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Cracking the Klamaths: Active Extension in the Southern Cascadia Forearc

By

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THESIS

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Masters of Science

in

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in the

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of the

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Davis

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2026

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Abstract

The Klamath Mountains province (KMP) of northern California and southern Oregon has traditionally been characterized as a rigid crustal block due to a lack of formally recognized active faults. However, the identification of active faults is hindered by the elevated, eroding, and densely forested character of the KMP, and assumptions of rigidity are inconsistent with the region's likely weak rheology and geodetic evidence of active transpressional deformation. I utilize recently released high-resolution lidar data to identify three active normal faults and one active dextral strike-slip fault in the KMP. I focus on the 47-km-long Scott Valley normal fault to provide the first geologic constraint on the rate of active deformation in the KMP. Fault scarp displacement analysis and Infrared Stimulated Luminescence dating indicate a minimum mean dip-slip rate of 0.7 ± 0.2 mm/yr since 21.2 ± 4.4 ka for the Scott Valley normal fault. I expect that ruptures on active normal faults in the KMP coincide with a stress field reversal induced by a CSZ megathrust event. Finally, I interpret north-south striking normal faults in the KMP as evidence of active east-west extensional deformation, which I attribute to southern Cascadia slab rollback and/or subcretion.

1. Introduction

The Klamath Mountains province (KMP) of northern California and southern Oregon forms a unique region with the highest average topography in the forearc of the Cascadia subduction zone (CSZ; Kelsey et al., 1994). Due to the northward migration of the Mendocino Triple Junction (MTJ), the KMP straddles the North American plate boundary transition from subduction along the CSZ to translation along the San Andreas fault system. South of the KMP, the San Andreas fault system drives northwest translation of the rigid Sierra Nevada-Great Valley (Sierran) block and strike-slip motion along the Walker Lane (Unruh et al., 2003; Hammond and Thatcher, 2005; Unruh and Humphrey, 2017). North of the KMP, the rigid Oregon Coast block (OCB) rotates clockwise to accommodate both northwest motion of the Sierran block and Basin and Range extension (Humphreys and Coblenz, 2007; McCaffrey et al., 2013). Given that the Sierran block and OCB move as rigid blocks in different directions at different rates, there must be deformation between them, across the KMP, to accommodate this differential motion (Unruh and Humphrey, 2017).

While geodetic models predict northwest-southeast contraction and/or dextral shear across the KMP (Hammond and Thatcher, 2005; Unruh and Humphrey, 2017), no active faults have been formally documented to accommodate this deformation, and the KMP has therefore been treated as rigid (McCaffrey et al., 2013; Unruh and Humphrey, 2017; McKenzie et al., 2022; Furlong et al., 2024). Yet the KMP should not behave rigidly; it is a collage of accreted terranes underplated by schist (Chapman et al., 2024; Figure 1), which should be rheologically weaker (Shinavar et al., 2015) than the strong, mafic crust of the Great Valley and OCB that bound it.

Although isolated reports have previously postulated the existence of active faults in the KMP (Mack, 1958; Yule, 1996), these structures have largely been overlooked in subsequent mapping and remain absent from formal databases (e.g., Petersen et al., 2023). In this study, I analyze recently released high-resolution lidar data to identify four active faults in this region. I combine fault scarp displacement estimates with Infrared Stimulated Luminescence (IRSL) geochronology to quantify the dip-slip rate and

hazard potential of the longest of these newly identified faults, the Scott Valley normal fault. These results offer the first geologic constraint on the rate and style of active deformation in the KMP that serve to test geodetic models of the plate-boundary transition in southernmost Cascadia.

2. Background

2.1 Geologic History of the Klamath Mountains Province

The KMP consists of a stack of east dipping and westward younging tectonostratigraphic terranes which correlate with those in the northern Sierra Nevada foothills and record sequential westward growth of the North American plate by several accretionary episodes from early Paleozoic through late Jurassic time (Snoke and Barnes, 2006; Figure 1). These metamorphosed terranes are separated by shallowly east-dipping faults and crosscut by predominantly calcic and calc-alkalic plutons ranging in age from Paleozoic to early Cretaceous (Snoke and Barnes, 2006; Piotraschke et al., 2015; Chapman et al., 2024).

The western margin of the KMP is marked by the eastward-dipping Coast Range fault, which places rocks of the KMP over Cretaceous and younger rocks of the Franciscan complex (Wagner and Saucedo, 1987; Snoke and Barnes, 2006). Tertiary and Quaternary volcanic rocks of the Cascade Range mark the eastern boundary of the KMP (Irwin, 1994; Snoke and Barnes, 2006). Rocks of the KMP may extend further east in the subsurface below these volcanics (Snoke and Barnes, 2006).

Terranes of the KMP can generally be subdivided by their age of accretion younging westward (Figure 1). From east to west, these include the Eastern Klamath, Central Metamorphic, Stuart Fork (also called the Fort Jones terrane), North Fork, Eastern Hayfork, Western Hayfork, Rattlesnake Creek, and Western Klamath terranes (Irwin, 1972; Wright, 1982; Wright and Fahan, 1988). The Eastern Klamath terrane is the oldest and most landward portion of the KMP upon which the remainder of the province was built (LaMaskin et al., 2021; Chapman et al., 2024). Neoproterozoic to Devonian basement rocks of the Eastern Klamath terrane are interpreted as the remnants of oceanic island arcs (LaMaskin et al., 2021; Chapman et al., 2024). In Silurian-Devonian time, oceanic assemblages of the Central Metamorphic terrane accreted to the Eastern Klamath terrane along an east-dipping subduction zone (Barrow and

Metcalf, 2006; LaMaskin et al., 2021; Chapman et al., 2024). After a hiatus, the Stuart Fork terrane accreted in Late Triassic time (LaMaskin et al., 2021; Chapman et al., 2024). The Stuart Fork terrane is generally interpreted as a subduction complex or accretionary prism and includes shale, chert, and volcanic rocks metamorphosed to blueschist facies (LaMaskin et al., 2021).

The KMP grew significantly during the Middle Jurassic Siskiyou orogeny, a contractional event which involved rapid sequential accretion of terranes (Wright and Fahan, 1988; Snoke and Barnes, 2006). Accretion of oceanic rocks of the North Fork and Eastern Hayfork terranes was followed by accretion of the Western Hayfork terrane, a continent-fringing oceanic arc, and its dismembered ophiolitic basement, the Rattlesnake Creek terrane (Wright and Fahan, 1988; Snoke and Barnes, 2006; Chapman et al., 2024; Figure 1). Following the Siskiyou orogeny, a Middle to Late Jurassic episode of in-situ extension led to the formation of basins floored by Josephine Ophiolite and blanketed with sediments of the Galice Formation. (Harper, 2003; LaMaskin et al., 2021; Chapman et al., 2024). Extension concluded with accretion of the Western Klamath terrane and the resumption of regional contraction associated with the Nevadan orogeny (LaMaskin et al., 2021; Chapman et al., 2024).

The Condrey Mountain Schist (CMS), exposed as a domal structural window through the overlying Rattlesnake Creek terrane, stands out as an unusual feature of the KMP (Chapman et al., 2024; Figure 1). The CMS resides at a lower metamorphic grade and shows younger cooling ages than adjacent rocks, making correlations with adjacent terranes untenable (Chapman et al., 2024). Chapman et al. (2024) propose that the modern far inboard position of the CMS can be attributed to ca. 135 Ma shallow-angle subduction which drove low angle underthrusting of the CMS beneath the KMP and removed the lower crust via tectonic erosion.

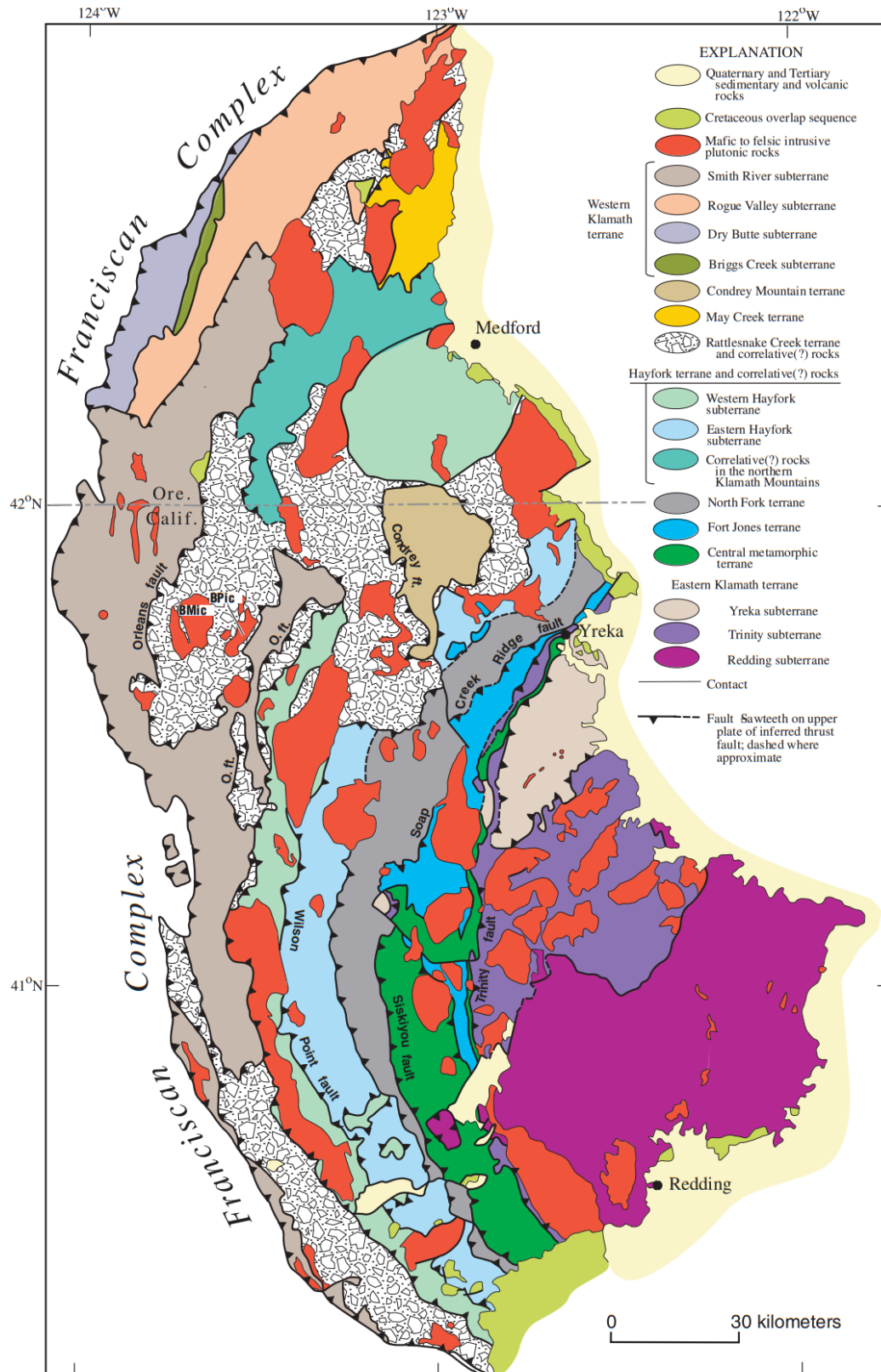


Figure 1. Simplified geologic map of the tectonostratigraphic terranes of the Klamath Mountains province (from Snoke and Barnes, 2006).

In early Cretaceous time, the KMP relocated ~200 km westward relative to correlative units in the Sierra Nevada, magmatism in the region abruptly ceased, and topography developed during the Nevadan orogeny was lost (Jones and Irwin, 1971; Batt et al., 2010a; Chapman et al., 2024). The KMP was subsequently blanketed by Great Valley Sequence and Hornbrook Formation deposits which record marine transgression from 140-90 Ma, with sedimentary thicknesses reaching 5 km in the northern KMP and 3.5-4 km in the southern KMP (Jones and Irwin, 1971; Batt et al., 2010a; Piotraschke, 2012; Michalak et al., 2024). This sedimentary cover was largely removed by tectonic modification and erosion from late Early Cretaceous to Paleogene time (Batt et al., 2010a; Piotraschke et al., 2015).

The southern KMP in Neogene time is characterized by moderate topographic relief and the formation of isolated grabens formed by extensional faulting and filled with terrestrial sediments of the Miocene Weaverville Formation (Michalak et al., 2024; Figure 2). Michalak et al. (2024) propose that extensional faulting at this time was driven by underplating of material from the downgoing slab to the overriding crust in a process referred to as subcretion (Delph et al., 2021).

The largest of these Miocene grabens, the Weaverville graben, is bounded by the shallowly south-dipping (5° or less) La Grange detachment fault, recognized by Irwin as one of the few Cenozoic-active faults in the KMP (1960; 1963; 1966; Schweickert and Irwin, 1989; Piotraschke et al., 2015; Figure 2). The La Grange detachment defines a prominent mountain front along the southern margin of the Trinity Alps, a region of locally high topography and relief. Piotraschke et al. (2015) estimate the La Grange detachment was active from ~45 Ma to ~15 Ma during which time it recorded ~60 km of north-south displacement at a rate of ~2 mm/yr.

Two smaller Miocene extensional basins occur within the southwestern KMP (Figure 2). The Hayfork graben is bounded on its north and south by normal faults and by east-side-up and right-lateral faulting on its eastern end (Michalak et al., 2024). The Hayfork graben is estimated to contain between 1 to 1.9 km of sedimentary fill (Michalak et al., 2024). An additional basin, the Hyampom basin, is estimated to contain between 120-180 m of Weaverville Formation sediments (Michalak et al., 2024). While a fault bounding the northeastern margin of the Hyampom basin appears on select geologic maps

(e.g., Strand, 1963; Jennings et al., 2010), its slip history and kinematics are poorly constrained. Notably, this mapped fault trace coincides with an active dextral strike-slip fault identified in this study (Appendix C).

Extension in the KMP concluded in the late Miocene and was followed by NW-striking dextral strike-slip and reverse faults active from the late Miocene to Pliocene (Michalak et al., 2024; Figure 2). Michalak et al. (2024) suggest these faults, which include the Browns Meadow fault and the Democrat Gulch fault, may have formed in response to northwestward motion of the Sierran block, which began influencing the KMP around this time.

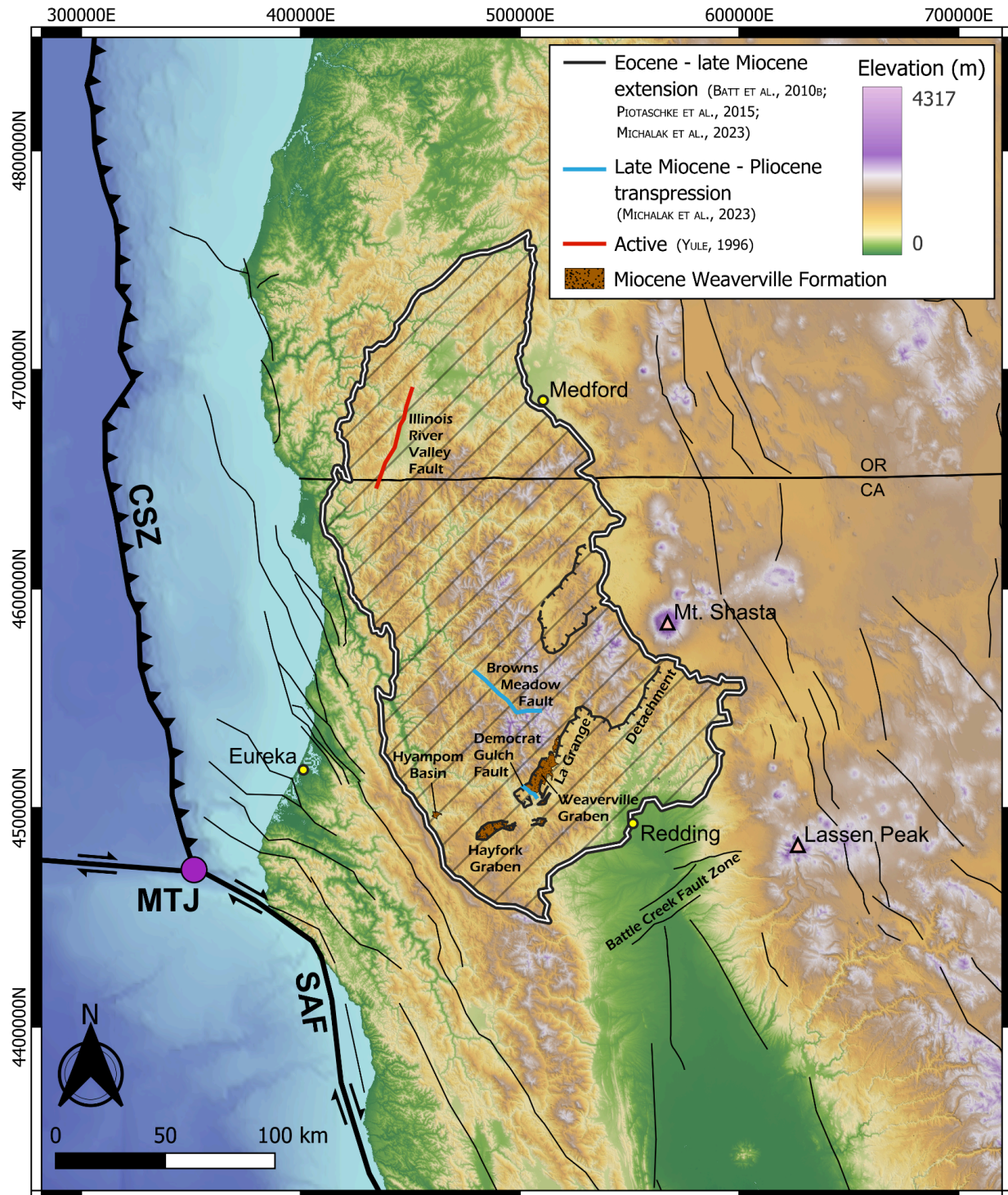


Figure 2. Cenozoic-active faults identified within the Klamath Mountains province (KMP) prior to this study overlain on modern hillshaded topography. KMP outlined in black with hashed interior. Contemporary-active features plotted around the KMP: thin black lines denote active faults from the 2023 USGS National Seismic Hazard Model (Petersen et al., 2023); thick black lines denote plate boundaries: SAF: San Andreas Fault, CSZ: Cascadia subduction zone.

2.2 Modern Tectonic Setting of the Klamath Mountains Province

Large-scale deformation in the broader region of western North America is driven by plate-boundary forces that translate and rotate microplates (Figure 3). The KMP is situated onshore of the leading edge of the northward migrating MTJ, which marks the replacement point of the Cascadia subduction zone with transform motion of the San Andreas fault. South of the KMP, the Sierran block, which comprises the Sierra Nevada and adjacent Great Valley, moves ~11-13 mm/yr northwest relative to stable North America, impinging upon the southern margin of the KMP (Unruh and Humphrey, 2017; Figure 3). North of the KMP, the rigid Oregon Coast block (OCB) rotates clockwise about an Euler pole in western Idaho to accommodate Sierran block motion and extension of the northern Basin and Range province (Humphreys and Coblenz, 2007; McCaffrey et al., 2013; Figure 3).

Given that the Sierran block and OCB move as rigid blocks that translate in different directions at different rates across their shared boundary, there must be deformation between them, across the KMP, to accommodate this differential motion (Unruh and Humphrey, 2017; Figure 3). McCaffrey (personal communication cited within Unruh and Humphrey, 2017) predicts the northern Sacramento Valley moves ~5 mm/yr west-northwest relative to the KMP in southern Oregon. Hammond and Thatcher (2005) find up to 4 mm/yr of north-south contraction near the Oregon-California border. Additionally, McKenzie et al. (2022) find a decay in the NNW-directed velocity field from ~12 mm/yr to 8 mm/yr across the KMP. Suggested mechanisms of this deformation include west-northwest-directed dextral shear (Unruh and Humphrey, 2017) or dextral transpression (Angster et al., 2021). Humphreys and Coblenz (2007) propose that east-west extension is necessary in this region to create space for Basin and Range dilation and Sierran block translation, and offer slab rollback as the driver of this extension. Because deformation in the KMP is directly related to the plate boundary transition at the MTJ, quantifying the rate and style of deformation in this region is key to developing geodynamic models which constrain absolute stress on faults (Humphreys and Coblenz, 2007). Due to the strong imprint of the locked subduction interface on the geodetic strain field (McCaffrey et al., 2013; McKenzie and Furlong, 2021; McKenzie et al., 2022),

sorting out the style and rate of permanent deformation here requires a geologic investigation of active faulting.

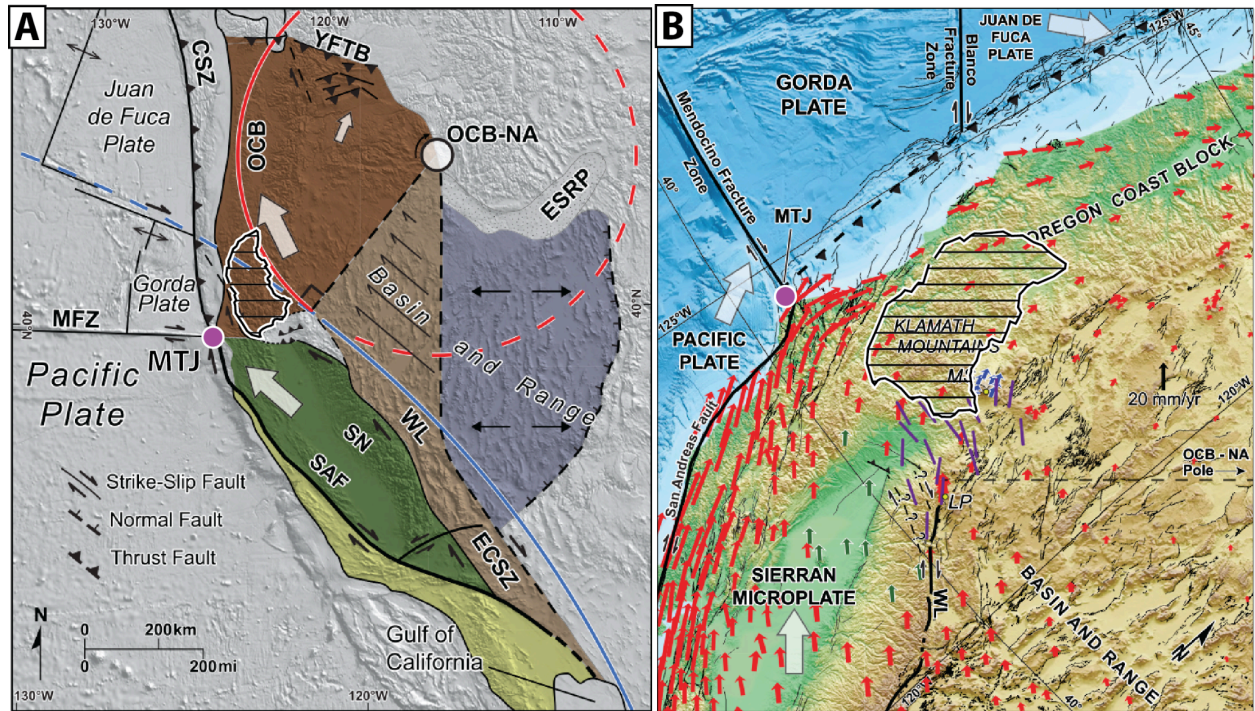


Figure 3. Tectonic setting of the MTJ region in northern California and southern Oregon. Figures modified from Unruh and Humphrey (2017). Klamath Mountains province (KMP) outlined in black with horizontal rule. **(A)** Active tectonic features of the broader Pacific-North America-Juan de Fuca plate boundary system. The Sierran block (SN) rotates counterclockwise toward the MTJ about an SN-NA Euler pole offshore to the SE; trajectory denoted by blue line. The Oregon Coast block (OCB) rotates clockwise about an Euler pole in western Idaho; trajectory denoted by red line. Walker Lane-Eastern California Shear Zone (WL-ECSZ) shear deformation diminishes northward as it is absorbed by Basin and Range extension. The remainder of this shear deformation is shunted westward toward the MTJ at the SN-OCB boundary, traversing the KMP. **(B)** Oblique Mercator projection about SN-NA Euler Pole shows northwestward-directed motion of the SN toward the MTJ. Purple lines show direction of shear displacement from inversion of microseismicity. GPS vectors (red arrows) plotted with respect to stable North America turn strongly eastward north of the MTJ, reflecting the shift from transform motion to oblique subduction. Note diminishing vector length across the KMP suggesting northwest-southeast contraction here.

Despite the presence of several mm/yr of deformation across the KMP, the absence of formally recognized active faults has led to assumptions that the region behaves as a rigid block (e.g., McKenzie et

al., 2022; Furlong et al., 2024) or as part of the OCB (McCaffrey et al., 2013; Unruh and Humphrey, 2017; Figure 3). Yet the crustal structure of the KMP indicates that it should not be rigid; both the western Great Valley and Siletz terrane, which underlie the Sierran block and OCB respectively, consist of mafic crust (Holbrook and Mooney, 1987; Tréhu et al., 1994), the lower crust of which should be highly viscous and resistant to deformation (Shinavar et al., 2015). In contrast, the KMP is composed of a collage of accreted terranes underplated by schist of predominantly felsic composition (Chapman et al., 2024). This structure, characterized by weaker, lower-viscosity intermediate composition middle crust (Shinavar et al., 2015) and numerous terrane-bounding shear zones, should make the KMP more susceptible to internal deformation than the stronger material underlying the Sierran block and OCB.

