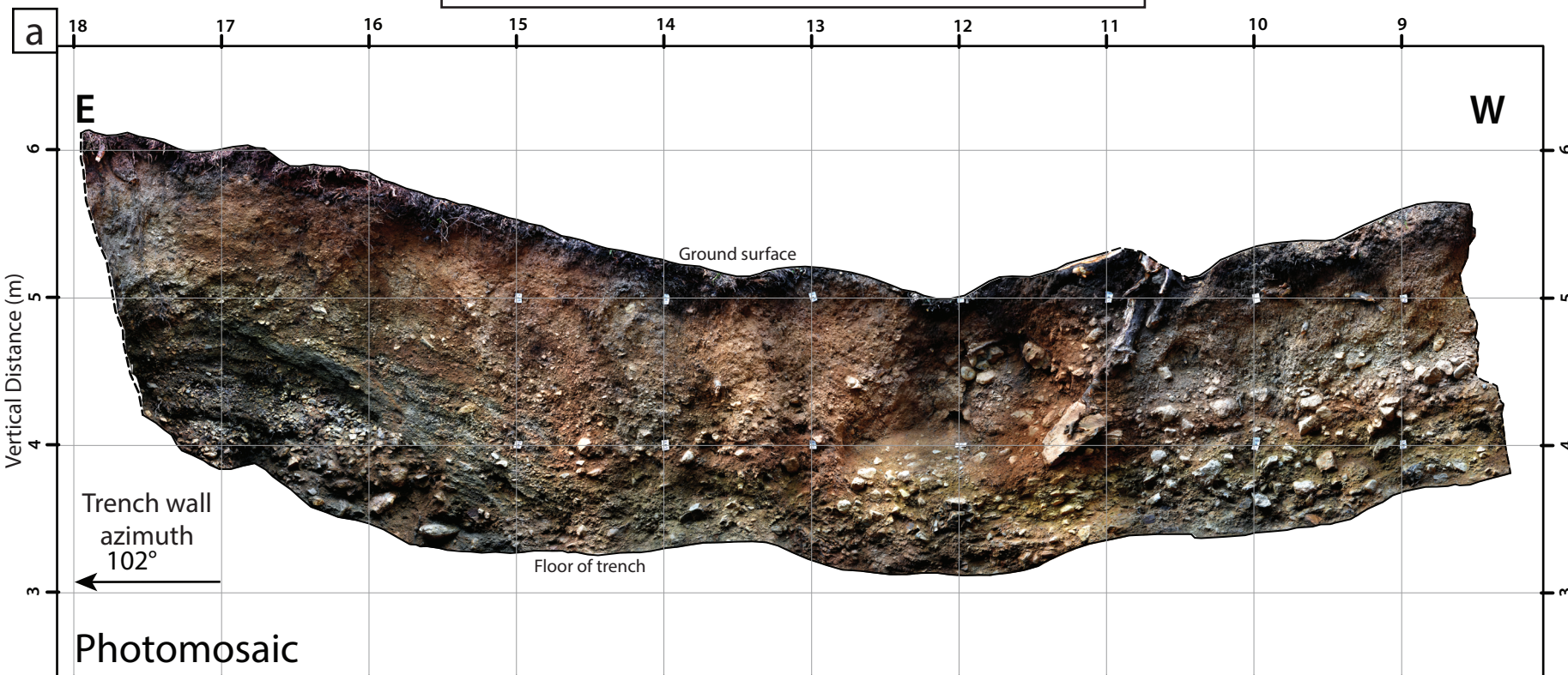
Structural Measurements & Samples

- Structural measurement, fault plane
- Structural measurement, depositional fabric
- Surveyed paleochannel (margin; thalweg)
- ¹⁴C Macro Sample, collected
- ¹⁴C Macro Sample, dated
- Bulk Sediment Sample for ¹⁴C, collected
- Bulk Sediment Sample for ¹⁴C, dated
- Sheared Sediment & Fissure Fill, Undivided
- Colluvium
- Scarp-derived Colluvium, Undivided
- Sandy Pebbles
- Debris Flow Deposit

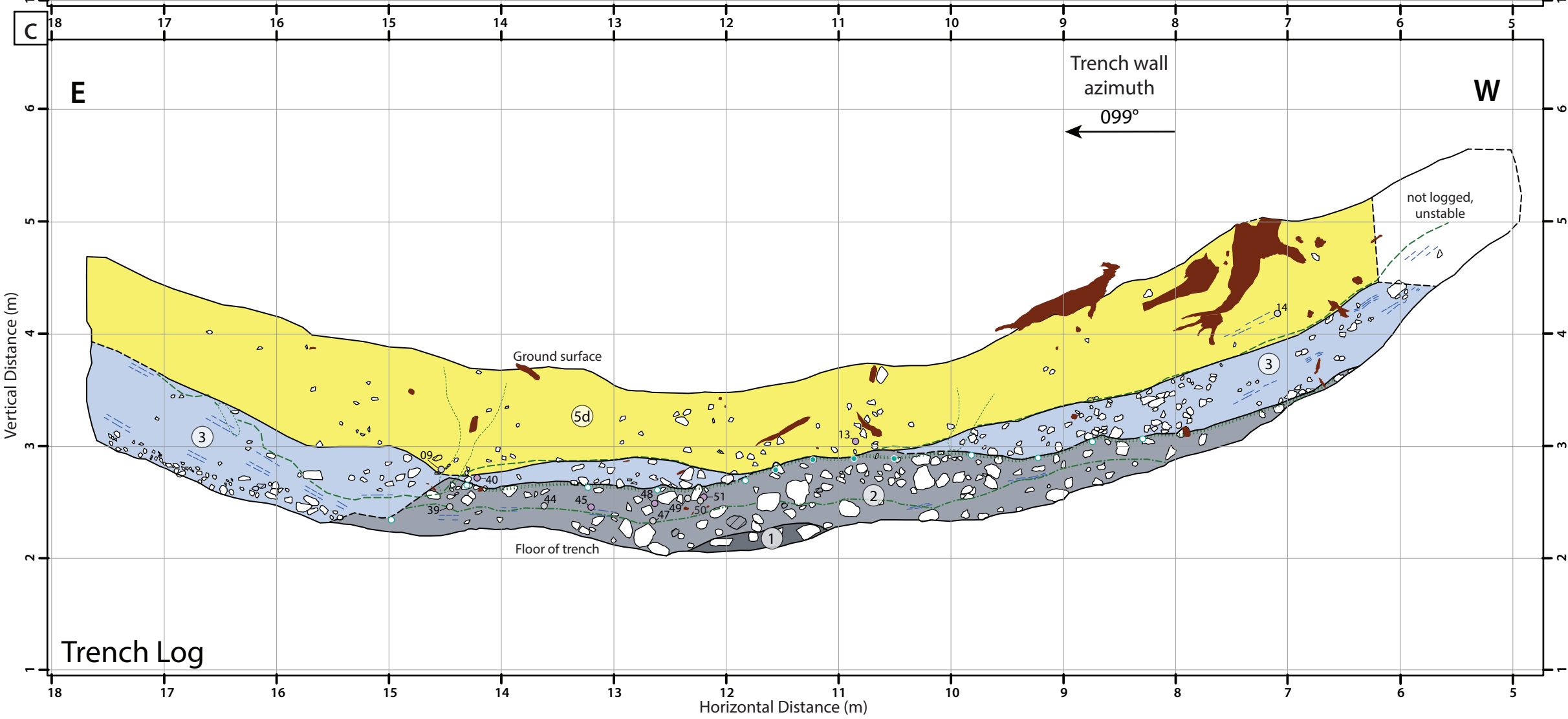
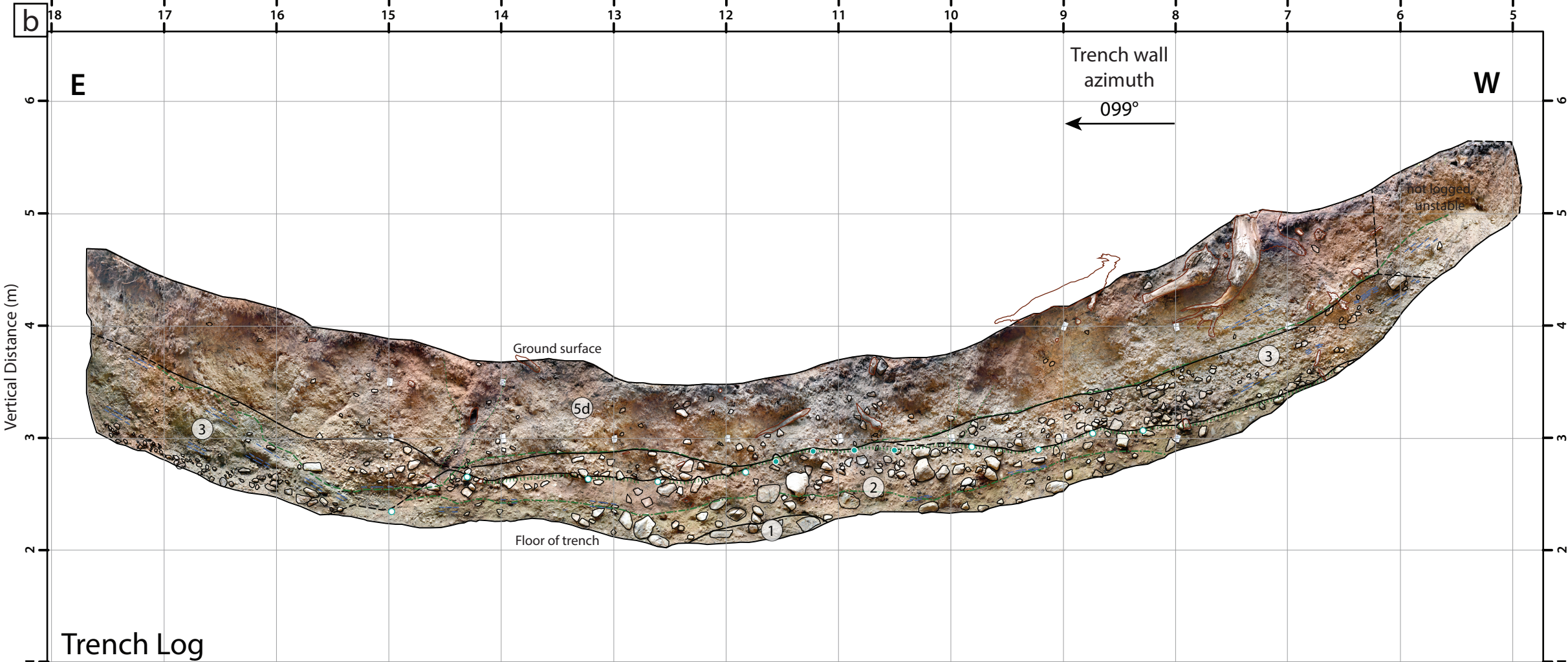
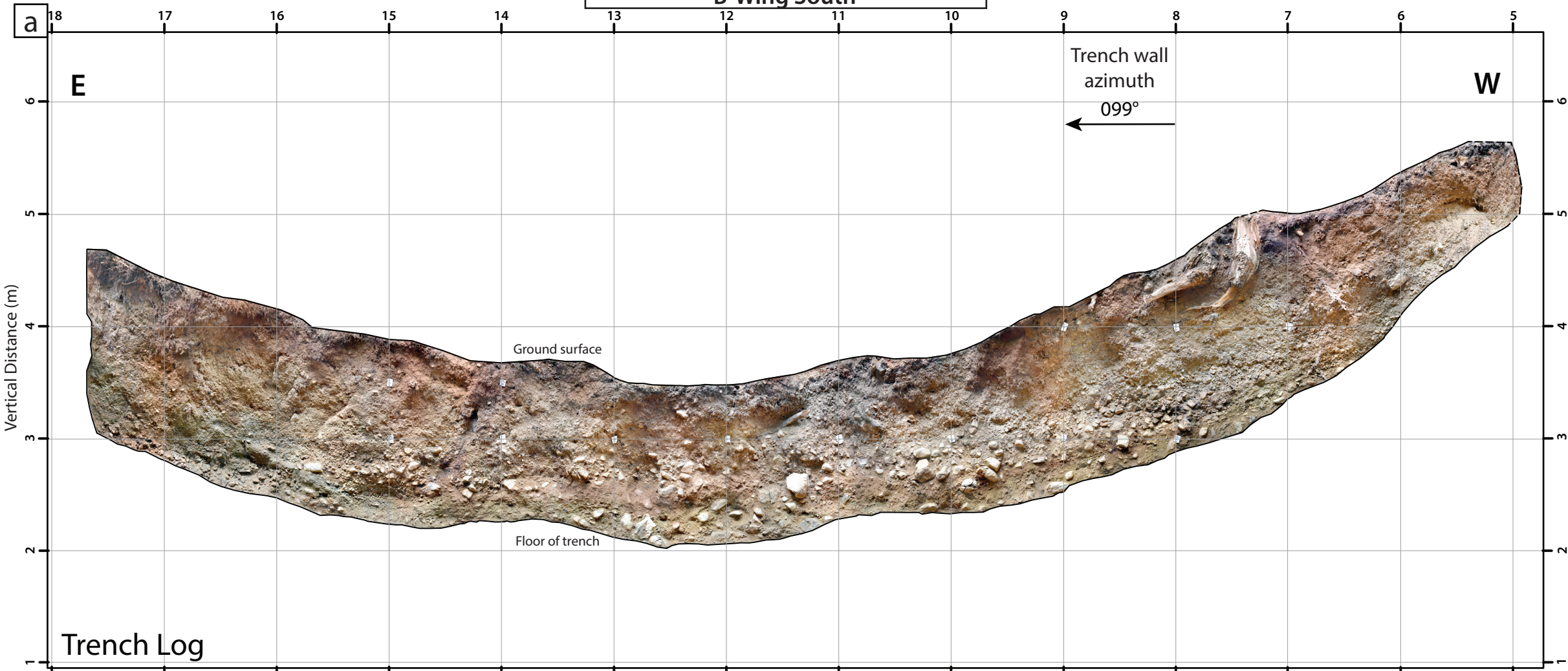
Beaufort Range Fault, Woodlot Trench A-Wing South



Beaufort Range Fault, Woodlot Trench
B-Wing N (reflected horizontally)



Beaufort Range Fault, Woodlot Trench
B-Wing South



Legend

Contacts

- Contact, certain
- Contact, approximate
- Fault, certain
- Fault, location uncertain
- Base of soil B horizon
- Bioturbated soil
- Base of buried soil developed on Unit 4
- Base of buried soil developed on Unit 6
- Base of buried soil developed on Units 1 & 2
- Strong orange oxidized horizon
- Depositional fabrics in colluvial wedges
- Depositional fabrics in other units
- Deformational fabrics

Clasts & Roots

- Clasts
- Missing clasts
- Clasts of till
- Roots

Trench Unit

- Colluvial Wedge 5
- Colluvial Wedge 4
- Silty Sand 3
- Colluvial Wedge 3
- Silty Sand 2
- Colluvial Wedge 2
- Silty Sand 1
- Colluvial Wedge 1
- Gravel
- Buried Colluvium developed on Unit 1
- Till and Paraglacial Deposits

Structural Measurements & Samples

- Structural measurement, fault plane
- Structural measurement, depositional fabric
- Surveyed paleochannel (margin; thalweg)
- ¹⁴C Macro Sample, collected
- ¹⁴C Macro Sample, dated
- Bulk Sediment Sample for ¹⁴C, collected
- Bulk Sediment Sample for ¹⁴C, dated
- Sheared Sediment & Fissure Fill, Undivided
- Colluvium
- Scarp-derived Colluvium, Undivided
- Sandy Pebbles
- Debris Flow Deposit

Legend includes all mapped units. Not all units crop out in all walls.

Major faults labeled F0.5-5

C.1.2 Captions for supplemental trench logs

Figure C.1: Photomosaic (a), interpreted photomosaic (b), and trench log (c) for Slice 1 (see Figure 3.2 for location). Photomosaic consists of two separate orthomosaics that were exported at different view angles and stitched together along the bend in the trench. Photomosaics and log are reflected horizontally for correlation with other walls. Major fault planes labeled F1-5 in red font. Radiocarbon samples listed in Table 3.1.

