

1 **Tectonic Evolution of the Condrey Mountain Schist: an**  
2 **Intact Record of Late Jurassic to Early Cretaceous**  
3 **Franciscan Subduction and Underplating**

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10 **Key Points:**

- 11 • The Condrey Mountain Schist (CMS) epidote-blueschist facies rocks were deposited  
12 and underplated in the Late Jurassic to Early Cretaceous.  
13 • Underplating postdates Klamath terrane assembly and predates coherent Fran-  
14 ciscan underplating, recording the early Franciscan history.  
15 • CMS pressure-temperature conditions are consistent with other estimates of early  
16 Franciscan subduction zone thermal structure.

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**Abstract**

The Klamath Mountains in northern California and southern Oregon are thought to record 200+ m.y. of subduction and terrane accretion, whereas the outboard Franciscan Complex records classic ocean-continent subduction along the North American margin. Unraveling the Klamaths' late history could help constrain this transition in subduction style. Key is the Mesozoic Condrey Mountain Schist (CMS), comprising, in part, a subduction complex that occupies a structural window through older, overlying central Klamath thrust sheets but with otherwise uncertain relationships to other, more outboard Klamath or Franciscan terranes. The CMS consists of two units (upper and lower), which could be correlated with 1) other Klamath terranes, 2) the Franciscan, or 3) neither based on regional structures and limited extant age data. Upper CMS protolith and metamorphic dates overlap with other Klamath terranes, but the lower CMS remains enigmatic. We used multiple geochronometers to constrain the timing of lower CMS deposition and metamorphism. Maximum depositional ages (MDAs) derived from detrital zircon geochronology of metasedimentary rocks are 153-135 Ma. Metamorphic ages from white mica K-Ar and Rb-Sr multi-mineral isochrons from intercalated and coherently deformed mafic lenses are 133-116 Ma. Lower CMS MDAs (<153 Ma) predominantly postdate the age of other Klamath terranes, but subduction metamorphism appears to predate the earliest coherent Franciscan underplating (ca. 123 Ma). The lower CMS thus occupies a spatial and temporal position between the Klamaths and Franciscan and preserves a non-retrogressed record of the Franciscan Complex's early history (>123 Ma), otherwise only partially preserved in retrogressed Franciscan high grade blocks.

**1 Introduction**

The Klamath Mountains of northern California and southern Oregon expose rocks that record protracted subduction from the Early Devonian to the Late Jurassic (Snoko & Barnes, 2006). The rocks of the Klamaths are bounded on the west by the younger Franciscan Complex, which records subduction from the Middle Jurassic to the Eocene (Fig. 1a-b) (e.g., Bailey et al., 1964; Evitt & Pierce, 1975; Dumitru et al., 2010; Morisani, 2006; Hopson et al., 2008; Shervais & Choi, 2012), and on the east and south by a Cretaceous to Tertiary onlap sequence (Nilsen, 1984; Snoko & Barnes, 2006). The Klamaths consist primarily of broadly eastward-dipping, westward-younging thrust sheets that represent suprasubduction zone terranes telescoped onto the North American margin (e.g., Saleeby, 1990; Hacker et al., 1995; Snoko & Barnes, 2006). Because terrane accretion plays a key role in long-term growth of continental crust, significant effort has gone into unraveling the complex tectonic history of the Klamaths, including, for example, the origin of ophiolitic terranes (Wright & Wyld, 1994; Gray, 1986; Yule et al., 2006; Harper et al., 1994), the ages and styles of arc magmatism (Allen & Barnes, 2006; Barnes et al., 2006; Bushey et al., 2006; Harper, 2006; McFadden et al., 2006), and the timing and mechanisms of terrane accretion and continental growth (Helper, 1986; Saleeby & Harper, 1993; Hacker et al., 1995; Gray, 2006; Pessagno, 2006; Snoko & Barnes, 2006) (Fig. 1b). The timing of deposition, subduction, and metamorphism of the structurally lowest unit in the central Klamaths, the Condrey Mountain Schist (Figs. 1b and 2), however, remains enigmatic.

How the Condrey Mountain Schist (CMS) fits into the complex regional tectonics of the Klamaths and the outboard Franciscan Complex is currently unresolved. The CMS principally occupies a large structural window through older greenschist to amphibolite facies Klamath terranes (Fig. 1c) (Helper, 1985, 1986; Saleeby & Harper, 1993; Snoko & Barnes, 2006), but subordinately also occurs within a small thrust sheet beneath these same rocks in the Klamath River canyon about 15 km west of the window (Hill, 1985). It consists of a greenschist- to epidote-amphibolite-facies unit (upper CMS), structurally underlain by epidote-blueschist facies rocks (lower CMS) that record the progressive underplating at 30-40 km depth of dominantly oceanic-affinity sedimentary protoliths,

with m- to km-scale lenses of mafic and ultramafic protoliths (Fig. 2) (Helper, 1986; Tewksbury-Christle et al., 2021). Previous researchers connected the CMS to forearc- and arc-related terranes in the Klamaths (Fig. 1b) based on regional relationships (e.g., Saleeby & Harper, 1993). However, previously published or cited emplacement or metamorphic ages for the lower CMS range from  $167 \pm 12$  Ma to  $118 \pm 2$  Ma (Hacker et al., 1995; Saleeby & Harper, 1993; Helper, 1986; Coleman et al., 1983), overlapping in time with both the youngest Klamath terranes and the oldest Franciscan units, therefore providing poor constraints on regional correlations and relationships. Resolving this issue will allow better constraints on the transition from outboard subduction and terrane accretion to the unimpeded ocean-continent subduction recorded by the Franciscan Complex.

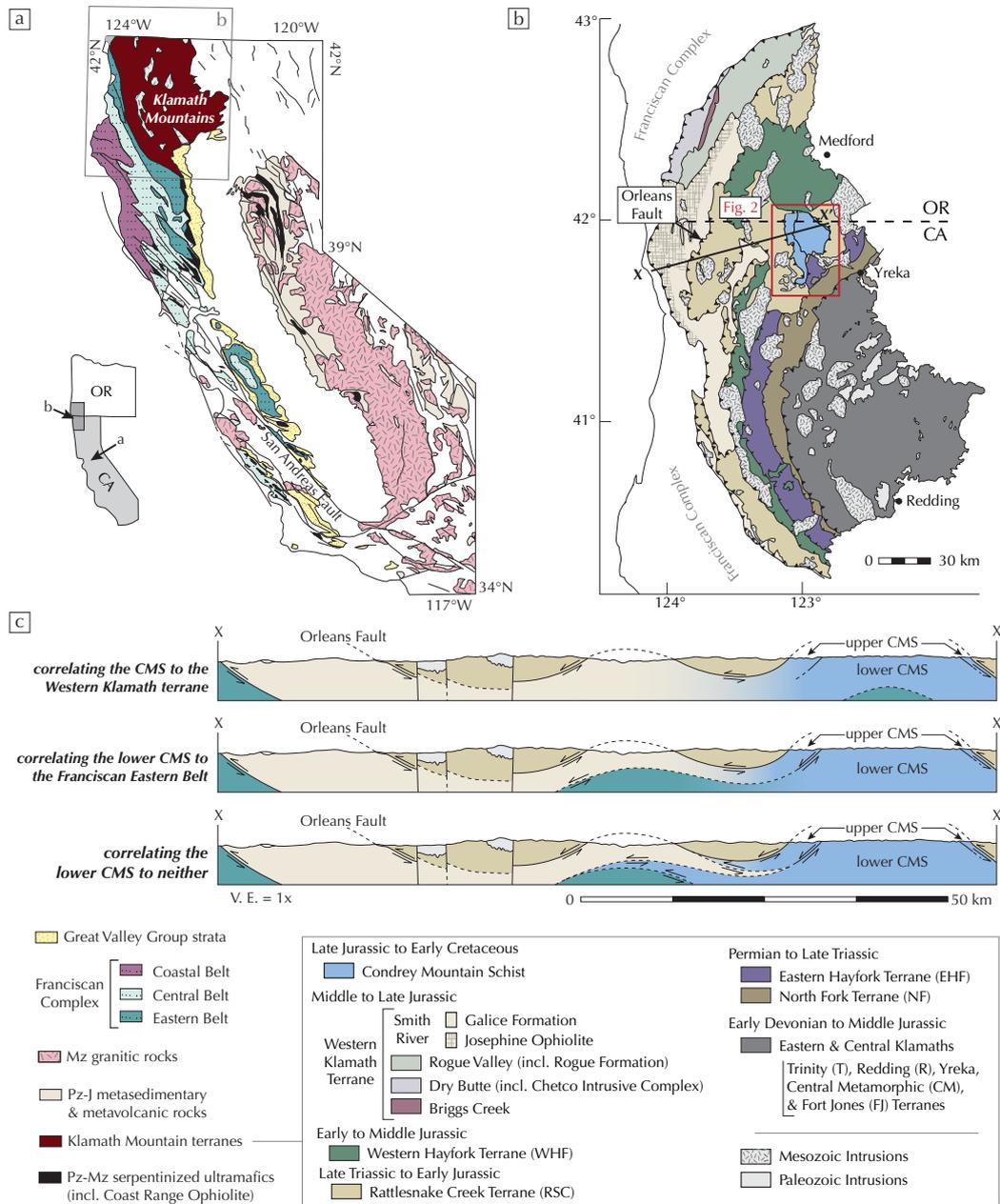
In order to resolve the tectonic evolution of the CMS, we employ multiple geochronometers to constrain the provenance and depositional ages (where applicable) of different protoliths within the lower CMS, along with the timing of metamorphism as a function of structural depth over the full thickness of the exposed lower CMS. We use these new datasets to place constraints on the timing of lower CMS deposition, subduction, and underplating, and we discuss the broader regional implications for the Klamaths and Franciscan Complex in the context of Western North America Cordilleran convergent tectonics.