At the base of the KMP, Delph et al. (2021) describe a ~10-15 km-thick zone of low density underplated metasedimentary material above the subduction interface, interpreted to have been emplaced by subcretion (accretion of material transported along the subduction interface). They propose that this buoyant material generates domal uplift and exhumation of the overriding crust, which they offer as a driver for the elevated topography, high uplift rates, and high erosion rates in the KMP (Delph et al., 2021). The Olympic Mountains, another region of elevated topography and high uplift rates in the Cascadia forearc, similarly show evidence of subcretion (Delph et al., 2021). Alternatively, McKenzie et al. (2022) propose that the steep channels and high relief observed in the KMP are a result of uplift due to intracrustal injection of Franciscan crust beneath the KMP.

Numerous studies have documented active faulting around the margins of the KMP (Figure 2). South of the KMP, Angster et al. (2021) estimate a north-south shortening rate of 1.3 ± 0.4 mm/year accommodated in the northern Sacramento Valley along structures of the Battle Creek fault zone and suggest that further shortening structures must exist to the north to accommodate the far-field contraction observed between northern California and Oregon. Adjacent to the western boundary of the KMP, Kelsey and Carver (1988) describe a narrow zone of NNW-trending active strike-slip faults that have accommodated several kilometers of right-slip since ~0.7 Ma. These faults spatially coincide with a zone of NNW-oriented right-lateral shear between the KMP and coastal Franciscan crust observed by

McKenzie et al. (2022). Situated immediately west of these dextral strike-slip faults lies a zone of subaerial forearc contraction accommodating at least 6.5 mm/yr of NE-SW shortening along a series of thrust faults, which Kelsey and Carver (1988) attribute to coupling of the internally contracting Gorda slab with the overlying North American plate.

Faults which have accumulated substantial offset often form prominent tectonic-geomorphic features such as a high relief mountain front and/or a low-lying sedimentary basin (Burbank and Anderson, 2001). The correlation of faults with these landforms makes them primary targets for active fault mapping. Within the KMP, several prominent basins have been formed by faults. The Weaverville and Hayfork grabens in the southern KMP were formed by Miocene extensional faulting (Cashman and Elder, 2002; Batt et al., 2010b; Piotraschke et al., 2015; Michalak et al., 2024; Figure 2). In the northern KMP, the Illinois River Valley is situated adjacent to a likely active fault recognized by Yule (1996; Figure 2; Appendix D). Similarly, Scott Valley is another topographic low in the KMP bounded by an active fault. First recognized by Mack (1958) as a Quaternary-active normal fault, then later overlooked or reinterpreted, this fault serves as the central focus of this study.

2.3 Scott Valley

2.3.1 Physiography of Scott Valley

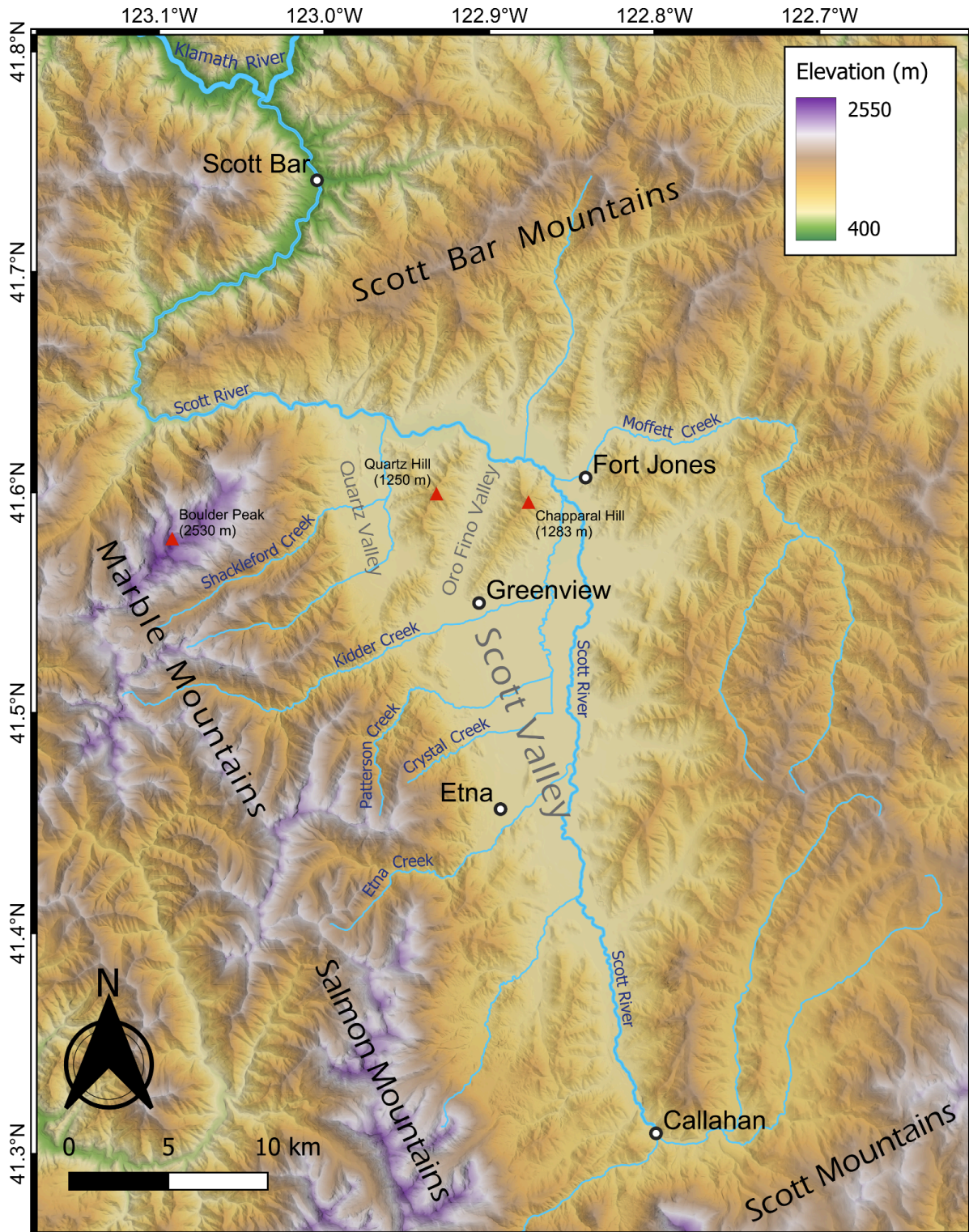


Figure 4. Hillshaded topography of Scott Valley with relevant geographic features labeled.

Scott Valley is an intramontane basin in the KMP approximately 50 km south of the Oregon border and 20 km southwest of Interstate 5 near Yreka (Figure 4). The valley is approximately 40 km long in a north-south direction and 16 km wide at its greatest east-west extent (Figure 4). The valley is widest in the north and narrows southward. It is drained by the Scott River, which flows from south to north through the eastern part of the valley before changing course near Fort Jones to flow from east to west across its northern flank to exit the valley at its northwest corner (Figure 4). Along the downstream course of the Scott River, the valley floor slopes from an elevation of 950 meters near Callahan to 800 meters where it exits the valley (Figure 5). Upon exiting the valley, the gradient of the Scott River dramatically steepens as it enters the Scott River canyon (Figure 5). Scott Valley is ringed by the Scott Bar, Marble, Salmon, and Scott Mountains, which bound its north, west, southwest, and southeast sides, respectively (Figure 4). The mountains to the south and the west of the valley are steeper and reach higher elevations (up to 2530 meters) than those to its east and north (1400-1850 meters; Siskiyou County Flood Control and Water District [SCFCWD], 2021; Figure 4).

The mountains bounding Scott Valley on its western margin are drained by perennial streams while those draining the eastern foothills are ephemeral (SCFCWD, 2021). An exception is Moffett Creek, a perennial stream which drains a large catchment area northeast of the valley (SCFCWD, 2021; Figure 4). The largest of these western tributaries have formed broad alluvial fans which extend nearly to the Scott River (Figure 6) and upon which the channels are braided or anastomosing (SCFCWD, 2021). In the wide central part of the valley, a meandering stream, which receives drainage from western tributaries, occupies a lowland area between the Scott River and the western mountain alluvial fans (Figure 6). During flood stages, the Scott River has apparently built up natural levees along its banks which prevent the western tributaries from joining the river via the shortest distance. Consequently, the western tributaries coalesce and flow parallel to the Scott River in a south-to-north direction for ~10 km before joining the Scott River near Fort Jones (Mack, 1958; Figure 6).

The alluvial valley floor, which encompasses approximately 269 km², is disrupted by 52 distinct bedrock hills. Most notable of these are Chaparral and Quartz Hills, which each rise ~400 meters above

their flat surroundings and delineate the boundaries of Oro Fino Valley and Quartz Valley, two subvalleys within the Scott River basin drained by north-flowing tributaries (Figures 4 & 6).

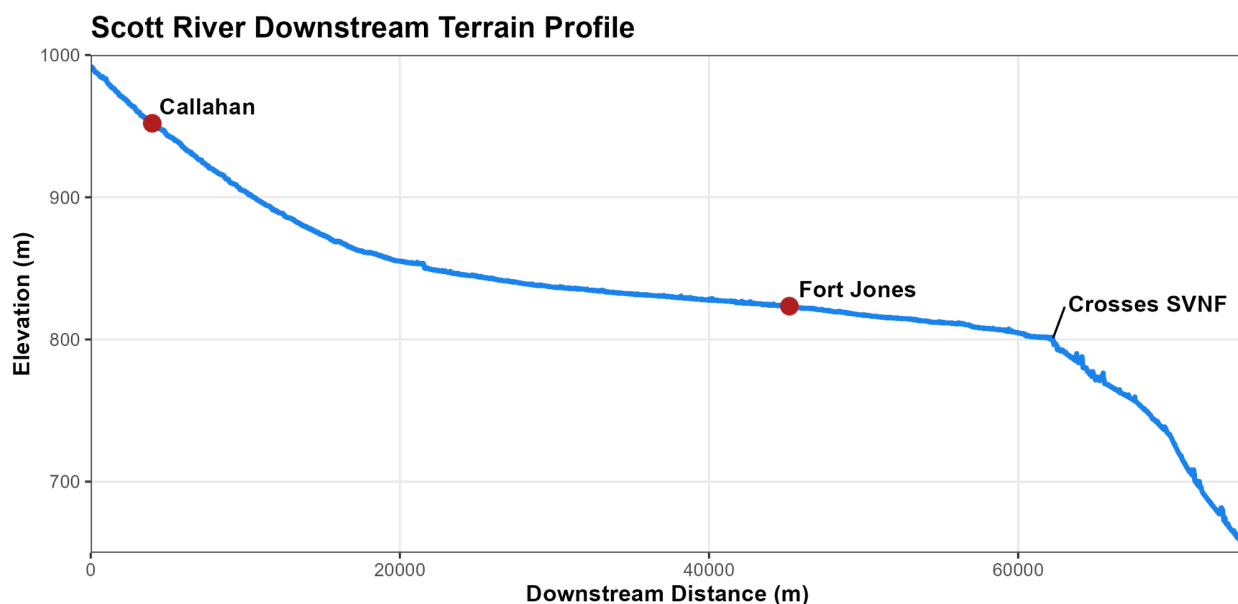


Figure 5. Downstream terrain profile of the Scott River through Scott Valley. Note shallowing gradient upon entering the valley and steepening gradient upon exiting the valley.

2.3.2 Geologic History of Scott Valley

Much of the prior work on the Quaternary geology and physiography of Scott Valley was conducted by Mack (1958), who recognized that the mountain front along the west side of Scott Valley is marked by an east-dipping and generally north-striking normal fault. Mack (1958) estimated the dip of this fault between 60° and 70° from measurements taken on a slickensided granodioritic outcrop near the outlet of Crystal Creek (Figure 4). Mack (1958) interprets that the trace of the fault south of Crystal Creek trends away from the mountain front before terminating near the town of Etna. This western fault was later reinterpreted as a reverse fault on more recent geologic maps of the region (e.g., Strand, 1963; Wagner and Saucedo, 1987).

2.3.3 Physiographic Development of Scott Valley

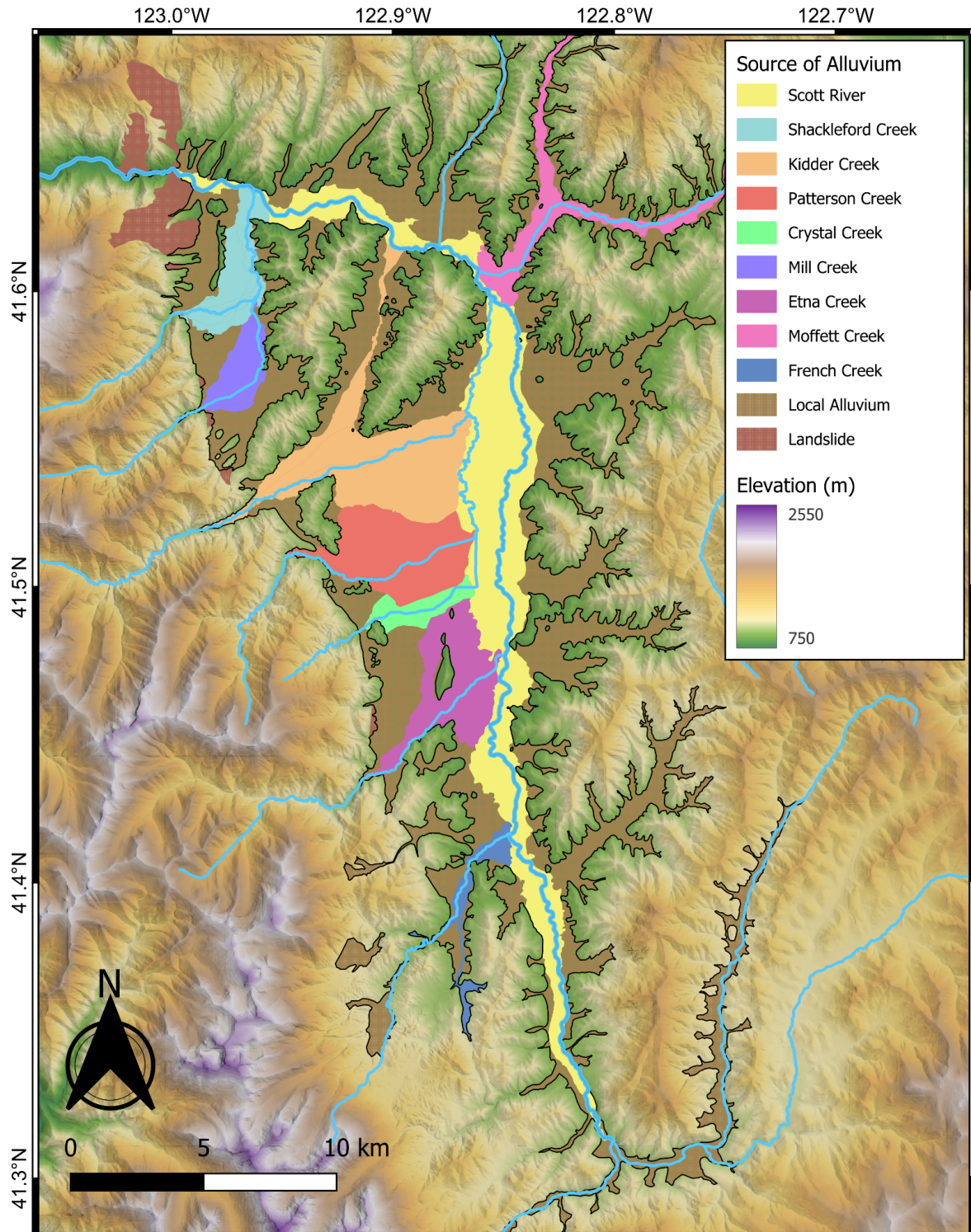


Figure 6. Hillshaded topography of Scott Valley with alluvium color-coded by source. Deposits not sourced from the Scott River or its major tributaries noted as Local Alluvium. Note relatively larger alluvial fans built by the western mountain tributaries.

Mack (1958) proposes that the course of the Scott River was initially established by the north-south orientation of relatively softer and more readily eroded sheared serpentine upstream of Fort Jones; and by a series east-west oriented fault and fracture systems downstream of Fort Jones.

For much of their early history, the Scott River and its tributaries were actively degrading streams which incised the landscape and established high-relief drainage divides. A subsequent shift to a depositional phase caused widespread aggradation across the valley, during which time a thick sequence of alluvium buried much of this ancestral topography. Mack (1958) interprets the isolated bedrock hills which currently protrude above the valley floor as relict high points of these buried drainage divides. The thickness of alluvial deposits in Scott Valley is not precisely known, though Mack (1958) suggests the alluvial sequence is thicker than 120 meters in the central part of the valley and around 60 meters at Fort Jones. Due to this spatial variation in alluvial thicknesses, which decrease upstream along the Scott River, Mack (1958) suggests that fault displacement is greatest in the central part of the valley between Etna and Greenview (Figure 4).

Mack (1958) proposes that several major tributaries of the Scott River have changed their course due to alluvial infilling of former mountainous topography. Etna Creek likely flowed several miles north of its present mouth before entering the valley, while Kidder Creek probably continued to flow to the northeast through Oro Fino Valley before joining the Scott River (Figure 4).

2.4 Fault Mapping with Lidar

The identification of displaced landforms is the basis for mapping active faults, and therefore, the location of potential future earthquakes. Faulted landforms, such as a fault scarp on a recent alluvial deposit or aligned offset stream channels, provide robust evidence of active movement along a fault. The recognition of faulted landforms is also important in determining fault length, an essential parameter used to estimate earthquake size and associated seismic hazard. Due to the subtlety of these features, which are often expressed by a few meters of offset, their identification requires the examination of high-resolution topography (Lin et al., 2013).

Airborne light detection and ranging (lidar) is conducted with a scanning laser rangefinder mounted in an aircraft (Haugerud et al., 2003). The rangefinder determines the distance to the ground surface by timing the round-trip travel time of a laser pulse. Post-processing of data can separate laser beams reflected off of the ground surface from those reflected off of vegetation, and enables detailed ground-surface topography even in densely vegetated regions.

The deployment of airborne lidar surveying in recent decades has provided the necessary high-resolution topography data to remotely identify and map active faults. Geomorphic mapping using lidar data was pioneered in the Puget Lowland region of Washington (Haugerud et al., 2003) and first applied in California along the San Andreas fault by Prentice et al. (2003). The recent release of lidar coverage of northern California and southern Oregon grants the opportunity to identify and map active faults in the KMP.

3. Methods

3.1 Lidar Processing and Analysis

This study has relied on high-resolution lidar data accessed online from recently released and pending datasets. Available lidar surveys are processed in QGIS to derive red relief image maps (RRIMs), which consist of a blend of slope shade and differential openness rasters (Chiba et al., 2008; Daxer, 2020). Differential openness is a metric which highlights prominent convexities and concavities in the landscape, including fault scarps (Yokoyama et al., 2002; Chiba et al., 2008). I utilize this derived map product to map active faults and surficial alluvial deposits.

I identify the location and style of active faults from their geomorphic surface expression. I identify dip-slip faults where geomorphic markers, such as alluvial fans or fluvial terraces, have been vertically displaced. I identify horizontal or oblique fault slip where geomorphic markers, such as a stream channel, have been deflected or beheaded by lateral displacement. To constrain fault dip direction, I observe the deflection of the fault trace as it crosses variable terrain.

The purpose of surficial mapping in this study is to map the extent of alluvial and fluvial deposits and to categorize these deposits by relative depositional age. I perform surficial mapping primarily through the remote analysis of RRIMs and elevation contours. To develop a relative age hierarchy of mapped surficial deposits, I examine surface roughness, contour directions, elevation above the mainstem channel, and edge sharpness. Where possible, I conduct field work to verify surficial maps created remotely.

3.2 Field Work

I undertake field investigations to verify remote findings, map and characterize Quaternary deposits adjacent to active faults, and identify and prepare sites for sampling for IRSL geochronology. In the field, I use TouchGIS on an iPad to map and record observations. This tool enables portable access to lidar data, elevation contours, GPS location, and offline notes. I use a Brunton compass and protractor to measure fault strike, dip, and rake.

3.3 IRSL Geochronology Sampling and Laboratory Procedures

The following summary is adapted largely from Murray and Olley (2002), Duller (2008), Wintle (2008), and Rhodes (2011). Infrared Stimulated Luminescence (IRSL) geochronology is a technique that estimates the time since mineral grains were last exposed to sunlight (the burial age). This technique relies on the accumulation of electrons in crystal lattice defects (traps) resulting from exposure to ambient ionizing radiation from the decay of uranium, thorium, and potassium-40, as well as cosmic rays. During sediment transport, exposure to sunlight releases the trapped electrons and resets the geochronometric clock to zero. Upon burial and shielding from light, the luminescence signal re-accumulates at a rate proportional to the environmental dose rate. In the laboratory, the burial age of a sample is estimated as the ratio of the energy accumulated during burial and the energy per year delivered by radioactive decay:

$$Age (yr) = Equivalent\ dose\ (Gy) / Dose\ rate\ (Gy/year)$$

When dosed minerals are then re-exposed to light or heat, they release the stored electrons, emitting a photon of light that is referred to as luminescence. Through controlled experiments using

special analytical equipment such as a Risø TL/OSL Reader, the emission of luminescence can be controlled and measured. This set of measurements is used to estimate the equivalent dose (D_e). The dose rate is the amount of energy absorbed per year from radiation in the environment surrounding the sample material. It can be estimated by measuring the amount of radioactivity directly or by chemically analyzing the surrounding material and calculating the concentration of radioisotopes.

I identify candidate sites for IRSL dating where well-preserved terrace surfaces of the same apparent depositional age have been displaced by faulting. I excavate holes to their target depths, at which time I clean and smooth the vertical sections with a putty knife. I describe and identify soil horizons based on variations in grain size, soil structure, and soil color using Munsell Color Charts. Samples for IRSL geochronology are collected from the walls of the pits and shielded from light. Samples collected in this study were processed and analyzed at the Desert Research Institute Luminescence Laboratory in Reno, Nevada by project collaborator Dr. Kathleen Rodrigues.

Detailed laboratory procedures are documented in Appendix E. All laboratory procedures were conducted under dim red light. Target K-feldspar grains were isolated using standard treatments to remove clays, carbonates, and organics, followed by wet sieving, magnetic separation, and heavy liquid density separation.

All samples were measured on a Lexsyg Research Instrument using continuous wave measurements of the luminescence signal. IRSL measurements were conducted using infrared (IR) laser diodes with peak emission at 850 nm and power of 145 mW/cm² at the sample position. Signals were detected through a filter with a 365 nm detection window using a H7360-02 (UV/VIS) photomultiplier tube. Artificial irradiation was administered using a calibrated ⁹⁰Sr beta source integrated into the Lexsyg Research Instrument.

Equivalent doses (D_e) were determined for 30 multi-grain aliquots per sample using a modified single aliquot regenerative-dose (SAR) technique (Murray and Wintle, 2000; 2003) for feldspar (Wallinga et al., 2000). Dose response curves were modeled using an exponential fit and aliquots which failed routine screening criteria were rejected.

Abanico plots were used to examine the D_e distribution for evidence of partial bleaching or post-depositional mixing and to determine the most appropriate age model to use for calculation (Figure E1). Final D_e values were calculated using the Central Age Model and Minimum Age Model (Galbraith et al., 1999). Fading rates for the IR_{50} signal were measured over the course of ~24 hours on 12 aliquots per sample using the approach of Auclair et al. (2003). Final ages were corrected utilizing a weighted mean fading rate for all samples ($2.96 \pm 0.29\%$ /decade; $t_c = 2$ days) and the correction model of Huntley and Lamothe (2001).

Samples for dose rate measurement were sent to ALS Geochemistry for measurement of U, Th (with ICP-MS), and K_2O (with ICP-AES). Dose rates (Gy/ka) were calculated using the conversion factors of Liritzis et al. (2013). Dose rate calculations assume radioactive equilibrium in the ^{238}U and ^{232}Th decay chains. Alpha dose rates assume an alpha efficiency value of 0.15 ± 0.05 based on the work of Balescu and Lamothe (1994). Beta dose attenuation factors were based on Guérin et al. (2012). Gamma dose rates were measured in-situ with a Canberra Inspector 1000 spectrometer. All sample dose rates were adjusted for estimated moisture content using the average laboratory measured gravimetric moisture content of all samples ($5\% \pm 5\%$). Cosmic dose rates (Gy/ka) were calculated according to Prescott and Hutton (1994). Dose rates and final age calculations were made using dose rate and age calculator (DRAC; Durcan et al., 2015) and include an internal dose rate component based on an internal K content of $10 \pm 2\%$ (Smedley et al., 2012). Ages are reported in thousands of years before 2025 and rounded to the nearest 10 years.

3.4 Estimating Fault Scarp Displacement

I derive fault scarp displacement from processed lidar-DEM ground surface elevations. I subset this data to define the location of the hangingwall, footwall, and fault scarp, and apply a linear regression to the surfaces of the footwall and hangingwall to model their pre-faulting surface geometries. To quantify fault throw, I subtract the hangingwall regression from the footwall regression at a randomly selected horizontal position along the fault scarp. I calculate dip-slip displacement from fault throw and dip, which

I randomly select from between 50° and 70° for normal faults. This process is simulated 10,000 times to generate a distribution from which I report a mean and 95% confidence interval.

3.5 Constructing the Subsurface Architecture of Scott Valley

I construct the subsurface basin architecture of Scott Valley by identifying the depth-to-bedrock and spatial location of 1591 wells drilled in Scott Valley. These wells are documented in the California Department of Water Resources Online System for Well Completion Reports (California Department of Water Resources, 2025). I verify well locations remotely by cross-referencing Siskiyou County parcel and address data (Siskiyou County GIS, 2024) with satellite imagery (Google, 2024). Wells which do not reach bedrock provide a minimum depth-to-bedrock constraint. I plot processed well data alongside its corresponding bedrock depth or minimum bedrock depth. From these plotted data, I draw depth-to-bedrock contours at 20 meter intervals. In areas where no wells reach bedrock, I extrapolate contours according to nearby bedrock surface gradients.