Figure C.2: Photomosaic (a), interpreted photomosaic (b), and trench log (c) for Slice 2L (see Figure 3.2 for location). Photomosaic and log are reflected horizontally for correlation with other walls. Major fault planes labeled F1-5 in red font. Radiocarbon samples listed in Table 3.1.

Figure C.3: Photomosaic (a), interpreted photomosaic (b), and trench log (c) for Slice 3L (see Figure 3.2 for location). Major fault planes labeled F1-5 in red font. Radiocarbon samples listed in Table 3.1.

Figure C.4: Photomosaic (a), interpreted photomosaic (b), and trench log (c) for Slices 2 (left) and 4 (right; see Figure 3.2 for locations). Photomosaics for Slices 2 and 4 were created separately. Slice 4 is ~3 m “behind” Slice 2 in this cross-section view. Major fault planes labeled F1-5 in red font. Radiocarbon samples listed in Table 3.1.

Figure C.5: Photomosaic (a), interpreted photomosaic (b), and trench log (c) for A-Wing north wall (see Figure 3.2 for location). Photomosaic and log are reflected horizontally for correlation with other walls. The southwestern part of the wall was not logged due to instability. Major fault planes labeled F1-5 in red font. Radiocarbon samples listed in Table 3.1.

Figure C.6: Photomosaic (a), interpreted photomosaic (b), and trench log (c) for A-Wing south wall (see Figure 3.2 for location). Major fault planes labeled F1-5 in red font. Radiocarbon samples listed in Table 3.1.

Figure C.7: Photomosaic (a), interpreted photomosaic (b), and trench log (c) for B-Wing north wall (see Figure 3.2 for location). Photomosaic and log are reflected horizontally for correlation with other walls. The eastern part of the wall was not logged due to instability. Major fault planes labeled F1-5 in red font. Radiocarbon samples listed in Table 3.1.

Figure C.8: Photomosaic (a), interpreted photomosaic (b), and trench log (c) for B-Wing south wall (see Figure 3.2 for location). The western part of the wall was not logged due to instability. Major fault planes labeled F1-5 in red font. Radiocarbon samples listed in Table 3.1.

C.1.3 Additional figures

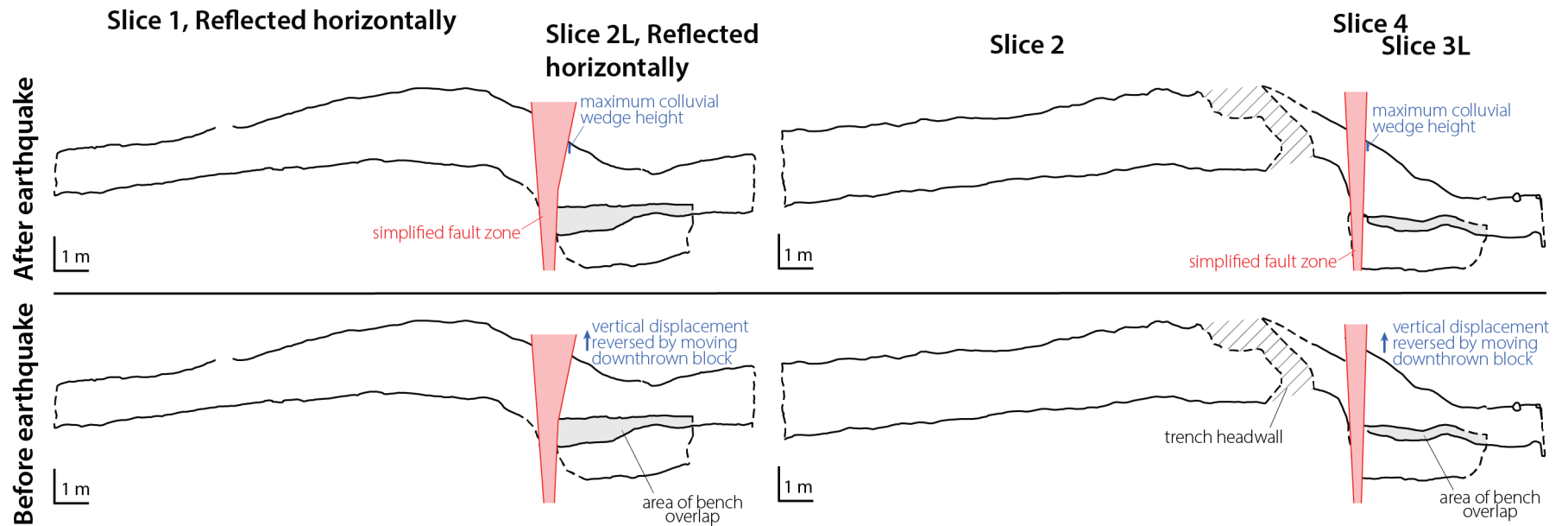


Figure C.9: Schematic showing retrodeformation methods. For each earthquake recorded in the Woodlot Trench, we reversed the dip slip estimated from colluvial wedge heights and recreated the stratigraphy that would have existed before the rupture. By reversing the deformation and deposition associated with all five ruptures, we recreated the stratigraphy that would have existed before the first event recorded in the Woodlot Trench.

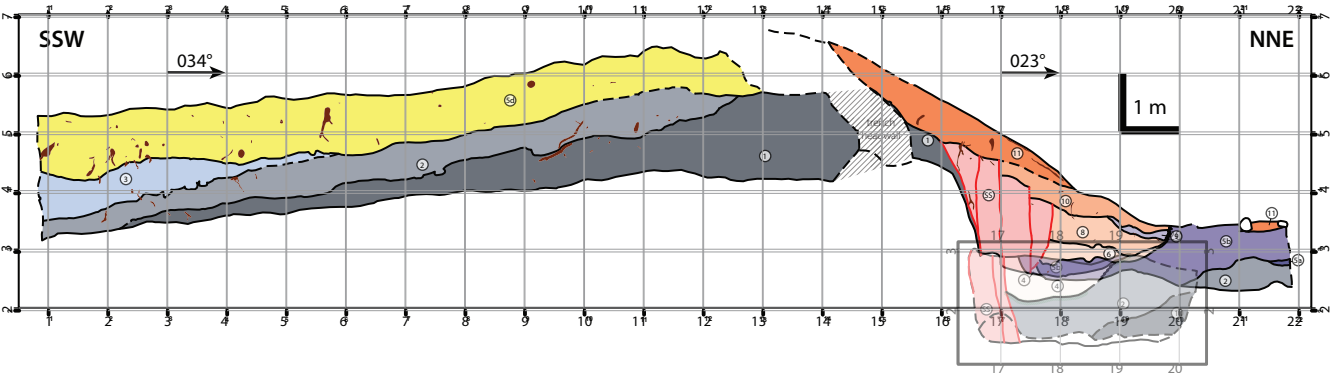
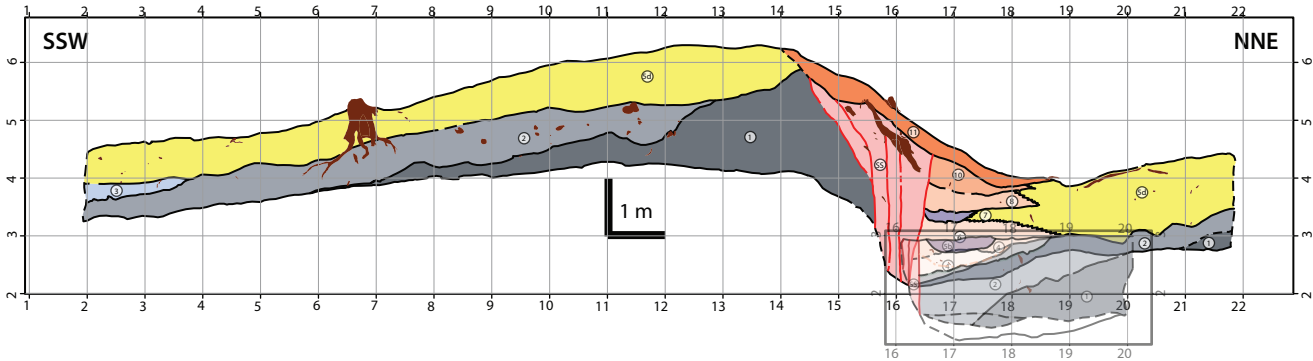
Slice 1, Reflected horizontally

Slice 2L, Reflected horizontally

Slice 2

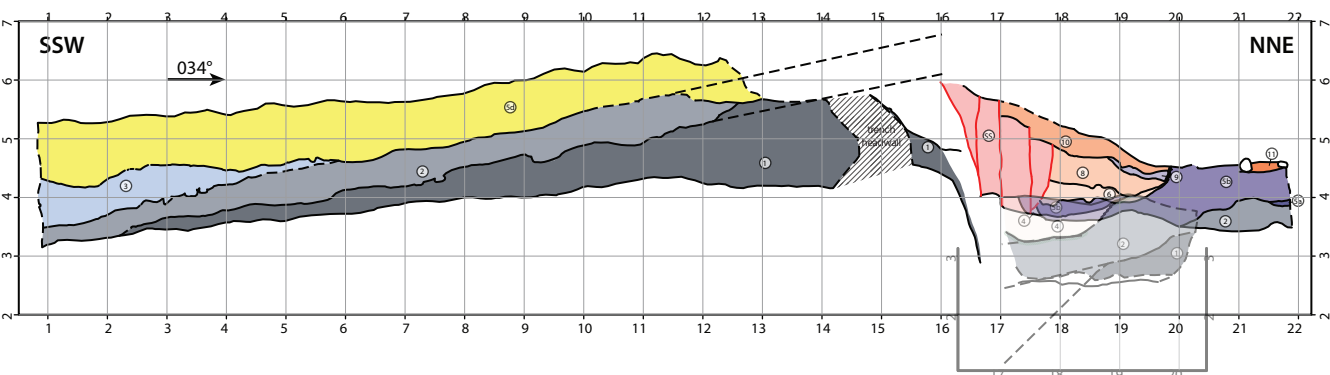
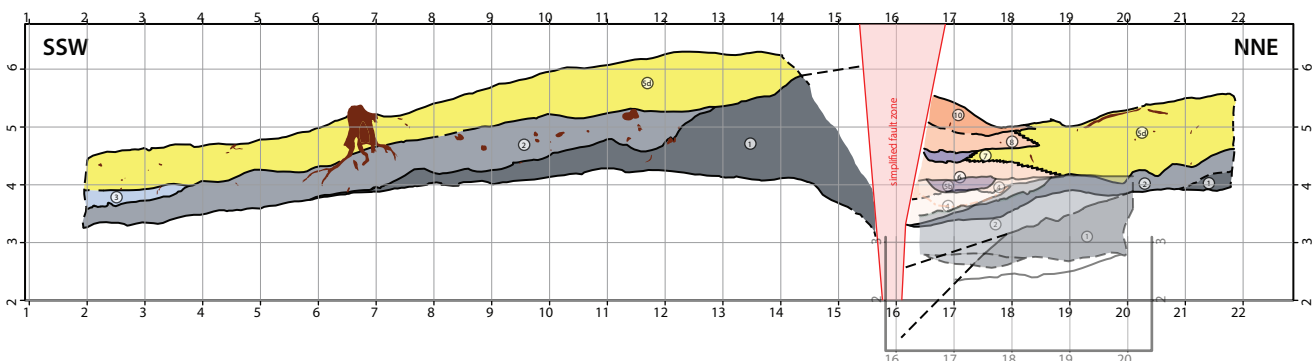
Slice 4
Slice 3L

Time 13



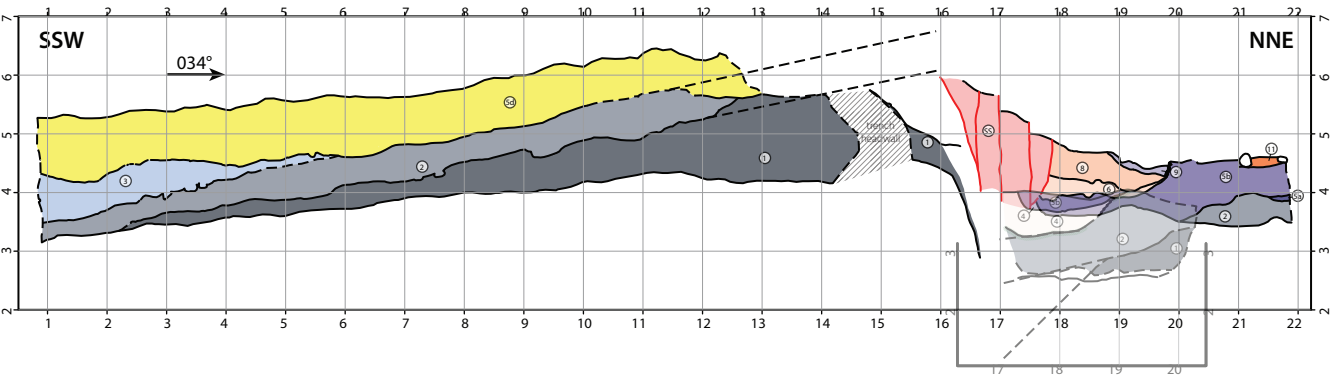
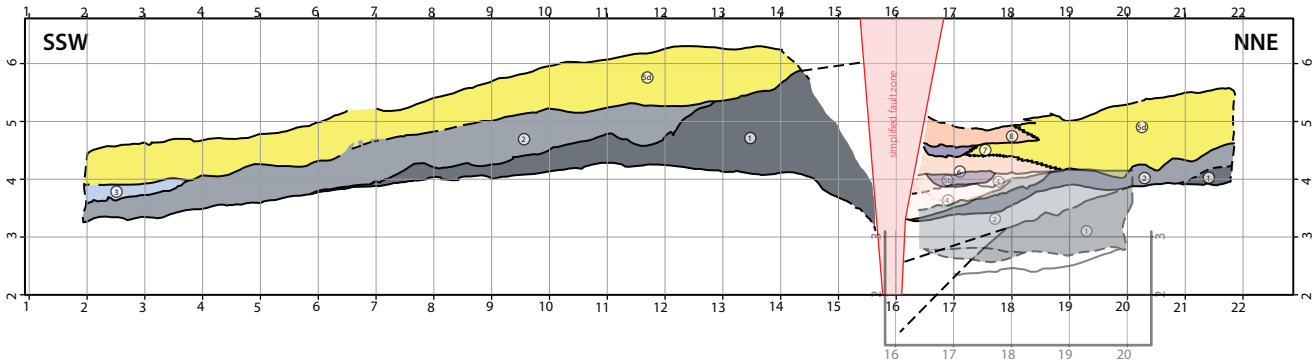
Deposition of colluvial wege unit 11 (~2.4-2.2 ka); all units and faults shown in present-day arrangement