## 2 Tectonic Setting

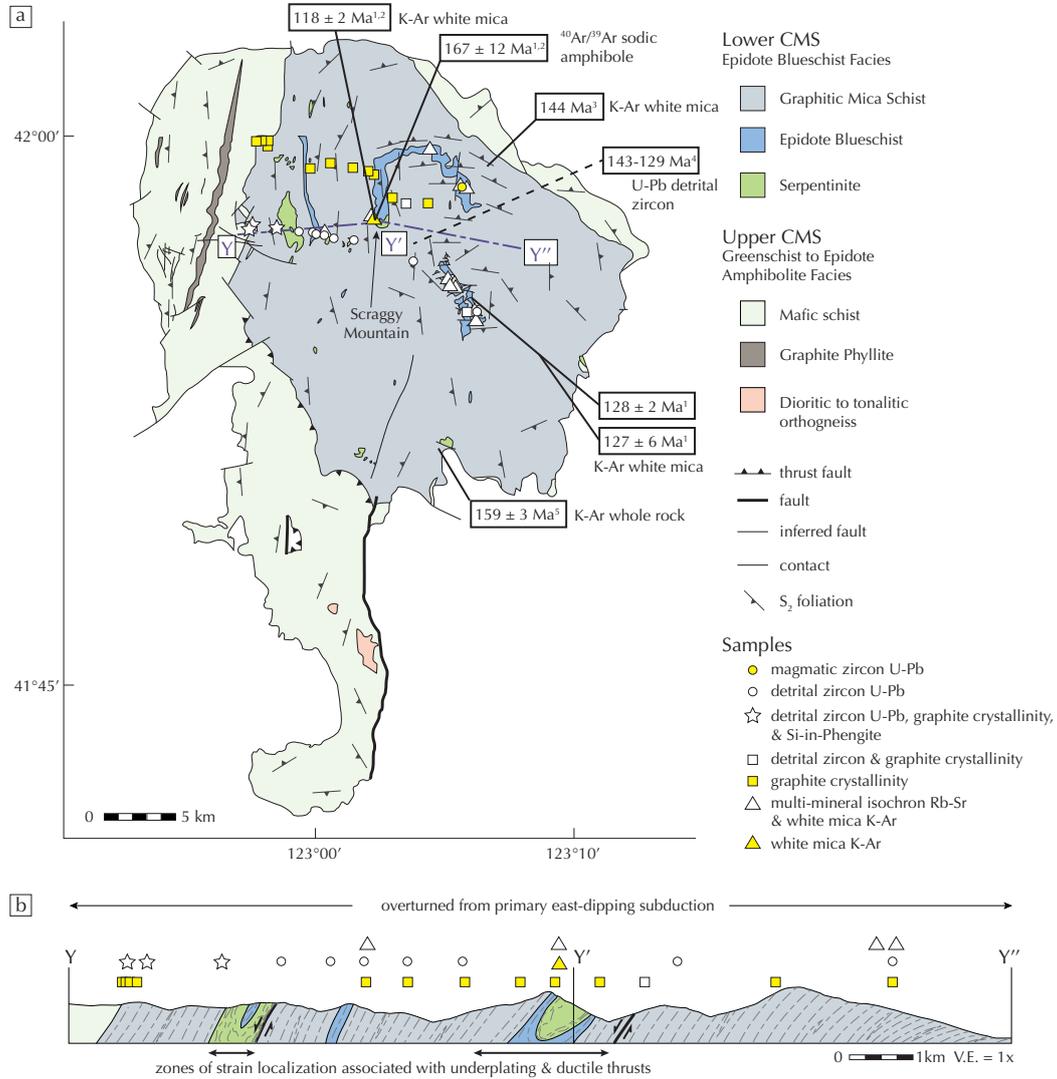
### 2.1 Klamath Mountains

The Klamath Mountains record Early Devonian through Early Cretaceous subduction and subsequent accretion of multiple fringing island arc systems onto the western margin of North America (Fig. 1b) (e.g., Snoke & Barnes, 2006; Saleeby, 1990; Hacker et al., 1995; Saleeby & Harper, 1993; Helper, 1986; Irwin, 1972). These terranes include arc-related units, forearc and backarc basinal deposits, ophiolites, and fossil accretionary wedges (e.g., Gray, 1986; Saleeby, 1990; Saleeby & Harper, 1993; Wright & Wyld, 1994; Hacker et al., 1995; Gray, 2006; Pessagno, 2006; Snoke & Barnes, 2006; Yule et al., 2006). Timing of formation and/or deposition of the Klamath terranes is well-constrained, as is motion on the suturing thrust faults, but similar age constraints do not exist for the CMS.

The CMS occupies a similar structural position as the Western Klamath terrane, with the majority of the window-bounding faults placing the Rattlesnake Creek terrane on top of the CMS, much like the Orleans fault places the Rattlesnake Creek on top of the Western Klamath terrane (Fig. 1b-c). The Rattlesnake Creek terrane (with components that range in age from 161-300 Ma) is floored by a Late Triassic ophiolitic melange, locally at amphibolite facies, with thin volcanic cover sequences that are overlain by the Western Hayfork arc volcanics (169-179 Ma) (Wright, 1982; Gray, 1986; Hacker et al., 1995; Snoke & Barnes, 2006; Barnes et al., 2006; Frost et al., 2006; LaMaskin et al., 2021). The Rattlesnake Creek/Western Hayfork terranes sutured to the inboard Klamath terranes ca. 170 Ma (Saleeby, 1990; Snoke & Barnes, 2006) and underwent transtensional forearc spreading that resulted in formation of the Josephine Ophiolite (161-165 Ma) (Saleeby & Harper, 1993; Hacker et al., 1995; Snoke & Barnes, 2006; Yule et al., 2006). The Galice Formation flysch (part of the Western Klamath terrane; Figure 1b) was deposited in the Josephine basin following spreading (151-158 Ma) (Frost et al., 2006; Macdonald et al., 2006; LaMaskin et al., 2021; Surpless et al., 2023). Onlap and interfingering of the Galice with the Rogue-Chetco arc volcanics (155-160 Ma) indicate a transition of the active arc from the Western Hayfork to a location farther outboard as a consequence of forearc spreading (Snoke & Barnes, 2006; Yule et al., 2006), transitioning the Josephine-Galice basins from the forearc to the backarc. The Western Klamath terrane was thrust under the Western Hayfork and Rattlesnake Creek terranes along the Orleans fault (ca. 153-150 Ma, well-constrained by rootless and cross-cutting plutons) (Snoke & Barnes, 2006; Frost et al., 2006; Saleeby, 1990). Although plutonic activity continued until ca.



**Figure 1.** a) Regional geologic map of the Klamaths and the younger, outboard accretionary wedge-forearc-arc system preserved in the Franciscan Complex, Great Valley Group, and Sierra Nevada (after Ernst, 2015). b) Geologic map of the Klamath terranes (after Snoke & Barnes, 2006; Tewksbury-Christle et al., 2021). c) Cross-sections along X-X' (after Saleeby & Harper, 1993) illustrating three possible correlations of the CMS based on regional relationships and limited extant age constraints.