4. Results

4.1 Active Faults in the Klamath Mountains Province

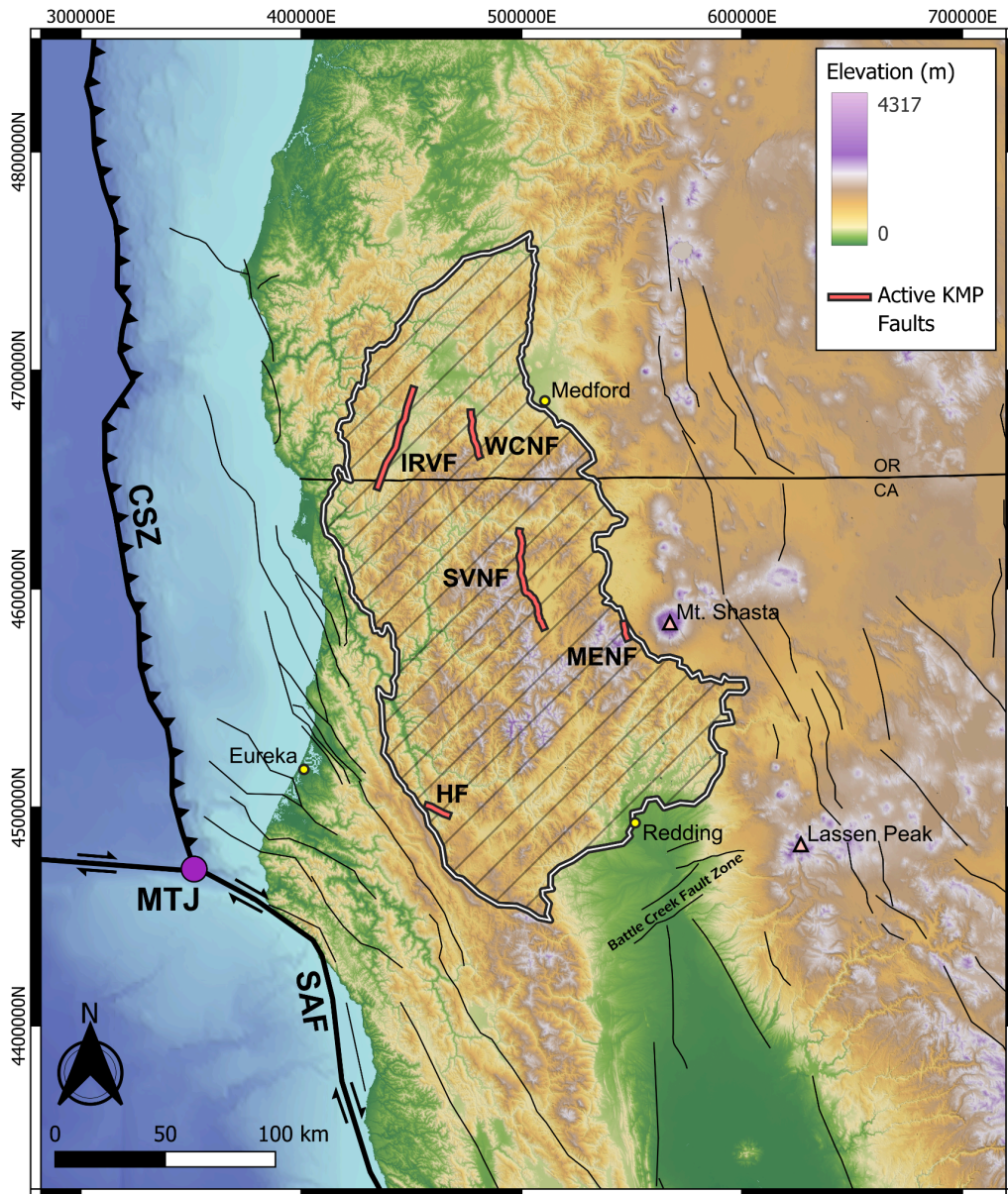


Figure 7. Active and potentially active faults in and around the Klamath Mountains province (KMP) overlain on modern hillshaded topography. KMP outlined in black with hashed interior. Thin black lines denote active faults from the 2023 USGS National Seismic Hazard Model (Petersen et al., 2023). Thick black lines denote plate boundaries: SAF: San Andreas fault; CSZ: Cascadia subduction zone. Red lines denote active faults identified in the KMP: WCNF: Williams Creek normal fault (Appendix A); IRVF: Illinois River Valley fault (Yule, 1996; Kirby et al., 2023; Appendix D); SVNF: Scott Valley normal fault; MENF: Mount Eddy normal fault (Appendix B); HF: Hyampom fault (Appendix C).

Analysis of high-resolution lidar data reveals four active fault zones in the KMP (Figure 7). These include one dextral strike-slip fault, the Hyampom fault (Appendix C), and three normal faults: the Williams Creek normal fault (Appendix A), the Mount Eddy normal fault (Appendix B), and the Scott Valley normal fault. An additional fault, the Illinois River Valley fault (Yule, 1996), appears to be a dip-slip fault, though its sense of slip is poorly constrained and lidar data do not show evidence for recent scarp-forming events to confirm activity (Appendix D). I thus consider this fault to be potentially active (Appendix D).

4.2 Scott Valley Normal Fault

4.2.1 Quaternary Geology of Scott Valley

The longest and most geomorphically prominent fault I identify in the KMP is the Scott Valley normal fault (SVNF), an east-dipping and north-striking normal fault marking the western edge of Scott Valley along much of its length (Figure 8). Based on high-resolution lidar mapping, the main strand of the SVNF is well-constrained over a strike length of 47 km (Figure 8). The fault stretches from a point approximately 7 km south of the town of Etna to a point approximately 4.5 km north of the town of Scott Bar (Figure 8). The fault may extend a further 16 km southward for a total fault length of 63 km along the western mountain front of Scott Valley; however, this stretch contains minimal geomorphic evidence for recent fault activity (Figure 8). Field measurements taken from a slickensided granodioritic outcrop adjacent to Crystal Creek show that the SVNF is a purely dip-slip fault with an eastward dip of 60°, validating earlier measurements from Mack (1958).

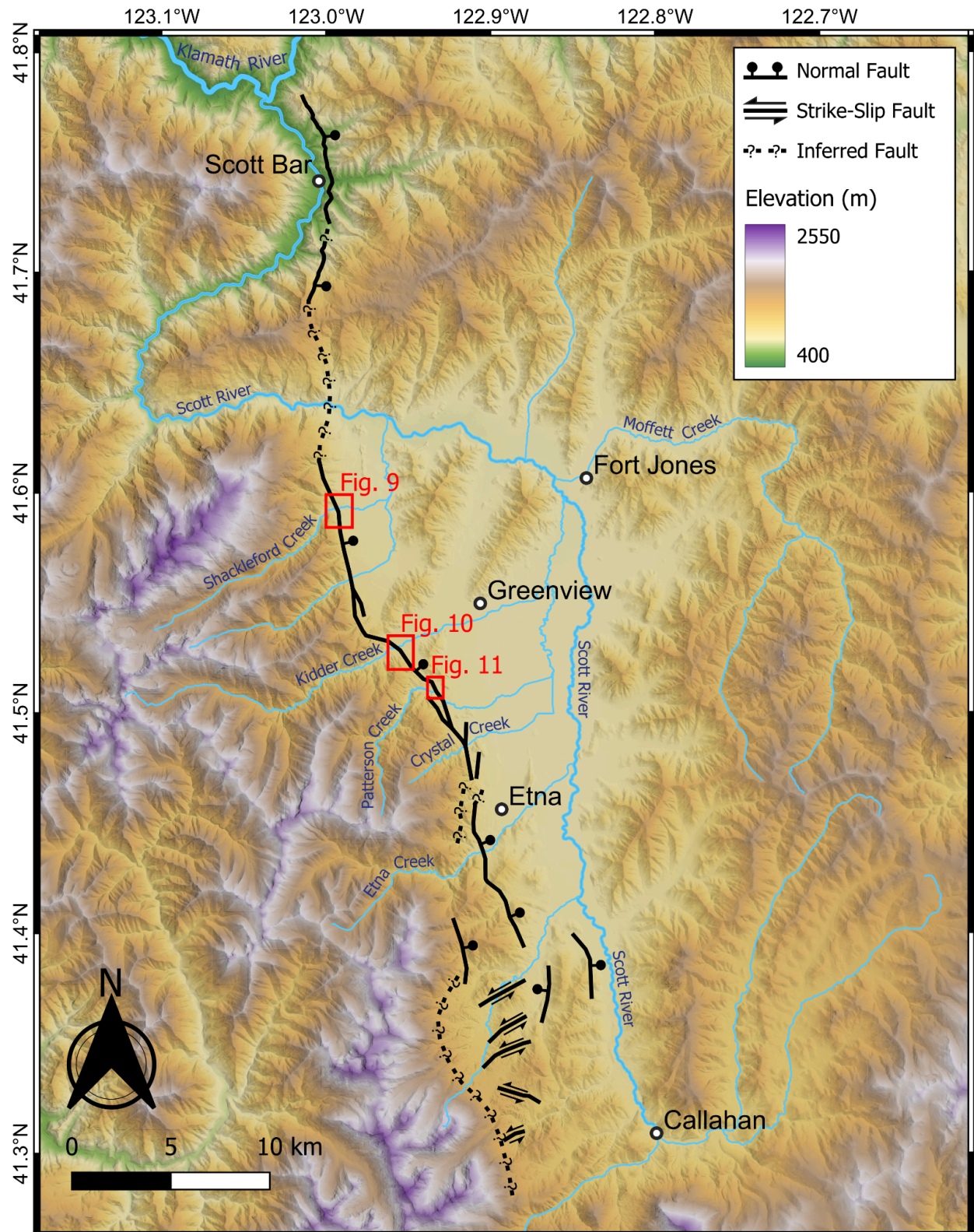


Figure 8. Hillshaded topography of the region encompassing the Scott Valley normal fault.

The geomorphic expression of the SVNf is most pronounced between the town of Etna and Shackleford Creek (Figure 8). Along this extent lie the best-preserved fault scarps and clearest evidence for recent fault activity. North of Shackleford Creek, geomorphic evidence for fault activity is absent for an 8 km-long stretch where the Scott River exits the valley (Figure 8). I interpret that geomorphic evidence for fault activity has been covered by a series of large landslides and removed by fluvial action of the Scott River (Figure 6).

North of this stretch, the fault scarp reappears and continues for 9 km over rugged terrain while displaying an eastward dip as it crosses ridges and gullies (Figure 8). Notably, this trace coincides with the east-dipping reverse fault separating the Hayfork terrane from the Rattlesnake Creek terrane and Condrey Mountain schist (Strand, 1963; Wagner and Saucedo, 1987; Figure 1). The northern terminus of the SVNf lies approximately ~4.5 km north of the town of Scott Bar where geomorphic evidence for faulting ceases. A field investigation of the northernmost extent of the SVNf was restricted by private land access.

In the southern Scott Valley, at a point approximately 7 km south of Etna, geomorphic evidence for activity along the main strand of the SVNf ceases. Broadly coincident with this point, the valley floor transitions to a mountainous upland, alluvial cover is diminished and restricted to isolated basins (Figure 6), and the east-facing mountain front steps westward by 3 km (Figure 8). The base of this secondary mountain front is marked by scarps of a separate east-dipping normal fault, which offsets recent alluvial deposits and strikes 3 km in a north-south direction (Figure 8). While the prominent western mountain front continues further southward, I do not observe geomorphic evidence for active faulting along its base. Should the trace of the SVNf continue along the base of the mountain front, its length would reach 63 km. This southward extent is mapped as inferred (Figure 8).

Numerous smaller active faults are present in the southern Scott Valley (Figure 8). These include two north-south striking normal faults and a series of subparallel ENE-WSW striking sinistral strike-slip faults (Figure 8). These faults stretch no more than 3 km in length and do not appear to have accumulated substantial long-term offset (Figure 8).

4.2.2 Surficial Mapping

I identify a sequence of six fluvial terrace surfaces (T1–T6) at the outlets of Shackleford, Kidder, and Patterson Creeks along the SVNf (Figures 9, 10, & 11). This numbering scheme represents a relative age hierarchy based on elevation above the active channel, surface roughness, and degree of riser degradation. Alluvial fan deposits sourced from smaller secondary tributaries are mapped but not categorized by relative age. While I focus on detailed characterization of deposits at the outlet of Shackleford Creek, I correlate surfaces at Kidder Creek and Patterson Creek based on these consistent geomorphic signatures.

At each site, T1 corresponds to the active fluvial deposits within modern channels. These deposits are characterized in the field by the presence of large, sub-rounded boulders reaching 2 meters in diameter. T1 deposits occupy the lowest elevations in the mapped areas and are restricted to areas immediately adjacent to the active channels.

T2 is situated slightly above T1. T2 is geomorphically distinct from older terraces due to its high surface roughness and sharp, well-defined terrace risers. Like T1, the surface of T2 is dominated by large boulders up to 2 meters in diameter. Remote mapping indicates that T2 lacks the degree of surface diffusion seen in older units.

T3 is found at the Shackleford and Kidder Creek sites situated ~5-7 meters above the active channels. In remote mapping, T3 shows a smoother surface texture and more diffuse terrace edges than T2. At the Shackleford Creek site, T3 is found predominantly on the hangingwall, where it forms a large contiguous surface north of the modern active channel (Figure 9b).

T4 is situated ~8-10 meters above the active channels of each site and is characterized by widespread, subrounded boulders on the surface and within eroded terrace edges. At the Shackleford and Patterson Creek sites, T4 extends across the SVNf and shows no sign of displacement by faulting. This observation suggests that T4 deposits post-date the most recent rupture along the SVNf.

At Shackleford Creek, T5 lies on the fault hangingwall, south of the active channel and adjacent to T4 (Figure 9b). T5 generally lacks the widespread surface cover of large boulders characteristic of

younger units, though isolated large clasts are present. The T4-T5 surface contact at Shackleford Creek is not defined by a change in relative elevation above the active channel but rather by a significant change in surface roughness, with T5 appearing substantially smoother than T4 in remote mapping. Because T5 is defined by fewer large surface clasts, and the T4-T5 contact is defined by a change in surface roughness rather than elevation, it is possible that T5 represents overbank deposits correlative in age with T4. T5 deposits are not continuously preserved across the SVNF at any of these three sites. Due to this, it cannot be determined if T5 has been displaced by the SVNF. Should the T5 deposits at Shackleford Creek represent the same depositional sequence as T4, which is not displaced across the SVNF, this would indicate that T5 deposits post-date the most recent rupture along the SVNF.

T6 is the highest and oldest preserved terrace surface mapped at these sites. T6 is restricted to the footwall of the SVNF and is situated ~22 meters above the active channels of Shackleford and Patterson Creeks. Remote surficial mapping indicates T6 is not present at the Kidder Creek site. The surface of T6 is characterized by a smooth morphology and diffuse edges, and it lacks the large surface clasts seen on younger deposits. T6 represents the only terrace to have been clearly displaced by the SVNF.

Local alluvium is mapped remotely where topographic contours appear to fan out in a different orientation than what would be expected by deposits sourced from mainstem active channels. Local alluvium at Shackleford Creek is characterized by generally smaller and more angular clasts. An exception is deposits sourced from Sniktaw Creek, a relatively large tributary north of Shackleford Creek, which has deposited boulders up to ~2 meters in diameter and built a large alluvial fan extending nearly to the active Shackleford Creek channel (Figure 9b). These Sniktaw deposits record as much as 8 meters of vertical displacement along the scarp of the SVNF.

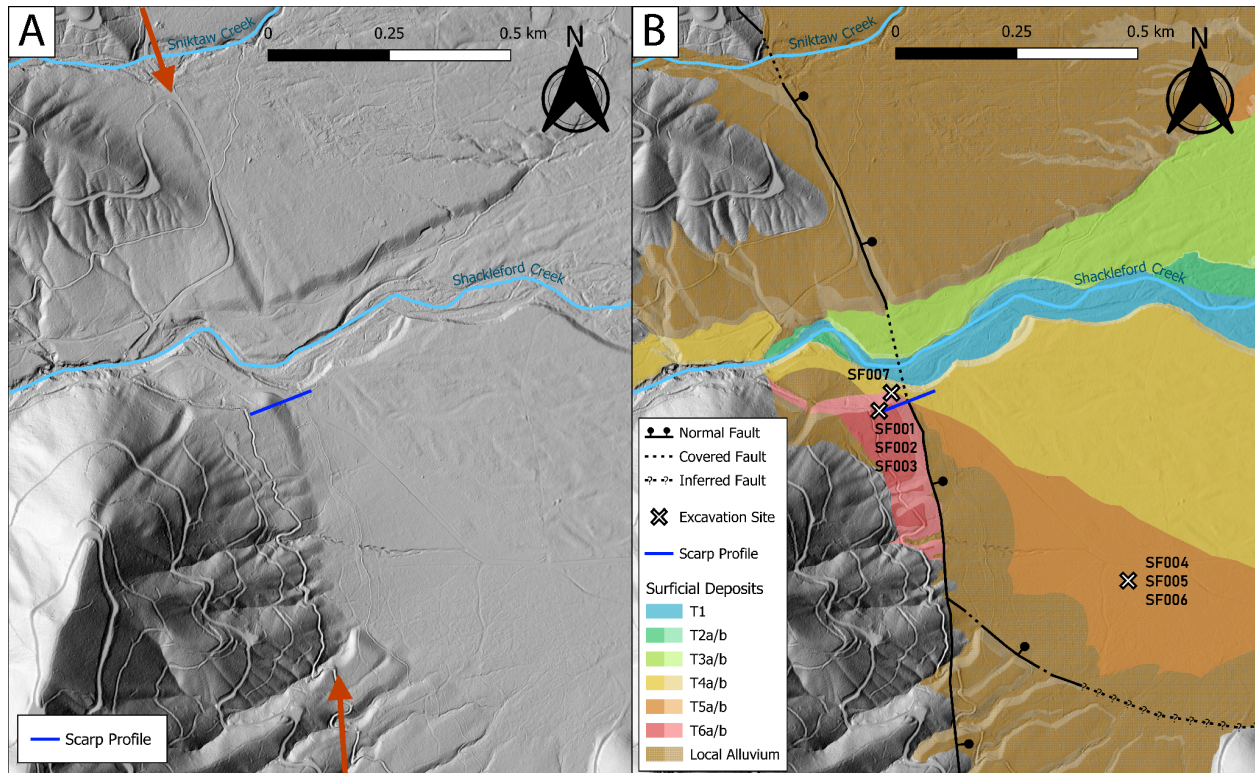


Figure 9. (A) Hillshade highlights the scarp of the Scott Valley normal fault at the outlet of Shackleford Creek. Red arrows point to the trace of the SVNF. Blue line indicates the scarp profile used in displacement analysis. **(B)** Surficial map at the outlet of Shackleford Creek with deposits sourced from Shackleford Creek categorized by depositional age from youngest to oldest from T1 to T6. Units labeled “a” denote relict depositional surfaces while units labeled “b” denote eroded terrace edges. Deposits not sourced from Shackleford Creek are noted as Local Alluvium. Locations of IRSL geochronology excavation sites are marked and labeled with corresponding sample IDs.

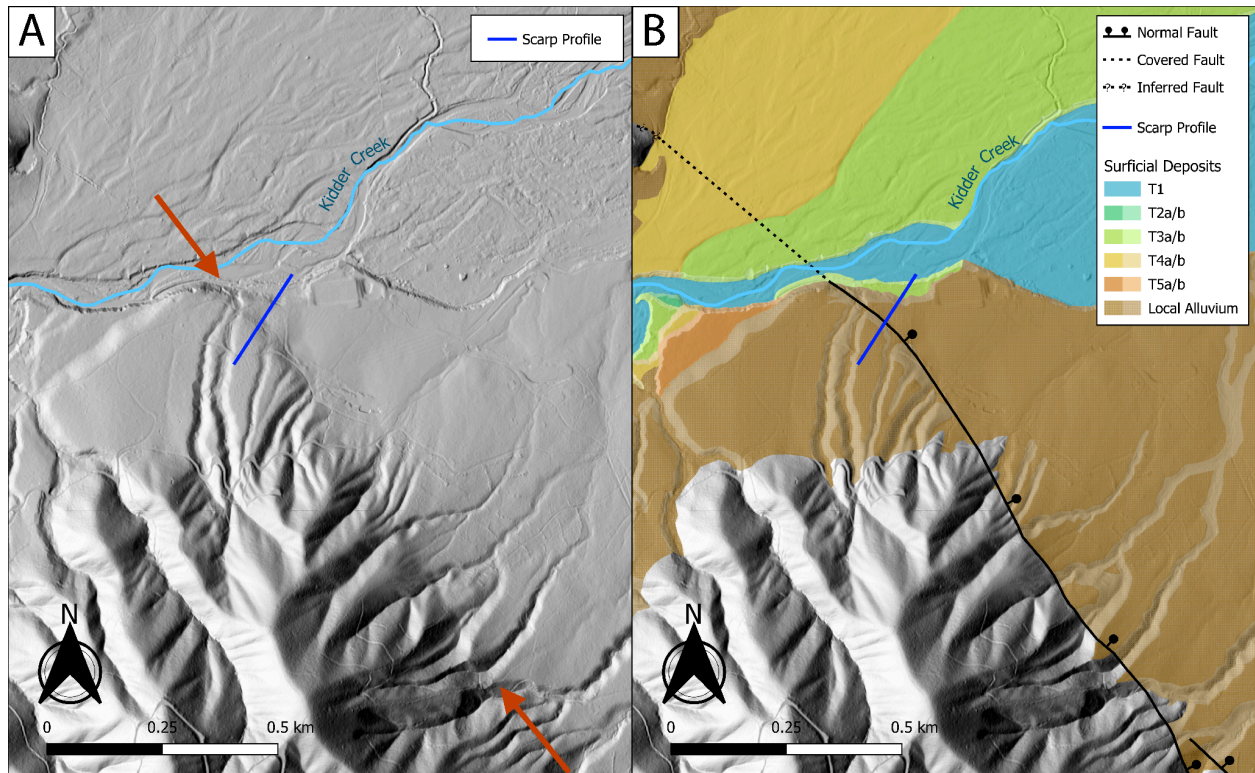


Figure 10. (A) Hillshade highlights the scarp of the Scott Valley normal fault at the outlet of Kidder Creek. Red arrows point to the trace of the SVNF. Blue line indicates the scarp profile used in displacement analysis. **(B)** Surficial map at the outlet of Kidder Creek with deposits sourced from Kidder Creek categorized by depositional age from youngest to oldest from T1 to T5. Units labeled “a” denote relict depositional surfaces while units labeled “b” denote eroded terrace edges. Deposits not sourced from Kidder Creek are noted as Local Alluvium.

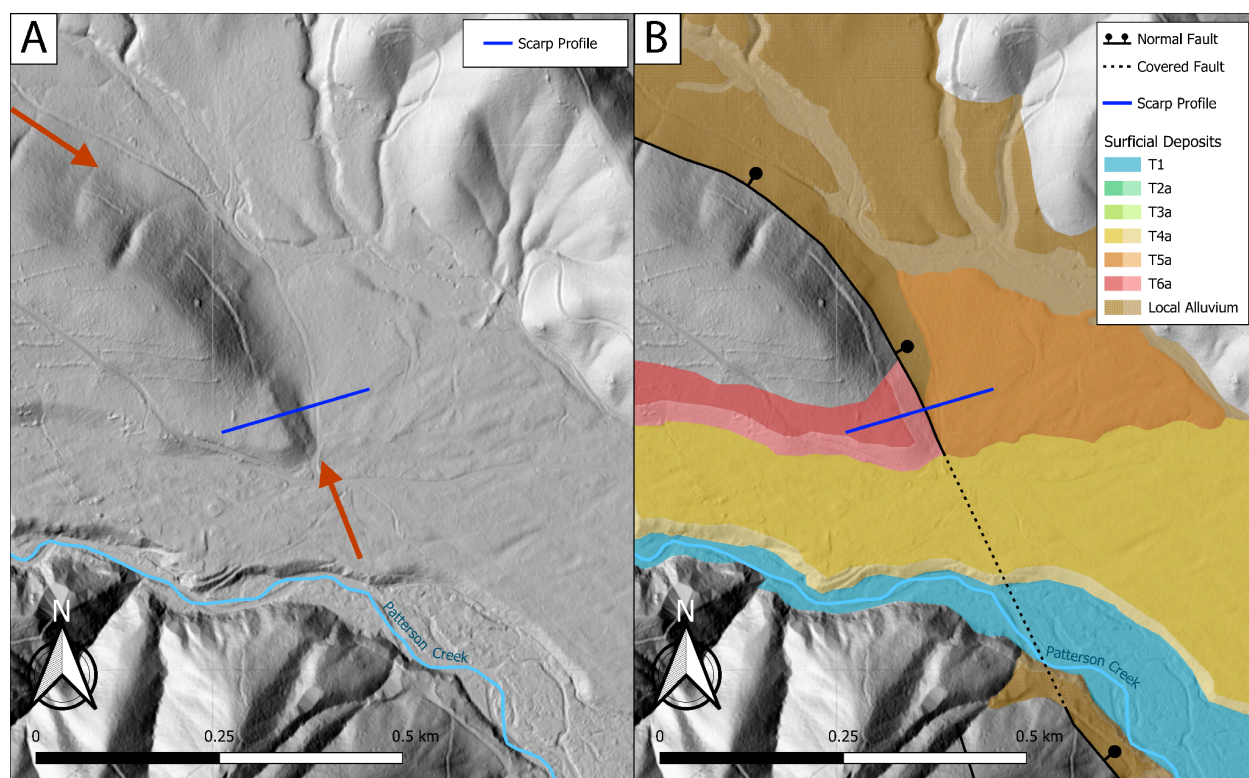


Figure 11. (A) Hillshade highlights the scarp of the Scott Valley normal fault at the outlet of Patterson Creek. Red arrows point to the trace of the SVNF. Blue line indicates the scarp profile used in displacement analysis. **(B)** Surficial map at the outlet of Patterson Creek with deposits sourced from Patterson Creek categorized by depositional age from youngest to oldest from T1 to T6. Units labeled “a” denote relict depositional surfaces while units labeled “b” denote eroded terrace edges. Deposits not sourced from Patterson Creek are noted as Local Alluvium.