Time 12



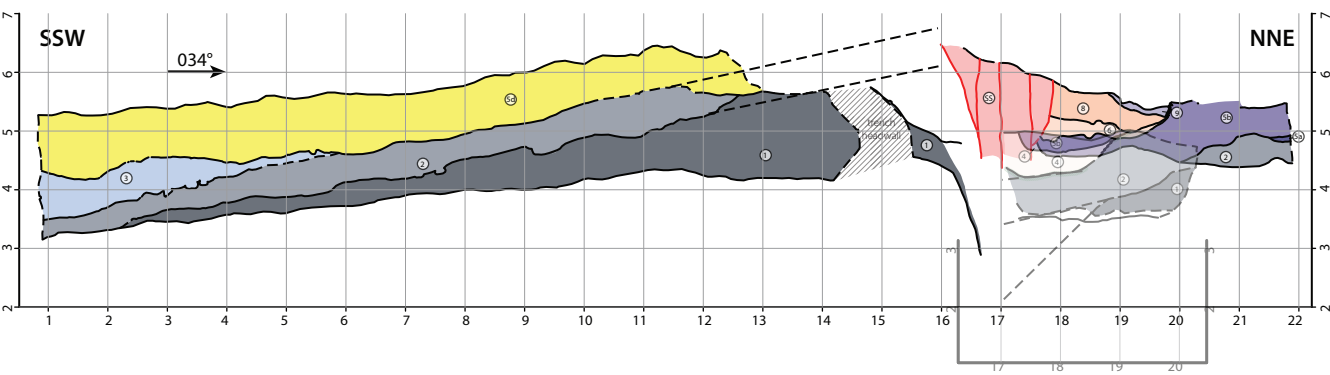
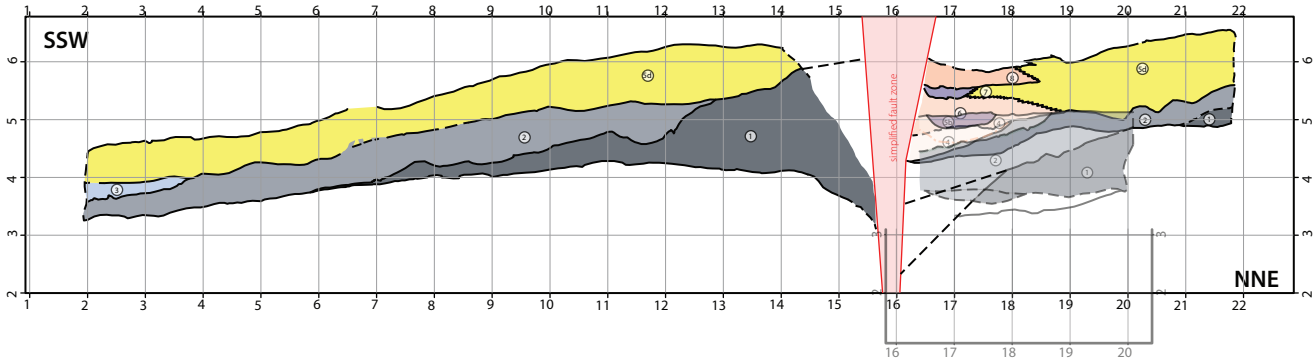
Deposition of colluvial wedge unit 10 (~3.0 ka)

Time 11



Event 4: ~4.2-3.0 ka

Time 10

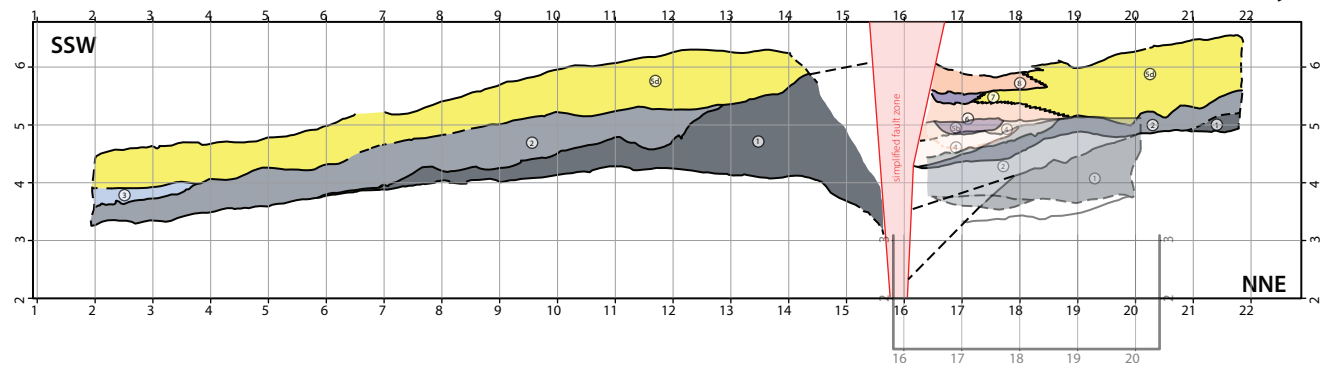


Deposition of channel deposit unit 9; continued deposition of hillslope deposit unit 5d, and channel deposit unit 5b

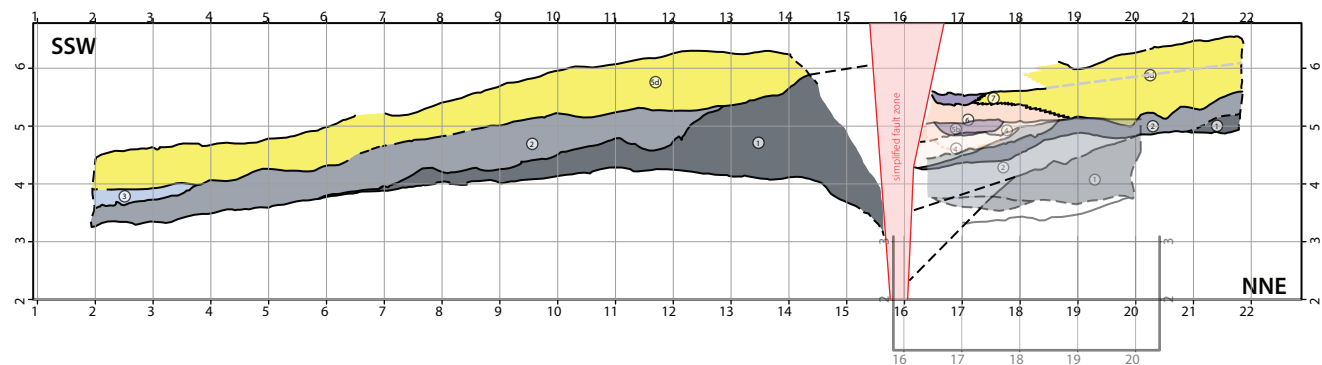
Slice 1, Reflected horizontally

Slice 2L, Reflected horizontally

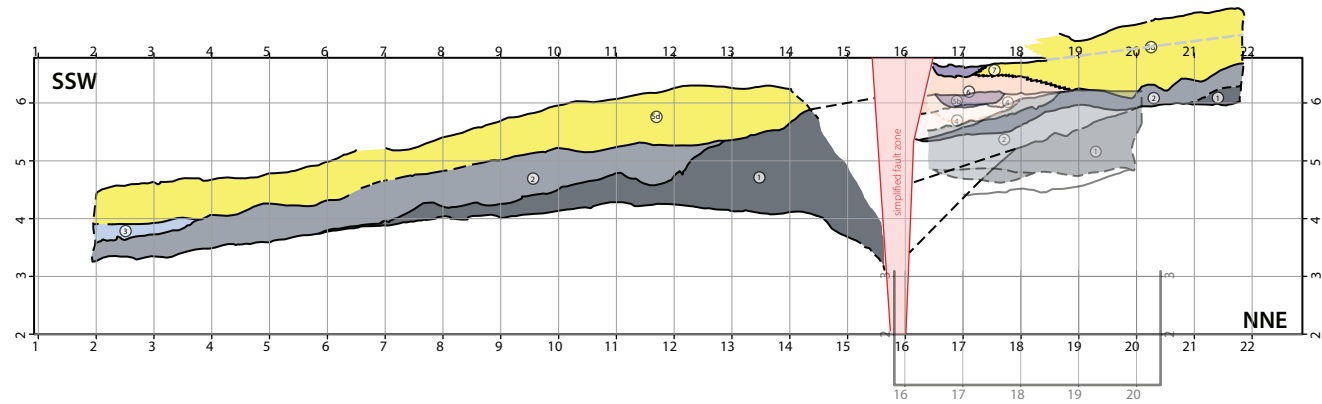
Time 9



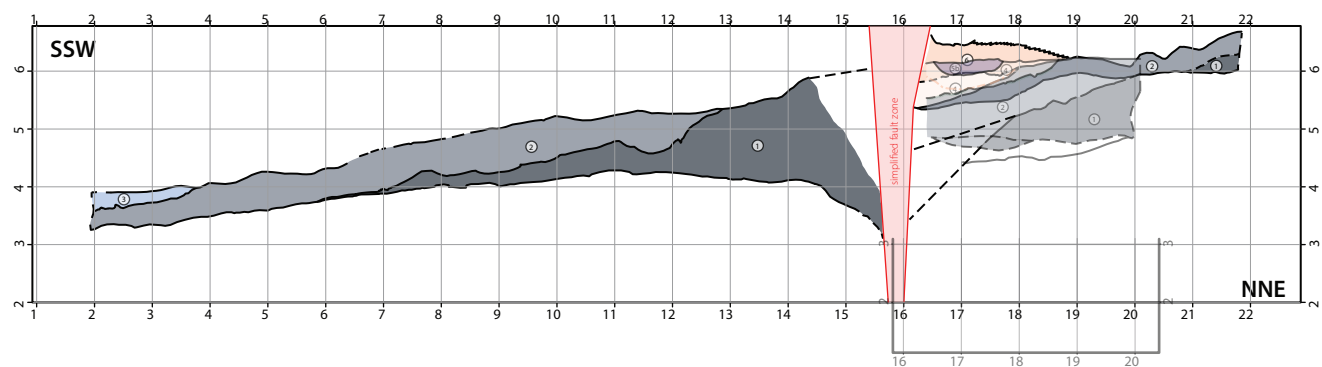
Time 8



Time 7



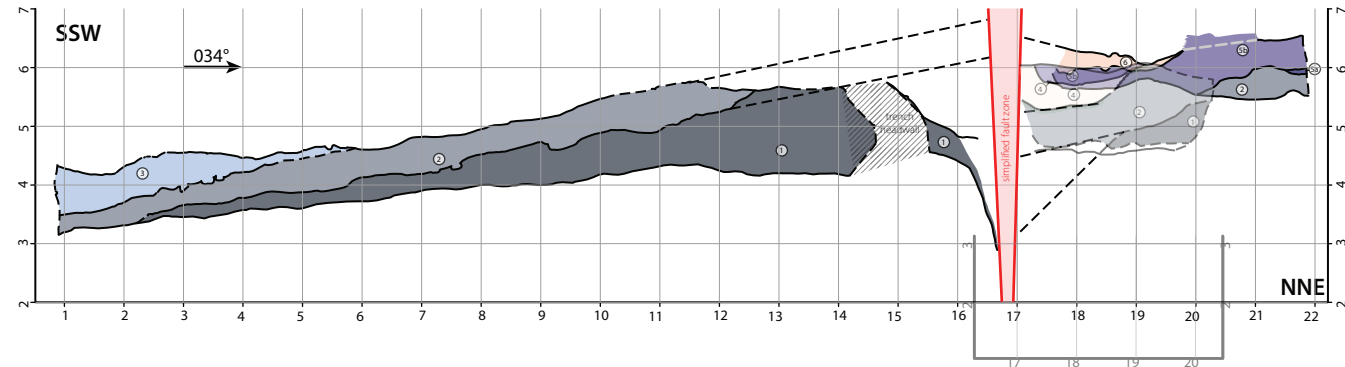
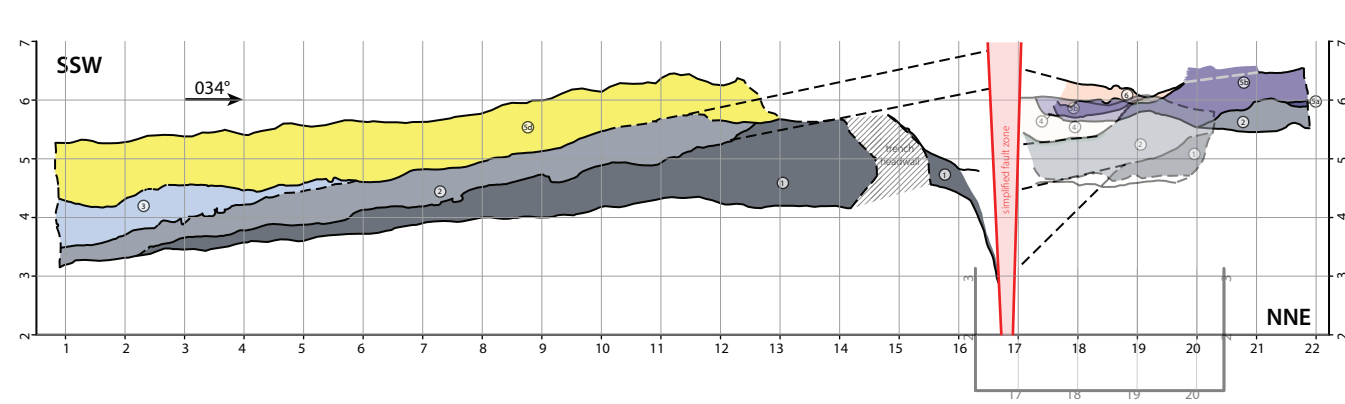
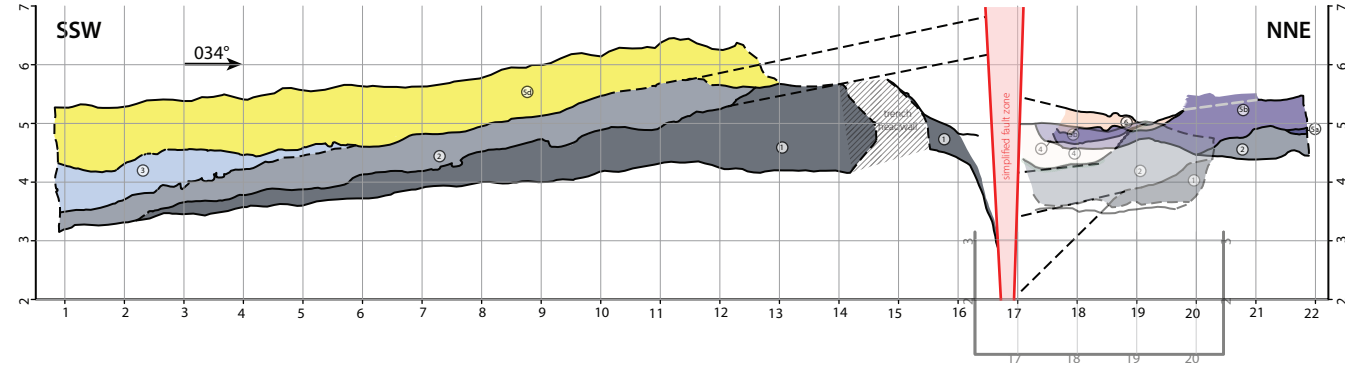
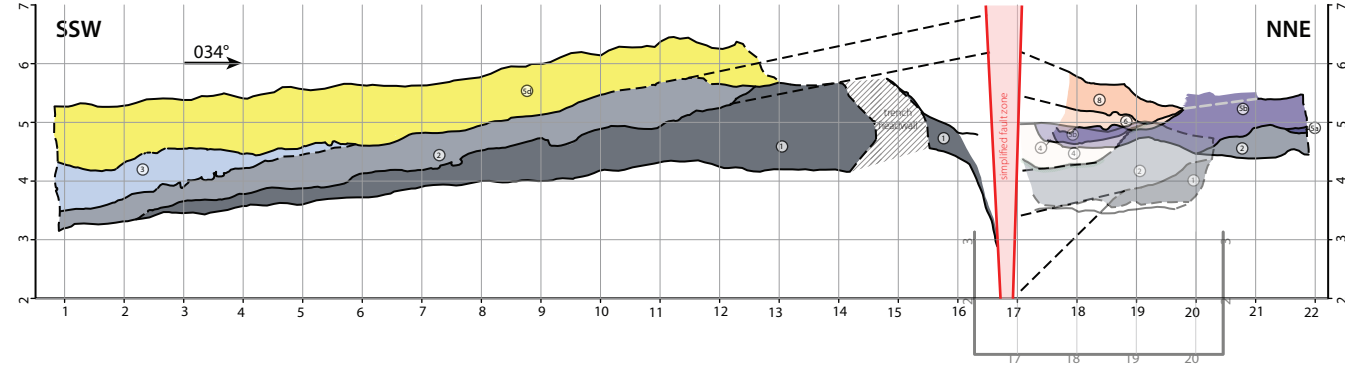
Time 6



Slice 2

Slice 4

Slice 3L



Deposition of colluvial
wedge unit 8 (~4.2-4.0
ka)

Event 3: 4.9-4.2 ka

Deposition of hillslope
deposit 5d and chan-
nel deposit 7, and
continued deposition
of channel deposit 5d

Deposition of colluvial
wedge unit 6 (~4.9 ka)