**Figure 2.** Geologic map (a) and cross section (b) of the CMS showing sample locations for magmatic and detrital zircon U-Pb, white mica K-Ar, multi-mineral isochron Rb-Sr, graphite crystallinity, and Si-in-phenigite (after Tewksbury-Christle et al., 2021). We projected samples onto the cross section where necessary using the pervasive S<sub>2</sub> foliation. Previously published CMS dates are from Helper (1986)<sup>1</sup>, Coleman et al. (1983)<sup>2</sup>, Lanphere et al. (1968)<sup>3</sup>, Chapman et al. (2021)<sup>4</sup> taken across a transect, and Suppe and Armstrong (1972)<sup>5</sup>.

120 136 Ma and migrated progressively eastward (Allen & Barnes, 2006), the Orleans fault  
 121 has traditionally been interpreted as the final suturing event of the Klamath terranes (e.g.,  
 122 Snoko & Barnes, 2006, and references therein).

123 Eastward subduction along the western margin of North America continued after  
 124 emplacement of the above-mentioned Klamath terranes, as recorded in the primarily younger,  
 125 outboard Franciscan Complex and the associated forearc deposition (Great Valley Group)  
 126 and arc magmatism (Sierra Nevada batholith) (Fig. 1a) (e.g., Bailey et al., 1964; Evitt  
 127 & Pierce, 1975; Dumitru et al., 2010; Morissani, 2006; Wakabayashi, 2015; Hopson et al.,  
 128 2008; Shervais & Choi, 2012; Hamilton, 1969; Orme & Surpless, 2019). The Franciscan  
 129 Complex transitions from east to west from coherently underplated, lawsonite-blueschist  
 130 facies sedimentary and mafic rocks (Eastern Belt) (e.g., Jayko & Blake Jr, 1986; Wak-  
 131 abayashi & Dumitru, 2007; Dumitru et al., 2010; Schmidt & Platt, 2018) to a mélangé  
 132 unit consisting of high grade blocks within a low grade shaley matrix (Central Belt) (e.g.,  
 133 Cloos, 1983; Ukar, 2012; Platt, 2015) to a very low grade accretionary wedge (Coastal  
 134 Belt) (e.g., Evitt & Pierce, 1975; Dumitru et al., 2013). The coherent blueschist facies  
 135 lenses of the Eastern Belt underplated <30 km deep from 111-123 Ma (Dumitru et al.,  
 136 2010; Apen et al., 2021). The Central Belt is a mélangé with a 75-95 Ma prehnite-pumpellyite  
 137 facies shaley matrix encompassing older, higher grade blocks (Morissani, 2006; Cloos,  
 138 1983; Platt, 2015). The Coastal Belt consists of very low grade (zeolite facies) metased-  
 139 imentary and rare metamafigs imbricated to form a classic accretionary wedge during  
 140 the Eocene to Miocene (Bachman, 1982; Dumitru et al., 2013). The Franciscan East-  
 141 ern Belt (Fig. 1a) is in thrust contact (ca. 123 Ma) with the Western Klamath and Rat-  
 142 tlesnake Creek terranes (Dumitru et al., 2010). A subunit of the Franciscan Eastern Belt,  
 143 the South Fork Mountain Schist, is dominated by distal, hemipelagic protoliths (Dumitru  
 144 et al., 2010; Schmidt & Platt, 2018) comparable to the CMS (Helper, 1986).

## 145 2.2 Condrey Mountain Schist (CMS)

146 The CMS records multiple generations of prograde ductile deformation in greenschist/epidote-  
 147 amphibolite to epidote-blueschist facies rocks associated with subduction along a con-  
 148 vergent margin (Fig. 2) (Helper, 1986; Tewksbury-Christle et al., 2021). The two main  
 149 units (upper and lower CMS) are exposed through a structural window in the Klamaths  
 150 due to regional doming, and the western limb of the dome is overturned from a primary  
 151 east-dipping orientation (seen along Y-Y'-Y'' cross section, Fig. 2b) (Helper, 1986). Up-  
 152 lift of onlap sequences constrains regional doming to the Neogene (Mortimer & Coleman,  
 153 1985), and low temperature thermochronology on plutons proximal to the CMS suggests  
 154 some component of uplift in the Oligocene (Piotraschke et al., 2015). This regional-scale  
 155 doming is roughly centered on the asymmetrical dome structure defined by the orien-  
 156 tation of pervasive transposition foliations in the CMS ( $S_2$ , Fig. 2) ( $S_T$  in Helper, 1986;  
 157 Mortimer & Coleman, 1985; Tewksbury-Christle et al., 2021).

158 The upper CMS is predominantly greenschist to epidote-amphibolite facies metavol-  
 159 canic rocks with intercalated graphite phyllites metamorphosed at 0.3-0.5 GPa, 300-400°C  
 160 (Helper, 1986). Concordant dioritic to tonalitic orthogneiss lenses crystallized at  $170 \pm$   
 161  $1$  Ma (magmatic zircon U-Pb crystallization ages in Saleeby & Harper, 1993), suggest-  
 162 ing that portions of the upper CMS are older than 170 Ma. Emplacement ages for the  
 163 upper CMS are  $156 \pm 1$  Ma, constrained by cross-cutting relationships with and cool-  
 164 ing ages in the overlying Rattlesnake Creek terrane (Hacker et al., 1995; Saleeby & Harper,  
 165 1993). The pre-170 Ma upper unit protolith age constraint places these rocks within the  
 166 age range of Western Hayfork arc volcanics, and emplacement beneath the base of the  
 167 Rattlesnake Creek terrane is roughly coeval with motion on the Orleans fault.

168 High angle, possibly late stage faults juxtapose the lower CMS against the upper  
 169 CMS, though lower angle faults marked by small metaserpentinite and metagabbro bod-  
 170 ies are locally preserved (Helper, 1986). In contrast to the upper CMS, the lower CMS

171 consists primarily of hemipelagic or pelagic sediments with m- to km-scale mafic and ser-  
 172 pentinized ultramafic intercalated lenses, all metamorphosed to epidote-blueschist facies  
 173 (0.7-1.1 GPa, 400-450°C) to form graphite mica schist, epidote blueschist, and metaser-  
 174 pentinite, respectively (Helper, 1986; Tewksbury-Christle et al., 2021). Although the sed-  
 175 iments entered the subduction zone on the downgoing plate, geochemical evidence sug-  
 176 gests that tectonic erosion of the overriding plate sourced the ultramafic and some of the  
 177 mafic lenses (Tewksbury-Christle et al., 2021).

178 Tewksbury-Christle et al. (2021) further characterized the fossil subduction inter-  
 179 face, prograde deformation, and underplating processes that assembled the lower CMS  
 180 rocks. These rocks record multiple generations of prograde ductile deformation. A per-  
 181 vasive transposition foliation that defines the CMS dome structure ( $S_2$ , Fig. 2) is coher-  
 182 ently developed across heterogeneous lithologies, suggesting assembly of the tectonically  
 183 eroded material and incoming sediments prior to coherent deformation and subsequent  
 184 underplating. Strain localization proximal to km-scale mafic + ultramafic lenses resulted  
 185 in a phase of locally developed overprinting structures ( $S_3$  and  $F_3$ , localization marked  
 186 in Fig. 2b). These structures broadly divide the lower CMS into three main underplated  
 187 packages. Thin incoming sediment packages and rare m-scale mafic + ultramafic lenses  
 188 underplated and were entrained during ongoing underplating and deformation. Intro-  
 189 duction of km-scale mafic + ultramafic lenses allowed for strain localization, abandon-  
 190 ment of the previously underplated package, and down-stepping of the subduction inter-  
 191 face. Subsequent underplating initiated below the ductile thrust zone. Multiple under-  
 192 plating phases suggest a possible protracted history that could provide insights into  
 193 evolution of this portion of the western North American margin.