4.2.3 IRSL Sampling and Analysis

I excavated holes on the footwall and the hangingwall of the SVNf at the outlet of Shackleford Creek to determine the timing of terrace formation at this site. The footwall excavation reached 1.60 meters below the preserved surface of T6 (Figures 12 & 13). Samples in this excavation were collected at depths of 1.47, 0.85, and 0.50 meters and assigned sample IDs of SF001, SF002, and SF003, respectively (Table 1). The hangingwall excavation reached 2.00 meters below the preserved surface of T5 (Figures 14 & 15). Samples in this excavation were collected at depths of 1.85, 1.25, and 0.76 meters and assigned sample IDs of SF004, SF005, and SF006, respectively (Table 1). An additional excavation was performed into a roadcut on the eroded edge of a T6 riser (Figure 12). One sample (ID: SF007) was collected from

this site at a depth of 0.40 meters below the roadcut surface, ~1.25 meters below the projected sloped surface of the terrace riser, and in a stratigraphic position ~12.5 meters below the surface of the primary hangingwall excavation site where samples SF001, SF002, and SF003 were collected (Table 1).

Parent material for terrace deposits at the outlet of Shackleford is derived from the Central Klamath Mountains and Rattlesnake Creek terranes and consists primarily of metavolcanic rocks and minor serpentized ultramafic rocks (Wagner and Saucedo, 1987; Irwin, 1994). At the primary excavation sites, terrace surfaces are flat or nearly flat and vegetated by a pine forest with a well-developed organic forest floor. Annual precipitation is approximately 550 mm/yr (PRISM Climate Group, 2026).

Deposits on the footwall are capped by a 25-cm-thick A horizon. The B horizon is approximately 110 cm thick and extends to a depth of 135 cm. I divide the B horizon into three units (Bt1, Bt2, Bt3). The Bt2 unit (40–100 cm) consists of mottled 10YR 5/6 to 4/6 clay/silt with abundant 7.5YR 4/6 clay films and manganese (Mn) coatings. Individual clasts of slate within this horizon are weathered and transformed into gray clay. The lower Bt3 unit is sandier and yellower, with clay films restricted primarily to clast surfaces. The soil transitions to a BC horizon at 135 cm, consisting of silty sand with gravel, no soil structure, and weak clay coatings.

Deposits on the hangingwall are capped by a 30-cm-thick A horizon. The B horizon is approximately 90 cm thick and extends to a depth of 120 cm. The Bt2 and Bt3 units (52–120 cm) are clay-rich. Colors in the Bt horizon are typically 10YR 5/6 to 4/6 (dry) and 10YR 4/6 to 3/6 (wet). At 120 cm, the soil grades into a BC horizon of silty sand. Below 165 cm, the material becomes much sandier. A narrow band of red, rounded, oblong lapilli tephra, between 1 mm and 1 cm in size and coated in white silt, was present near the surface of both excavations.

Grain size analysis shows broadly similar sedimentological characteristics for each of the excavation sites, with silt-dominated textures, minor clay, and median grain sizes in the coarse silt to very fine sand range (Table 1). Based on these characteristics, the T5 and T6 deposits at Shackleford Creek are most consistent with overbank deposition.

Sample ID	Terrace	Collection Depth (m)	% Sand Size (2000-63 μm)	% Silt Size (63-2 μm)	% Clay Size (<2 μm)
SF001	T6	1.47	46	50	4
SF002	T6	0.85	46	50	4
SF003	T6	0.50	34	60	6
SF004	T5	1.85	41	54	5
SF005	T5	1.25	33	62	6
SF006	T5	0.76	30	65	6
SF007	T6	1.25 (~12.50)	40	56	4

Table 1. Basic summary of IRSL sample IDs, corresponding terrace, and sample collection depth, and sample grain size distributions. Collection depth is from the surface; in parentheses is the depth from the terrace surface.

T6 (Footwall) Soil Horizons

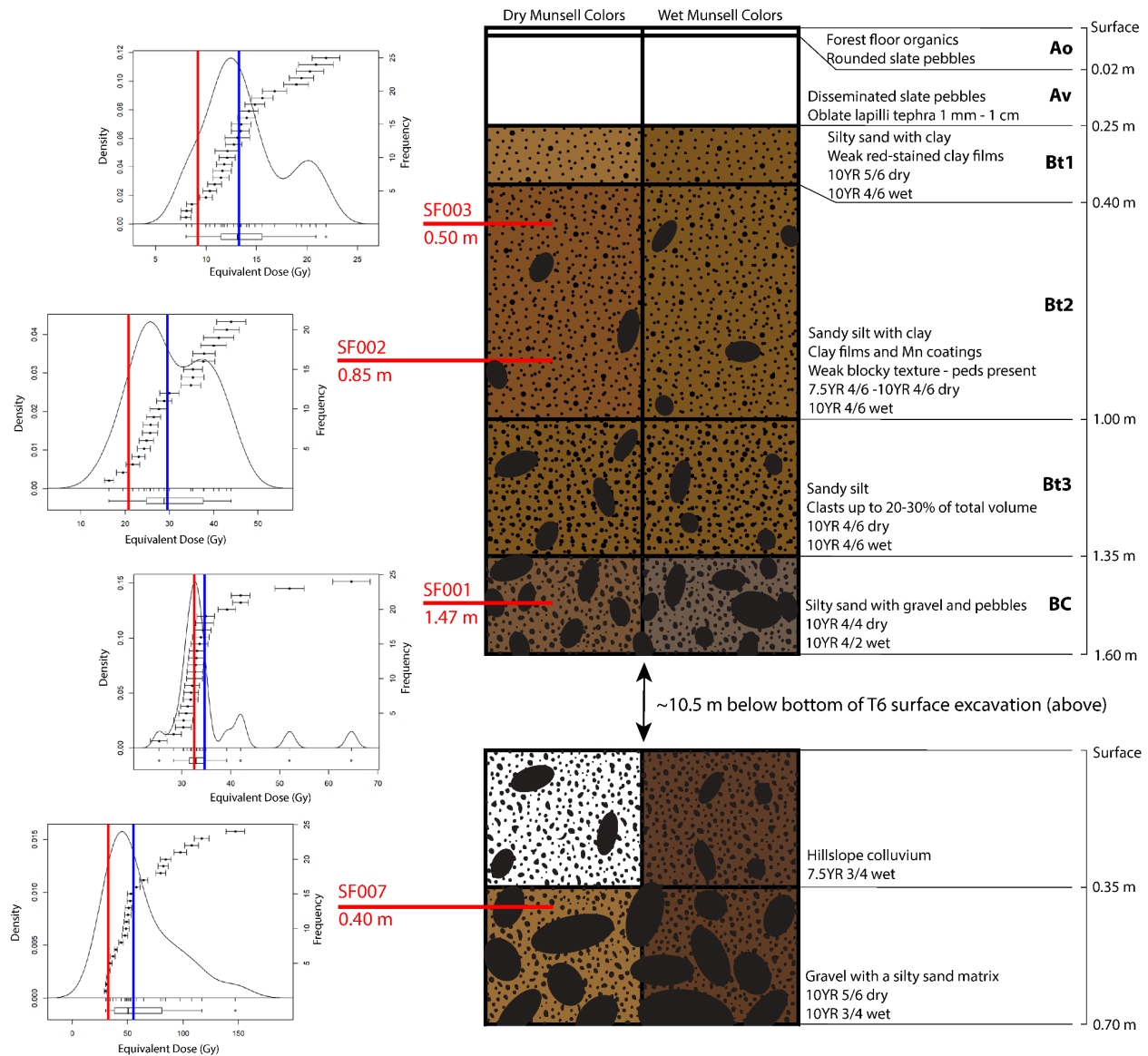


Figure 12. Soil colors and particle sizes exposed in T6 at Shackleford Creek. Soil horizons marked and labeled. Note ~10.5 meter stratigraphic break separating T6 sample localities. Sample collection depths marked and labeled adjacent to their corresponding kernel density estimate plots of measured equivalent dose (D_e) populations. Kernel density estimate plots include a boxplot depicting median as a thick central line, first and third quartiles as box limits, whiskers denoting ± 1.5 interquartile ranges, and dots further outliers. The vertical blue line represents the Central Age Model D_e estimate; the vertical red line represents the Minimum Age Model D_e estimate. See Figure E1 for Abanico plots.



Figure 13. Images displaying upper (left) and lower (right) sections of the 1.6-meter-deep T6 footwall excavation with measuring tape for scale (cm).

T5 (Hangingwall) Soil Horizons

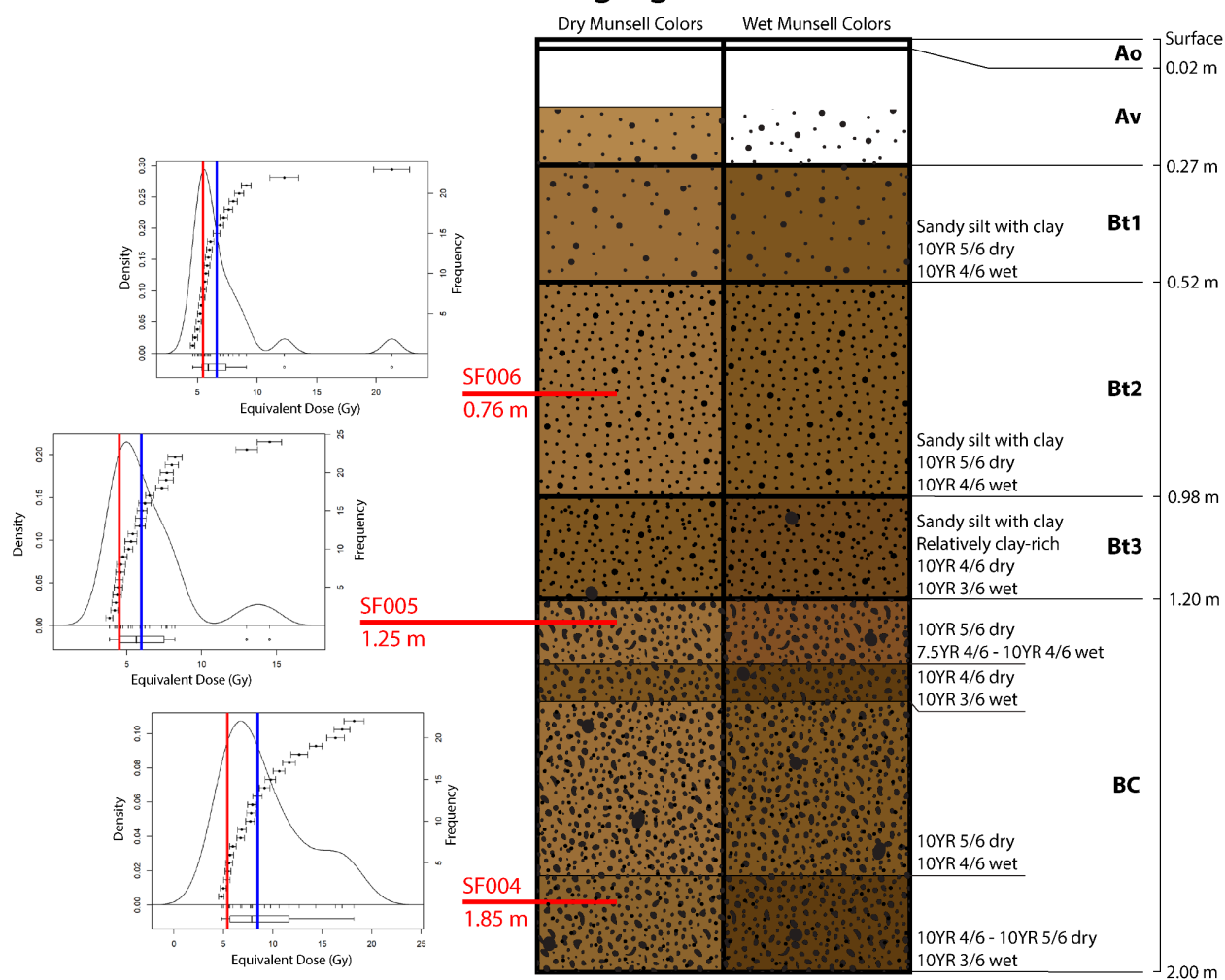


Figure 14. Soil colors and particle sizes exposed in T5 at Shackleford Creek. Soil horizons marked and labeled. Sample collection depths marked and labeled adjacent to their corresponding kernel density estimate plots of measured equivalent dose (D_e) populations. Kernel density estimate plots include a boxplot depicting median as a thick central line, first and third quartiles as box limits, whiskers denoting ± 1.5 interquartile ranges, and dots further outliers. The vertical blue line represents the Central Age Model D_e estimate; the vertical red line represents the Minimum Age Model D_e estimate. See Figure E1 for Abanico plots.



Figure 15. Images displaying upper (left) and lower (right) sections of the 2.0-meter-deep T5 hangingwall excavation with measuring tape for scale (cm).

The IR_{50} signal intensity for all samples was low, which prevented the use of single-grain IRSL measurements. Consequently, even though small aliquots were measured, some degree of signal averaging across grains is possible. In several cases, the equivalent dose (D_e) distributions appear multimodal, likely reflecting a combination of incomplete bleaching prior to burial and possible post-depositional sediment mixing.

Several samples (SF001, SF002, SF003, and SF006) yield relatively low overdispersion values (< 30%; Table 2), which might ordinarily suggest well-bleached populations. However, because the measurements were conducted on small aliquots rather than individual grains, some degree of averaging of luminescence signals across grains is inevitable. As a result, a portion of the true scatter in the D_e distributions may not be fully represented in the calculated overdispersion values.

Sample ID	N	OD (%)	CAM D _e (Gy)	MAM D _e (Gy)	Total dose rate (Gy/ka)	Uncorrected CAM Age (ka)	Fading-corrected CAM Age (ka)	Uncorrected MAM Age (ka)	Fading-corrected MAM Age (ka)
SF001	24 (30)	15 ± 4	34.61 ± 1.33	32.52 ± 1.01	1.81 ± 0.17	19.11 ± 1.92	22.53 ± 2.41	17.96 ± 1.75	21.16 ± 2.18
SF002	21 (30)	24 ± 5	29.51 ± 1.73	20.69 ± 1.85	2.29 ± 0.18	12.90 ± 1.29	15.16 ± 1.67	9.04 ± 1.09	10.59 ± 1.23
SF003	25 (30)	25 ± 4	13.24 ± 0.75	9.18 ± 0.75	2.11 ± 0.17	6.27 ± 0.63	7.32 ± 0.79	4.34 ± 0.50	5.05 ± 0.62
SF004	22 (30)	39 ± 6	8.46 ± 0.70	5.40 ± 0.47	1.67 ± 0.16	5.06 ± 0.63	5.90 ± 0.75	3.23 ± 0.41	3.75 ± 0.46
SF005	24 (30)	76 ± 11	5.95 ± 0.41	4.49 ± 0.35	1.85 ± 0.16	3.22 ± 0.36	3.74 ± 0.47	2.43 ± 0.29	2.81 ± 0.40
SF006	23 (30)	32.9 ± 5	6.59 ± 0.46	5.46 ± 0.23	1.84 ± 0.16	2.96 ± 0.29	3.43 ± 0.36	2.96 ± 0.29	3.43 ± 0.36
SF007	26 (30)	91 ± 13	55.09 ± 4.98	32.37 ± 2.63	1.51 ± 0.15	36.45 ± 4.95	43.23 ± 5.89	21.42 ± 2.78	25.28 ± 3.27

Table 2. Summary of equivalent dose and luminescence age data based on the IR₅₀ signal from multigrain feldspar aliquots. N is the number of equivalent dose determinations accepted after screening; in parentheses are the total number of aliquots measured. Preferred ages are in bold-face type. Age uncertainties are reported at the 1 σ confidence level.

Sample ID	U (ppm)	Th (ppm)	K (%)	Rb (ppm)	Alpha dose rate (Gy/a)	Beta dose rate (Gy/ka)	Gamma dose rate (Gy/ka)	Cosmic dose rate (Gy/ka)	Total dose rate (Gy/ka)
SF001	0.11	0.19	0.98	19.90	0.01 ± 0.01	0.71 ± 0.08	0.25 ± 0.03	0.21 ± 0.02	1.81 ± 0.17
SF002	0.51	1.29	1.27	30.20	0.03 ± 0.01	0.98 ± 0.11	0.41 ± 0.04	0.22 ± 0.02	2.29 ± 0.18
SF003	0.50	1.24	1.06	30.80	0.03 ± 0.01	0.83 ± 0.09	0.36 ± 0.03	0.25 ± 0.03	2.11 ± 0.17
SF004	0.47	0.96	0.68	18.20	0.03 ± 0.01	0.56 ± 0.06	0.25 ± 0.02	0.20 ± 0.02	1.67 ± 0.16
SF005	0.52	1.21	0.81	23.60	0.03 ± 0.01	0.66 ± 0.07	0.30 ± 0.03	0.21 ± 0.02	1.85 ± 0.16
SF006	0.41	0.96	0.85	22.90	0.02 ± 0.01	0.67 ± 0.07	0.29 ± 0.03	0.23 ± 0.02	1.84 ± 0.16
SF007	0.44	0.96	0.56	13.30	0.02 ± 0.01	0.47 ± 0.05	0.22 ± 0.02	0.15 ± 0.02	1.51 ± 0.15

Table 3. Summary of dose rate data used for age calculation. Dose rate uncertainties are reported at the 1 σ confidence level.

The Central Age Model (CAM) is generally best suited for well-bleached deposits and may yield age overestimates when incompletely bleached grains are present in the D_e distribution. In contrast, the Minimum Age Model (MAM) isolates the lowest-dose population in the distribution and is typically better suited for partially bleached deposits. It remains possible that some younger grains may have been introduced post-depositionally, for example, through bioturbation or other sediment mixing processes, which could bias the model toward anomalously young values. Nevertheless, the MAM ages likely represent a robust minimum constraint on the timing of deposition and provide the most defensible

chronological interpretation given the limitations of the dataset. Ages and dose rate data for all samples are listed in Tables 2 and 3 and should be evaluated in conjunction with independent age information and field observations.

Samples SF001, SF002, SF003, and SF007 were collected on the footwall of the SVNf from T6, the oldest preserved terrace at the Shackleford Creek site. Sample SF001, sampled at a depth of 1.47 m, yields a fading-corrected MAM age of 21.2 ± 4.4 ka (Table 2). Samples SF002 and SF003 yield significantly younger ages of 10.6 ± 2.5 ka and 5.1 ± 1.2 ka, respectively (Table 2). Sample SF007, collected from a stratigraphic position ~12.5 meters below the surface of T6, yields an older age of 25.3 ± 6.5 ka (Table 2).

Samples SF004, SF005, and SF006 were collected from T5, the oldest preserved terrace surface on the fault hangingwall at the Shackleford Creek site. These samples yield consistent fading-corrected MAM ages between 2.8 ± 0.8 ka and 3.8 ± 0.9 ka (Table 2).

Age constraints were not obtained at the Kidder Creek and Patterson Creek sites.

4.2.4 Fault Scarp Displacement Analysis

I analyze scarps of the SVNf at Shackleford Creek, Kidder Creek, and Patterson Creek to estimate the fault throw and dip-slip displacement preserved at these sites. At the Shackleford Creek site, the scarp of the SVNf displaces the T6 surface on the footwall (Figure 9b). Here, I estimate 12.5 ± 0.5 meters of throw and $14.5^{+1.9}_{-1.5}$ meters of dip-slip displacement (Figure 16). The absence of a T6 surface marker on the hangingwall constrains these to minimum values. At the Kidder Creek site, the scarp of the SVNf displaces alluvial deposits derived from minor tributaries adjacent to Kidder Creek (Figure 10b). Here, I estimate 6.6 ± 1.1 meters of throw and $7.7^{+1.8}_{-1.5}$ meters of dip-slip displacement (Figure 17). At the Patterson Creek site, the scarp of the SVNf displaces the T6 surface on the footwall (Figure 11b). Here, I estimate 13.9 ± 0.1 meters of throw and $16.2^{+1.8}_{-1.4}$ meters of dip-slip displacement (Figure 18). The absence of a T6 surface marker on the hangingwall constrains these to minimum values.

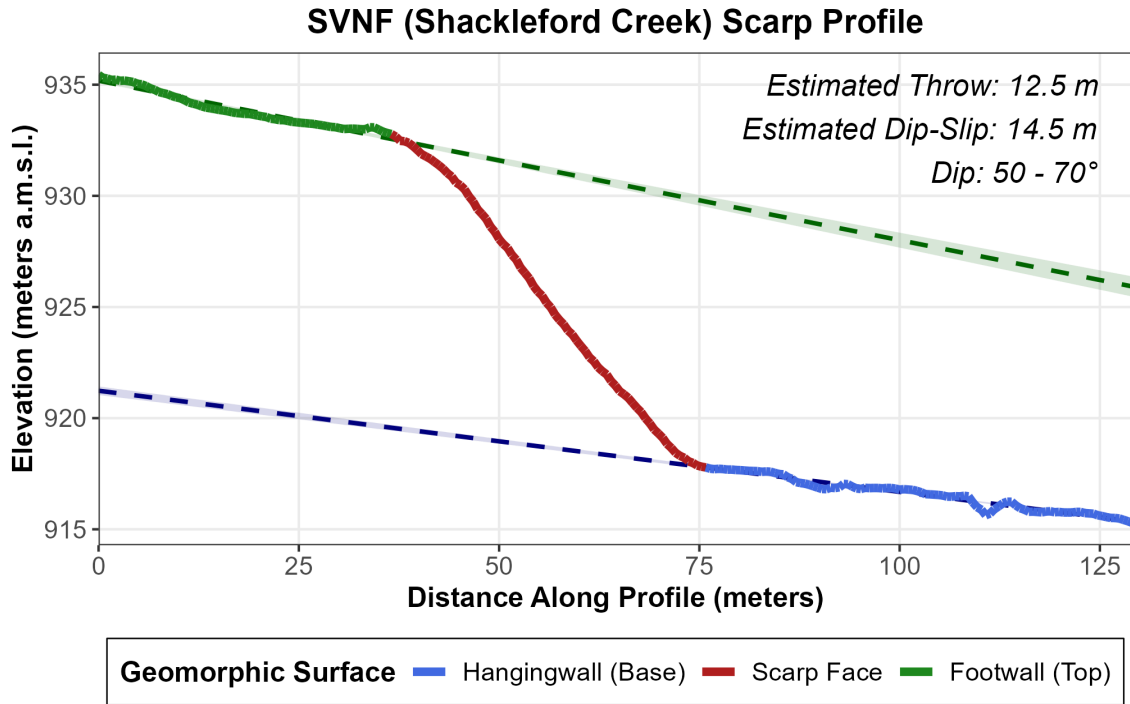


Figure 16. Terrain profile across the scarp of the Scott Valley normal fault at Shackleford Creek. This scarp preserves an estimated 12.5 ± 0.5 meters of throw and $14.5^{+1.9}_{-1.5}$ meters of dip-slip displacement.

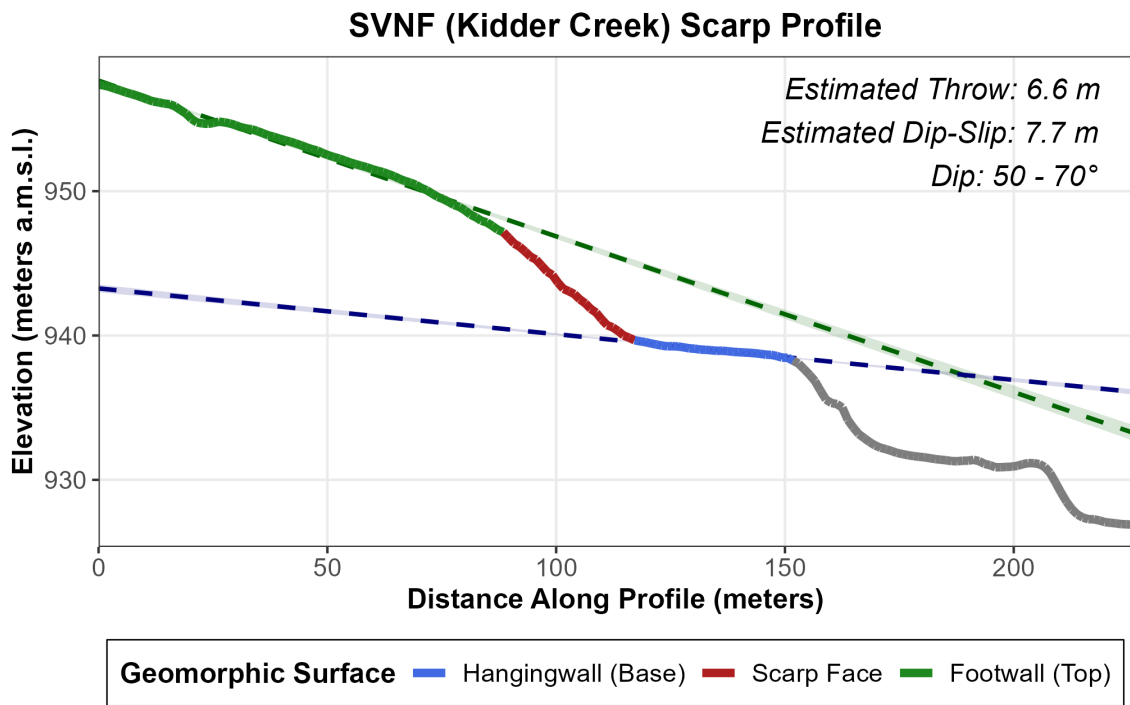


Figure 17. Terrain profile across the scarp of the Scott Valley normal fault at Kidder Creek. This scarp preserves an estimated 6.6 ± 1.1 meters of throw and $7.7^{+1.8}_{-1.5}$ meters of dip-slip displacement.