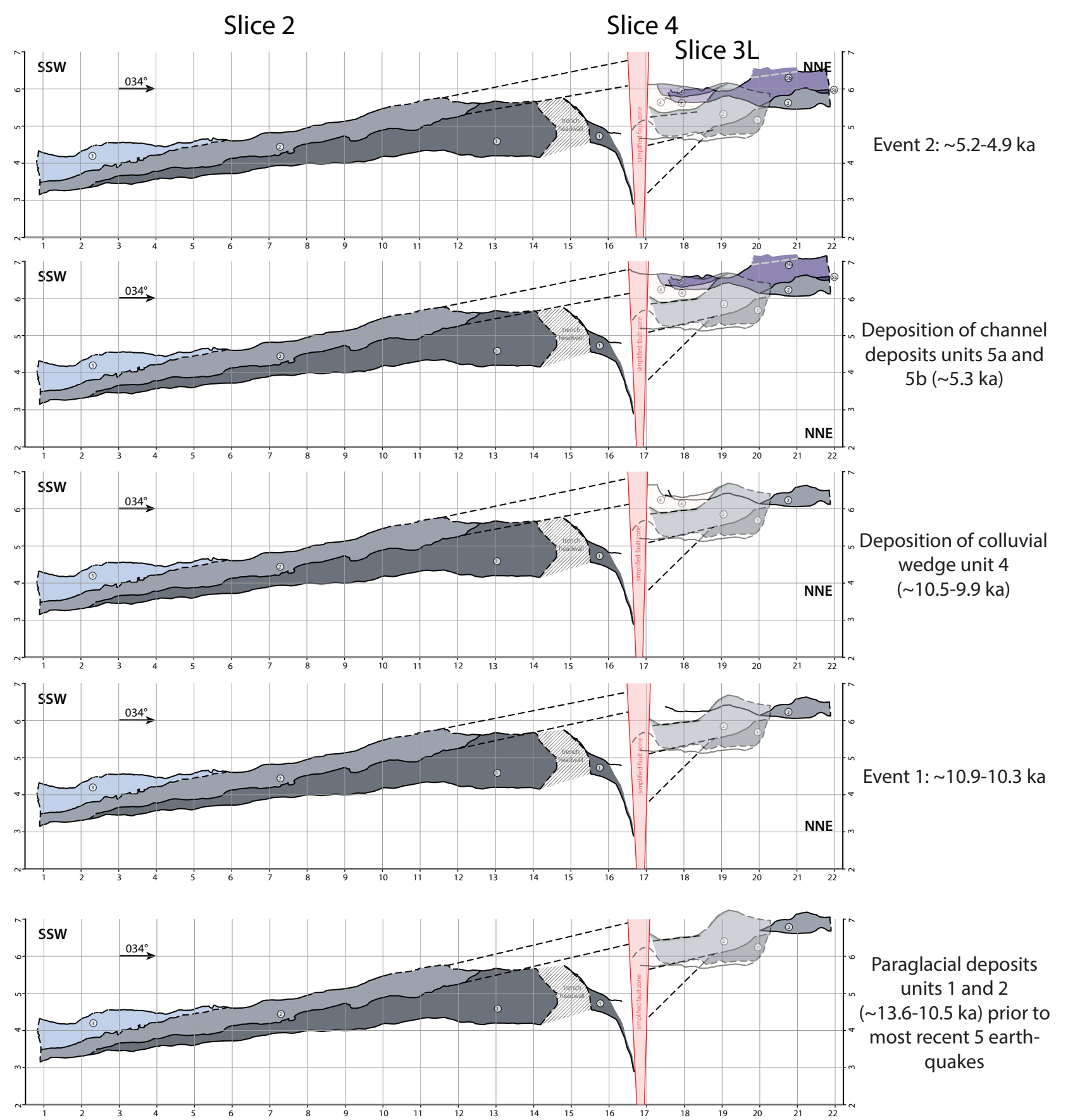
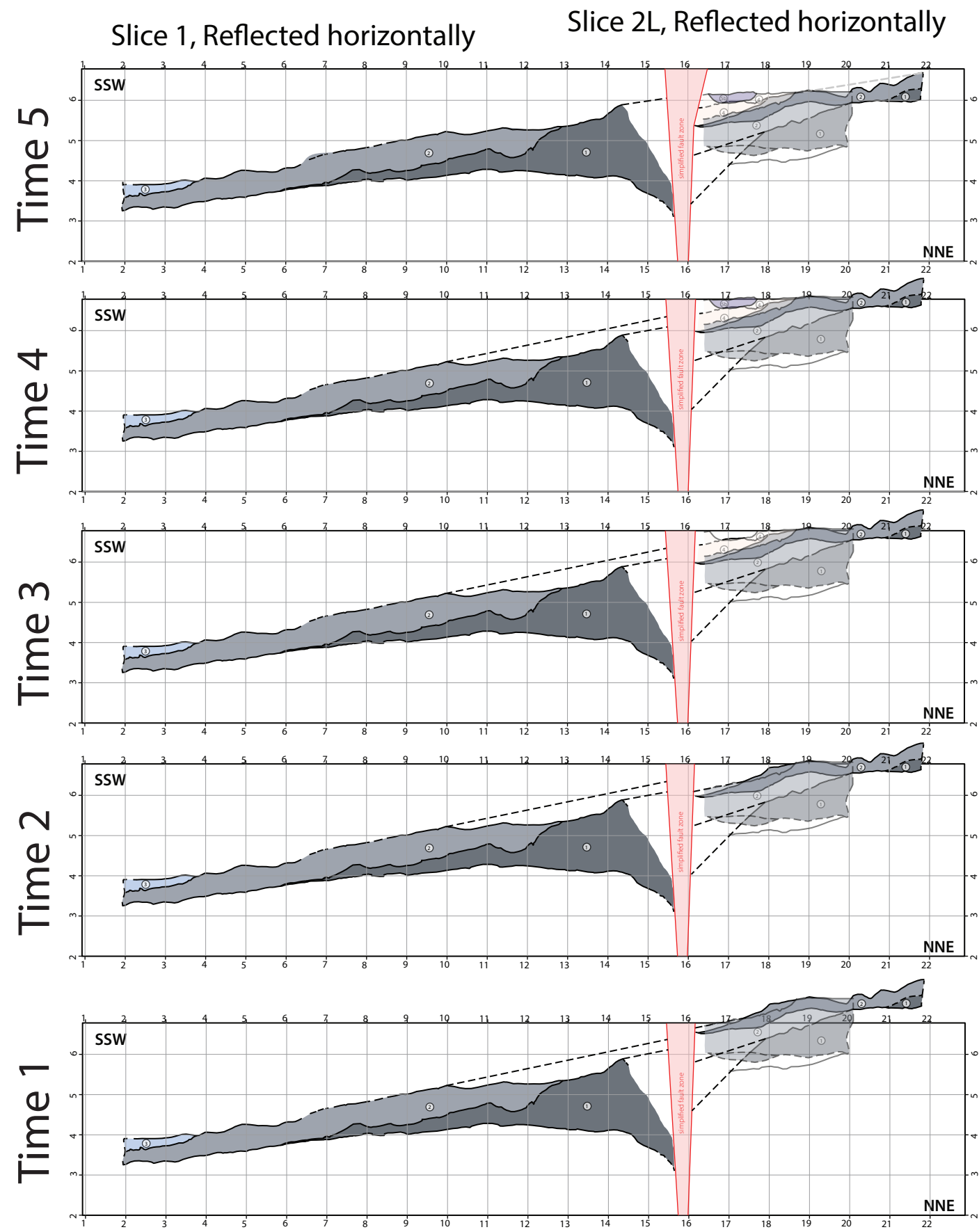


Figure C.10: Retrodeformation of displaced units as mapped in the two main walls of the trench: Slices 1 and 2L (left), and Slices 2, 3L, and 4 (right). Time 1 represents the original, undeformed deposits. Time 2 is immediately after the earliest event, Event 1, with deposition of the colluvial wedge Unit 4 in Time 3 shortly following. Times 2, 5, 8, 11, and 13 are immediately following Events 1, 2, 3, 4, and 5, respectively, while Times 3, 6, 9, 12, and 13 show the deposition of colluvial wedges following those events. Times 4, 7, and 10 show channel ponding against the scarp between events. Channel activity slows through time, and the channel is abandoned by Time 11. The offsets retrodeformed in Times 2, 5, 8, 11, and 13 are the vertical displacement alone, following the wedge heights in Table 3.2. These reconstructions do not include lateral slip or potential rotation of the downthrown block. Except where slip on individual faults can be attributed to specific events (Events 4 and 5 in Slice 4), the fault zone is simplified. Light grey dashed lines indicate how much of a long-lived deposit may have accumulated at that Time. We would expect the deposits that existed before the Event 1 (paraglacial deposits units 1 and 2) to be continuous across the fault zone at Time 1. The pre-Event 1 stratigraphy we recreated by reversing the displacement of each individual structure includes units 1 and 2 that project nearly continuously across the fault zone. This suggests that the total amount of vertical displacement in the Woodlot Trench is consistent with the minimum per-event dip slip estimated from colluvial wedge heights.

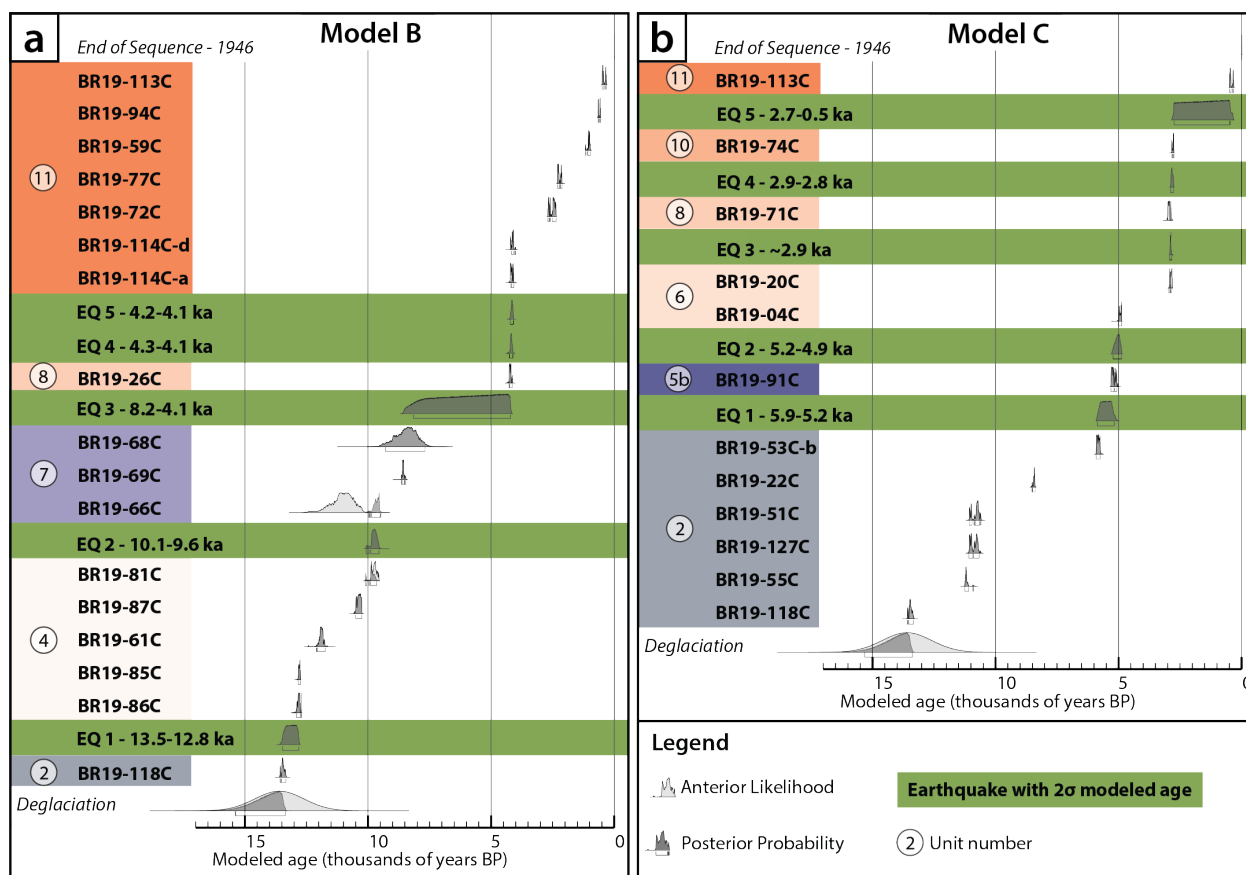


Figure C.11: Synthesis figure showing results of OxCal models B and C. The calibrated (anterior likelihood) and modeled (posterior probability) age PDFs are shown for each radiocarbon sample. See Section 3.3.5 for OxCal methods. A: OxCal Model B with modeled earthquake age PDFs and 2σ earthquake age ranges highlighted in green. Model B includes 17 samples and excludes 21 samples. B: OxCal model C with modeled earthquake age PDFs and 2σ earthquake age ranges highlighted in green. Model C includes 12 samples and excludes 26 samples. Results of OxCal models A.1 and A.2 are in Figure 3.5.

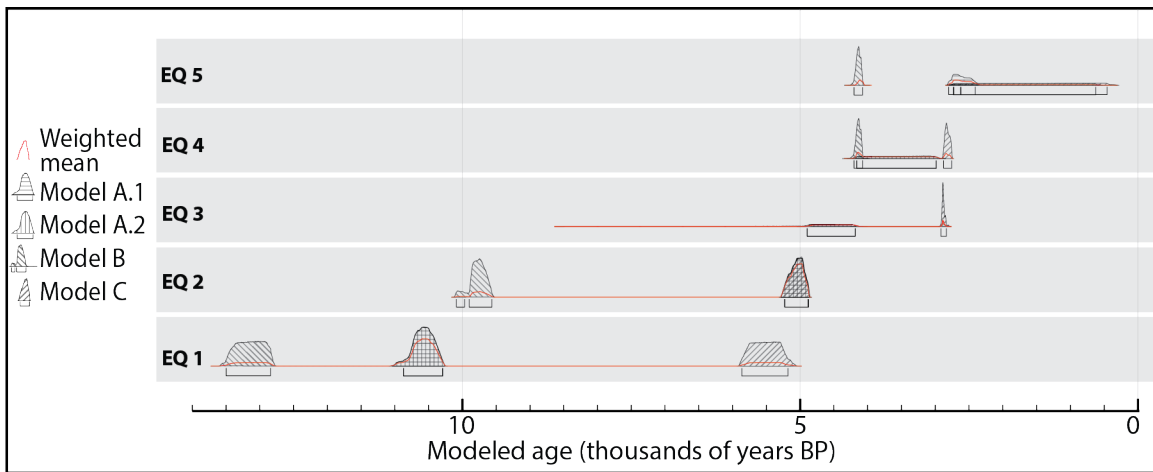


Figure C.12: Modeled earthquake age PDFs for OxCal models A.1, A.2, B, and C, as well as the weighted average PDF for each event. Model A.1 (preferred), model A.2 (younger MRE), model B (oldest events), and model C (youngest events) are shown with different hatched fill, while the weighted average PDF is shown in red.

Figure C.13 (following page): Likelihood and posterior PDFs for radiocarbon samples and event ages from the Woodlot Trench. Anterior likelihood (calibrated but unmodeled) PDFs are shown for all dated samples (Table 3.1). Modeled posterior probability PDFs for the four geologically plausible models shown with different colors—model A.1 (preferred), model A.2 (younger MRE), model B (oldest events), and model C (youngest events). Sample names in grey italics were not included in any of the OxCal models. See trench logs in Figures 3.3 and C.1-C.8, and unit descriptions in Table C.1.