194 Because of limited age data, however, regional correlations amongst the subduc-  
 195 tion record of the CMS, other Klamath terranes, and the Franciscan Complex are cur-  
 196 rently poorly constrained. Extant CMS dates are consistent with regional correlations,  
 197 in whole or in part, between the CMS and either 1) the Western Klamath terrane (e.g.,  
 198 Klein, 1975; Saleeby & Harper, 1993), 2) the Franciscan Eastern Belt (e.g., Brown & Blake,  
 199 1987; Chapman et al., 2021) or 3) none of the Klamath terranes or Franciscan Complex  
 200 (e.g., Hill, 1985). Correlation to the Western Klamath terrane is based on lithologic sim-  
 201 ilarities to the Galice Formation within the Western Klamath terrane and regional re-  
 202 lationships, which show the CMS and Western Klamath terrane are both structurally  
 203 beneath and in fault contact with the Rattlesnake Creek terrane (Fig. 1b and c) (Saleeby  
 204 & Harper, 1993). Comparable emplacement timing of the upper CMS and the Western  
 205 Klamaths further supports this correlation. Previously cited ages for the lower CMS, how-  
 206 ever, include white mica K-Ar cooling ages ( $118 \pm 2 \text{ Ma}^1$  and  $128 \pm 2 \text{ Ma}^2$ ), a sodic am-  
 207 phibole K-Ar cooling age ( $127 \pm 6 \text{ Ma}^2$ ) (Coleman et al. (1983)<sup>1</sup>; Helper (1986)<sup>2</sup>), and  
 208 detrital zircon maximum depositional ages (MDAs, 143-129 Ma) (Chapman et al., 2021),  
 209 overlapping with, or closely predating, emplacement of the Franciscan Eastern Belt ca.  
 210 123 Ma, suggesting that the lower CMS might instead be a down-dip continuation of the  
 211 Franciscan Eastern Belt (c.f., Brown & Blake, 1987) and that the Western Klamath ter-  
 212 rane either correlates with the upper CMS only or pinches out between the two (Fig. 1c).  
 213 A third, largely unaddressed alternative, is that the lower CMS correlates with neither  
 214 the Western Klamath terrane nor the Franciscan Eastern Belt. Although Hill (1985) of-  
 215 fered this alternative, he did so on the basis of an upper CMS occurrence west of the win-  
 216 dows along the Klamath River, treating the upper and lower CMS as a single terrane. In  
 217 so doing, he concluded that the Condrey Mountain Terrane pre-dated formation and as-  
 218 sembly of the Western Klamath Terrane and was emplaced prior to it, after 162 and be-  
 219 fore 150 Ma (Fig. 1c).

220 In addition to helping regional correlations, improved age data can help better con-  
 221 strain global mass and volatile recycling estimates. Modern tectonically erosive margins  
 222 are estimated to contribute significantly to supplying continental material and carbon  
 223 to the upper mantle at 65% and 30% of the total, respectively (Clift, 2017; Clift et al.,

224 2009). These budgets, however, assume no underplating processes in erosive margins.  
 225 Tewksbury-Christle et al. (2021) demonstrated that the lower CMS subducted along a  
 226 margin undergoing shallow tectonic erosion and deep underplating that preserved an es-  
 227 timated 10-60% of the incoming sediment from being recycled into the upper mantle. The  
 228 significant uncertainty in their estimates comes from limited age constraints that can be  
 229 improved with additional depositional and metamorphic ages for the lower CMS.

### 230 3 Methods

#### 231 3.1 Dating lower CMS crystallization and deposition

232 *Magmatic zircon U-Pb geochronology.* U-Pb zircon dates from the Scraggy Moun-  
 233 tain epidote blueschist lens (Fig. 2a) were collected by Helper and N. Walker at Univer-  
 234 sity of Austin’s TIMS laboratory (1988), but these legacy data were previously unpub-  
 235 lished (data, standards, and detailed methods in Dataset S1).

236 *Detrital zircon U-Pb geochronology.* We sampled a transect across the structural  
 237 thickness of the lower CMS (Fig. 2) to map depositional age variation with structural  
 238 depth. Samples were prepared by mechanical crushing and density and magnetic sep-  
 239 aration. Zircons were sprinkle mounted on double-sided tape and unpolished zircon grains  
 240 were U-Pb dated using depth profiling laser ablation technique (e.g., Marsh & Stockli,  
 241 2015) at the University of Texas at Austin’s geochronology lab (UTChron). Depth-profile  
 242 analysis allows for spatial recovery of multiple zircon age domains during progressive ab-  
 243 lation. The LA-ICP-MS system, which consists of an Analyte G.2 193 nm Excimer laser  
 244 ablation system with a Helix sample cell attached to an Element2 HR-ICP-MS. For each  
 245 U-Pb analysis, a 30  $\mu\text{m}$  laser spot with a nominal energy of 4 mJ, an average fluence of  
 246 1.98 J/cm<sup>2</sup>, and a pulse rate of 10 Hz was used to ablate zircons at a depth of 15  $\mu\text{m}$ .  
 247 This allowed for a <0.5  $\mu\text{m}$  depth resolution of different age domains. For this study, GJ1  
 248 (Jackson et al., 2004) was the primary standard to correct for downhole, elemental, and  
 249 isotopic fractionation, and two secondary standards (Plešovice, Sláma et al. (2008) and  
 250 91500, Wiedenbeck et al. (1995)) were analyzed to monitor data quality. Data reduc-  
 251 tion was performed using the Iolite (Paton et al., 2010) and VisualAge data reduction  
 252 scheme (Petrus & Kamber, 2012) within the WaveMetric IgorPro software package. Best  
 253 zircon U-Pb ages were obtain based on a zircon’s age with respect to 850 Ma and dis-  
 254 cordance between <sup>206</sup>Pb/<sup>238</sup>U and <sup>207</sup>Pb/<sup>235</sup>U ages. <sup>206</sup>Pb/<sup>238</sup>U ages were used for grains  
 255 younger than 850 Ma, including two-sigma internal error, while <sup>207</sup>Pb/<sup>206</sup>Pb ages were  
 256 used for grains older than 850 Ma. Zircon ages were eliminated if grains contained greater  
 257 than 10% analytical error, the <sup>206</sup>Pb/<sup>238</sup>U age error was greater than 10%, or the <sup>207</sup>Pb/<sup>235</sup>U  
 258 age errors was greater than 10% and had greater than 10% discordance.

259 For detrital provenance and maximum depositional age determinations, we ana-  
 260 lyzed 120+ zircon grains per sample to ensure the capture of any age component >5%  
 261 (Vermeesch, 2004). We calculated maximum depositional age (MDA) for each sample  
 262 as the youngest single grain (YSG), youngest 2+ grains that overlap at 1 $\sigma$  (YC1 $\sigma$ ), and  
 263 youngest 3+ grains that overlap at 2 $\sigma$  (YC2 $\sigma$ ) (Dickinson & Gehrels, 2009) after omit-  
 264 ting grains with <sup>206</sup>Pb/<sup>238</sup>U discordance and/or error >10% (Dataset S2). We report  
 265 the pooled 1 $\sigma$  and number of grains for each MDA. We present YC1 $\sigma$  as the best con-  
 266 straint of lower CMS depositional ages. YSG is more likely than the other methods to  
 267 result in an MDA younger than the true depositional age, and YC2 $\sigma$  is more likely to  
 268 overestimate the depositional age (Dickinson & Gehrels, 2009).

269 Depth profile LA-ICP-MS analysis on detrital zircons allows for dating of thin meta-  
 270 morphic or igneous zircon rims (e.g., Poulaki et al., 2021). Where rims are present and  
 271 measurable, we report an age for both the zircon rims and cores and use the word ‘rim’  
 272 throughout to denote presence of overgrowths without interpretation (e.g., metamorphic  
 273 or igneous growth). In addition, we employed split-stream LA-ICP-MS analysis, using

274 two Element2 instruments, to simultaneously acquire U-Pb dates, trace element, and rare  
 275 earth element (REE) data over depth profiles for selected samples (Section 4.1). We also  
 276 gathered cathodoluminescence (CL) images on selected depth-profiled grains using a Deben  
 277 Centaurus panchromatic CL system on the JEOL JSM-6390 LA scanning electron mi-  
 278 croscope (SEM) at ETH Zurich’s Electron Microscopy Lab and a Gatan MiniCL on the  
 279 FEI Quanta 200F SEM at ETH Zurich’s ScopeM.

### 280 3.2 Dating lower CMS metamorphism

281 *K-Ar and Rb-Sr geochronology.* K-Ar and Rb-Sr dates were collected at the Uni-  
 282 versity of Texas at Austin by F. McDowell (1988-89) and Helper (1987-89), respectively,  
 283 but these legacy data, similar to the U-Pb dates on magmatic zircons in Section 3.1, were  
 284 previously unpublished. K-Ar dates presented below and in Dataset S3 are for white mica.  
 285 Rb-Sr multi-mineral isochrons are defined by apatite-whole rock-white mica, unless oth-  
 286 erwise noted (Dataset S4 and Fig. S1). Both K-Ar and Rb-Sr closure temperatures (350-  
 287 450°C and 550°C, respectively) (e.g., Jäger, 1979; Ruffet et al., 1997; Scaillet, 1998) are  
 288 near to or higher than the lower CMS peak temperatures (Helper, 1986; Tewksbury-Christle  
 289 et al., 2021), so these geochronometers constrain either timing of metamorphism or co-  
 290 eval metamorphism/cooling.