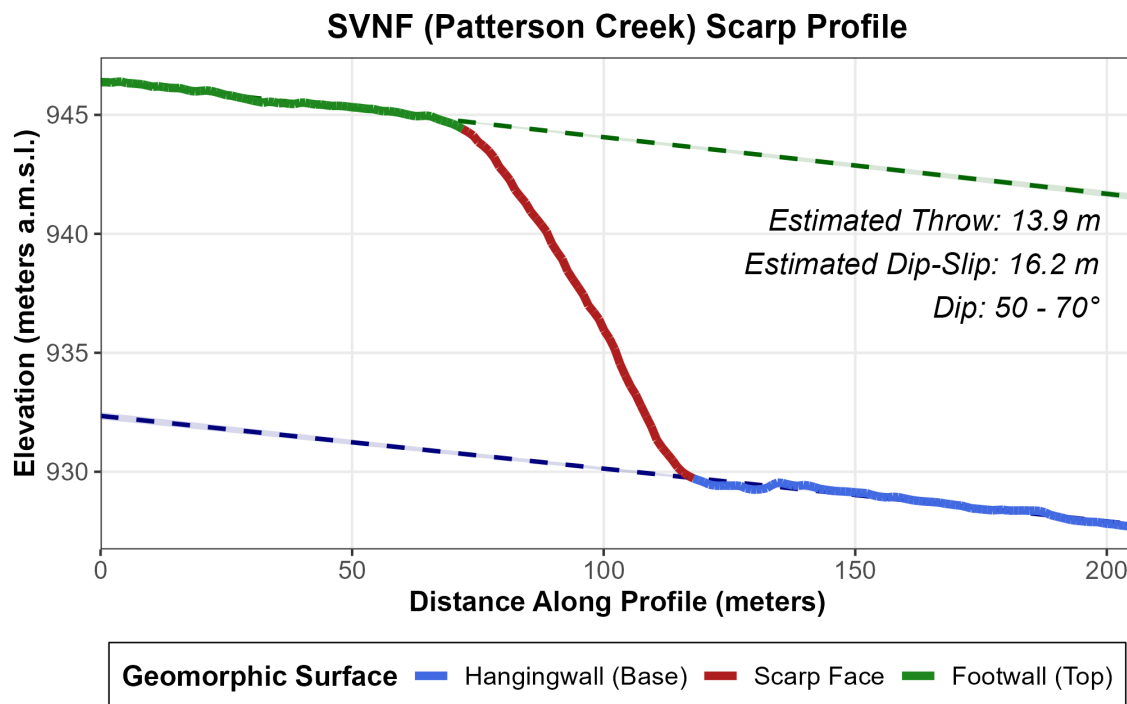


Figure 18. Terrain profile across the scarp of the Scott Valley normal fault at Patterson Creek. This scarp preserves an estimated 13.9 ± 0.1 meters of throw and $16.2^{+1.8}_{-1.4}$ meters of dip-slip displacement.

4.2.5 Subsurface Basin Architecture of Scott Valley

Of the 1591 documented wells in Scott Valley, 1089 were usable for the purposes of depth-to-bedrock mapping. Of these, 559 encountered bedrock (Figure 19). The deepest drilled well reaches a depth of 142 meters without encountering bedrock. The deepest bedrock encountered is at a depth of 122 meters. Both of these superlative wells are located in Quartz Valley.

The scarcity of wells drilled deeper than 100 meters limits the accuracy of subsurface geometries beyond this point. However, based on extrapolating the gradient of the bedrock surface, I estimate that bedrock depths reach 320 meters in the wide central part of Scott Valley, over 142 meters in Quartz Valley, and 60 meters in Oro Fino Valley (Figure 20). South of Etna, bedrock depths remain under 40 meters and drop below 20 meters in the southernmost portion of the valley. Following the Scott River downstream, bedrock depths reach 320 meters in the wide central part of the valley, under 80 meters near Fort Jones, and over 160 meters west of Fort Jones (Figure 20).

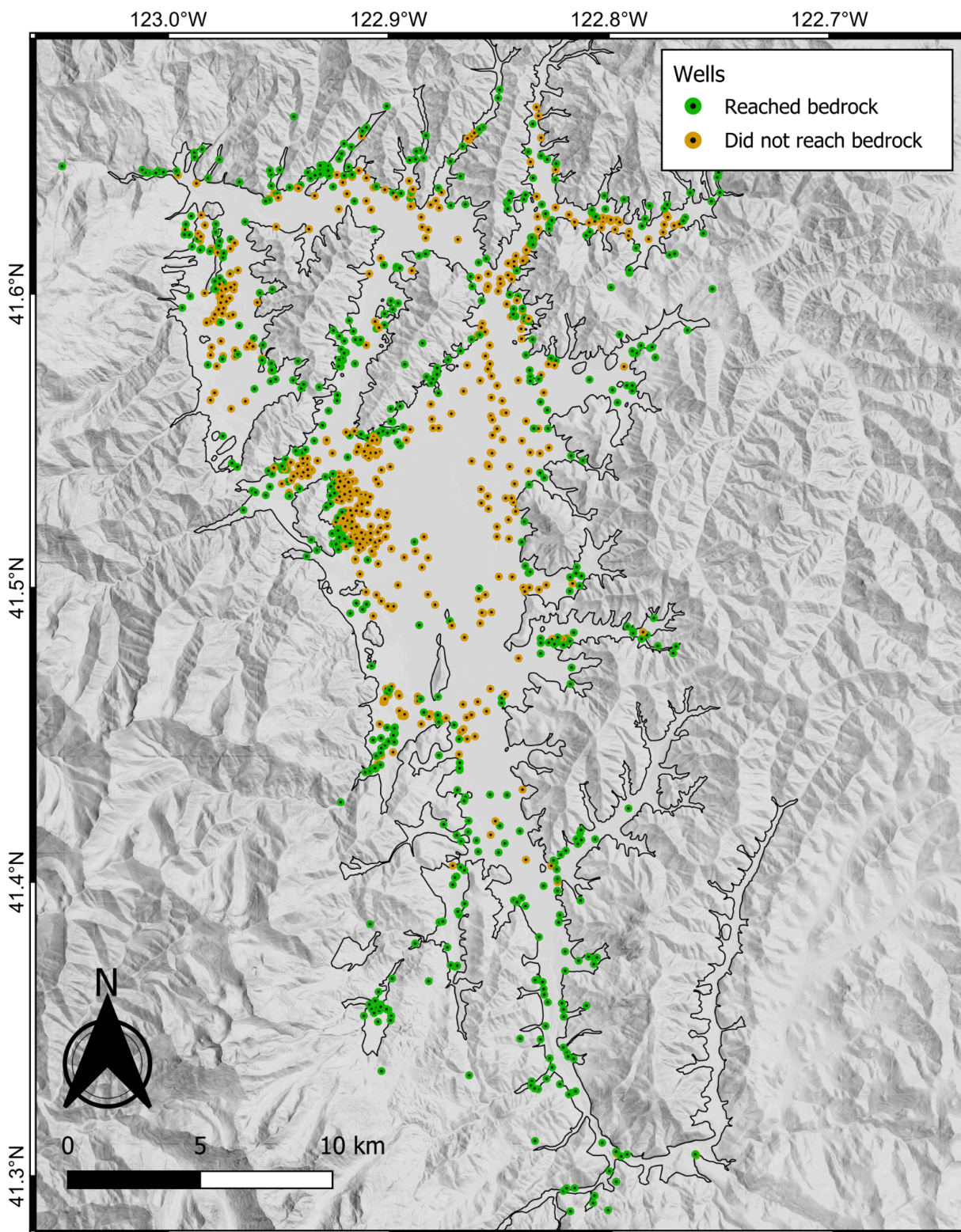


Figure 19: Hillshade of Scott Valley with well locations plotted as points. Wells encountering bedrock marked green. Wells not encountering bedrock marked yellow. Black line denotes the extent of valley fill.

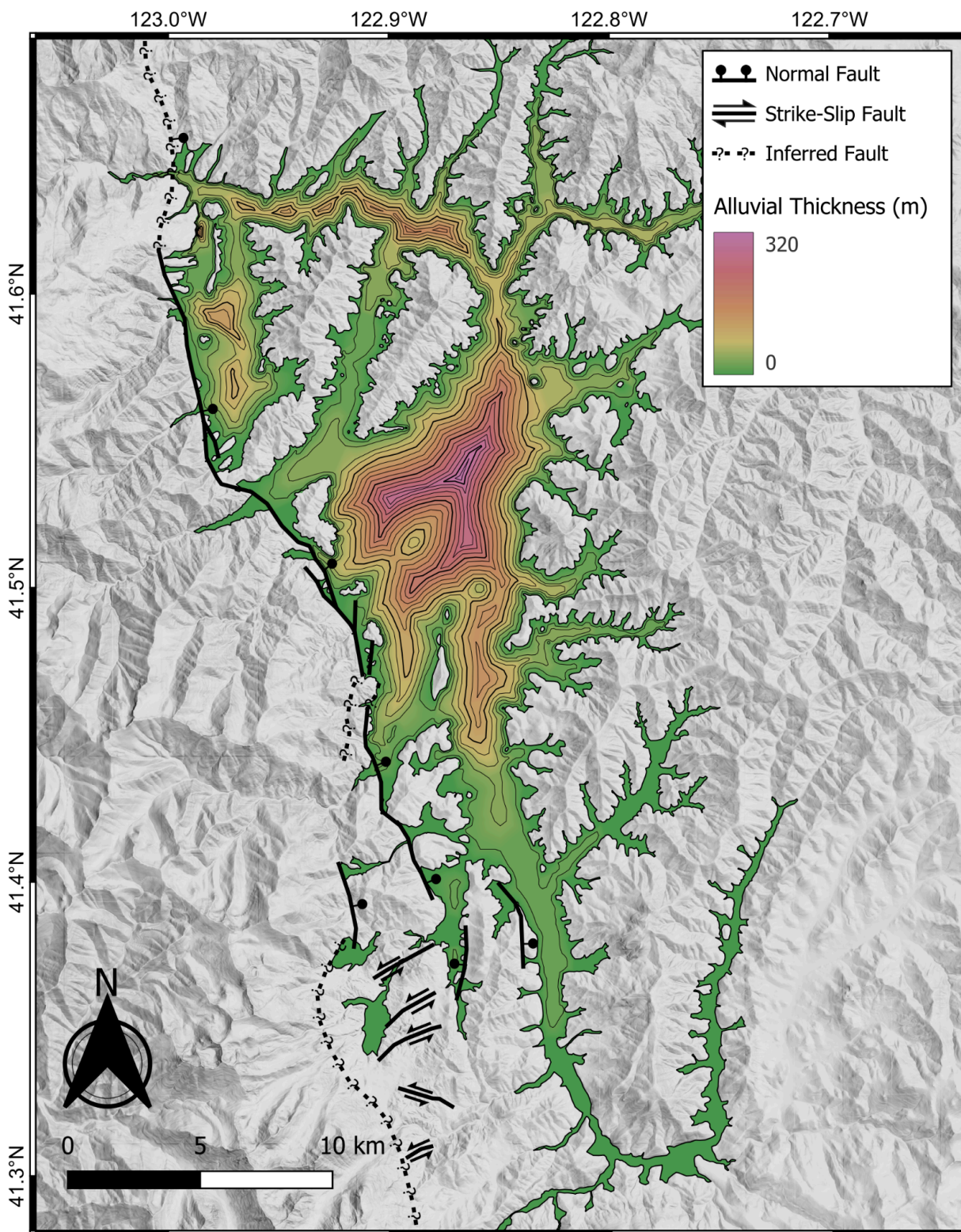


Figure 20: Hillshade of Scott Valley with mapped active faults. Alluvial thicknesses color-coded by depth. Depth-to-bedrock contours plotted at 20 meter intervals with 100 meter index contours. I estimate alluvium reaches a maximum thickness of 320 meters in Scott Valley.

5. Discussion

5.1 Minimum Dip-Slip Rate of the Scott Valley Normal Fault

Surficial mapping and scarp displacement analysis at the outlet of Shackleford Creek indicate a minimum dip-slip displacement of the T6 surface of $14.5^{+1.9}_{-1.5}$ meters. I interpret the age of the T6 surface to be best represented by the 21.2 ± 4.4 ka from the MAM estimate obtained from sample SF001, collected at a depth of 1.47 meters from the T6 surface. This age is broadly synchronous with the timing of mid-Last Glacial Maximum (LGM), a period that may have been characterized by relatively higher sediment flux, dated at $19.9^{+0.3}_{-1.6}$ ka in the Klamath Mountains by Heermance et al. (2026). Furthermore, this age is consistent with the age obtained from sample SF007 (25.3 ± 6.5 ka) collected within T6 from a stratigraphic position ~12.5 meters below the T6 surface and ~11 meters below SF001. The more advanced age of SF007 represents an earlier part of the aggradation of the T6 deposit, and its lower stratigraphic position provides a weaker constraint on determining the displacement of the T6 surface.

I attribute the younger ages of 10.6 ± 2.5 ka and 5.1 ± 1.2 ka obtained from samples SF002 and SF003, respectively, as the result of post-depositional sediment mixing (bioturbation). I base this interpretation on the shallower sampling depths, bimodal equivalent dose distributions, and relatively higher overdispersion values of IRSL aliquot ages from these samples.

T5, the oldest preserved terrace surface on the hangingwall of the SVNf at the Shackleford Creek site, is considerably younger than T6, yielding ages between 2.8 ± 0.8 ka and 3.8 ± 0.9 ka across its three samples.

I expect that T6 deposits have been covered by T5 deposits on the hangingwall of the SVNf in a younger aggradational stage. For this reason, the estimated mean dip-slip rate of the SVNf at Shackleford Creek provided in this study is a minimum. As the excavation of T5 did not uncover the surface of T6 to a depth of 2 meters, it may be interpreted that the SVNf has vertically displaced T6 by at least 2 meters more than estimated.

Combining the estimated minimum dip-slip displacement of the T6 surface ($14.5^{+1.9}_{-1.5}$ meters) with its IRSL-derived age of 21.2 ± 4.4 ka yields an estimated minimum mean dip-slip rate of 0.7 ± 0.2 mm/yr for the SVNf at Shackleford Creek.

5.2 Earthquake Potential of the Scott Valley Normal Fault

To estimate the moment magnitude (M_w) and average displacement (AD) of an earthquake on the SVNf, I utilize the scaling relationships of Thingbaijam et al. (2017) with fault rupture area (RA) as the predictor. I utilize the scaling relationships of Thingbaijam et al. (2017) rather than those of Wells and Coppersmith (1994) as the normal faulting dataset used by Thingbaijam et al. (2017) encompasses greater potential RAs, up to 29,531 km² compared to the 900 km² limit of Wells and Coppersmith (1994). Thingbaijam et al. (2017) additionally provide a direct regression for AD as a function of RA, which I prioritize as the predictor in these analyses due to the likely high-aspect-ratio geometry of the SVNf. The RA- M_w and RA-AD relationships for normal faults, from Thingbaijam et al. (2017), are as follows:

$$M_w = (\log(RA) + 2.551) / 0.808$$

$$\log(AD) = -2.779 + (0.858 * \log(RA))$$

RA represents the total surface area of the fault plane and is defined as the product of the surface rupture length (SRL) and downdip width (W). While the SRL of the SVNf is directly quantifiable through mapping, determining W, and by extension, RA, requires a well-constrained estimate of the seismogenic depth (H) in this region.

H is governed by the depth of the brittle-ductile transition, where creep begins to dominate over brittle deformation, and depends on rock composition, temperature, stress regime, and strain rate (Brace and Kohlstedt, 1980; Zeng et al., 2022). Assuming a constant eastward dip of 60°, the SVNf would reach this depth at 25 km according to the models of Zeng et al. (2022), which define lower seismogenic depths for the western US by averaging the deepest 10% of seismicity with magnitudes greater than 1.0.

I evaluate the earthquake potential of the SVNf using two scenarios based on the fault's well-constrained and inferred SRL. While lidar analysis confirms a minimum SRL of 47 km, the fault has

the potential to reach a total SRL of 63 km. This higher estimate depends on the southern terminus of the fault, which remains enigmatic. South of Etna in Scott Valley, both the fault and the eastward-facing mountain front appear to step westward, with the fault transferring slip to a separate, 3-km-long normal fault. While this mountain front continues southward for 16 km, geomorphic indicators of active faulting at its base are absent south of the northern 3 km. It is possible that slip along a formerly active range-front-bounding normal fault in the south may have altogether ceased. Alternatively, an active fault may be present here with longer interevent times than the northern SVNf. Because the base of this mountain front consists of undisturbed LGM moraines, this suggests that no slip has occurred here since at least the LGM. However, because I cannot discount the possibility that the SVNf is capable of rupturing up to 63 km, I include this SRL in my analyses to account for a potential maximum SRL. In either case, deformation in the southern Scott Valley is distributed across a series of relatively smaller faults rather than concentrated on one primary fault strand.

SRL	M_w	Lower	Upper	AD (m)	Lower (m)	Upper (m)
47 km	7.0	6.6	7.4	0.8	0.2	3.6
63 km	7.2	6.8	7.6	1.0	0.2	4.6

Table 4: Predicted moment magnitude (M_w) and average displacement (AD) for a rupture on the Scott Valley normal fault. Lower and upper bounds of the 95% confidence intervals provided. Predictions are made for the well-constrained (47 km) and inferred (63 km) surface rupture length (SRL) of the Scott Valley normal fault assuming a seismogenic depth of 25 km (Zeng et al., 2022).

Predicted moment magnitudes (M_w) and average fault displacement (AD) for a rupture on the SVNf are summarized in Table 4. These results indicate the SVNf is capable of producing M_w 7.0 to 7.2 earthquakes with an AD of $0.8^{+2.8}_{-0.6}$ meters. This estimated earthquake magnitude is somewhat elevated compared to normal faults of a similar length in the Basin and Range province due to the greater seismogenic thickness of the KMP. Zeng et al. (2022) attribute this deep brittle-ductile transition to subduction of the Gorda plate, which lowers temperatures in the crust and upper mantle and results in a

more brittle structural setting. However, if the SVNf is listric, similar to the model proposed by Platt (1986) for forearc normal faulting, the seismogenic depth may be shallower. This is because seismogenic depths decline to the east, toward the Cascade arc (Zeng et al., 2022). In this case, however, the corresponding increase in fault RA with a listric geometry may ultimately increase M_w and AD estimates.

To estimate the earthquake recurrence interval of the SVNf, I simulate the number of rupture events required to accumulate the minimum $14.5^{+1.9}_{-1.5}$ meters of dip-slip displacement preserved at the Shackelford Creek site. Using a Monte Carlo approach based on the calculated AD distribution (Table 4), I estimate that this scarp was built by 13^{+7}_{-6} rupture events. Integrating this result with the 21.2 ± 4.4 ka age of the T6 surface yields an estimated mean recurrence interval of 1700^{+1300}_{-700} years for the SVNf.

5.3 Evidence of Sustained Faulting in Scott Valley

Scott Valley is a half-graben basin formed by the active SVNf on its western margin. Dip-slip displacement along the SVNf has generated as much as 1700 meters of vertical relief between the valley floor and the western mountains. This combination of footwall uplift and hangingwall subsidence has similarly created the accommodation space for an estimated 320-meter-thick package of alluvium to blanket the valley floor.

I expect that prior to the formation of the SVNf, topography in what is now the fault hangingwall resembled the mountainous terrain of surrounding areas. As hangingwall subsidence accumulated, the Scott River responded by depositing sediment, effectively burying much of the pre-existing landscape and establishing the modern flat valley floor. Today, only the highest peaks of this ancestral topography remain exposed at the surface as isolated bedrock outcrops surrounded by valley fill.

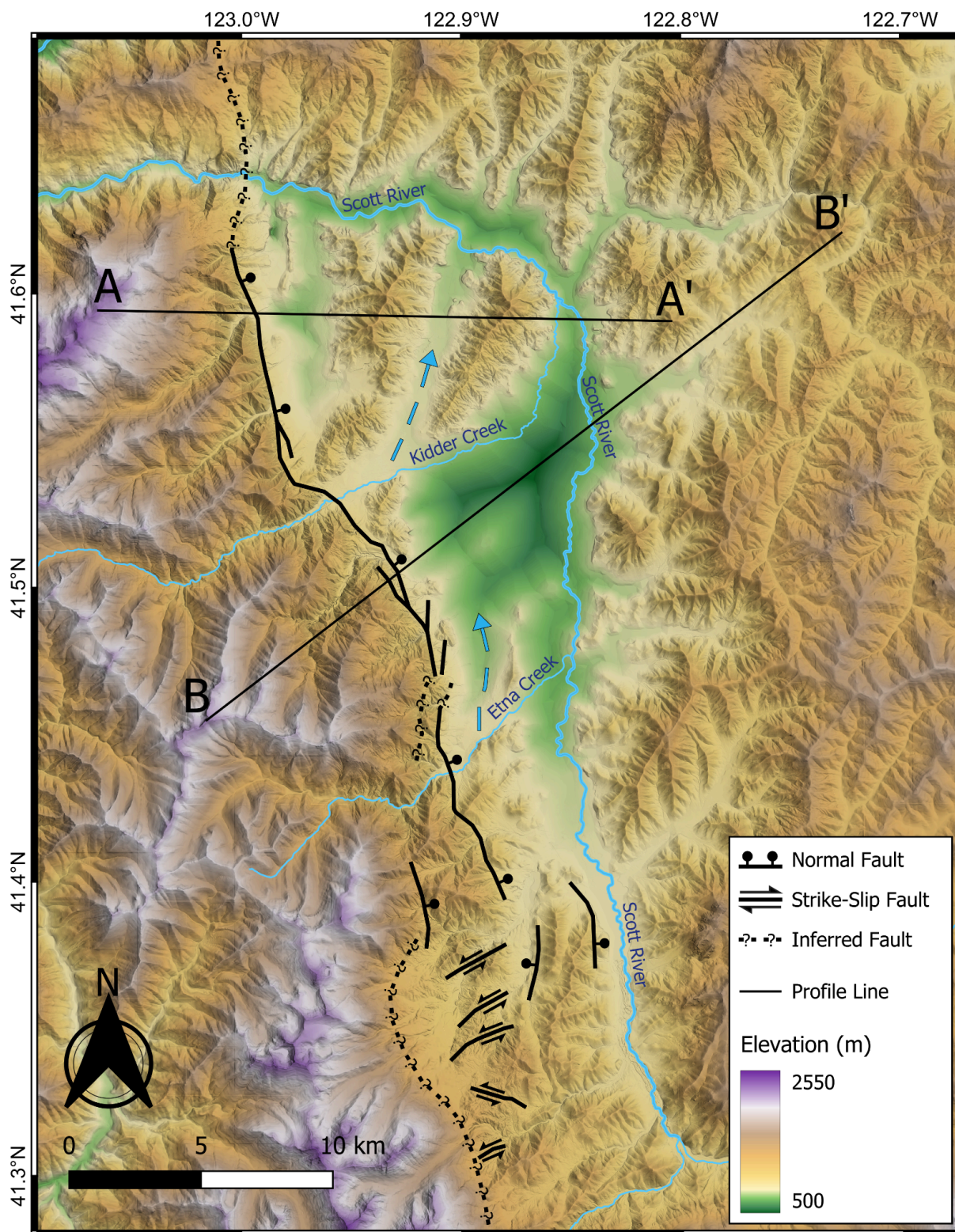


Figure 21: Hillshaded topographic map of Scott Valley with valley floor alluvium removed. Terrain profiles (Figures 22 & 23) shown. Dashed blue arrows denote likely ancestral flow directions of Kidder Creek and Etna Creek.

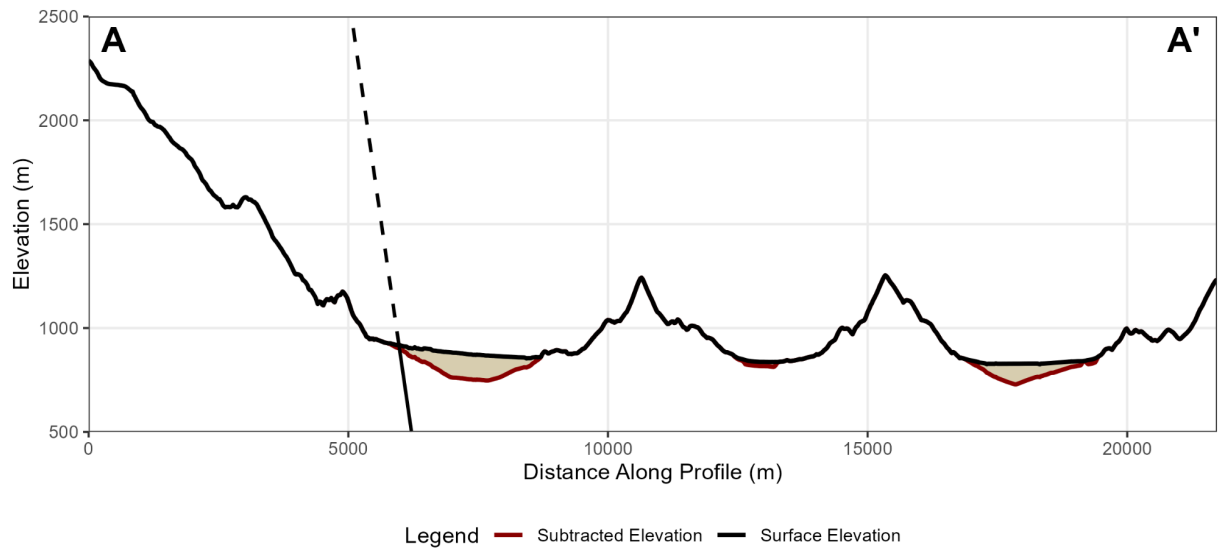


Figure 22. East-west terrain profile A - A' across the northern Scott Valley. Black profile line denotes true ground surface elevation above sea level; red profile line denotes elevation of bedrock above sea level. Brown shading indicates alluvium of Scott Valley. Scott Valley normal fault plotted assuming a 60° dip as a dashed line in the air and a solid line beneath the surface. Plotted with 4:1 vertical exaggeration.