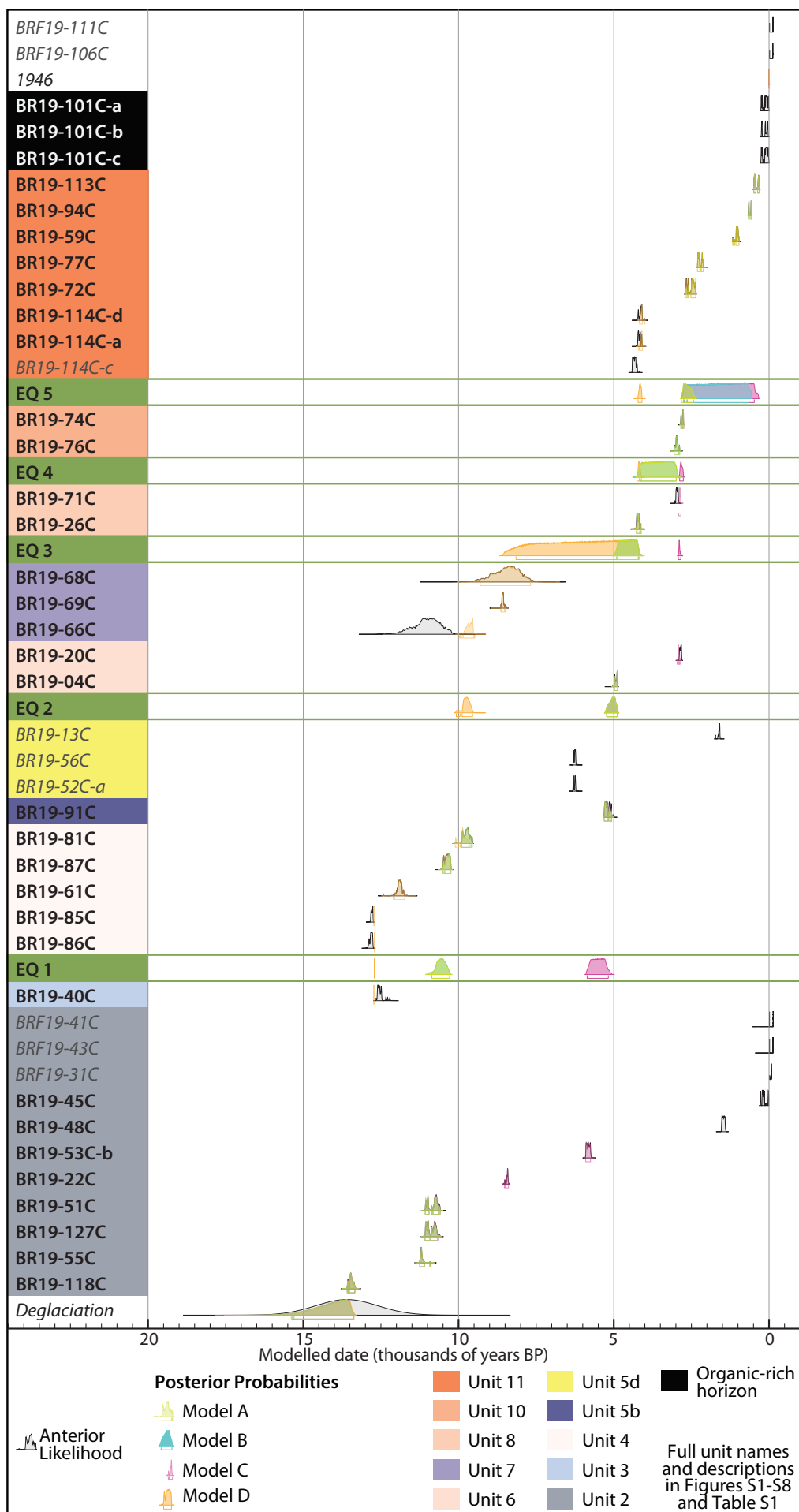




Figure C.14: Free face exposed along Strand Ew at Site 1D.

C.2 Tables

Table C.1: Sedimentary descriptions of units and soils mapped in the Woodlot Trench.

Unit	Interpretation	%C/ %M ¹	Support ²	Matrix ³	Matrix color ⁴	Sorting ⁵	%B/%C/ %P ⁶	Average clast size ⁷	Minimum clast size ⁷	Maximum clast size ⁷	Angular-ity ⁸	Clasts	Lower contact	Unit thickness	Unit shape	Soils and organics	Root size	Post-depositional deformation
11	Scarp-derived colluvial wedge; sheet wash facies	55/45	C	Poorly sorted SA-SR clay to coarse lithic sand with a sandy loam texture	Dry: 7.5YR2.5/3; Moist: 5YR3/4	M	0/2/98	Fine pebbles	Very fine pebbles	Small cobbles	A-R	Basalt, primarily slope-parallel in weak stratification and at the toe of the deposit	Clear, 25-mm-thick smooth contact, defined by burial of soil	0-24 cm	Tabular	Weak Bt and A horizons described separately, with pieces of charcoal and few roots throughout	<1-4 mm	—
	Scarp-derived colluvial wedge; debris facies	70/30	C	Poorly sorted SA-SR silt to very fine lithic pebble with a loamy sand texture	Dry: 10YR4/4; Moist: 7.5YR3/3	P	1/15/84	Medium pebbles	Fine pebbles	Small boulders	Pebbles: SA-SR; Cobbles and boulders: SR-R	Basalt and clasts of till, primarily in one clump above F1	Clear, 60-mm-thick wavy contact, defined by contact with fault and till and the projection of fault terminations	27-50 cm	Wedge; tips out on up-thrown block	Weak Bt and A horizons described separately, with bioturbation by common roots including large roots above and within fault zone	<1 mm-15 cm	—
10	Scarp-derived colluvial wedge	55/45	C	Moderately sorted SR clay to medium lithic sand with a sandy texture	Moist: 2.5YR3/3	P	0/7/93	Medium pebbles	Very fine pebbles	Small cobbles	Pebbles: A-SR, Cobbles: SR	Basalt, primarily at the base and toe of the deposit	Clear, 25-mm-thick smooth contact, defined by an influx of coarse cobbles and a transition to sandier matrix than the unit below	<9-57 cm	Wedge	Bt horizon described separately, with some bioturbation by common roots including larger roots near the fault zone. Charcoal throughout	<1.5cm	Sharp truncation by F4, indistinct back-tilting
9	Ponding against scarp	25/75	M	Moderately sorted SR clay to medium lithic sand with a loamy texture	2.5YR2.5/4	W	0/0/100	Fine pebbles	Very fine pebbles	Coarse pebbles	SA-SR	Basalt	Clear, 25-mm-thick smooth contact defined by a loss of coarse material and gain of a clay-dominated matrix	0-10 cm	Lenticular	Roots and charcoal throughout	1-10 mm	—

(continued).

Unit	Interpretation	%C/ %M ¹	Support ²	Matrix ³	Matrix color ⁴	Sorting ⁵	%B/%C/ %P ⁶	Average clast size ⁷	Minimum clast size ⁷	Maximum clast size ⁷	Angular-ity ⁸	Clasts	Lower contact	Unit thickness	Unit shape	Soils and organics	Root size	Post-depositional deformation
8	Scarp-derived colluvial wedge	60/40	C	Poorly sorted SA-SR clay to coarse lithic sand with a loamy sand texture	Moist: 2.5YR3/3; Increasing orange near the fault zone	P	5/15/80	Coarse pebbles	Very fine pebbles	Small boulders	Pebbles: A-R; Cobbles and boulders: SA-SR	Basalt and minor exotics ⁹	Clear, 25-mm-thick contact defined by a loss of clays, burial of a line of cobbles, and a change in matrix color and texture	0-57 cm	Wedge	Minor soil development and roots	1-15 mm	Truncated by fault, back-tilting
7	Ponding against scarp	30/70	M	Moderately sorted SR clay to coarse lithic sand with a loamy texture	Moist: 2.5YR2.5/3; Black staining on clasts	P	0/1/99	Fine pebbles	Very fine pebbles	Small cobbles	SA-SR	Basalt	Clear, 25-mm-thick contact defined by a loss of cobbles and a sudden change to a clay-dominated matrix	8-17 cm	Lenticular	Roots	1-5 mm	–
6	Scarp-derived colluvial wedge	65/35	C	Moderately sorted SR clay to medium sand with a sandy clay loam texture	Some orange mottle; black staining on clasts	P	0/35/65	Coarse pebbles	Fine pebbles	Large cobbles	Pebbles: A-R; Cobbles: SA-SR	Basalt along the base of the deposit and at the fault; toe of deposit almost entirely cobbles	Abrupt, 10-mm-thick contact defined by an influx of cobbles on top of ponded sediments	0-60 cm	Wedge	Roots	1-2 cm	Truncated by fault, back-tilted
5d	Hillslope colluvium	40/60	M	Poorly sorted SR silt to very coarse quartz & lithic sand with a loamy sand texture	Dry: 10YR5/8; Moist: 10YR3/6; Mottle development in soils	P	2/18/80	Medium pebbles	Very fine pebbles	Large boulder	A-SR	Unconsolidated with primarily basalt in chaotic clast orientations	Gradual, 60-mm-thick wavy contact defined by loss of any induration, increasingly chaotic texture, and more matrix than the units below	25-113 cm	Tabular	Bt horizon described separately, C horizon and some bioturbation from a large tree root in an area <30 cm by >1 m	<15 cm	–
5c	Possible scarp-derived colluvial deposit	90/10	C	Primarily silt	–	M-W	0/7/40/53 ¹⁰	Very fine pebbles	Very coarse sand	Large cobbles	Pebbles and sand: A-SR; Cobbles: A-SA	Basalt cobbles at the base and lateral terminations of deposit	Clear, 50-mm-thick smooth contact defined by a line of flat-lying cobbles burying a clay-rich deposit	0-35 cm	Lenticular-Tabular	Within the modern Bt horizon described separately, with common rootlets and a 3 cm burnt root	1-2 mm	Unclear

(continued).

Unit	Interpretation	%C/ %M ¹	Support ²	Matrix ³	Matrix color ⁴	Sorting ⁵	%B/%C/ %P ⁶	Average clast size ⁷	Minimum clast size ⁷	Maximum clast size ⁷	Angular- ity ⁸	Clasts	Lower contact	Unit thickness	Unit shape	Soils and organics	Root size	Post- depositional deformation
5b	Channel deposit ponding against scarp	20/80	M	Moderately sorted clay to fine sand with a silty clay loam texture	Moist: 2.5YR2.5/4	M	5/15/80	Coarse to very coarse pebbles	Fine pebbles	Medium boulders	Pebbles: SA-SR; Cobbles and boulders: R	Basalt pebbles at the base of deposit, grading laterally into coarser grains towards the SW	Gradual, 50-mm-thick contact defined by an influx of clay and a clay-dominated matrix, loss of most cobbles, and a color change	10-43 cm	Lenticular	Roots	1-30 mm	—
5a	Channel deposit	95/5	C	Clay at the base of the deposit	—	M	0/2/58/40 ¹⁰	Fine pebbles	Coarse sand	Small cobbles	SA-SR	Basalt and lithic clasts in normal grading with interbedded sands and pebbles	Clear, 25-mm-thick smooth contact, defined by an influx of sand and pebbles, loss of coarse cobbles, and burial of paleosol	0-66 cm	Lenticular	Few roots	<6 mm	—
4	Scarp-derived colluvial wedge	70/30	C	Moderately sorted SA clay to medium lithic sand with a sandy loam texture	Moist: 7.5YR3/4	P	1/30/69	Fine to medium pebbles	Very fine pebbles	Small boulder (30 cm)	A-SR	Basalt and minor exotics are coarser at the base of the deposit and near the fault, and cobbles form a line at the toe of the deposit	Clear, 25-mm-thick contact defined by an influx of coarse material, burial of paleosol, and decreased induration	0-40 cm	Wedge	Paleo Bt horizon described separately	1-2 mm	Truncated by fault, back-tilted
3a	Debris flow	40/60	M	Moderately sorted SA-SR silt to medium lithic sand with a sandy loam texture	Moist: 10YR3/6	P	0/15/85	Coarse gravel	Very fine pebbles	Small cobbles	SA	Basalt and minor exotics in fairly distinct reverse grading	Clear, 25-mm-thick channel-shaped contact, defined by an influx of coarse material, a loss of stratification and fine pebbles, and a color change	7-56 cm	Lenticular	Modern Bt horizon overprinting on paleo Bt horizon, described separately	1-2 mm	—

(continued).

Unit	Interpretation	%C/ %M ¹	Support ²	Matrix ³	Matrix color ⁴	Sorting ⁵	%B/%C/ %P ⁶	Average clast size ⁷	Minimum clast size ⁷	Maximum clast size ⁷	Angular- ity ⁸	Clasts	Lower con- tact	Unit thick- ness	Unit shape	Soils and organics	Root size	Post- depositional deformation
3	Channel deposit	95/5	C	—	—	M	0/40/60	Coarse pebbles	Fine pebbles	Large cobbles	SA-R	Basalt and minor exotics in normal grading with interbedded pebbles and gravels	Clear, 25-mm-thick wavy contact defined by an influx of coarse, better-sorted material, a loss of matrix, and burial of paleosol	0-100 cm	Lenticular	Bt horizon described separately, some smaller roots and larger tree roots	< 7 cm	—
2	Weathering profile through Unit 1	55/45	C	Moderately sorted SA-SR fine to coarse lithic sands with clay coatings, with a loamy sand texture	Dry: 2.5Y5/4; Mottle: 10YR3/4	P	0/10/90	Coarse pebbles	Fine pebbles	Large cobbles	Pebbles: SA; Cobbles: SR	Primarily basalt clasts, evenly distributed throughout deposit	Indistinct, 100-mm-thick gradational contact, described by hillslope creep texture and loss of induration	81-86 cm	Tabular	Paleo Bt horizon described separately, with pods of indurated material and few roots	2 mm	Offset by fault
1	Paraglacial debris cones	40/60	M	Poorly sorted SR clay to coarse lithic sand, with a loamy sand texture	Moist: 10YR4/4	P	5/35/60	Small cobbles	Fine pebbles	Small boulders	SR; Boulders: R	Basalt and minor exotics, evenly distributed throughout deposit in weak stratification	Buried	>4 m	—	Soil developed (see Unit 2); very indurated	1 mm-5 cm	Offset by fault, slight synform
Soils ¹¹																		
Modern A horizon		10/90	M	Fairly well sorted silt and organics with a loamy texture	Dry: 7.5YR2.5/2; Moist: 7.5YR2.5/1	—	0/0/100	Very fine pebbles	—	—	SR	—	Abrupt, 10-mm-thick smooth contact	1-3 cm, up to 10 cm	—	Abundant roots	—	—
Modern Bt horizon		40/60	M	Poorly sorted SA-SR clay to medium lithic sand with a loamy sand texture	2.5YR2.5/4; 7.5YR3/4; 10YR4/6	—	0/10/90	Coarse pebbles	Very fine pebbles	Small cobbles	A-SR	Primarily basalt	Diffuse, >120-mm-thick wavy contact	<100 cm	—	Abundant roots	—	—

(continued).

Unit	Interpre- tation	%C/ %M ¹	Sup- port ²	Matrix ³	Matrix color ⁴	Sort- ing ⁵	%B/%C/ %P ⁶	Average clast size ⁷	Minimum clast size ⁷	Maximum clast size ⁷	Angular- ity ⁸	Clasts	Lower con- tact	Unit thick- ness	Unit shape	Soils and organics	Root size	Post- depositional deformation
Modern Bt horizon on Unit 11		40/60	M	Very poorly sorted A- SR silt to very coarse lithic sand with a sandy loam texture	5YR3/3	–	0/5/95	Medium pebbles	Very fine pebbles	Small cob- bles	SA-SR	Basalt and others	Gradual, 60-mm-thick irregular contact	–	–	Abundant roots	–	
Modern E horizon		60/40	C	Moderately sorted silt with a sandy loam texture	Dry: 7.5YR4/4; Moist: 7.5YR3/2	–	0/0/100	Fine pebbles	Very fine pebbles	Very coarse pebbles	SA-SR	Basalt and exotics	Clear, 15- mm-thick broken con- tact	0-30 cm	Pod	Pods around roots	–	
Paleo Bt horizon		40/60	M	Poorly sorted SR clay to coarse lithic sand with a sandy loam texture	10YR3/6	–	0/20/80	Medium pebbles	Very fine pebbles	Large cob- bles	SA-R	Basalt and exotics	Gradual, 80-mm- thick wavy contact	10-30 cm	–	Roots	–	Truncated by fault

¹% cobbles, % matrix

²C = clast-supported, M = matrix=supported

³Texture described following soil classification system of Thien (1979).

⁴Colors described following Munsell soil color charts (Inc., 2010).

⁵W = well-supported, M = moderately sorted, P = poorly sorted

⁶% boulders, % cobbles, % pebbles

⁷Clast sizes following modified Wentworth scale after Lane (1947).

⁸A = angular, SA = sub-angular, SR = sub-rounded, R = rounded

⁹Exotics consist primarily of granodiorite, found in the Island Intrusions and West Coast Crystalline Complex west of Alberni Valley (DeBari et al., 1999).