291 To better interpret spatial variations in depositional and/or metamorphic ages, we  
 292 characterized peak temperatures using graphite crystallinity, where jumps in tempera-  
 293 ture might be indicative of cryptic structures that would affect the spatial distribution  
 294 of ages. Graphite Raman spectra vary systematically with peak temperature (330-650°C)  
 295 and are not sensitive to pressure or retrogression (e.g., Beyssac et al., 2002). We ana-  
 296 lyzed samples from a transect across the lower CMS structural thickness (Fig. 2) on ETH  
 297 Zurich’s DILOR Labram micro-Raman spectrometer, used Igor Pro for baseline correc-  
 298 tion and peak fitting, and applied Beyssac et al. (2002)’s temperature calibration. Re-  
 299 ported temperatures average results from 10 analyses per sample (average peak fits and  
 300 temperatures in Dataset S5), and we report the standard error.

301 Three detrital zircon samples (stars, Fig. 2) exhibit complex spatial relationships  
 302 with respect to their MDAs. To aid in interpretation of these results, we collected graphite  
 303 crystallinity data on these samples, as well as Si content in two generations of white mica  
 304 growth (D<sub>1</sub> and D<sub>2</sub>, Tewksbury-Christle et al., 2021). Although the CMS lacks the lim-  
 305 iting assemblage needed to calculate an absolute pressure using the Si-in-phengite geo-  
 306 barometer (Massonne & Schreyer, 1987), Si concentrations can constrain relative pres-  
 307 sures amongst these samples. We measured Si on the ETH Zurich’s JEOL JXA-8230 Elec-  
 308 tron Probe Microanalyser and calculated Si p.f.u. assuming all ferrous Fe (calibration  
 309 standards and calculated Si per formula unit detailed in Dataset S6).

## 310 4 Results

### 311 4.1 Lower CMS crystallization and deposition ages

312 Magmatic zircons from the Scraggy Mountain epidote blueschist lens (Fig. 2a) date  
 313 to  $169 \pm 2$  (magmatic zircon TIMS U-Pb). This age is comparable to the Western Hay-  
 314 fork arc in the Klamaths.

315 Detrital zircon U-Pb ages from from the lower CMS are characterized by a large  
 316 component of Western Hayfork arc ages, but the youngest zircon components, and hence  
 317 the MDAs, significantly postdate the age of crystallization of the Scraggy Mountain lens.  
 318 Detrital zircon spectra plotted as Kernel Density Estimations (KDEs, Figures S2 and  
 319 S3 with color bins that follow Sharman et al. (2015)) are dominated by <500 Ma grains  
 320 with only minor older components of up to 3.0 Ga. The bulk of the detrital zircons yield  
 321 dates <170 Ma with peaks between 145 Ma and 155 Ma, coeval with the Rogue-Chetco

arc in the Klamaths (Yule et al., 2006). The oldest MDA using any method is  $155 \pm 3$  Ma, but 8 of the 11 samples have an MDA younger than 150 Ma, regardless of the method used. Employing  $YC1\sigma$  method, MDAs range from 135-150 Ma with limited correlation between age discontinuities and mapped structures (Figs. 3 and 4). CT-CMW36, for example, yields a younger MDA ( $135 \pm 1$  Ma) than the samples both structurally above and below it ( $149 \pm 1$  Ma and  $153 \pm 2$  Ma, respectively), but all three are within the same thrust-bounded package (Tewksbury-Christle et al., 2021). Grains from CT-CMW36 have sectoral or poorly developed zonation, whereas grains from other samples have clear oscillatory zonation in the cores (CL images in Fig. 5 and S4).

Zircon rims show a range of behavior with respect to trace elements and REEs. Cores primarily have  $Th/U > 0.1$  (Fig. 5), consistent with igneous zircons (Hoskin & Schaltegger, 2003; Rubatto, 2017; Yakymchuk et al., 2018) or recrystallized igneous zircon (e.g., Poulaki et al., 2021). Rims, however, have a wide range of  $Th/U$  values, with populations of rims with  $Th/U > 0.1$  that both pre- and postdate the MDA. CL images show that some grains with low  $Th/U$  rims have irregular rims that cut oscillatory zoning in the cores (e.g., CT-CMW42 and CT-CMW55, Fig. 5), while others have more conformable planar boundaries between a bright core and dark rims (CT-CMW35, Fig. 5). Grains with high  $Th/U$  rims have both planar oscillatory zoning (CT-CMW55) and more diffuse irregular boundaries (CT-CMW42). Additional CL images are presented in Fig. S4. For the two samples with REE analyses (CT-CMW36 and CT-CMW53),  $Ce/Ce^*$  versus  $Sm/La$  show that most single ages and cores plot in the igneous field (e.g., Hoskin, 2005), but some plot in the metamorphic field or between the two fields, and a similar spread occurs in the zircon rims (Fig. 6).  $Th/U$  ratios appear to be poorly correlated to  $Ce/Ce^*$  versus  $Sm/La$ , with low  $Th/U$  rims plotting predominantly in or close to the igneous field (Fig. S5).

## 4.2 Lower CMS metamorphic ages, peak temperatures, and relative pressures

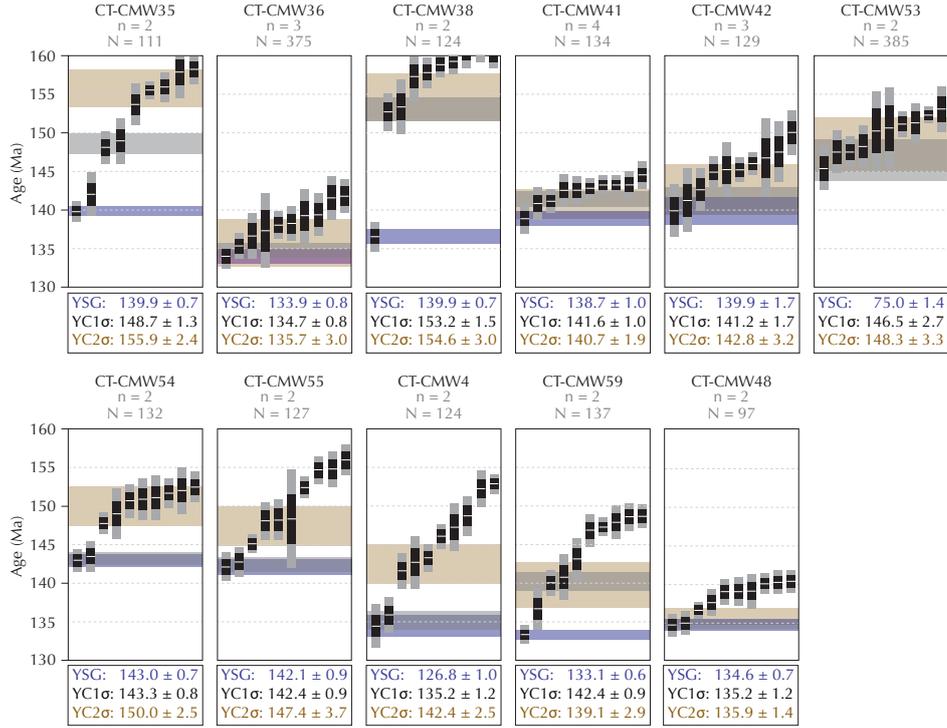
Multi-mineral isochron Rb-Sr dates and white mica K-Ar dates postdate the MDAs (Fig. 4a). Rb-Sr dates range from 119-132 Ma and average  $124 \pm 7$  Ma. K-Ar dates range from 124-133 Ma and average  $128 \pm 2$  Ma. These dates have a wide range but young broadly downwards.

Peak temperatures from graphite crystallinity show little variation across the structural thickness. The temperatures average  $460 \pm 10^\circ\text{C}$  with a minor inverted metamorphic gradient (approximately  $1.5^\circ\text{C}/\text{km}$ ). Temperature variations do not correlate with mapped structures. Peak temperatures from graphite crystallinity (Fig. 4b) and relative pressures from Si-in-Phengite (Fig. S6) are not significantly different amongst the top three samples collected for detrital zircon geochronology. Pressures from multiple generations of white mica growth are also not significantly different from each other.