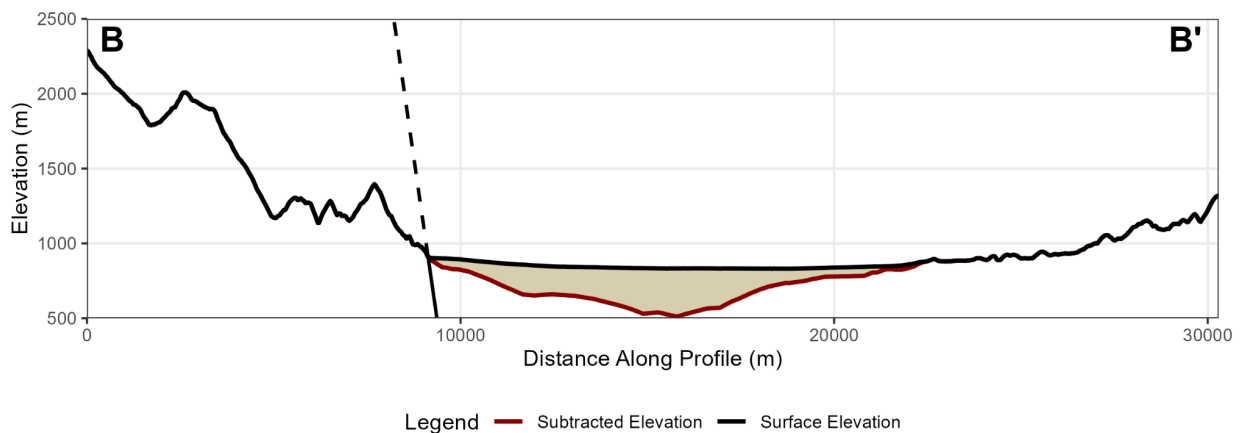


Figure 23. Southwest-northeast terrain profile B - B' across the central Scott Valley. Black profile line denotes true ground surface elevation above sea level; red profile line denotes elevation of bedrock above sea level. Brown shading indicates alluvium of Scott Valley. Scott Valley normal fault plotted assuming a 60° dip as a dashed line in the air and a solid line beneath the surface. Plotted with 4:1 vertical exaggeration.

Depth-to-bedrock mapping confirms the hypotheses of Mack (1958) that alluvial infilling of the valley floor has led to the reorganization of Scott River tributaries. Etna Creek, which presently flows due east upon reaching the valley floor, once extended ~5 km northward of its present mouth before joining

the Scott River (Figure 21). It seems likely that Kidder Creek once flowed further northeast than its present course and occupied what is now Oro Fino Valley prior to joining the Scott River (Figure 21).

Contrary to a typical half-graben, where tilting of the hangingwall focuses sedimentation adjacent to the fault, the thickest package of alluvium in Scott Valley lies nearly 10 km east of the fault (Figure 20). I attribute this depocenter location to the ancestral eastward-plunging paleocanyons in the hangingwall. I propose that while the hangingwall has undergone westward tilting, tilting has not yet overcome the original eastward gradient of these paleocanyons. This interpretation is evidenced by a comparison of modern and buried stream gradients: while modern western tributaries descend as much as 1500 m over a 10 km distance in the footwall, the buried paleocanyons in the hangingwall descend no more than 320 m over a comparable distance.

An additional feature of the basin depth is that alluvial thicknesses are not greatest in the northern part of the valley, where the downstream reach of the Scott River paleocanyon once occupied a lower elevation. This observation may suggest that the greatest cumulative offset along the SVNf lies between the towns of Etna and Greenview, in which case the slip rate of the fault along this stretch likely exceeds the minimum mean dip-slip rate of 0.7 ± 0.2 mm/yr estimated further to the north at Shackleford Creek. Alternatively, slip along the northern segment of the fault may be partitioned such that the majority of slip contributes to footwall uplift rather than hangingwall subsidence, potentially due to high erosion rates in the Scott and Klamath River canyons focusing isostatic uplift in the north.

I calculate a preliminary estimate of long-term vertical displacement on the SVNf based on the topographic relief between prominent high-points in the hangingwall and footwall. Chaparral and Quartz Hills, the two most significant hills within Scott Valley on the fault hangingwall, reach elevations of 1283 m and 1250 m, respectively, while the Marble Mountains in the footwall are as high as 2530 meters (Figure 4). This observed relief suggests a minimum cumulative throw of ~1300 m. When combined with an estimated minimum mean throw rate of 0.6 ± 0.1 mm/yr at Shackleford Creek, this times the initiation of slip on the SVNf to ~2.2 Ma.

This estimate contains considerable uncertainty. First, the estimated ~1300 m cumulative throw is likely a minimum due to potentially higher erosion rates on the footwall in the Marble Mountains, which show evidence of extensive past glaciation. Second, this age calculation assumes the SVNf has slipped at a constant rate throughout its history. While further work is required to better constrain the long-term slip history of the SVNf, this preliminary estimate aligns with the documented history of deformation in the KMP.

5.4 Active Deformation in the Klamath Mountains Province

Active deformation in the KMP contrasts distinctly with documented deformation on its margins. To the south, deformation is characterized by north-south shortening (Angster et al., 2021), while the western margin is characterized by dextral shear (Kelsey and Carver, 1988). Within the KMP, however, the presence of three north-south-striking normal faults indicates that deformation is dominated by east-west extension. In this way, active deformation in the KMP most closely resembles that of the Basin and Range province, to its east.

Newly identified normal faults in the KMP are unique in the Cascadia forearc; outside the KMP, the dextral oblique Beaufort Range fault on Vancouver Island comprises the only documented active normal fault in the Cascadia forearc (Lynch et al., 2025). Furthermore, the presence of generally north-south striking normal faults accommodating east-west extension in the KMP contrasts with geodetic models that predict northwest-southeast oriented contraction and/or regional dextral shear (Hammond and Thatcher, 2005; Unruh and Humphrey, 2017). While the dextral strike-slip Hyampom fault (Appendix C), situated east of and adjacent to a narrow zone of NNW-trending active strike-slip faults (Kelsey and Carver, 1988), provides evidence for dextral shear in the southwestern KMP, the limited length of this fault and the absence of similar structures imply that extension is dominant in the KMP. This discrepancy suggests that geodetic observations are not manifested in active faulting but rather primarily reflect elastic strain accumulation from the locked subduction interface.

The modern east-west extensional regime of the KMP is a relatively recent phenomenon. The KMP has experienced three distinct episodes of deformation since the late Miocene: north-south extension, dextral transpression, and modern east-west extension. Miocene extension was apparently more dramatic than at present; I observe no modern feature comparable to the high slip rate (~ 2 mm/yr) and ~ 60 km of cumulative displacement estimated on the La Grange detachment (Piotraschke et al., 2015). Around ~ 10 Ma, this north-south Miocene extension was replaced by dextral strike-slip and reverse faults, a transition Michalak et al. (2024) attribute to the northwestward-directed motion of the Sierran block. I find no evidence that these Miocene-to-Pliocene structures have been reactivated. I expect this is due to the near-perpendicular strike of these features to the east-west orientation of modern extension.

Transpressional deformation ceased in the KMP around ~ 2 Ma, at which time the region likely shifted to its current phase of east-west extension. This timing aligns with the initiation of slip on the SVNF, which I preliminarily date at ~ 2.2 Ma. I propose that this ~ 2 Ma shift to extension may have occurred due to a recent pulse of subcretion (Delph et al., 2021) and likely compounded by increased slab rollback from northward migration of the MTJ (Dickinson and Snyder, 1979; Humphreys and Coblenz, 2007).

Humphreys and Coblenz (2007) propose that southern Cascadia slab rollback creates the space necessary to accommodate both east-west extension in the northern Basin and Range and northwestward translation of the Sierran block. In this case, normal faults in the KMP represent the surface expression of southern Cascadia forearc extension. Domal uplift due to forearc subcretion beneath the KMP, proposed by Delph et al. (2021), offers an additional explanation for active extension. Forearc subcretion leads to crustal thickening and a decrease in bulk crustal density near the forearc-mantle corner of the overriding plate, resulting in isostatic uplift (Delph et al., 2021). Additionally, subcreted material can lift overlying crustal material as it detaches from the downgoing slab (Delph et al., 2021). Platt (1986) proposes that listric normal faults will form in an overriding plate when crustal thickening, due to subcretion, causes gravitational instability. Michalak et al. (2024) similarly offer subcretion as the driver for Miocene

extension in the KMP. The east-west orientation of modern extension, which differs from the north-south orientation of Miocene extension, is likely due to north-south compression induced by the northwestward translation of the Sierran block and corresponding reorientation of the least principal stress.

5.5 Rupture Coupling with Cascadia

Analysis of the strain tensor for the region encompassing the SVNf, which I derive from a two-dimensional least-squares strain tensor inversion of GPS/GNSS velocity data compiled for the 2023 National Seismic Hazard Model (Zeng, 2022), reveals a regime characterized by ENE-WSW-oriented shortening at a rate of $-2.54 \times 10^{-8} \text{ yr}^{-1}$ (Figure 24). This regional compression is due to interseismic locking of the Cascadia subduction zone (CSZ) interface and is broadly observed across the forearc of the CSZ (McCaffrey et al., 2013; McKenzie and Furlong, 2021).

This ENE-WSW-oriented compressive stress field is unfavorable for the activation of north-south striking normal faults in the KMP, and implies that ruptures on the SVNf, and other normal faults in the KMP, are coupled to the earthquake cycle of the CSZ. Large subduction zone earthquakes produce major stress field reversals during their coseismic and postseismic phases, during which time forearc extension and the activation of normal faults have been observed (Perfettini et al., 2005; Kato et al., 2011). For example, following the 2011 Mw 9.0 Tohoku event, a normal-faulting seismic sequence in the overlying plate was observed in NE Japan, a region dominated by reverse faulting prior to the megathrust event (Kato et al., 2011). This abrupt change from compression to extension is attributed to a reduction in trench-normal compressive stress and extension of the overriding plate as it rebounded trenchward (Kato et al., 2011). In another example following the 2001 Mw 8.4 Arequipa earthquake on the Peru-Chile subduction zone, a GPS station situated 180 km inland of the trench recorded 50 cm of trenchward motion during the coseismic phase and a further 15 cm of trenchward motion in the following 2 years (Perfettini et al., 2005). Perfettini et al. (2005) conclude that the period of extension of the overlying plate following a major subduction earthquake can last for several decades to over a century before returning to a state of trench-normal compressive stress.

Due to the subparallel orientation of normal faults in the KMP with the CSZ, I expect these faults are locked throughout the interseismic period of the CSZ when the region is subject to steady trench-normal east-west compression as presently observed. Furthermore, I expect that ruptures on these normal faults coincide with the stress field reversal brought about during the coseismic or postseismic phase of a CSZ megathrust event.

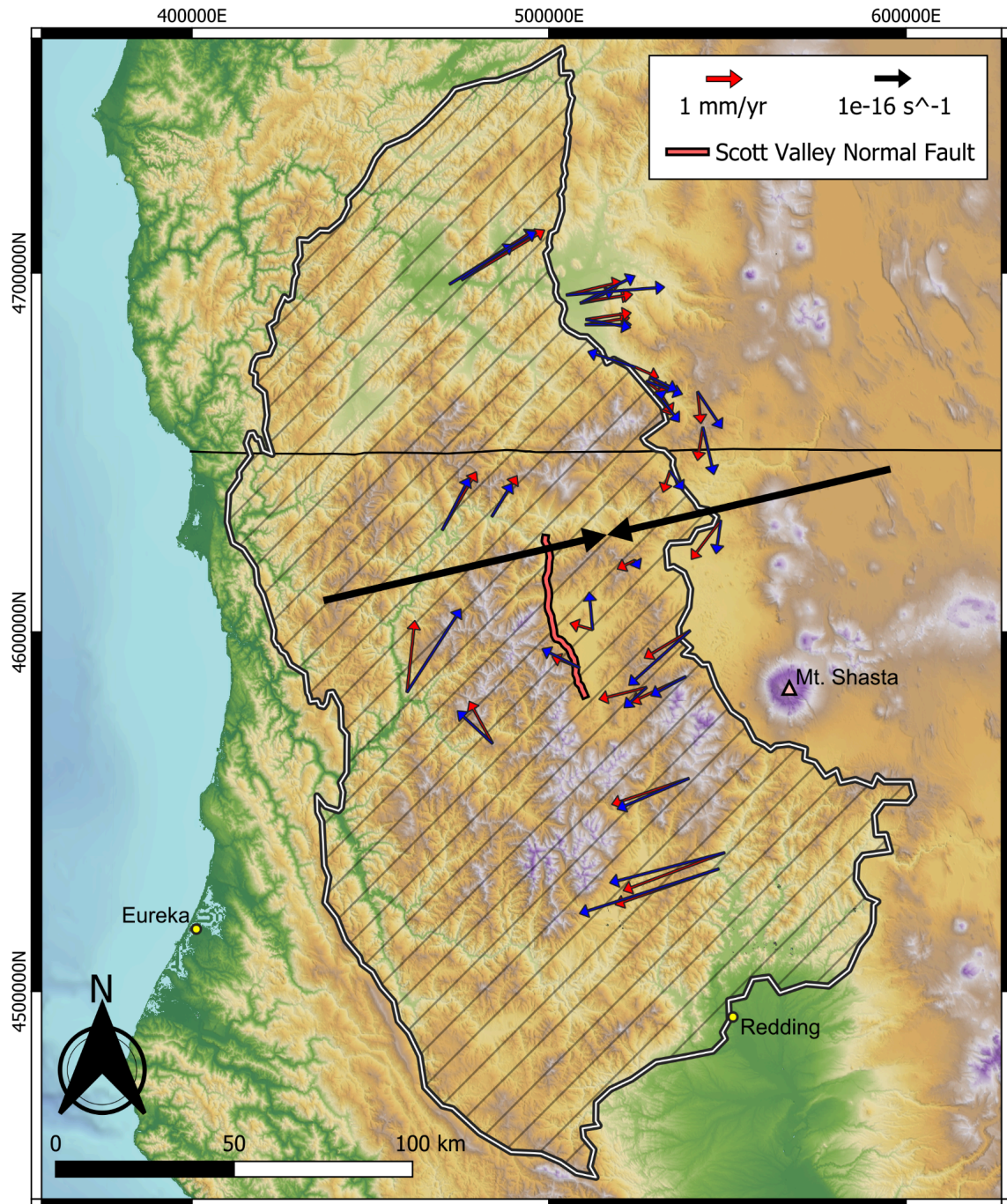


Figure 24. GPS/GNSS velocity field and derived regional strain tensor for the region surrounding the Scott Valley normal fault. Blue arrows represent observed horizontal velocities from the 2023 National Seismic Hazard Model GPS/GNSS compilation (Zeng, 2022); red arrows signify the modeled velocity field derived from a uniform strain inversion. Inward-pointing black arrows signify the orientation and magnitude of the principal shortening strain rate. All vectors are plotted relative to the mean velocity of the selected stations. Note presence of clockwise rigid-body rotation in addition to ENE-WSW shortening.

6. Conclusions

I identify three active normal faults and one dextral strike-slip fault in the Klamath Mountains province (KMP) and provide the first constraint on the rate of active faulting in this region. The longest fault I identify is the 47-km-long Scott Valley normal fault (SVNF), which I estimate possesses a minimum mean dip-slip rate of 0.7 ± 0.2 mm/yr based on the $14.5^{+1.9}_{-1.5}$ meters of displacement since 21.2 ± 4.4 ka estimated at the Shackleford Creek site. Scaling relationships suggest the SVNF is capable of producing a M_w 7.0 to 7.2 earthquake with an average displacement of $0.8^{+2.8}_{-0.6}$ meters per event. I estimate a median of 13^{+7}_{-6} events have occurred on the SVNF since 21.2 ± 4.4 ka, which indicates a mean earthquake recurrence interval of 1700^{+1300}_{-700} years. Preliminary analysis indicates a dip-slip rate of 0.3 ± 0.1 mm/yr for the Mount Eddy normal fault (Appendix B), though a more thorough investigation of this fault and the Williams Creek normal fault (Appendix A) is required to quantify the total rate of extension in the KMP. The subparallel trace of the CSZ with normal faults in the KMP indicates that these faults are coupled to the earthquake cycle of the CSZ.

The identification of these normal faults is noteworthy due to their location in the Cascadia forearc, a region containing just one other documented normal fault (Lynch et al., 2025). These findings challenge long-standing assumptions that the KMP behaves rigidly and, contrary to geodetic models that predict northwest-southeast oriented contraction and/or regional dextral shear, indicate that deformation in

this region is dominated by east-west extension. I propose that extension in the KMP is best explained by subcretion (Delph et al., 2021) and/or rollback of the Gorda slab (Humphreys and Coblenz, 2007).

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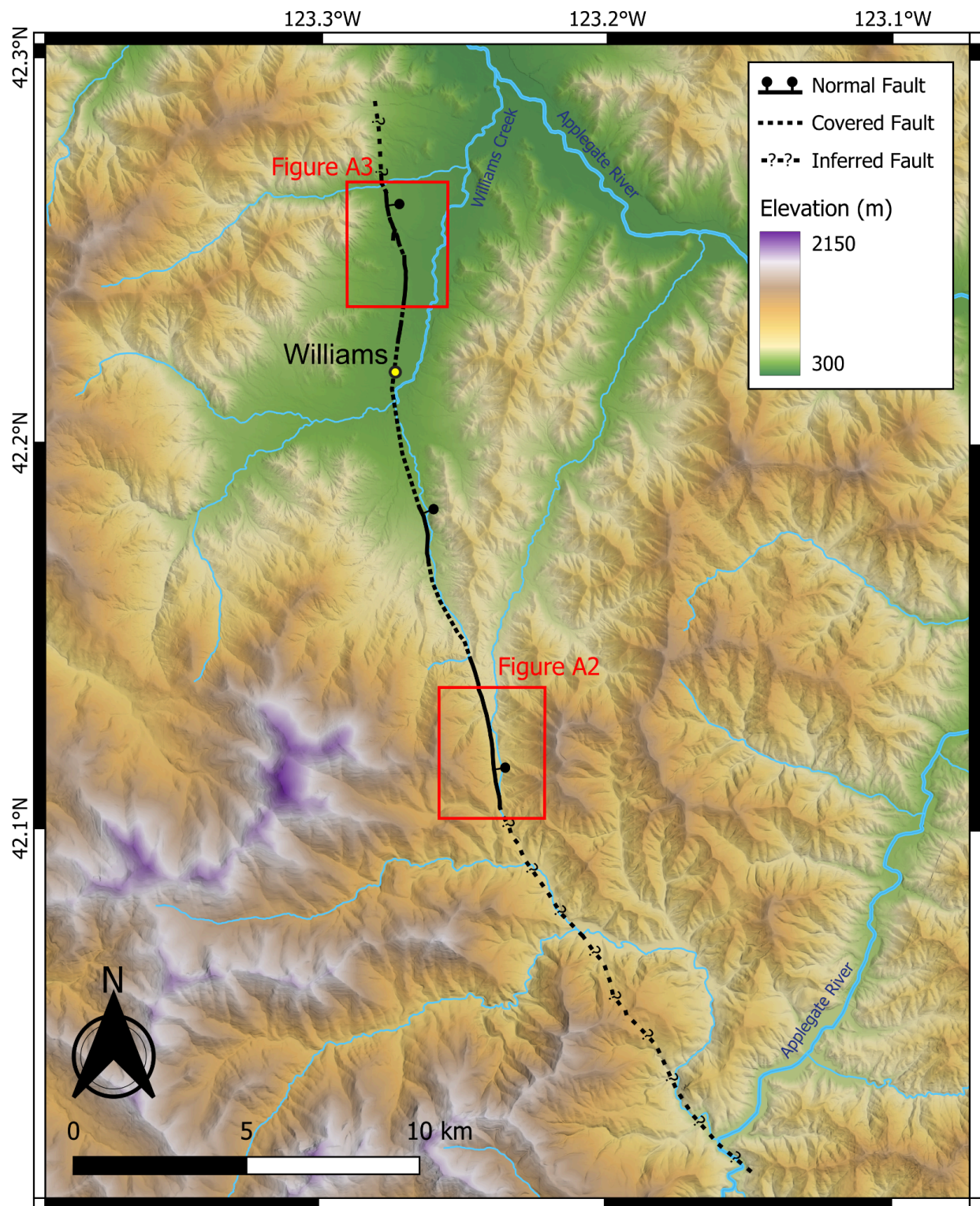
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Appendix A: Williams Creek Normal Fault



The Williams Creek normal fault (WCNF) is a newly identified NNW-striking and east-dipping active fault located in southern Oregon (Figure A1). The fault is named for its proximity to Williams Creek, which flows subparallel to the fault trace along much of its length. Based on lidar analysis, the WCNF possesses a minimum well-constrained length of 19 km and may be as long as 34 km when incorporating inferred extensions to the north and south. Activity along the WCNF is evidenced by a series of aligned fault scarps which have vertically displaced alluvium (Figures A2 & A3). I infer the WCNF to possess a normal sense of slip due to its relatively linear trace and alignment with the SVNf, to its south.

I infer the WCNF to extend an additional 13 km south of its last clearly defined scarp. The trace of this southern extension is coincident with the ancestral east-dipping reverse fault separating the Western Hayfork and Rattlesnake Creek terranes (Irwin, 1994). While a distinct topographic lineament is visible along this trend, there is an absence of clear geomorphic markers preserving recent fault offset. I additionally infer the WCNF to extend an additional 2 km to the north. Along this reach, the inferred fault trace follows a prominent mountain front that defines the contact between the valley alluvium and the bedrock of the footwall. However, I observe no geomorphic evidence for faulting along this 2 km stretch.

Remote surficial mapping along the WCNF reveals displaced alluvial terrace surfaces (Figure A3b). Terrace surfaces are classified by relative depositional age from youngest to oldest as T1 through T7. Based on remote surficial mapping, I identify T5 as the youngest alluvial terrace surface clearly displaced by the WCNF (Figure A3b). No geochronology was performed to constrain the depositional age of features displaced by the WCNF.

Direct field verification of the WCNF was not conducted during this study. Based on the lack of significant topographic relief between the fault hangingwall and footwall, I preliminarily suggest that the WCNF has not yet accumulated substantial long-term offset.

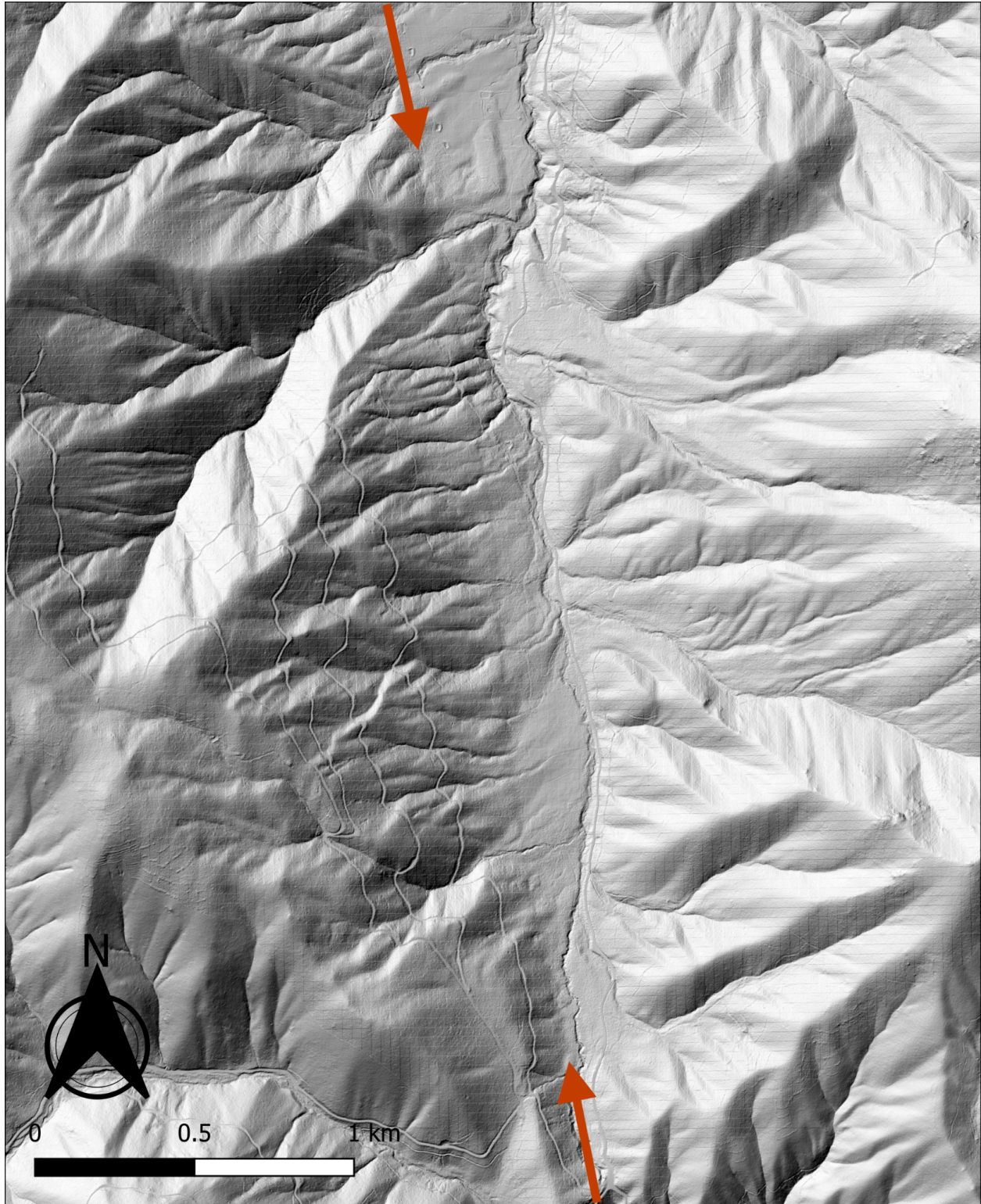


Figure A2. Hillshade highlights the scarp of the Williams Creek normal fault displacing alluvium south of the town of Williams.