¹⁰% boulders, % cobbles, % pebbles, % sand

¹¹Soils mapped and described following Schoeneberger et al. (2012); Thien (1979).

Table C.2: Physical characteristics used to assess whether units could confidently be described as a scarp-derived colluvial wedge. R - Robust; M - Moderate; W - Weak; N - Nonexistent.

Colluvial wedge unit	Wall ¹	Wedge-shaped geometry	Distinct Lower Contact ²	Presence of slope-parallel stratification ³	Texture distinct from neighboring units	Color distinct from neighboring units	Prevalence of dispersed organics ⁴	Soil development at top of unit ⁵	Burial of soil below unit	Presence of fault terminations ⁶	Back-rotation of wedge base ⁷
11	1	N	R	R	M	R	M	W	R	R	N/A
	4	R	W	W	W	M	M	W	R	R	N/A
10	1	R	R	W	M	M	R	R	M	N	W
	4	R	M	M	M	M	R	R	R	R	W
8	1	M	R	N	R	R	R	M	R	N	W
	4	R	R	N	R	R	M	M	R	N	M
6	1	M	M	M	M	M	M	R	M	N	M
	4	R	R	N	R	R	R	M	R	N	R
4	1	R	R	N	R	R	W	R	R	N	R
	2L	R	M	N	R	M	M	R	R	N	R
	3L	R	R	W	R	R	M	R	R	N	R

¹Trench walls labeled in Figure 3.2.
²Lower contact assessed using the width, abruptness, and distinctness of the contact.
³Slope-parallel texture assessed on whether stratification and aligned clasts were parallel to the modern ground-surface slope (or the assumed ground-surface slope prior to back-rotation).
⁴Dispersed organics include both charcoal and plant material throughout the unit.
⁵Soil development assessed on coloration and texture changes developed at the top of the unit.
⁶Fault terminations include distinct fault planes terminating at the base of or within the wedge.
⁷Back-rotation assessed using the angle between the modern scarp face and the base of the wedge.

Table C.3: Anterior likelihood and posterior modeled calibrated ages for the Woodlot Trench radiocarbon samples included in OxCal models A.1, A.2, B, and C.

Sample	Unit	Median calibrated age (cal yr BP)	Model A ¹ and A.1 ² (cal yr BP)	Model A.2 ² (cal yr BP)	Model B ² (cal yr BP)	Model C ² (cal yr BP)
1946 Vancouver Island earthquake ³		1946.5 CE	1946.5 CE	1946.5 CE	1946.5 CE	1946.5 CE
BR19-106C	Boulder ⁴	1956 CE	1180-970 2690-2350	1946.5 CE	1180-970 2700-2350 4210-4080 4180-4000	500-320
BR19-111C	4	1956 CE				
BR19-101C-a	5b ⁵	102				
BR19-101C-b	5b ⁵	120				
BR19-101C-c	5b ⁵	106				
BR19-59C	11	1021				
BR19-72C	11	2474				
BR19-114C-A	11	4181				
BR19-114C-C	11	4339				
BR19-114C-D	11	4129				
BR19-113C	11	451				
BR19-94C	11	590				
BR19-77C	11	2250				
EQ 5			2830-2430	2760-650	4230-4100	2760-480
BR19-76C	10	2977	3070-2880	2970-2860		2850-2750
BR19-74C	10	2977	2850-2750	2850-2750		
EQ 4			4170-2990 2990-2880 ¹	2990-2880	4270-4120	2900-2770
BR19-71C	8	2918	3000-2900 ¹	3000-2890	4300-4510	2920-2840
BR19-26	8	4192	4250-4090	4250-4090		
EQ 3			4900-4180	4900-4180	8170-4220	2940-2850
BR19-68C	7	8412	4970-4850	4970-4850	9300-7670	2960-2850 4970-4850
BR19-66C	7	11011			9990-9490	
BR19-69C	7	8565			8640-8450	
BR19-56C	5d	6234				
BR19-52C-a	5d	6237				
BR19-13C	5d	1556				
BR19-20C	6	2814				
BR19-04C	6	4905				
EQ 2			5230-4870	5230-4870	10090-9560	5220-4870
BR19-91C	5b	5184	5290-5040	5290-5040	10120-9660 12740-12710 12100-11730 12740-12720 10500-10240	5290-5040
BR19-81C	4	9739	9900-9550	9900-9550		
BR19-86C	4	12811	10500-10240	10500-10240		
BR19-61C	4	11893				
BR19-85C	4	12789				
BR19-87C	4	10363				
EQ 1			10870-10280	10870-10280	13460-12810	5860-5170
BR19-40C	3	12547	11250-10890	11250-10890		11250-10890 5910-5740 11070-10580
BR19-55C	2	11188				
BR19-53C-b	2	5820				
BR19-51C	2	10758	11080-10590	11080-10590		

(continued).

Sample	Unit	Median calibrated age (cal yr BP)	Model A ¹ and A.1 ² (cal yr BP)	Model A.2 ² (cal yr BP)	Model B ² (cal yr BP)	Model C ² (cal yr BP)
BR19-48C	2	1467	11080-10680	11080-10680		8520-8380 11080-10660
BR19-45C	2	187				
BR19-22C	2	8427				
BR19-127C	2	10859				
BR19-31C	2	1965 CE				
BR19-41C	2	1955 CE				
BR19-43C	2	1956 CE	13590-13330	13590-13330	13590-13330	13590-13330
BR19-118C	2	13462				
Deglaciation ⁶			15310-13370	15340-13370	15420-13370	15320-13370

¹Single sample included in model A but excluded from model A.1 and model A EQ 4 age listed in italics.

²Modeled ages are reported as the modeled 95% confidence intervals as calculated by OxCal (calibrated years Before Present, where present = 1950).

³Minimum model age is the occurrence of the June 1946 Vancouver Island earthquake.

⁴Samples BR19-106C and BR190111C were collected from beneath boulders on the ground surface (see Fig. 3.3).

⁵Samples BR19-101C-a, BR19-101C-c, and BR19-101C-b are from an organic-rich horizon at the top of unit 5b (see Fig 3.4).

⁶Maximum model age based on timing of local ice retreat (Mosher & Hewitt, 2004; Easterbrook, 1992; Clague & James, 2002).

Table C.4: Structural measurements collected in the Woodlot Trench.

No. ¹	Unit ²	Wall ³		Strike	Dip	Trend	Plunge
<i>Fault planes</i>							
1	–	1	F1	286	77	N	–
2	–	1	F1	281	83	N	–
3	–	1	F1	125	86	S	–
4	–	2L	F3	305	90	–	–
5	–	1	F0.5	290	55	NW	–
6	–	4	F1	305	57	NE	–
7	–	3L	F4	281	80	NE	–
8	–	3L	F3	300	66	NE	–
9	–	3L	F3	280	84	NE	–
10	–	Headwall	F1	304	84	NE	–
11	–	Headwall	F1	297	90	NE	–
12	–	Headwall	F1	293	89	NE	–
13	–	2L	F4	110	84	S	–
14	–	1	F2	309	87	N	–
15	–	1	F4	80	76	S	–
16	–	1	F4	95	81	S	–
17	–	1	F3	109	85	S	–
17a	–	1	F3	100	89	S	–
	<i>Average</i>		<i>F1</i>	<i>294</i>	<i>80</i>	<i>NE</i>	–
	<i>Average</i>		<i>F2</i>	<i>300</i>	<i>82</i>	<i>NE</i>	–
	<i>Average</i>		<i>F3</i>	<i>315</i>	<i>57</i>	<i>NE</i>	–
	<i>Average</i>		<i>F4</i>	<i>92</i>	<i>71</i>	<i>S</i>	–
<i>Slickenlines and Striations</i>							
	SS	4	F2	290	78	N	90
	1	Headwall	F1	283	79	N	Rake: 73° from E
	1	Headwall	–	–	–		9
<i>Depositional Fabrics</i>							
1	3	AN	aligned cobbles	45	25	SE	–
2	2	AN	stratification	150	23	SW	–
3	2	AN	stratification	–	–		–
4	2	1	apparent dip	–	–		203
5	2	1	apparent dip	–	–		207
6	2	1	apparent dip	–	–		198
7	1	2L	stratification or fissility	76	40	S	–
8	SS	1	stratification	325	45	NE	–
9	SS	1	stratification	327	50	NE	–
10	1	1	stratification	75	9	S	–
11	1	1	stratification	106	38	S	–
<i>Clast Orientations</i> ⁴							
	2	1		–	–		142
	2	1		49	22	SE	–

(continued).

No. ¹	Unit ²	Wall ³		Strike	Dip	Trend	Plunge	
	2	BN		148	43	SW	—	—
	1	BN		—	—		80	11
	2	BN		291	27	N	—	—
	2	BS		90	22	S	—	—
	2	BS		—	—		238	20
	2	BS		240	24	NW	—	—
	2	BS		85	24	S	—	—
	2	BS		255	49	N	—	—
	2	BS		265	48	N	—	—
	1	2		280	36	N	—	—
	1	2		317	25	NE	—	—
	1	2		—	—		285	25
	2	1		—	—		100	16
	2	3L		240	46	NW	—	—
	2	BN		—	—		5	10
	2	BN		—	—		64	12
	SS	3L		—	—		234	30
	SS	3L		—	—		310	49

¹Numbers correspond to locations marked on the trench logs (Figures C.1-C.8). Fault planes are in red, and depositional fabrics are in blue.

²Units numbered and described in Table C.1. SS refers to sheared sediment within the fault zone.

³1 - Slice 1; 2 - Slice 2 (upper); 2L - Slice 2L (lower); 3L - Slice 3L (lower); 4 - Slice 4; AN - A-Wing North; AS - A-Wing South; BN - B-Wing North; BS - B-Wing South. See Figure 3.2 for trench footprint.

⁴Strike and Dip of ab-planes; trend and plunge of long axes.

Table C.5: Estimates of earthquake magnitude.

Event	Vertical displacement from wedge height (m)	Strike Slip (m)		Total slip (m)		Corresponding length(km)		Moment Magnitude	
		Constant slip ¹	Constant ratio ²	Constant slip	Constant ratio	(N) ³	(SS) ⁴	(N) ⁵	(SS) ⁶
Event 5	1.14	0.9	0.9-1.03	1.45	1.54	48-51	145-154	6.9	7.2-7.5
Event 4	0.96	0.9	0.86	1.32	1.29	43-44	129-132	6.9	7.1-7.4
Event 3	1.08	0.9	0.97	1.41	1.45	47-48	141-145	6.9	7.1-7.4
Event 2	0.61	0.9	0.55	1.09	0.82	27-36	82-109	6.7-6.9	7.0-7.3
Event 1	0.56	0.9	0.5	1.06	0.75	25-35	75-106	6.7-6.8	6.9-7.3

¹Right-lateral displacement of paleochannel thalweg of 4.57 m, average 0.9 m/event (see Section 4.5.3).

²Ratio of right-lateral to vertical of ~0.8:1 from paleochannel thalweg, and median ratio of 0.9:1 from Chapter 1 (see Section 4.5.3).

³Calculated using the regression for length from average geologic slip for normal faults, from Wesnousky (2008).

⁴Calculated using the regression for length from average geologic slip for strike-slip faults, from Wesnousky (2008).

⁵Calculated using the regressions for moment magnitude from average displacement for normal faults from Wells and Coppersmith (1994), and the regression for moment magnitude from length for normal faults from Wesnousky (2008).

⁶Calculated using the regression for moment magnitude from average displacement for strike-slip faults from Wells and Coppersmith (1994), and the regression for moment magnitude from length for strike-slip faults from Wesnousky (2008).

Table C.6: Interval slip rates at the Woodlot Trench site given multiple earthquake age models.

Interval	Displacement (m)	Model A.1		Model A.2		Model B		Model C	
		RI ² (kyrs)	Slip rate (mm/yr)	RI ² (kyrs)	Slip rate (mm/yr)	RI ² (kyrs)	Slip rate (mm/yr)	RI ² (kyrs)	Slip rate (mm/yr)
Event 4 to 5	1.45-1.54	0.2-1.8	0.8-7.7	0.2-2.3	0.6-7.7	<0.1	>14.5	<2.4	>0.6
Event 3 to 4	1.29-1.5	<1.9	>0.7	1.2-2	0.4-1.3	<4.1	>0.3	<0.2	>6.5
Event 2 to 3	1.41-1.5	<1	>1.4	<1	>1.4	1.4-5.9	0.2-1.1	1.9-2.4	0.6-0.8
Event 1 to 2	0.82-1.5	5.1-6	0.2	5.1-6	0.2	2.7-3.9	0.2-0.6	<1	>0.8

C.3 Woodlot Trench OxCal Models

C.3.1 Model A.1

```
Options()
{
    BCAD=FALSE;
};

//this model uses all samples determined to be the youngest non-bioturbated ages in
each unit, and all samples not excluded by those ages. Also excluded is sample BR19-71C
to avoid over-constraining Event 4. "up" = up-thrown block; "down" = down-thrown block.
All colluvial wedge and ponding units are only found on the down-thrown block.

Plot()
{
    Sequence("BRF Trench Model A.1")
    {
        Zero_Boundary("base of sequence");
        C_Date("deglaciation", 13600, 1000);
        Phase("Unit 2, Paleo Hillslope Colluvium")
        {
            R_Date("BR19-118C (down)", 11600, 50);
            R_Date("BR19-55C (up)", 9740, 40);
            R_Date("BR19-127C (down)", 9525, 40);
        }
    }
}
```



```

//R_Date("BR19-22C (down)", 7655, 20);
//R_Date("BR19-45C (up)", 165, 15);
//R_Date("BR19-48C (up)", 1590, 25);
R_Date("BR19-51C (up)", 9495, 40);
//R_Date("BR19-53C-b (up)", 5060, 25);
};
Phase("Unit 3, Washed Cobbles")
{
    //R_Date("BR19-40C (up)", 10500, 40);
};
Date("EQ 1");
Phase("Unit 4 - Colluvial Wedge")
{
    R_Date("BR19-87C", 9205, 45);
    //R_Date("BR19-85C", 10875, 25);
    //R_Date("BR19-61C", 10220, 45);
    //R_Date("BR19-86C", 10910, 45);
    R_Date("BR19-81C", 8755, 35);
};
Phase("Unit 5b - Ponding")
{
    R_Date("BR19-91C", 4480, 15);
};
Phase("Unit 5d - Modern Hillslope Colluvium")
{
    //R_Date("BR19-13C (up)", 1670, 20);
    //R_Date("BR19-52C-a (up)", 5450, 25);
    //R_Date("BR19-56C (up)", 5430, 20);
};
Date("EQ 2");
Phase("Unit 6 - Colluvial Wedge")
{
    R_Date("BR19-04C", 4345, 20);
};

```

```

    //R_Date("BR19-20C", 2735, 15);
};
Phase("Unit 7 - Ponding")
{
    //R_Date("BR19-69C", 7790, 30);
    //R_Date("BR19-66C", 9640, 290);
    //R_Date("BR19-68C", 7540, 380);
};
Date("EQ 3");
Phase("Unit 8 - Colluvial Wedge")
{
    R_Date("BR19-26C", 3805, 15);
    //R_Date("BR19-71C", 2820, 15);
};
Date("EQ 4");
Phase("Unit 10 - Colluvial Wedge")
{
    R_Date("BR19-76C", 2860, 20);
    R_Date("BR19-74C", 2700, 15);
};
Phase("Organic-Rich Horizon")
{
    //R_Date("BR19-101C-c", 125, 20);
    //R_Date("BR19-101C-b", 95, 15);
    //R_Date("BR19-101C-a", 130, 15);
};
Date("EQ 5");
Phase("Unit 11 - Colluvial Wedge")
{
    R_Date("BR19-72C", 2440, 20);
    R_Date("BR19-77C", 2185, 15);
    R_Date("BR19-59C", 1145, 15);
    R_Date("BR19-94C", 665, 15);
};

```

```

R_Date("BR19-113C", 375, 20);
//R_Date("BR19-114C-d", 3770, 20);
//R_Date("BR19-114C-c", 3895, 20);
//R_Date("BR19-114C-a", 3795, 20);
};
Boundary("Sequence ends", 1946);
};
};