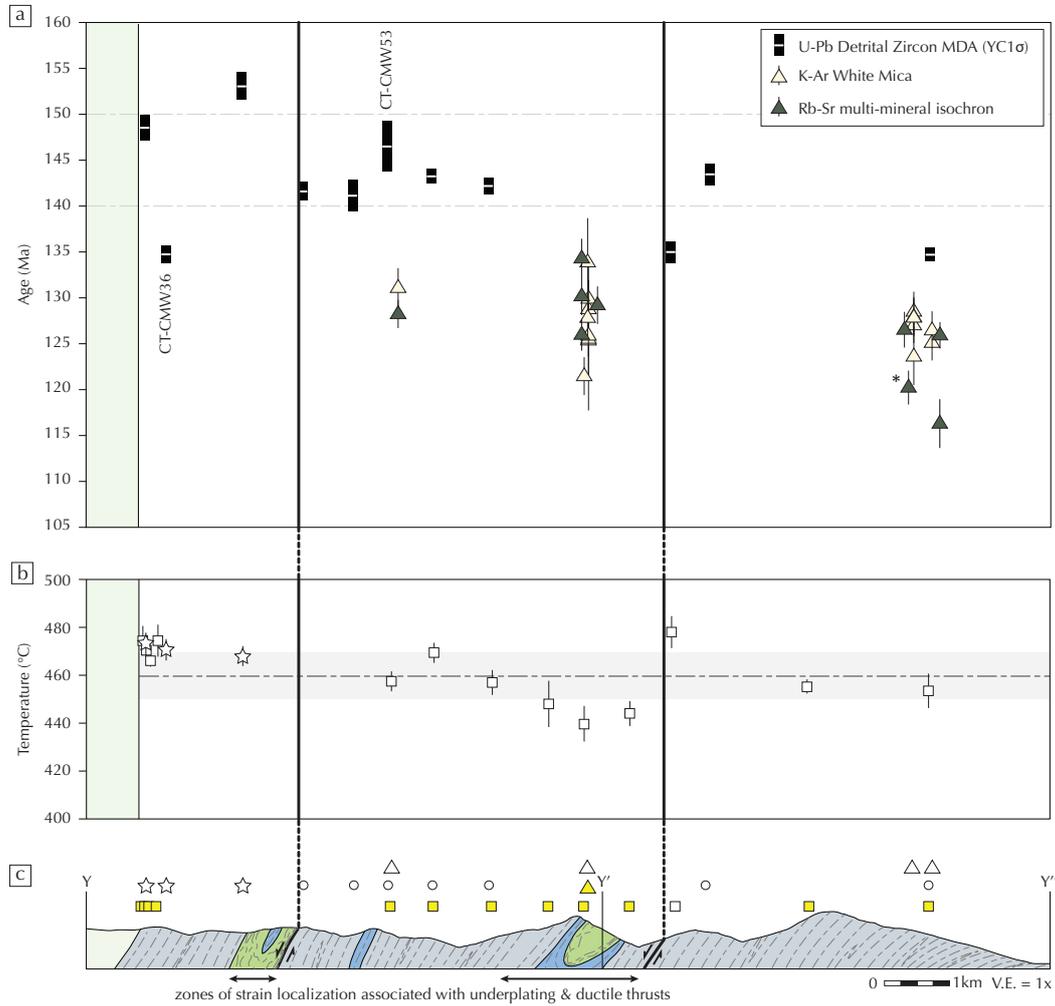
## 5 Synthesis of data to determine timing of Lower CMS emplacement

### 5.1 Interpretations of calculated MDAs

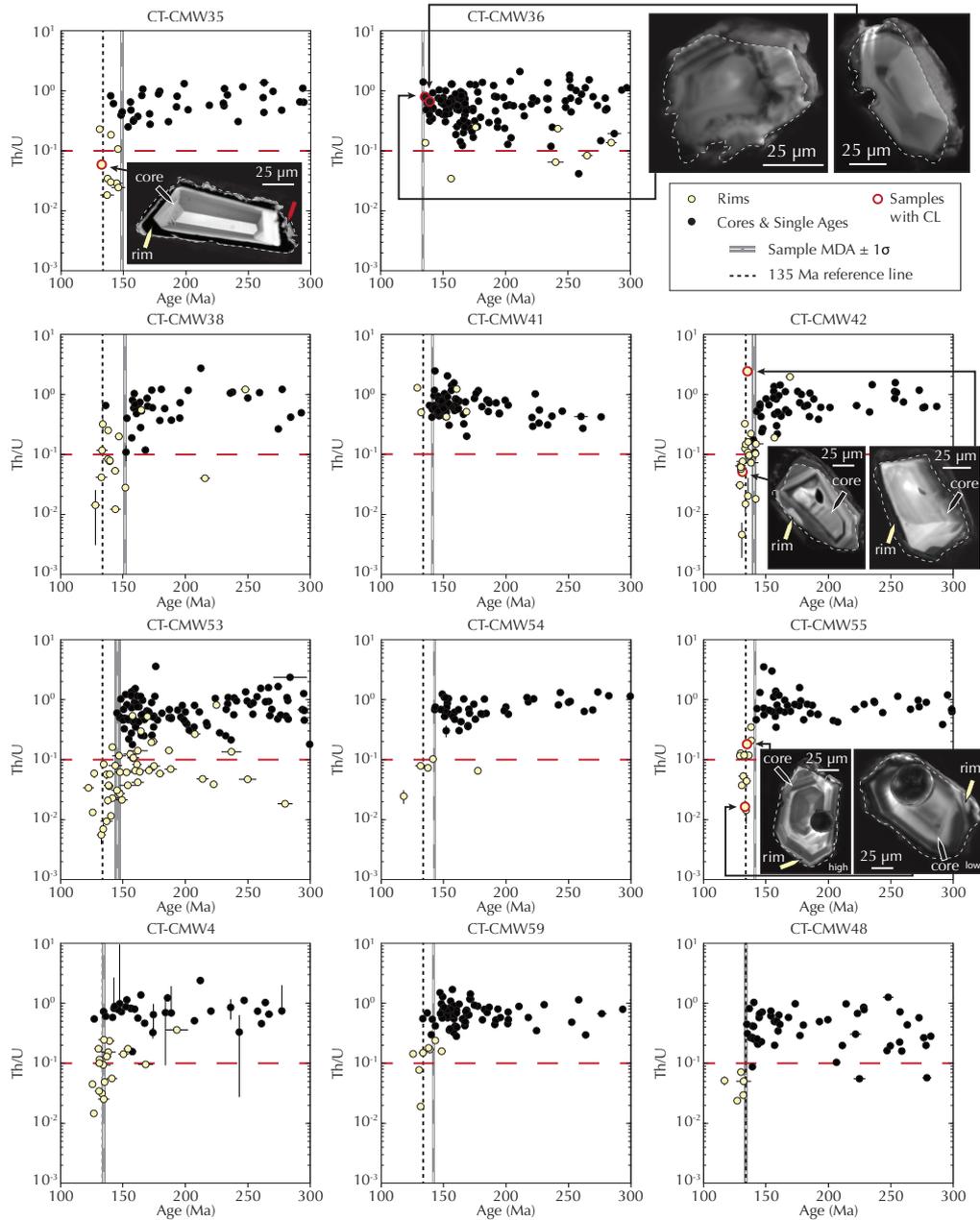
All MDAs are 135-153 Ma with complex spatial behavior across increasing structural depth, where MDA discontinuities are not always correlated with inferred and/or observed structures. MDA discontinuities of ca. 5-18 Ma occur within packages imbricated by ductile thrusts. Out of sequence thrusting or large scale folding could explain the repetition of the ca. 135 Ma MDA towards the top of the structural package (CT-CMW36) that results in the largest MDA discontinuity (Fig. 4), but structural observations do not support either explanation. We observed no evidence of folding at wavelengths comparable to the structural thickness, brittle faulting between these samples, or strain localization that could be indicative of ductile thrusting (Tewksbury-Christle



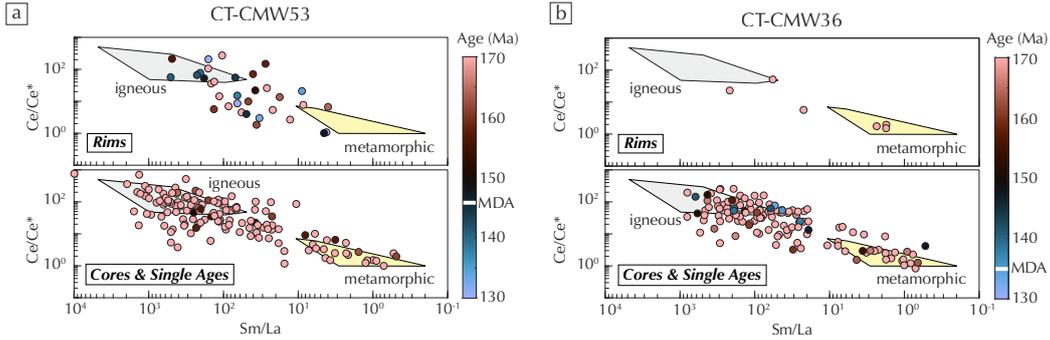
**Figure 3.** Age of the youngest ten grains in each sample ( $\pm 1\sigma$  in black,  $\pm 2\sigma$  in gray), with increasing structural depth from left to right, top to bottom. Colored bars mark the calculated MDA for the three different methods: youngest single grain (YSG; violet), youngest two or more grains that overlap at  $1\sigma$  (YC1 $\sigma$ ; gray), and youngest three or more grains that overlap at  $2\sigma$  (YC2 $\sigma$ ; brown). N is the number of concordant grains out of the 120+ analyzed. n is the number of grains used to calculate the MDA using YC1 $\sigma$ . CT-CMW36 and CT-CMW53 were also analyzed for REEs, resulting in a larger number of total grains.