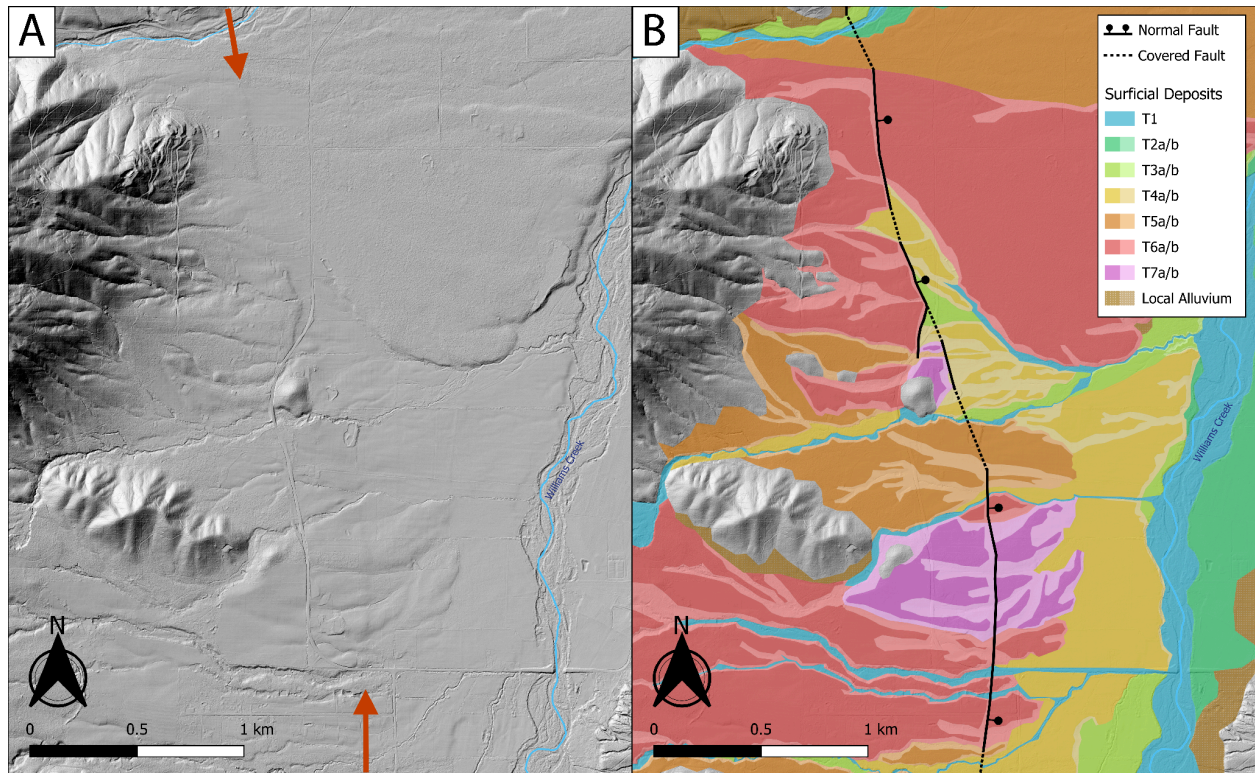


Figure A3. (A) Hillshade highlights the scarp of the Williams Creek normal fault north of the town of Williams. (B) Surficial map of alluvial deposits along the Williams Creek normal fault. Deposits are categorized by depositional age from youngest to oldest from T1 to T7. Units labeled “a” denote relict depositional surfaces while units labeled “b” denote eroded terrace edges. Deposits sourced from minor tributaries noted as Local Alluvium.

Appendix B: Mount Eddy Normal Fault

The Mount Eddy normal fault (MENF) is a newly identified north-striking and east-dipping active normal fault in the eastern KMP approximately 6 km southwest of the city of Weed. I name this fault after Mount Eddy, the highest peak in the KMP situated on the footwall of the fault (Figure B1). Based on lidar analysis, the MENF possesses a minimum well-constrained length of 9.5 km. The MENF marks the base of a mountain front with up to 1500 meters of vertical relief between its uplifted footwall and subsided hangingwall. I infer the MENF possesses a normal sense of slip due to its relatively linear trace and subparallel orientation with the SVNF, to its east.

Activity along the MENF is geomorphically evidenced by fault scarps which have displaced deposits I interpret as LGM-age glacial outwash (Figure B2). Heermance et al. (2026) find a Mid-LGM age of $19.9^{+0.3}_{-1.6}$ ka in the Klamath Mountains.

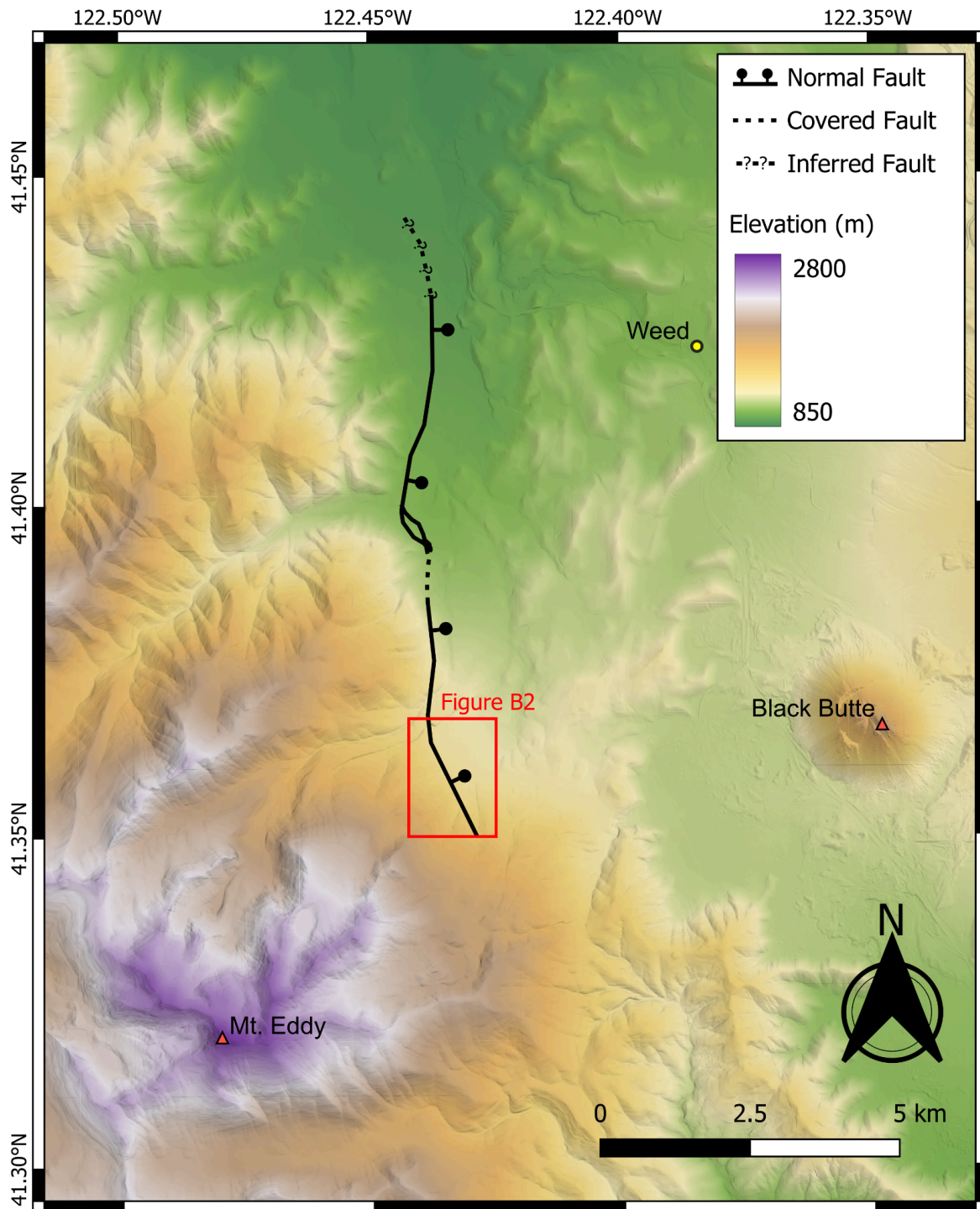


Figure B1. Hillshaded topographic map of the region encompassing the Mount Eddy normal fault.

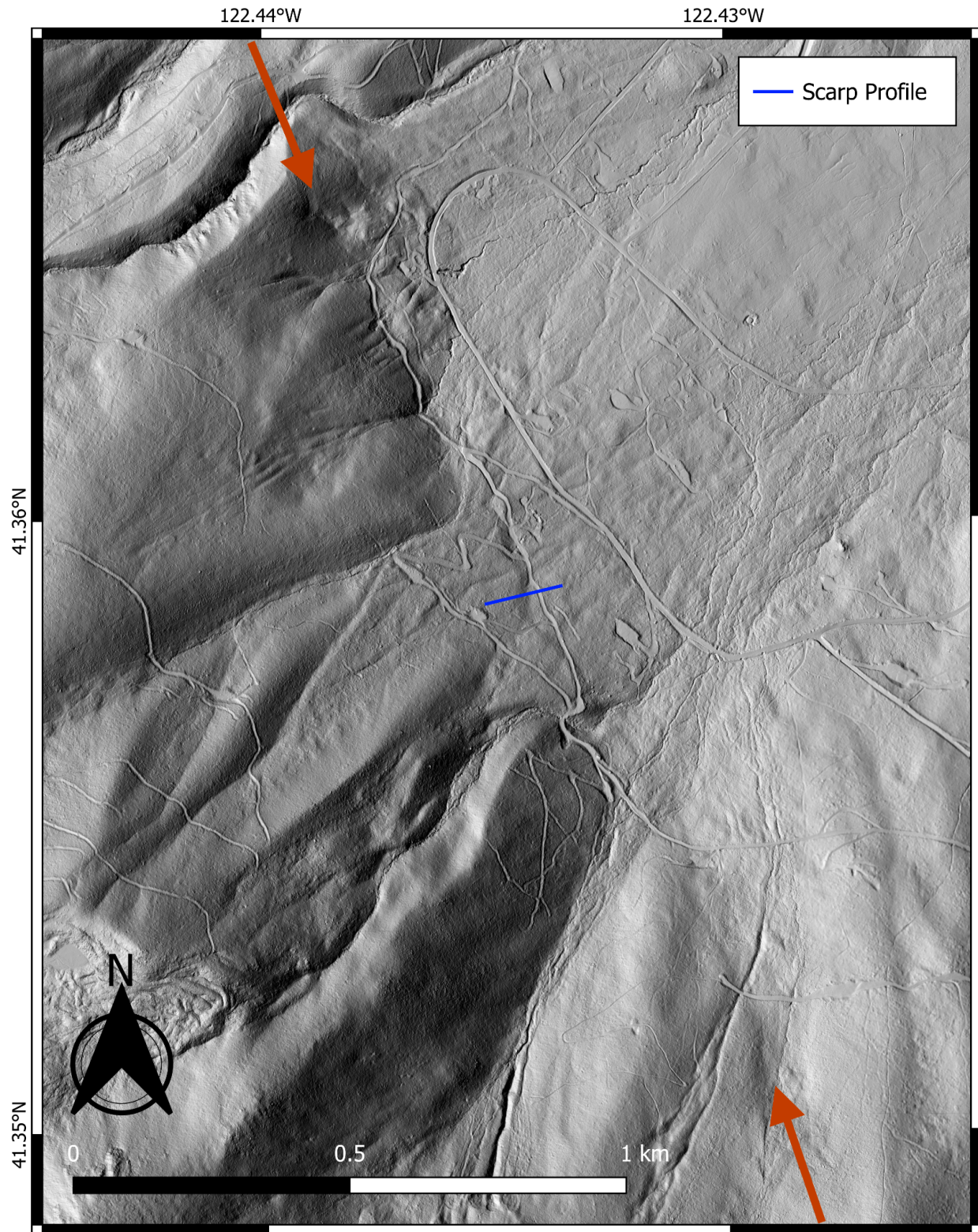


Figure B2: Hillshade highlights the scarp of the Mount Eddy normal fault. Red arrows point to the trace of the Mount Eddy normal fault. Blue line indicates the scarp profile used in displacement analysis (Figure B3).

I analyze a scarp of the MENF to estimate the fault throw and dip-slip displacement of LGM-age deposits at this site. From the scarp profile, I estimate 4.5 ± 0.9 meters of throw and $5.2^{+1.4}_{-1.3}$ meters of dip-slip displacement (Figure B3). Assuming the MENF has displaced $19.9^{+0.3}_{-1.6}$ ka glacial outwash deposits by $5.2^{+1.4}_{-1.3}$ meters indicates a mean dip-slip rate of 0.3 ± 0.1 mm/yr for the MENF. A more thorough investigation of the MENF is required to further constrain its dip-slip rate.

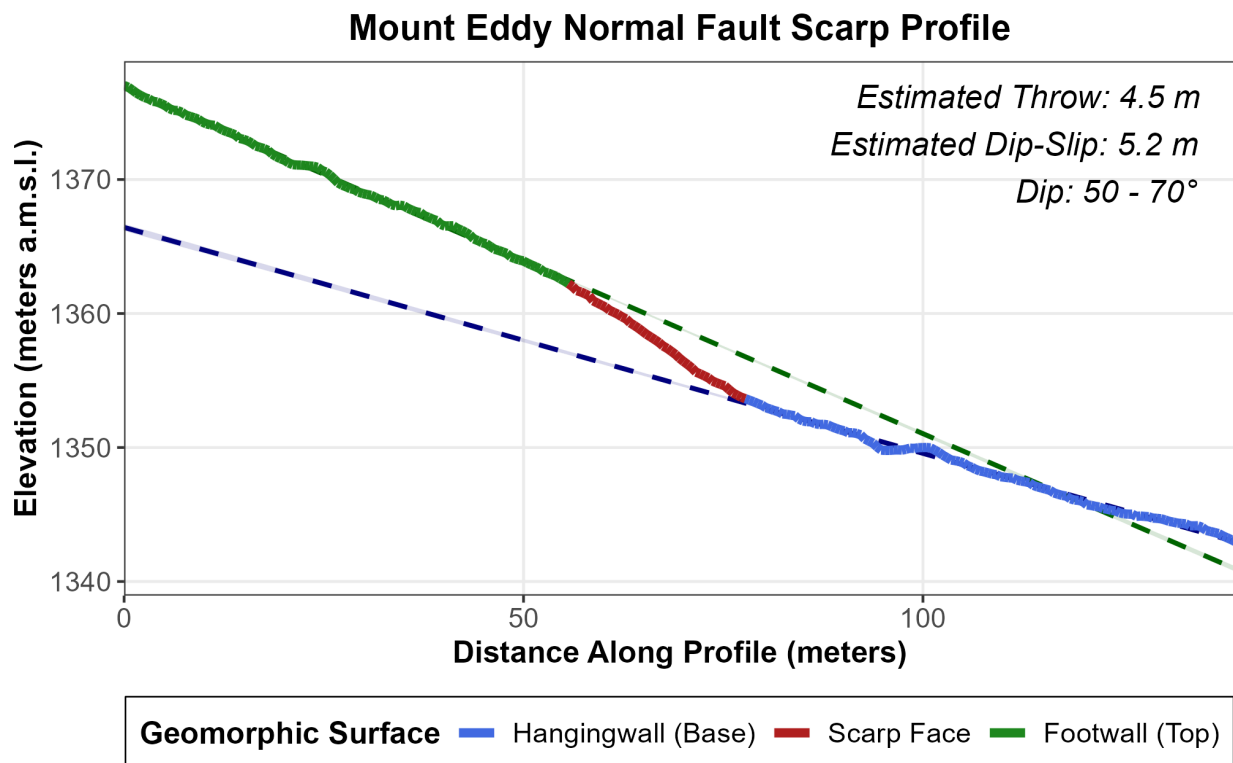


Figure B3. Terrain profile across the scarp of the Mount Eddy normal fault displacing LGM outwash deposits. This scarp preserves an estimated 4.5 ± 0.9 meters of throw and $5.2^{+1.4}_{-1.3}$ meters of dip-slip displacement.

Appendix C: Hyampom Fault

The Hyampom fault is a newly identified dextral strike-slip fault located in the southwestern KMP I name for the adjacent Hyampom valley along the South Fork Trinity River. This linear, west-northwest-striking feature appears to be a dextral strike-slip fault which has deflected and beheaded

several small tributaries along its trace (Figure C1). The uphill facing scarp of the Hyampom fault ponds locally derived alluvium on the northeast side of the fault (Figure C1). Based on lidar analysis, the Hyampom fault possesses a well-constrained length of 6.5 km.

A shallow trench was excavated across the trace of the Hyampom fault to uncover evidence of faulting (Figure C2). This excavation reached a maximum depth of 1 meter and extended 2 meters in length perpendicular to the fault strike. The exposure revealed four distinct units, from top to bottom: a soil containing rounded clasts of the Miocene Weaverville Formation (Strand, 1963; Irwin, 2010; Michalak et al., 2024), a clay-rich layer, a buried paleosol, and weathered bedrock (Figure C2).

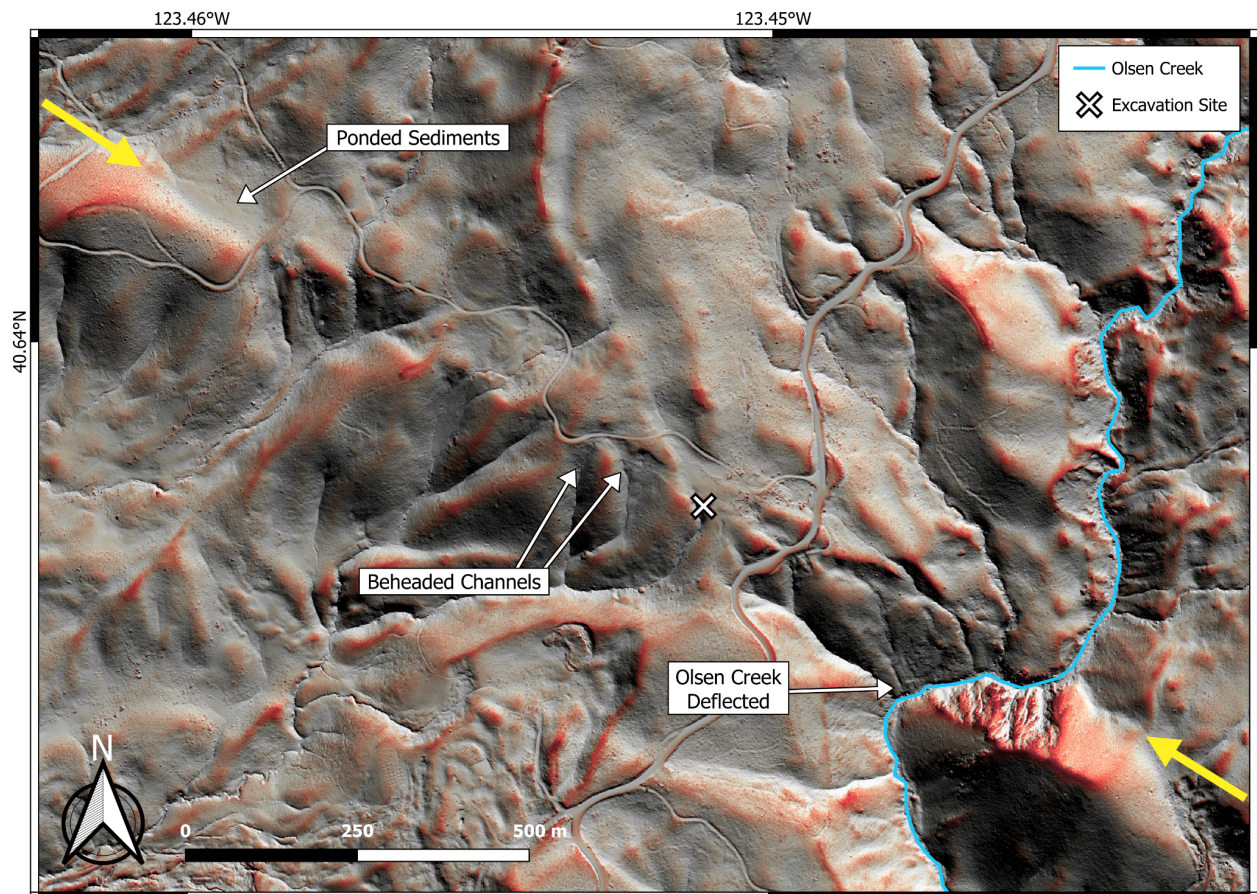


Figure C1. Hillshade of the Hyampom fault near Olsen Creek with red-shade differential openness raster overlay (Yokoyama et al., 2002; Chiba et al., 2008). Yellow arrows point to the trace of the Hyampom fault. The location of the excavation site (Figure C2) marked. Note ponded sediments, beheaded channels, and right-deflected channels which indicate right-lateral slip.

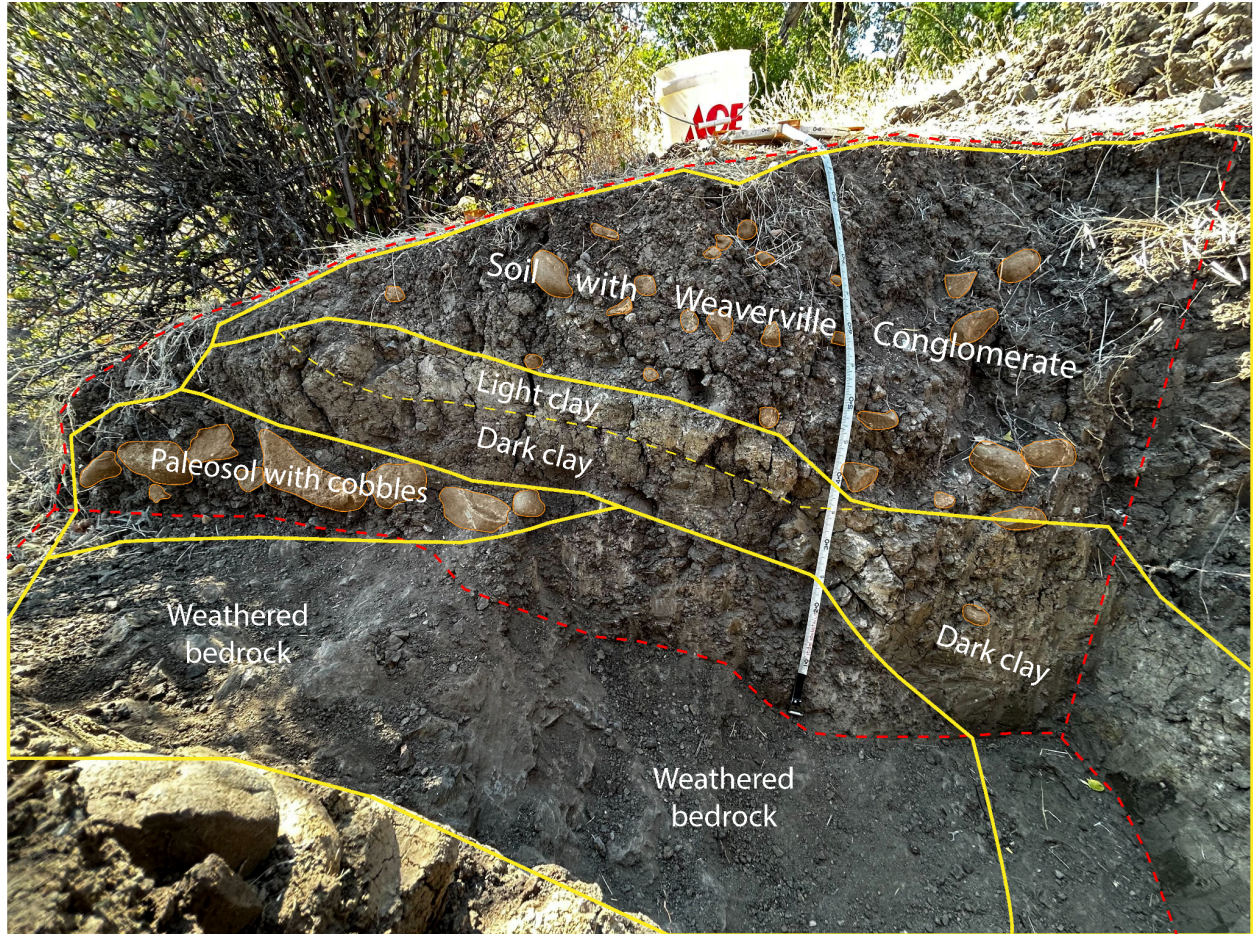


Figure C2: Labeled and interpreted field photo Hyampom fault excavation with measuring tape for scale (cm), facing west-northwest. Yellow lines denote unit contacts. Dashed red lines denote the geometry of the excavation cross-section. Note near-surface thrusting of soil containing reworked Weaverville Formation conglomerate over paleosol and bedrock. The clay-rich layer marks the plane of the Hyampom fault, which steepens to the north with depth.