```

C.3.2 Model A.2

```

Options()
{
  BCAD=FALSE;
};

//this model uses all samples determined to be the youngest non-bioturbated ages in
each unit, and all samples not excluded by those ages. Also excluded are sample
BR19-71C to avoid over-constraining Event 4, and samples BR19-72C, -77C, and -59C
to allow for a young Event 5 consistent with geomorphic observations. "up" = up-thrown
block; "down" = down-thrown block. All colluvial wedge and ponding units are only found
on the down-thrown block.

Plot()
{
  Sequence("BRF Trench Model \A.2)
  {
    Zero_Boundary("base of sequence");
    C_Date("deglaciation", 13600, 1000);
    Phase("Unit 2, Paleo Hillslope Colluvium")
    {
      R_Date("BR19-118C (down)", 11600, 50);
      R_Date("BR19-55C (up)", 9740, 40);
      R_Date("BR19-127C (down)", 9525, 40);
      //R_Date("BR19-22C (down)", 7655, 20);
    }
  }
}

```

```

//R_Date("BR19-45C (up)", 165, 15);
//R_Date("BR19-48C (up)", 1590, 25);
R_Date("BR19-51C (up)", 9495, 40);
//R_Date("BR19-53C-b (up)", 5060, 25);
};
Phase("Unit 3, Washed Cobbles")
{
    //R_Date("BR19-40C (up)", 10500, 40);
};
Date("EQ 1");
Phase("Unit 4 - Colluvial Wedge")
{
    R_Date("BR19-87C", 9205, 45);
    //R_Date("BR19-85C", 10875, 25);
    //R_Date("BR19-61C", 10220, 45);
    //R_Date("BR19-86C", 10910, 45);
    R_Date("BR19-81C", 8755, 35);
};
Phase("Unit 5b - Ponding")
{
    R_Date("BR19-91C", 4480, 15);
};
Phase("Unit 5d - Modern Hillslope Colluvium")
{
    //R_Date("BR19-13C (up)", 1670, 20);
    //R_Date("BR19-52C-a (up)", 5450, 25);
    //R_Date("BR19-56C (up)", 5430, 20);
};
Date("EQ 2");
Phase("Unit 6 - Colluvial Wedge")
{
    R_Date("BR19-04C", 4345, 20);
    //R_Date("BR19-20C", 2735, 15);
};

```

```

};
Phase("Unit 7 - Ponding")
{
    //R_Date("BR19-69C", 7790, 30);
    //R_Date("BR19-66C", 9640, 290);
    //R_Date("BR19-68C", 7540, 380);
};
Date("EQ 3");
Phase("Unit 8 - Colluvial Wedge")
{
    R_Date("BR19-26C", 3805, 15);
    //R_Date("BR19-71C", 2820, 15);
};
Date("EQ 4");
Phase("Unit 10 - Colluvial Wedge")
{
    R_Date("BR19-76C", 2860, 20);
    R_Date("BR19-74C", 2700, 15);
};
Phase("Organic-Rich Horizon")
{
    //R_Date("BR19-101C-c", 125, 20);
    //R_Date("BR19-101C-b", 95, 15);
    //R_Date("BR19-101C-a", 130, 15);
};
Date("EQ 5");
Phase("Unit 11 - Colluvial Wedge")
{
    //R_Date("BR19-72C", 2440, 20);
    //R_Date("BR19-77C", 2185, 15);
    //R_Date("BR19-59C", 1145, 15);
    R_Date("BR19-94C", 665, 15);
    R_Date("BR19-113C", 375, 20);
};

```

```

//R_Date("BR19-114C-d", 3770, 20);
//R_Date("BR19-114C-c", 3895, 20);
//R_Date("BR19-114C-a", 3795, 20);
};
Boundary("Sequence ends", 1946);
};
};

```

C.3.3 Model B

```

Options()
{
    BCAD=FALSE;
};

//this model makes all EQs as old as possible. "up" = up-thrown block;
"down" = down-thrown block. All colluvial wedge and ponding units are only found
on the down-thrown block.

Plot()
{
    Sequence("BRF Trench Model B")
    {
        Zero_Boundary("base of sequence");
        C_Date("deglaciation", 13600, 1000);
        Phase("Unit 2, Paleo Hillslope Colluvium")
        {
            R_Date("BR19-118C (down)", 11600, 50);
            //R_Date("BR19-127C (down)", 9525, 40);
            //R_Date("BR19-22C (down)", 7655, 20);
            //R_Date("BR19-45C (up)", 165, 15);
            //R_Date("BR19-48C (up)", 1590, 25);
            //R_Date("BR19-51C (up)", 9495, 40);
            //R_Date("BR19-53C-b (up)", 5060, 25);
            //R_Date("BR19-55C (up)", 9740, 40);

```

```

};
Phase("Unit 3, Washed Cobbles")
{
    //R_Date("BR19-40C (up)", 10500, 40);
};
Date("EQ 1");
Phase("Unit 4 - Colluvial Wedge")
{
    R_Date("BR19-87C", 9205, 45);
    R_Date("BR19-85C", 10875, 25);
    R_Date("BR19-61C", 10220, 45);
    R_Date("BR19-86C", 10910, 45);
    R_Date("BR19-81C", 8755, 35);
};
Phase("Unit 5b - Ponding")
{
    //R_Date("BR19-91C", 4480, 15);
};
Phase("Unit 5d - Modern Hillslope Colluvium")
{
    //R_Date("BR19-13C (up)", 1670, 20);
    //R_Date("BR19-52C-a (up)", 5450, 25);
    //R_Date("BR19-56C (up)", 5430, 20);
};
Date("EQ 2");
Phase("Unit 6 - Colluvial Wedge")
{
    //R_Date("BR19-04C", 4345, 20);
    //R_Date("BR19-20C", 2735, 15);
};
Phase("Unit 7 - Ponding")
{
    R_Date("BR19-69C", 7790, 30);

```



```

R_Date("BR19-66C", 9640, 290);
R_Date("BR19-68C", 7540, 380);
};
Date("EQ 3");
Phase("Unit 8 - Colluvial Wedge")
{
  R_Date("BR19-26C", 3805, 15);
  //R_Date("BR19-71C", 2820, 15);
};
Date("EQ 4");
Phase("Unit 10 - Colluvial Wedge")
{
  //R_Date("BR19-74C", 2700, 15);
  //R_Date("BR19-76C", 2860, 20);
};
Phase("Organic-Rich Horizon")
{
  //R_Date("BR19-101C-c", 125, 20);
  //R_Date("BR19-101C-b", 95, 15);
  //R_Date("BR19-101C-a", 130, 15);
};
Date("EQ 5");
Phase("Unit 11 - Colluvial Wedge")
{
  R_Date("BR19-77C", 2185, 15);
  R_Date("BR19-94C", 665, 15);
  R_Date("BR19-113C", 375, 20);
  R_Date("BR19-114C-d", 3770, 20);
  //R_Date("BR19-114C-c", 3895, 20);
  R_Date("BR19-114C-a", 3795, 20);
  R_Date("BR19-72C", 2440, 20);
  R_Date("BR19-59C", 1145, 15);
};

```

```

    Boundary("Sequence ends", 1946);
};
};

```

C.3.4 Model C

```

Options()
{
    BCAD=FALSE;
};

//this model makes all the EQs as young as possible. "up" = up-thrown block;
"down" = down-thrown block. All colluvial wedge and ponding units are only found
on the down-thrown block.

Plot()
{
    Sequence("BRF Trench Model \C)
    {
        Zero_Boundary("base of sequence");
        C_Date("deglaciation", 13600, 1000);
        Phase("Unit 2, Paleo Hillslope Colluvium")
        {
            R_Date("BR19-118C (down)", 11600, 50);
            R_Date("BR19-127C (down)", 9525, 40);
            R_Date("BR19-22C (down)", 7655, 20);
            //R_Date("BR19-45C (up)", 165, 15);
            //R_Date("BR19-48C (up)", 1590, 25);
            R_Date("BR19-51C (up)", 9495, 40);
            R_Date("BR19-53C-b (up)", 5060, 25);
            R_Date("BR19-55C (up)", 9740, 40);
        };
        Phase("Unit 3, Washed Cobbles")
        {
            //R_Date("BR19-40C (up)", 10500, 40);

```

```

};
Date("EQ 1");
Phase("Unit 4 - Colluvial Wedge")
{
    //R_Date("BR19-87C", 9205, 45);
    //R_Date("BR19-85C", 10875, 25);
    //R_Date("BR19-61C", 10220, 45);
    //R_Date("BR19-86C", 10910, 45);
    //R_Date("BR19-81C", 8755, 35);
};
Phase("Unit 5b - Ponding")
{
    R_Date("BR19-91C", 4480, 15);
};
Phase("Unit 5d - Modern Hillslope Colluvium")
{
    //R_Date("BR19-13C (up)", 1670, 20);
    //R_Date("BR19-52C-a (up)", 5450, 25);
    //R_Date("BR19-56C (up)", 5430, 20);
};
Date("EQ 2");
Phase("Unit 6 - Colluvial Wedge")
{
    R_Date("BR19-04C", 4345, 20);
    R_Date("BR19-20C", 2735, 15);
};
Phase("Unit 7 - Ponding")
{
    //R_Date("BR19-69C", 7790, 30);
    //R_Date("BR19-66C", 9640, 290);
    //R_Date("BR19-68C", 7540, 380);
};
Date("EQ 3");

```

```

Phase("Unit 8 - Colluvial Wedge")
{
  //R_Date("BR19-26C", 3805, 15);
  R_Date("BR19-71C", 2820, 15);
};
Date("EQ 4");
Phase("Unit 10 - Colluvial Wedge")
{
  R_Date("BR19-74C", 2700, 15);
  //R_Date("BR19-76C", 2860, 20);
};
Phase("Organic-Rich Horizon")
{
  //R_Date("BR19-101C-c", 125, 20);
  //R_Date("BR19-101C-b", 95, 15);
  //R_Date("BR19-101C-a", 130, 15);
};
Date("EQ 5");
Phase("Unit 11 - Colluvial Wedge")
{
  //R_Date("BR19-77C", 2185, 15);
  //R_Date("BR19-94C", 665, 15);
  R_Date("BR19-113C", 375, 20);
  //R_Date("BR19-114C-d", 3770, 20);
  //R_Date("BR19-114C-c", 3895, 20);
  //R_Date("BR19-114C-a", 3795, 20);
  //R_Date("BR19-72C", 2440, 20);
  //R_Date("BR19-59C", 1145, 15);
};
Boundary("Sequence ends", 1946);
};
};

```

References

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Appendix D

Supporting Information for Chapter 4

D.1 Figures

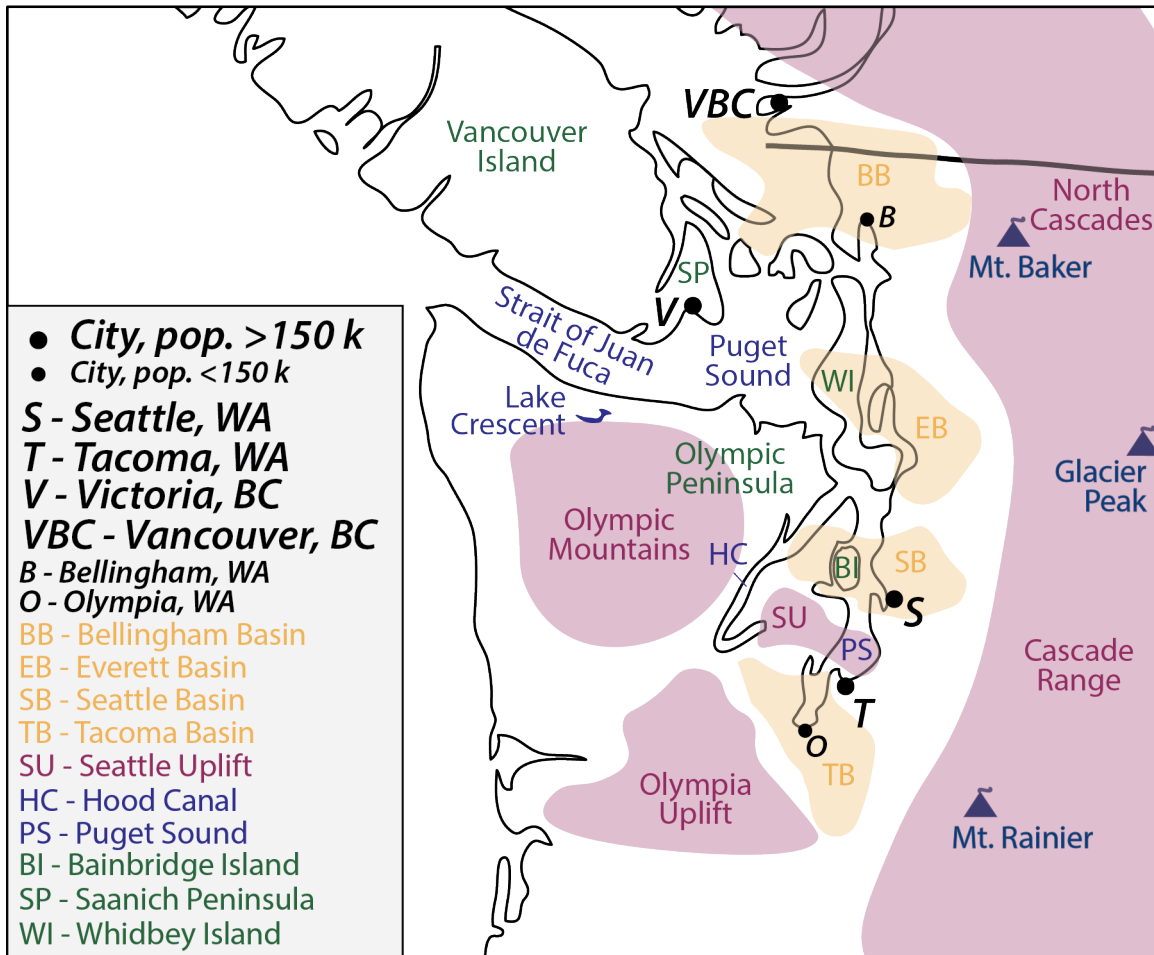


Figure D.1: Locations referenced in Chapter 4. Basins are shown in orange. Mountain ranges and topographic uplifts are shown in magenta.