**Figure 4.** a) Variation of MDAs ( $\pm 1\sigma$ ) and metamorphic dates across the lower CMS structural thickness. Sample locations follow Fig. 2, and black lines mark ductile thrusts recognized by Tewksbury-Christle et al. (2021). Sample marked with (\*) was calculated using a titanite-whole rock-phengite isochron. All others use an apatite-whole rock-phengite isochron. MDA discontinuities do not typically correlate with mapped ductile thrusts. Metamorphic ages young broadly downward and closely postdate the youngest MDAs. REE analyses on labeled samples presented in Fig. 6a-b. b) Peak temperatures from graphite crystallinity show limited variation with structural depth. c) Sample locations projected onto the Y-Y'-Y'' cross section line in Fig. 2. Symbols in (b) and (c) follow Fig. 2.



**Figure 5.** Th/U variations by sample for rims and cores/single ages. Structural depth of the samples increases from left to right, top to bottom. Where not shown, error bars are smaller than symbol size. Most samples show a slight decrease in Th/U when considering both cores and rims approximately coeval with the calculated MDA (gray bar showing  $\pm 1\sigma$ ). Most rims, cores, and single ages are older than 135 Ma (dashed reference line), but younger low Th/U rims are present throughout the structural package. Example CL insets from CT-CMW35, CT-CMW42, and CT-CMW55 show cores with oscillatory zoning and predominantly irregular rims. Grains from CT-CMW36 have sector or poorly developed zoning. Yellow and black arrows mark rims and cores, respectively, and dashed gray lines outline the polished surface of the zircons. Bright irregular zones at the grain margins in CT-CMW35 are tape residue (red arrow). Additional CL images from these samples are in Fig. S4.



**Figure 6.** REE variations in CT-CMW53 (a) and CT-CMW36 (b) (data available in Dataset S7). Fields after Hoskin (2005). The calculated MDA is plotted as a white line in the color bar. Rims in CT-CMW53 show a range of affinity, whereas cores/single ages plot predominantly in or near the igneous field. Rare rims in CT-CMW36 plot in the metamorphic field or in the area between the two fields.

371 et al., 2021). Peak temperatures are also consistent across the structural thickness, with  
 372 no significant variations in temperature correlated with MDA discontinuities. Relative  
 373 pressures across the three topmost samples are comparable amongst the samples and de-  
 374 formation phases. Together, these data suggest that the lower CMS protoliths were sub-  
 375 ducted and assembled along the same prograde path. In the case of later juxtaposition  
 376 via cryptic thrusts, folding, or melange mixing, we would expect samples with different  
 377 depositional ages to have different pressure and/or temperature histories.

378 We can interpret the relatively anomalously young CT-CMW36 MDA (Fig. 4) as  
 379 either representative of the true maximum depositional age or as younger metamorphic  
 380 zircon recrystallization and overprinting. Peak temperatures constrained by graphite crys-  
 381 tallinity ( $460 \pm 10^\circ\text{C}$ ) are lower than temperatures needed to form new zircon grains ( $\geq$   
 382  $580^\circ\text{C}$ , Watson et al., 2006). Graphite crystallinity is not sensitive to retrogression, and  
 383 the graphitic mica schist records prograde deformation across the phengite + graphite  
 384 + quartz + chlorite  $\pm$  albite assemblage as indicated by both temperature constraints  
 385 and Si-in-phengite compositions. Young dates in CT-CMW36 are associated with high  
 386 Th/U more characteristic of magmatic zircons and with igneous affinity Ce/Ce\* versus  
 387 Sm/La. Metamorphic zircons can have high Th/U (Hoskin & Schaltegger, 2003; Rubatto,  
 388 2017; Yakymchuk et al., 2018; Poulaki et al., 2021), however, and all REE analyses from  
 389 these samples have a range of affinities. Rims in CT-CMW42, for example, appear to  
 390 be metamorphic based on the irregular rims cross-cutting oscillatory zonation (yellow  
 391 arrows, Fig. 5) but have both high and low Th/U. Furthermore, internal zonation is poorly  
 392 developed in the CL images of CT-CMW36 and morphologically different from the in-  
 393 ternal structure of zircons from other samples (Fig. 5 and S4), supporting a possible meta-  
 394 morphic origin. Tomaschek et al. (2003) proposed topotactic fluid-assisted dissolution  
 395 and replacement of igneous zircons from metabasites in Syros, Greece, that reset the U-  
 396 Pb ages during prograde metamorphism at  $480^\circ\text{C}$  and 1.6 GPa. Although still higher grade  
 397 than the lower CMS, this temperature is much lower than zircon crystallization temper-  
 398 atures and might explain the internal morphology of the CT-CMW36 zircons. Poulaki  
 399 et al. (2021) demonstrated similar conditions in the Cycladic subduction complex for both  
 400 fluid-absent and fluid-assisted metamorphic zircon recrystallization that resulted in REE  
 401 signatures that were partially inherited from the crystallized zircon core.

402 Because these zircon trace element data are non-diagnostic, we present two differ-  
 403 ent possible interpretations of lower CMS deposition and emplacement. Regardless of

404 how we interpret CT-CMW36, the MDAs of the other samples still bracket the oldest  
 405 possible deposition of the lower CMS at  $\leq 153$  Ma, with the majority of samples  $< 150$   
 406 Ma (Fig. 3 and 4). These results definitively preclude correlation with the Western Klamath  
 407 terrane. The interpretation of CT-CMW36 simply determines the duration of deposition  
 408 of the lower CMS. The possible scenarios for CT-CM36 are:

- 409 1. *The MDA is representative of the lower CMS true maximum depositional age.* In  
 410 the absence of structural repetition, or melange mixing and assuming no growth  
 411 of metamorphic zircons, this complex spatial behavior across increasing structural  
 412 depth (illustrated in Fig. 4) is best explained as true maximum depositional ages,  
 413 where samples with MDAs older than ca. 135 Ma are missing the youngest pop-  
 414 ulation of detrital zircons. The lower CMS graphite schist protolith was hemipelagic  
 415 (Helper, 1986), indicative of distal, deep water deposition with possible concomi-  
 416 tant heterogeneous distributions of zircon populations in protolith sediment. For  
 417 example, 24% of the grains in CT-CMW41 are  $< 150$  Ma, compared to  $\leq 10\%$  for  
 418 the majority of the other samples. If the youngest population is not present in all  
 419 samples due to heterogeneous source input, the MDA could overestimate the true  
 420 depositional age but still bracket the oldest possible deposition. In this circum-  
 421 stance, we would therefore interpret the youngest MDAs (ca. 135 Ma) as best con-  
 422 straining deposition of the lower CMS, with subduction and emplacement post-  
 423 dating 135 Ma.