I interpret the uppermost unit revealed by this excavation as a soil containing reworked clasts of the Weaverville Formation. I base this interpretation on the high degree of rounding exhibited by the clasts, which is inconsistent with the limited transport distance of modern local streams. The paleosol unit at this site is distinguished from the Weaverville Formation by relatively larger, sub-rounded to sub-angular clasts up to 30 cm in diameter I interpret as locally sourced. Separating these units is a curved band of north-dipping sticky clay that thickens and steepens with depth. This clay layer is present across the excavated section and can be segregated into relatively lighter and darker bands. A thin (~10 cm) band

of lighter clay occurs near the surface and tapers out at a depth of 50 cm. A darker clay band thickens to 30 cm at the bottom of the excavation.

Though I interpret the Hyampom fault as primarily strike-slip along its length, the observed structural juxtaposition of units indicates a local component of near-surface north-side-up reverse motion which has overthrust a soil containing reworked Weaverville conglomerate above bedrock and a now-buried paleosol. I interpret the clay-rich layer to represent the fault zone of the Hyampom fault, which dips gently near the surface and steepens with depth. I expect that this steepening of the fault plane continues in the subsurface until the fault reaches a near-vertical dip.

Appendix D: Illinois River Valley Fault

An additional active fault, the Illinois River Valley fault (IRVF), resides in the northwest KMP in Oregon. This fault was recognized by Yule (1996) as a southwest-striking, westward dipping thrust and reverse fault with quaternary offset. The IRVF marks the base of the mountain front bounding the Illinois River Valley on its west.

Yule (1996) proposes that the IRVF has experienced Quaternary, and possibly Holocene, offset expressed by the presence of 1) broad alluvial plains in the footwall, 2) several levels of bench river gravels in the hangingwall elevated as much as ~800 feet (~240 m) above the present streams, and 3) a mountainous hangingwall adjacent to a flat, alluvium-covered footwall. Yule (1996) expects the IRVF is a reverse, rather than normal, fault due to structural data and the occurrence of deeper level rocks on fault hangingwall.

Though I observe no direct geomorphic evidence of recent fault slip along the IRVF, lidar analysis confirms the evidence presented by Yule (1996) to suggest that the mountains west of the fault are upthrown relative to the eastern valley. Though Yule (1996) proposes a reverse sense of slip on the IRVF, the dip and dip direction of this fault is poorly constrained. Due to this, and the presence of three similarly striking normal faults in the region, the IRVF may be a normal fault. The IRVF is presently being studied by E. Kirby and colleagues (Kirby et al., 2020; Kirby et al., 2023).

Appendix E: IRSL Geochronology Laboratory Procedures

All laboratory procedures were conducted under dim red (> 660 nm) light specific to luminescence preparation and analysis.

Equivalent dose

Preparation

A small portion of the sample was reserved for dose rate and water content measurements. The remaining portion of the sample was prepared for analysis as follows:

- Treatment with sodium pyrophosphate decahydrate ($\text{Na}_4\text{P}_2\text{O}_7 \cdot 10\text{H}_2\text{O}$) to deflocculate clays.
- Isolation of the target grain size fraction by wet sieving.
- Treatment with 10% HCl to remove carbonates.
- Treatment with 30% H_2O_2 to remove organics.
- Separation of magnetic sub-fraction using a hand magnet.
- Mineral density separation with heavy liquid (lithium heteropolytungstate) to isolate target minerals:
 - K-feldspar: $\rho < 2.58 \text{ g/cm}^3$
- Grains were mounted using silicone adhesive sprayed through a 2 mm mask.

Instrument configuration for measurement

All samples were measured on a Lexsyg Research Instrument using continuous wave measurements of the luminescence signal. IRSL measurements were conducted using infrared (IR) laser diodes with peak emission at 850 nm and power of 145 mW/cm^2 at the sample position. Signals were detected through a filter with a 365 nm detection window using a H7360-02 (UV/VIS) photomultiplier tube. Artificial irradiation was administered using a calibrated ^{90}Sr beta source integrated into the Lexsyg Research Instrument.

Sample measurement and analysis

All measurements were conducted on small multi-grain aliquots as noted above. The equivalent dose (D_e) of all samples was measured using a modified single aliquot regenerative-dose (SAR)

technique (Murray and Wintle, 2000, 2003) for feldspar (Wallinga et al., 2000). The dating experiments were conducted using the measurement parameters outlined in Table E1 that were determined on the basis of dose recovery tests.

Table E1. The SAR measurement protocol used in this study¹.

Step	Treatment
1	Natural/Regenerative Dose
2	Preheat (160°C, 10 s)
3	IR diodes (50°C, 200 s) → Ln, Lx
4	Test dose (~10 Gy)
5	Preheat (160°C, 10 s)
6	IR diodes (50°C, 200 s) → Tn, Tx
7	IR diodes (180°C, 100 s)
8	<u>Return to step 1.</u>

¹Ln = natural signal, Lx = regenerative dose signal. Tn = test dose signal measured after Ln, Tx = test dose signal measured after Lx. All steps of the protocol constitute one SAR cycle. The first SAR cycle measures the sensitivity corrected natural signal, and subsequent SAR cycles measure the Lx/Tx value after a series of successively increasing laboratory radiation doses (regeneration doses) administered to the sample. The regeneration doses are used to generate a dose-response curve (where ‘x’ values are the given radiation doses, and ‘y’ values are measured Lx/Tx signals) onto which the natural signal (Ln/Tn) is plotted to calculate the D_e value. Regeneration doses include: 1) one “zero-dose” point where no radiation dose is given to check for recuperation, and 2) one “repeat dose” point, where a previous regeneration dose is given a second time to perform a recycling ratio test.

An exponential fit was used to fit dose response curves for each aliquot. Routine screening criteria included rejection of aliquots that exhibited the following behavior:

- Failure to reproduce, to within 20%, the same sensitivity-corrected luminescence signal from identical regeneration doses given at the beginning and end of the SAR sequence, which suggests inaccurate sensitivity correction (recycling test).
- A test dose error of >20%.
- Recuperation (i.e., the Lx/Tx value measured after the “zero-dose” point regenerative dose) that is >10% of the sensitivity-corrected natural signal (Ln/Yn).

- Poor signals as judged from net natural test dose (Tn) signals that are less than three times the standard deviation of the background signal.

De modelling

A total of 30 aliquots were measured for each sample. Abanico plots (Figure E1) were used to examine the D_e distribution for evidence of partial bleaching or post-depositional mixing and to determine the most appropriate age model to use for age calculation. Both the Central Age Model and Minimum Age Model (Galbraith et al., 1999) D_e values were calculated and are reported in Table 2.

Anomalous fading tests

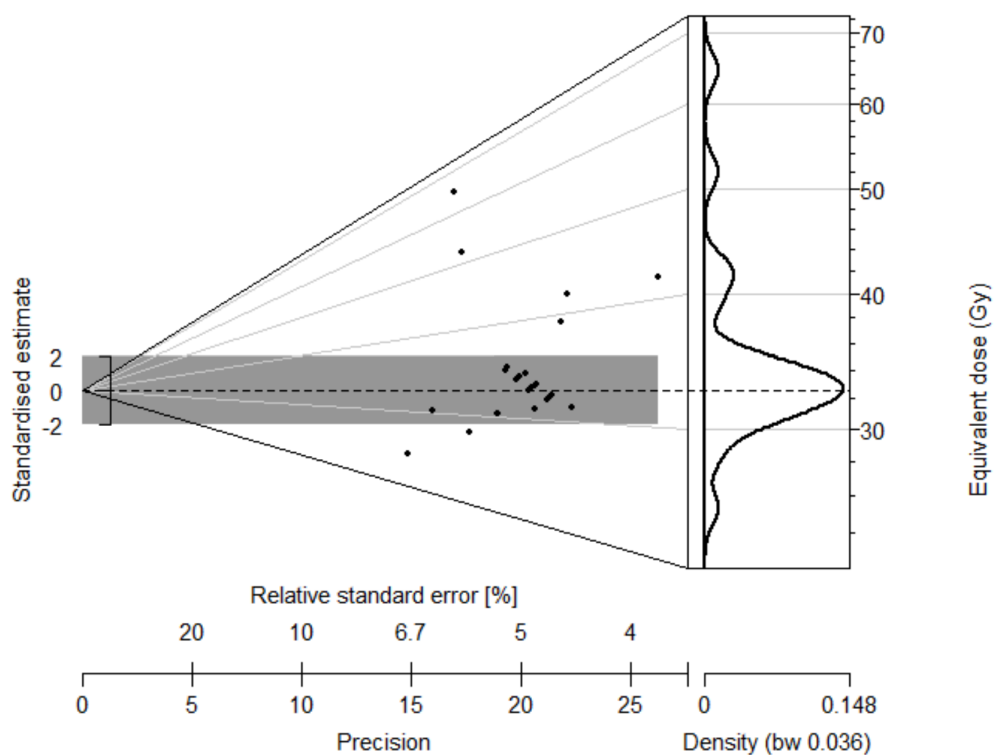
Because the IRSL signal from feldspar is known to fade over geological time periods, this fading rate must be measured and corrected for (Huntley and Lamothe, 2001). Fading rates for the IR50 signal were measured over the course of ~24 hours on 12 aliquots of each sample using the approach of Auclair et al. (2003) after D_e measurements were completed. Final ages were corrected using the weighted mean fading rate for all samples ($2.96 \pm 0.29\%$ /decade; $t_c = 2$ days) and the correction model of Huntley and Lamothe (2001).

Dose rate and age calculations

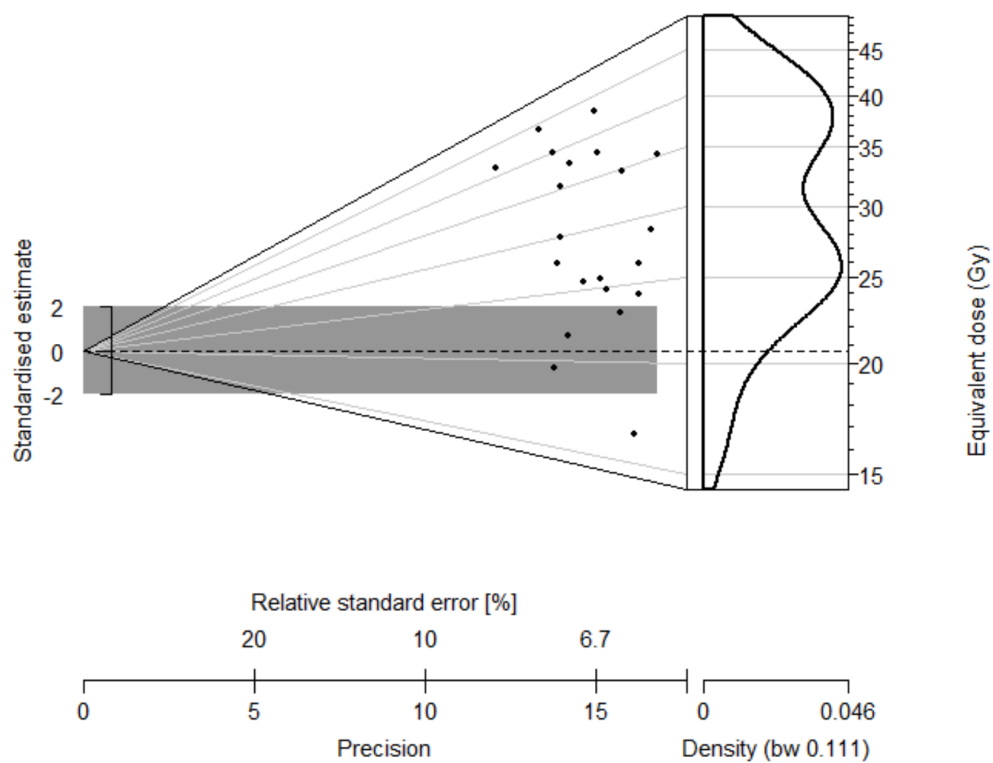
Samples for dose rate measurement were dried and milled to a fine, flour consistency and sent to ALS Geochemistry for geochemical analysis of U, Th, and K_2O . Subsamples used for U and Th measurement were fused with lithium borate and measured with ICP-MS. K_2O was measured from the bulk sample with ICP-AES and converted to %K. Dose rates (Gy/ka) were calculated using the conversion factors of Liritzis et al. (2013). Dose rate calculations assume radioactive equilibrium in the ^{238}U and ^{232}Th decay chains. Alpha dose rates assume an alpha efficiency value of 0.15 ± 0.05 based on the work of Balescu and Lamothe (1994). Beta dose attenuation factors were based on Guérin et al. (2012). Gamma dose rates were estimated from *in-situ* gamma spectrometry measurements (made using a calibrated Canberra Inspector 1000). All sample dose rates were adjusted for estimated moisture content using the average laboratory measured gravimetric moisture content of all samples ($5\% \pm 5\%$). Cosmic

dose rates (Gy/ka) were calculated according to Prescott and Hutton (1994). Dose rate and final age calculations were made using DRAC (Tables 2 & 3; Durcan et al., 2015) and include an internal dose rate component based on an internal K content of $10 \pm 2\%$ (Smedley et al., 2012). Ages reported in Table 2 are expressed as thousands of years before 2025 and rounded to the nearest 10 years.

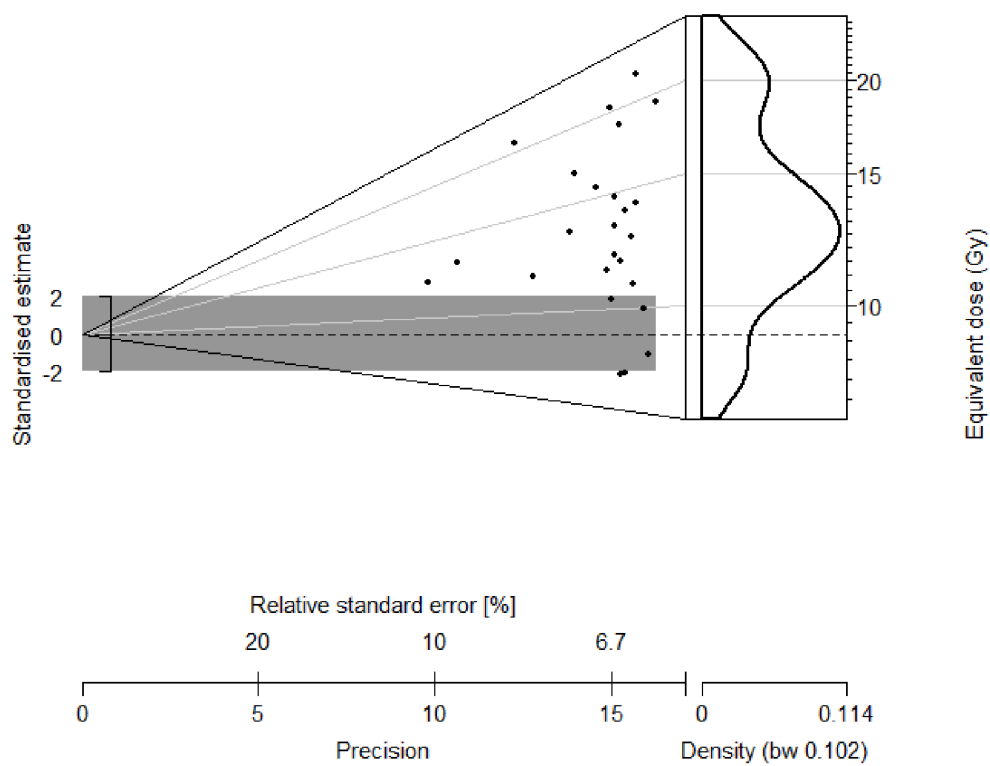
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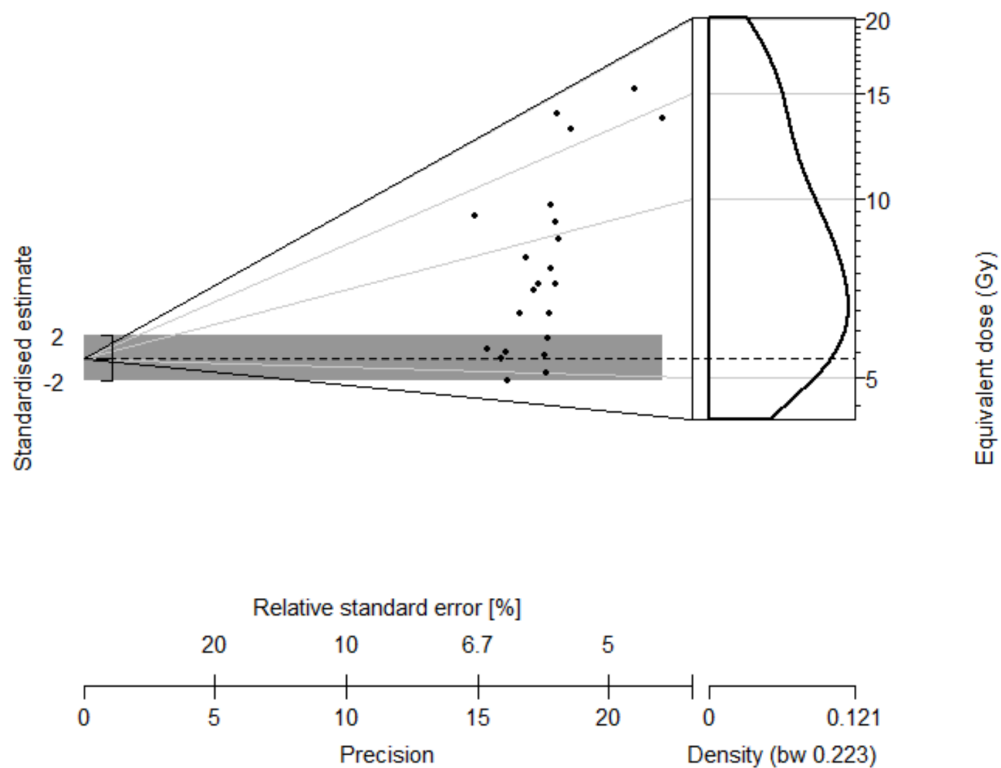
SF002



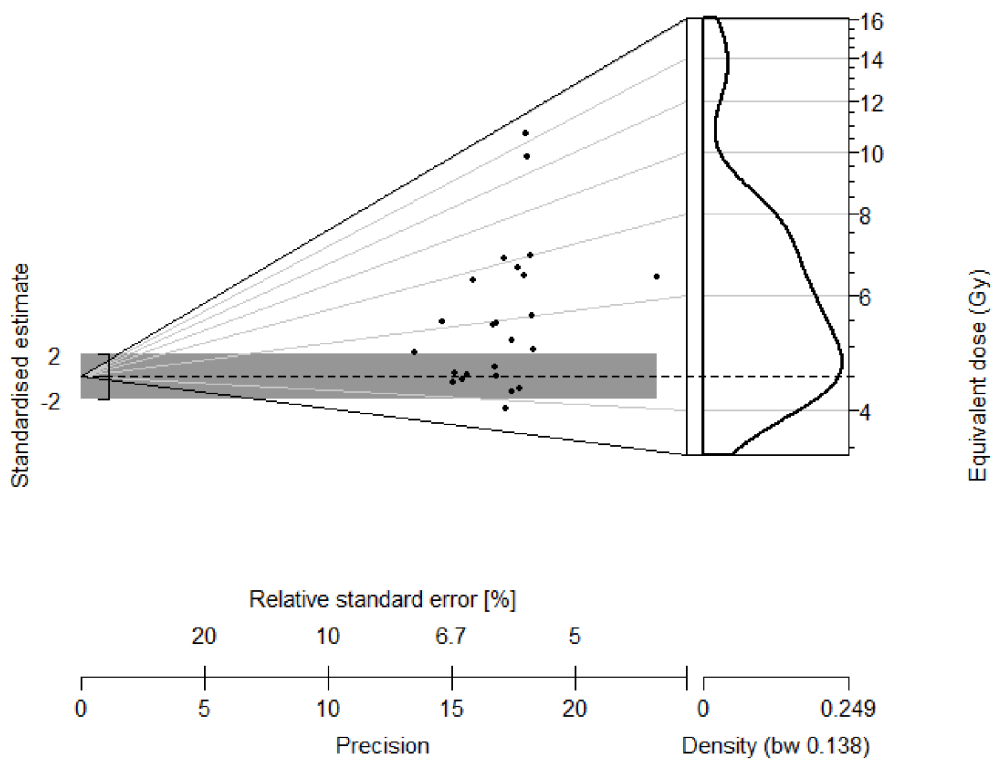
SF003



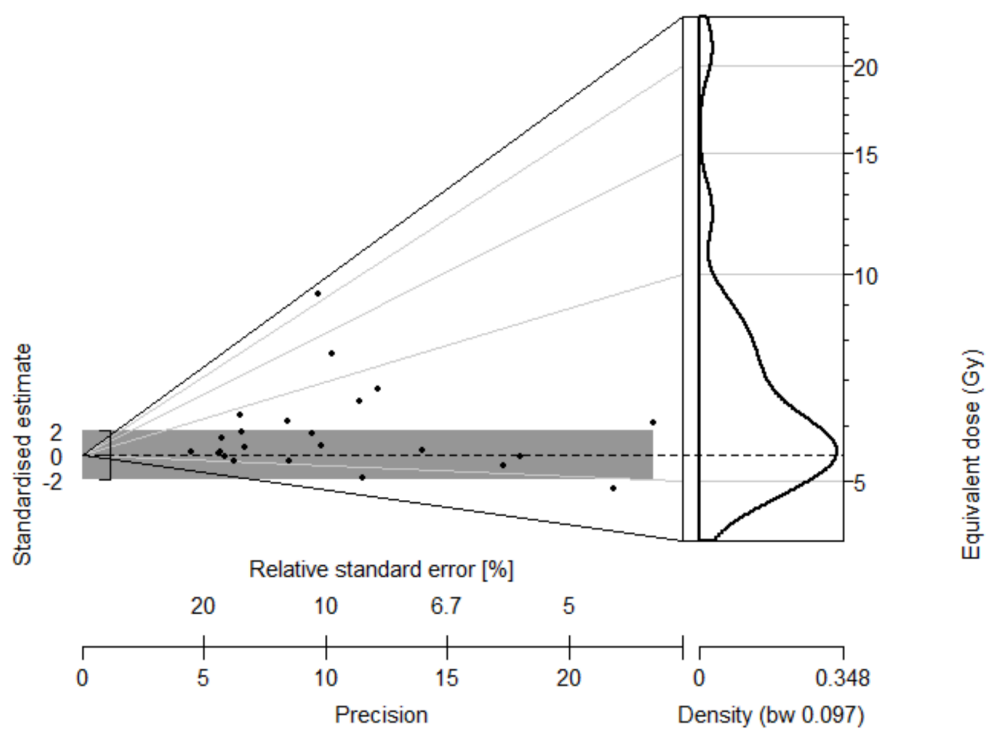
SF004



SF005



SF006



SF007

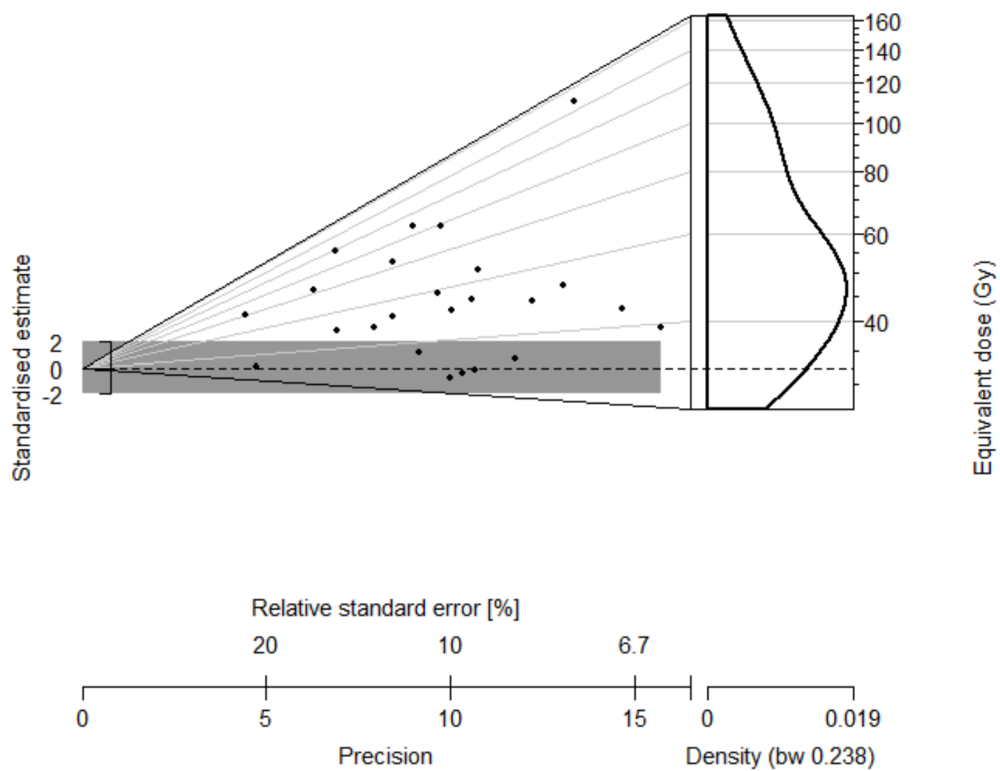


Figure E1. Abanico plots showing measured D_e populations for all samples. The dashed horizontal lines denote the Minimum Age Model (MAM) D_e estimate, with 2σ confidence intervals shown as shaded bands.