D.2 Monte Carlo script for calculating slip rates

```
# Calculating interval slip rates from trench data
# written by Emerson Lynch 22 Feb 2023
# updated 14 Apr 2023 - changed to 2 sigma
# based on RISer python script by Zinke et al 2017
#### currently returns very large uncertainty for individual intervals; use with caution

## set working directory to folder where script & profiles are kept
setwd("/file-path")

# install.packages("matlib")
library(matlib)

rm(list=ls()) #clear workspace if you have already run script in this session

# prepared survey csv or txt file (with extension)
mydata <- read.csv(file="./file-name.csv", header=TRUE, sep=",")
# #calculate middles of ranges
# mean_slip = ((mydata$minDisp+mydata$maxDisp)/2) # middle of slip range
# slip_sd = (mean_slip-mydata$minDisp) # set sd for cumulative displacement as
# the size of the range
#
# mean_age = ((mydata$minAge+mydata$maxAge)/2) # middle of slip range
# age_sd = (mean_age-mydata$minAge) # set sd for cumulative displacement as
# the size of the range

#turn units into mm/yr
mydata$minDisp<-mydata$minDisp*1000 #turn m disps into mm disps if needed
mydata$maxDisp<-mydata$maxDisp*1000
mydata$minCumDisp<-mydata$minCumDisp*1000
mydata$maxCumDisp<-mydata$maxCumDisp*1000
n <- 10000 #number of runs
```

```

##### youngest interval slip rate #####
interval<-"interval_name"
event1<-mydata[1,]
event2<-mydata[2,]

mean_age1 = ((event1$minAge+event1$maxAge)/2) # middle of age range
age_sd1 = (mean_age1-event1$minAge)/2 # set sd for cumulative displacement as half the range

mean_age2 = ((event2$minAge+event2$maxAge)/2) # middle of age range
age_sd2 = (mean_age2-event2$minAge)/2 # set sd for cumulative displacement as half the range

mean_slip = ((event1$minDisp+event1$maxDisp)/2) # middle of slip range
slip_sd = ((mean_slip-event1$minDisp)/2) # set sd for cumulative displacement as
# half the range

all_runs=data.frame() #create a blank data frame to store your outputs

i<-c(1:n)
  for (val in i){
    event_age1 <- rnorm(1, mean_age1, sd=age_sd1) # choose 1 random event age from a
    # normal distribution with mean and sd defined above

    event_age2 <- rnorm(1, mean_age2, sd=age_sd2) # choose 1 random event age from a
    # normal distribution with mean and sd defined above

    time_range=event_age2-event_age1
    if (time_range < 0){
      event_age1 <- rnorm(1, mean_age1, sd=age_sd1)
      event_age2 <- rnorm(1, mean_age2, sd=age_sd2)
      time_range=event_age2-event_age1

      if (time_range < 0){

```

```

    event_age1 <- rnorm(1, mean_age1, sd=age_sd1)
    event_age2 <- rnorm(1, mean_age2, sd=age_sd2)
    time_range=event_age2-event_age1
  }
}
else
  incremental_slip <- rnorm(1,mean_slip,sd=slip_sd) # choose 1 random slip mag from
# a normal distribution with mean and sd defined above
  if (incremental_slip<0){
    incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)
    if (incremental_slip<0){
      incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)
      if(incremental_slip<0){
        incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)
      }
    }
  }
}
else
  slip_rate=incremental_slip/time_range

#### save it all ####
modelrun <-data.frame(interval,event_age1,event_age2,incremental_slip,
  time_range,slip_rate)

all_runs <- rbind(all_runs,modelrun)
}

IntervalName <- all_runs #'[enter the name of the interval]

save(IntervalName,file="./IntervalName.Rda") #'[re-enter the name of the profile as
#'[the first arg, specify file location/name in second]

event_age1_m = mean(all_runs$event_age1)

```

```

event_age1_sd = sd(all_runs$event_age1)

event_age2_m = mean(all_runs$event_age2)
event_age2_sd = sd(all_runs$event_age2)

incremental_slip_m = mean(all_runs$incremental_slip)
incremental_slip_sd = sd(all_runs$incremental_slip)

time_range_m = mean(all_runs$time_range)
time_range_sd = sd(all_runs$time_range)

slip_rate_m = mean(all_runs$slip_rate)
slip_rate_sd = sd(all_runs$slip_rate)

starting_disp_m = 0
starting_disp_sd = 0

cum_slip_m = incremental_slip_m
cum_slip_sd = incremental_slip_sd

#create dataframe on first run
rates<-data.frame(event_age1_m,event_age1_sd,event_age2_m,event_age2_sd,incremental_slip_m,
                  incremental_slip_sd,time_range_m,time_range_sd,slip_rate_m,slip_rate_sd,
                  starting_disp_m,starting_disp_sd,cum_slip_m,cum_slip_sd)
row.names(rates)<-interval

#save data frame to be able to load later in 3D regression
save(rates,file="./GBrates.Rda") #'[give file a useful name]

##### 2nd interval #####
interval<-"interval_name2"
event1<-mydata[2,]
event2<-mydata[3,]

```

```

mean_age1 = ((event1$minAge+event1$maxAge)/2) # middle of age range
age_sd1 = (mean_age1-event1$minAge)/2 # set sd for cumulative displacement as
# half the range

mean_age2 = ((event2$minAge+event2$maxAge)/2) # middle of age range
age_sd2 = (mean_age2-event2$minAge)/2 # set sd for cumulative displacement as
# half the range

mean_slip = ((event1$minDisp+event1$maxDisp)/2) # middle of slip range
slip_sd = ((mean_slip-event1$minDisp)/2) # set sd for cumulative displacement as
# half the range

all_runs=data.frame() #create a blank data frame to store your outputs
i<-c(1:n)
for (val in i){
  event_age1 <- rnorm(1, mean_age1, sd=age_sd1) # choose 1 random event age from a
  # normal distribution with mean and sd defined above

  event_age2 <- rnorm(1, mean_age2, sd=age_sd2) # choose 1 random event age from a
  # normal distribution with mean and sd defined above

  incremental_slip <- rnorm(1,mean_slip,sd=slip_sd) # choose 1 random slip mag from
  # a normal distribution with mean and sd defined above
  if (incremental_slip<0){
    incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)
    if (incremental_slip<0){
      incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)
      if(incremental_slip<0){
        incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)
      }
    }
  }
}

```

```

else
time_range=event_age2-event_age1
if (time_range < 0){
  event_age1 <- rnorm(1, mean_age1, sd=age_sd1)
  event_age2 <- rnorm(1, mean_age2, sd=age_sd2)
  time_range=event_age2-event_age1

  if (time_range < 0){
    event_age1 <- rnorm(1, mean_age1, sd=age_sd1)
    event_age2 <- rnorm(1, mean_age2, sd=age_sd2)
    time_range=event_age2-event_age1
  }
}
else
slip_rate=incremental_slip/time_range

#### save it all ####
modelrun <-data.frame(interval,event_age1,event_age2,incremental_slip,time_range,slip_rate)

all_runs <- rbind(all_runs,modelrun)
}

IntervalName2 <- all_runs #'[enter the name of the interval]

save(IntervalName2,file="./IntervalName2.Rda") #'[re-enter the name of the profile as
#'[the first arg, specify file location/name in second]

event_age1_m = mean(all_runs$event_age1)
event_age1_sd = sd(all_runs$event_age1)

event_age2_m = mean(all_runs$event_age2)
event_age2_sd = sd(all_runs$event_age2)

```

```

incremental_slip_m = mean(all_runs$incremental_slip)
incremental_slip_sd = sd(all_runs$incremental_slip)

time_range_m = mean(all_runs$time_range)
time_range_sd = sd(all_runs$time_range)

slip_rate_m = mean(all_runs$slip_rate)
slip_rate_sd = sd(all_runs$slip_rate)

# save rates on subsequent runs
load("./GBrates.Rda")
starting_disp_m=rates$cum_slip_m
starting_disp_sd=rates$cum_slip_sd
cum_slip_m = incremental_slip_m + rates$cum_slip_m
cum_slip_sd = incremental_slip_sd + rates$cum_slip_sd

new_rates<-data.frame(event_age1_m,event_age1_sd,event_age2_m,event_age2_sd,incremental_slip_m,
                      incremental_slip_sd,time_range_m,time_range_sd,slip_rate_m,slip_rate_sd,
                      starting_disp_m,starting_disp_sd,cum_slip_m,cum_slip_sd)
row.names(new_rates)<-interval
rates<-rbind(rates,new_rates)

save(rates,file="./GBrates.Rda")

##### 3rd interval #####
interval<-"IntervalName3"
event1<-mydata[3,]
event2<-mydata[4,]

mean_age1 = ((event1$minAge+event1$maxAge)/2) # middle of age range
age_sd1 = (mean_age1-event1$minAge)/2 # set sd for cumulative displacement as
# half the range

```



```

mean_age2 = ((event2$minAge+event2$maxAge)/2) # middle of age range
age_sd2 = (mean_age2-event2$minAge)/2 # set sd for cumulative displacement as
# half the range

mean_slip = ((event1$minDisp+event1$maxDisp)/2) # middle of slip range
slip_sd = ((mean_slip-event1$minDisp)/2) # set sd for cumulative displacement as
# half the range

all_runs=data.frame() #create a blank data frame to store your outputs

i<-c(1:n)
for (val in i){
  event_age1 <- rnorm(1, mean_age1, sd=age_sd1) # choose 1 random event age from a
  # normal distribution with mean and sd defined above

  event_age2 <- rnorm(1, mean_age2, sd=age_sd2) # choose 1 random event age from a
  # normal distribution with mean and sd defined above
  incremental_slip <- rnorm(1,mean_slip,sd=slip_sd) # choose 1 random slip mag from
  # a normal distribution with mean and sd defined above
  if (incremental_slip<0){
    incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)
    if (incremental_slip<0){
      incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)
      if(incremental_slip<0){
        incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)
      }
    }
  }
  else
    time_range=event_age2-event_age1
  if (time_range < 0){
    event_age1 <- rnorm(1, mean_age1, sd=age_sd1)

```

```

event_age2 <- rnorm(1, mean_age2, sd=age_sd2)
time_range=event_age2-event_age1

if (time_range < 0){
  event_age1 <- rnorm(1, mean_age1, sd=age_sd1)
  event_age2 <- rnorm(1, mean_age2, sd=age_sd2)
  time_range=event_age2-event_age1
}
}
else
slip_rate=incremental_slip/time_range

#### save it all ####
modelrun <-data.frame(interval,event_age1,event_age2,incremental_slip,time_range,slip_rate)

all_runs <- rbind(all_runs,modelrun)
}

GBint3 <- all_runs #'[enter the name of the interval]

save(GBint3,file="./GBint3.Rda") #'[re-enter the name of the profile as
#'[the first arg, specify file location/name in second]

event_age1_m = mean(all_runs$event_age1)
event_age1_sd = sd(all_runs$event_age1)

event_age2_m = mean(all_runs$event_age2)
event_age2_sd = sd(all_runs$event_age2)

incremental_slip_m = mean(all_runs$incremental_slip)
incremental_slip_sd = sd(all_runs$incremental_slip)

time_range_m = mean(all_runs$time_range)

```

```

time_range_sd = sd(all_runs$time_range)

slip_rate_m = mean(all_runs$slip_rate)
slip_rate_sd = sd(all_runs$slip_rate)

# save rates on subsequent runs
load("./GBrates.Rda")
starting_disp_m=rates$cum_slip_m[2]
starting_disp_sd=rates$cum_slip_sd[2]
cum_slip_m = incremental_slip_m + starting_disp_m
cum_slip_sd = incremental_slip_sd + starting_disp_sd

new_rates<-data.frame(event_age1_m,event_age1_sd,event_age2_m,event_age2_sd,
                      incremental_slip_m,incremental_slip_sd,time_range_m,time_range_sd,
                      slip_rate_m,slip_rate_sd,starting_disp_m,starting_disp_sd,cum_slip_m,
                      cum_slip_sd)
row.names(new_rates)<-interval
rates<-rbind(rates,new_rates)

save(rates,file="./GBrates.Rda")

##### 4th interval #####
interval<-"BInterval4"
event1<-mydata[4,]
event2<-mydata[5,]

mean_age1 = ((event1$minAge+event1$maxAge)/2) # middle of age range
age_sd1 = (mean_age1-event1$minAge)/2 # set sd for cumulative displacement as half the range

mean_age2 = ((event2$minAge+event2$maxAge)/2) # middle of age range
age_sd2 = (mean_age2-event2$minAge)/2 # set sd for cumulative displacement as half the range

mean_slip = ((event1$minDisp+event1$maxDisp)/2) # middle of slip range

```

```

slip_sd = ((mean_slip-event1$minDisp)/2) # set sd for cumulative displacement as half the range

all_runs=data.frame() #create a blank data frame to store your outputs

i<-c(1:n)
for (val in i){
  event_age1 <- rnorm(1, mean_age1, sd=age_sd1) # choose 1 random event age from a
  # normal distribution with mean and sd defined above

  event_age2 <- rnorm(1, mean_age2, sd=age_sd2) # choose 1 random event age from a
  # normal distribution with mean and sd defined above
  incremental_slip <- rnorm(1,mean_slip,sd=slip_sd) # choose 1 random slip mag from
  # a normal distribution with mean and sd defined above
  if (incremental_slip<0){
    incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)
    if (incremental_slip<0){
      incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)
      if(incremental_slip<0){
        incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)
      }
    }
  }
  else
  time_range=event_age2-event_age1
  if (time_range < 0){
    event_age1 <- rnorm(1, mean_age1, sd=age_sd1)
    event_age2 <- rnorm(1, mean_age2, sd=age_sd2)
    time_range=event_age2-event_age1

    if (time_range < 0){
      event_age1 <- rnorm(1, mean_age1, sd=age_sd1)
      event_age2 <- rnorm(1, mean_age2, sd=age_sd2)
      time_range=event_age2-event_age1
    }
  }
}

```

```

    }
  }
  else
    slip_rate=incremental_slip/time_range

  #### save it all ####
  modelrun <-data.frame(interval,event_age1,event_age2,incremental_slip,time_range,slip_rate)

  all_runs <- rbind(all_runs,modelrun)
}

BLint4 <- all_runs #'[enter the name of the interval]

save(BLint4,file="./BLint4.Rda") #'[re-enter the name of the profile as
#'[the first arg, specify file location/name in second]

event_age1_m = mean(all_runs$event_age1)
event_age1_sd = sd(all_runs$event_age1)

event_age2_m = mean(all_runs$event_age2)
event_age2_sd = sd(all_runs$event_age2)

incremental_slip_m = mean(all_runs$incremental_slip)
incremental_slip_sd = sd(all_runs$incremental_slip)

time_range_m = mean(all_runs$time_range)
time_range_sd = sd(all_runs$time_range)

slip_rate_m = mean(all_runs$slip_rate)
slip_rate_sd = sd(all_runs$slip_rate)

# save rates on subsequent runs
load("./BLrates.Rda")

```

```

starting_disp_m=rates$cum_slip_m[3]
starting_disp_sd=rates$cum_slip_sd[3]
cum_slip_m = incremental_slip_m + starting_disp_m
cum_slip_sd = incremental_slip_sd + starting_disp_sd

new_rates<-data.frame(event_age1_m,event_age1_sd,event_age2_m,event_age2_sd,
                      incremental_slip_m,incremental_slip_sd,time_range_m,time_range_sd,
                      slip_rate_m,slip_rate_sd,starting_disp_m,starting_disp_sd,cum_slip_m,
                      cum_slip_sd)
row.names(new_rates)<-interval
rates<-rbind(rates,new_rates)

save(rates,file="./BLrates.Rda")

##### final event -----
interval<-"final-event"
event1<-mydata[4,]
mean_age1 = ((event1$minAge+event1$maxAge)/2) # middle of age range
age_sd1 = (mean_age1-event1$minAge)/2 # set sd for cumulative displacement as half the range

mean_slip = ((event1$minDisp+event1$maxDisp)/2) # middle of slip range
slip_sd = (mean_slip-event1$minDisp)/2 # set sd for cumulative displacement as half the range

all_runs=data.frame() #create a blank data frame to store your outputs

i<-c(1:n)
for (val in i){
  event_age1 <- rnorm(1, mean_age1, sd=age_sd2) # choose 1 random event age from a
  # normal distribution with mean and sd defined above
  incremental_slip <- rnorm(1,mean_slip,sd=slip_sd) # choose 1 random slip mag from
  # a normal distribution with mean and sd defined above
  if (incremental_slip<0){
    incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)

```

```

    if (incremental_slip<0){
      incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)
      if(incremental_slip<0){
        incremental_slip <- rnorm(1,mean_slip,sd=slip_sd)
      }
    }
  }
}

else
#### save it all ####
event_age2=NaN
time_range=NaN
slip_rate=NaN
modelrun <-data.frame(interval,event_age1,event_age2,incremental_slip,time_range,slip_rate)

all_runs <- rbind(all_runs,modelrun)
}

event_age1_m = mean(all_runs$event_age1)
event_age1_sd = sd(all_runs$event_age1)

event_age2_m = NaN
event_age2_sd = NaN

incremental_slip_m = mean(all_runs$incremental_slip)
incremental_slip_sd = sd(all_runs$incremental_slip)

time_range_m = NaN
time_range_sd = NaN

slip_rate_m = NaN
slip_rate_sd = NaN

load("./GBrates.Rda")

```



```

starting_disp_m=rates$cum_slip_m[3]
starting_disp_sd=rates$cum_slip_sd[3]
cum_slip_m = incremental_slip_m + starting_disp_m
cum_slip_sd = incremental_slip_sd + starting_disp_sd

new_rates<-data.frame(event_age1_m,event_age1_sd,event_age2_m,event_age2_sd,
                      incremental_slip_m,incremental_slip_sd,time_range_m,time_range_sd,
                      slip_rate_m,slip_rate_sd,starting_disp_m,starting_disp_sd,cum_slip_m,
                      cum_slip_sd)
row.names(new_rates)<-interval
rates<-rbind(rates,new_rates)
save(rates,file="./GBrates.Rda")

##### final step to save -----
# if you want to save your file to csv to open in another program
write.csv(rates,file="./nofz-gb-rates.csv")

```