424 If the CT-CMW36 MDA is the true depositional age, zircon rims from the other  
 425 samples can be interpreted in multiple ways. Zircon rim ages predominantly post-  
 426 date the MDA in individual samples, but many predate 135 Ma (dashed lines, Fig.  
 427 5). In CL, the morphology of these rims appears to be consistent with both ig-  
 428 neous (CT-CMW35, despite low Th/U) and metamorphic (CT-CMW42, despite  
 429 low and high Th/U) overgrowths. The morphology, therefore, appears to have lim-  
 430 ited correlation to Th/U values. REE analyses are similarly non-diagnostic, show-  
 431 ing a range of affinities in Ce/Ce\* versus Sm/La that are commonly contradic-  
 432 tory to Th/U (e.g., rims with igneous affinity but low Th/U, Fig. S5). Although  
 433 igneous rims are rarely  $< 0.1$  (Hoskin & Schaltegger, 2003; Rubatto, 2017; Yakym-  
 434 chuk et al., 2018), magmatic zircons grown in fractionated magmas can have low  
 435 Th/U (Kirkland et al., 2015), consistent with the tonalite-trondhjemite-granodiorite  
 436 and granodiorite suites that characterized late stage plutonic activity in the Klamath  
 437 (Allen & Barnes, 2006). Recrystallized rims formed during metamorphism,  
 438 however, may inherit the REE and Th/U signature of the cores (e.g., Poulaki et  
 439 al., 2021), which might account for the range of REE affinities. If we interpret the  
 440 rims as igneous, then our MDAs overestimate the true depositional age. If we in-  
 441 terpret the rims as metamorphic, then many of the dated zircons would need to  
 442 be subducted prior to the minimum depositional age of the lower CMS at 135 Ma.  
 443 Because the lower CMS subducted along a tectonically erosive margin (Tewksbury-  
 444 Christle et al., 2021), these metamorphic rims could be the record of older, pre-  
 445 viously underplated blocks that were stripped off during subduction erosion, in-  
 446 tercalated into the younger incoming sediment, ductilely deformed along the pro-  
 447 grade path, and subsequently underplated at depth after 135 Ma. Given the CL  
 448 imagery, the best explanation is likely a mix of the two cases, where some zircons  
 449 have igneous rims that would result in a lower MDA for the sample and others have  
 450 metamorphic rims that would necessitate intercalation of older sediments already  
 451 in the subduction interface.

- 452 2. *Young CT-CMW36 zircon cores are metamorphic.* In this case, the young CT-CMW36  
 453 zircon ages need to be excluded from the calculated MDA, and individual MDAs  
 454 of other samples are representative of their true maximum depositional ages. The  
 455 lower CMS maximum depositional ages, therefore, would young broadly downwards  
 456 from 149 to 135 Ma (Fig. 3), and older MDAs towards the bottom of the struc-  
 457 tural package could be due to heterogeneous zircon provenance, as discussed above.

458 This younging downward pattern is consistent with five MDAs discussed in Chapman  
 459 et al. (2021). Age discontinuities spatially correlated with ductile thrusts (black  
 460 lines, Fig. 4) could represent brief periods with no underplating during strain lo-  
 461 calization to the km-scale mafic+ultramafic lenses that facilitated assembly of the  
 462 lower CMS (Tewksbury-Christle et al., 2021).

## 463 5.2 Timing of lower CMS emplacement and underplating

464 Metamorphic white mica K-Ar (Dataset S3) and multi-mineral-isochron Rb-Sr dates  
 465 (Dataset S4) place constraints on subduction and emplacement of the lower CMS, in-  
 466 dependent of the detrital zircon ages. By this measure, metamorphism in the lower CMS  
 467 ranged from 119-133 Ma with a downward-younging trend, closely post-dating the youngest  
 468 MDAs (<5 Ma; Fig. 3 and 4). This age range is consistent with cessation of arc mag-  
 469 matism west of or in the vicinity of the CMS window (Allen & Barnes, 2006). Before ca.  
 470 135 Ma, active plutonism was either outboard and west of the CMS or fell along approx-  
 471 imately the same longitude, so older lower CMS emplacement would have required un-  
 472 derplating below an arc or in the backarc. After ca. 135 Ma, there is no active arc sig-  
 473 nature within the Klamaths (Allen & Barnes, 2006). Remnants of the arc active dur-  
 474 ing lower CMS emplacement are presumably buried under the Cretaceous onlap sequences  
 475 to the east, along the structural trend of the Sierra Nevada batholith.

## 476 6 Implications

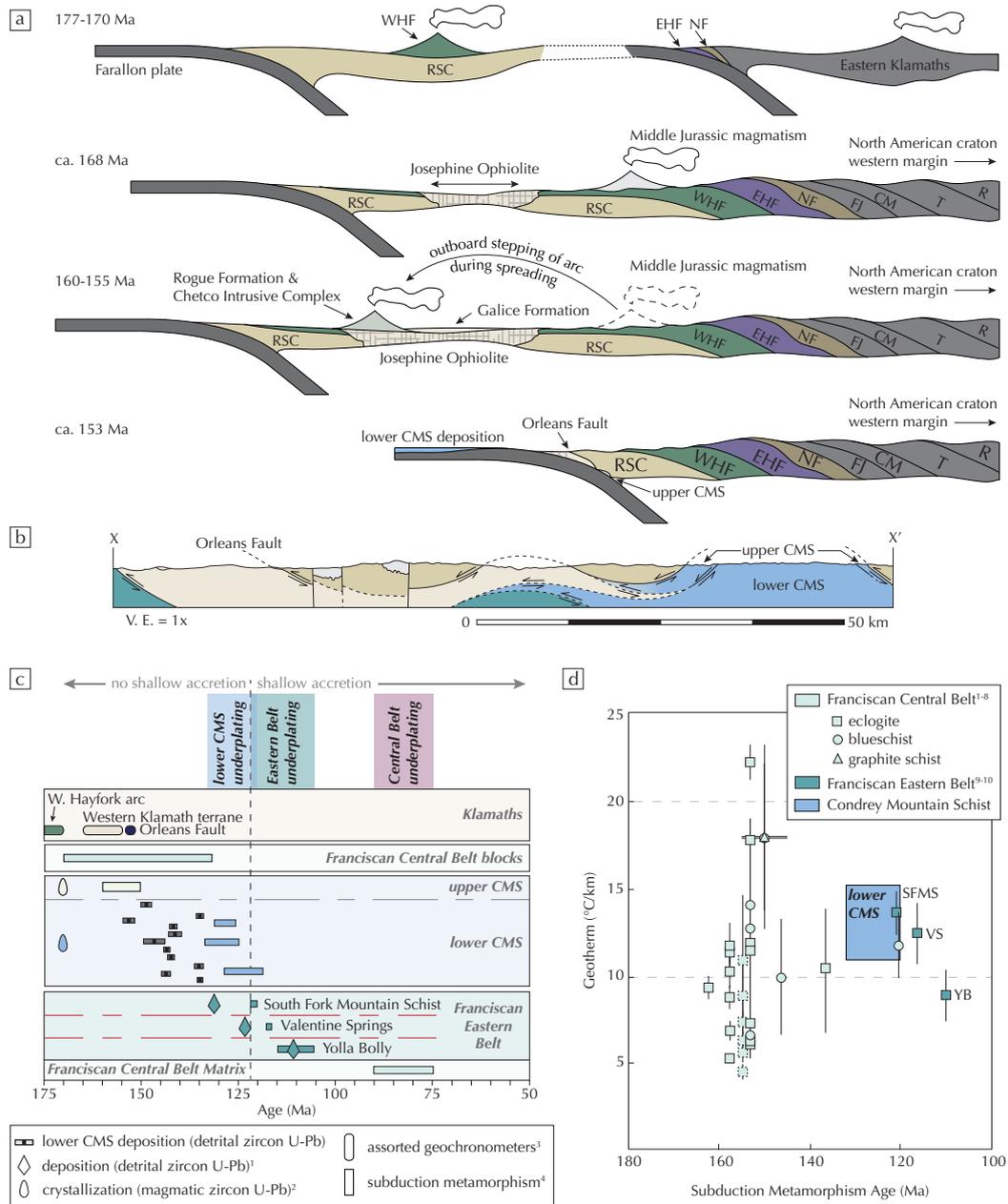
### 477 6.1 Implications for CMS terrane affinity within the Klamaths and Fran- 478 ciscan

479 Our new geochronologic results preclude correlation of the lower CMS with either  
 480 the Western Klamath terrane or the South Fork Mountain Schist (Fig. 7a-b). The lower  
 481 CMS depositional ages are younger than the Western Klamath terrane (Fig. 7c). Even  
 482 if we consider the most conservative estimate of MDA ( $YC2\sigma$ ), the majority of samples  
 483 were deposited after 150 Ma and thus postdate deposition of the Western Klamath ter-  
 484 rane (Frost et al., 2006; Macdonald et al., 2006). Furthermore, onset of metamorphism  
 485 in the lower CMS predates and overlaps with subduction and metamorphism of the South  
 486 Fork Mountain Schist (Dumitru et al., 2010).

487 The transition of the western North American margin from outboard subduction  
 488 and accretion of fringing island arcs to an along-strike coherent Andean-style margin oc-  
 489 curred during the early Franciscan history. The Klamaths record sporadic terrane ac-  
 490 cretion, whereas the Franciscan Complex preserves development of an accretionary wedge  
 491 at a well-developed, Andean-style subduction margin. Based on our new geochronologic  
 492 constraints, the lower CMS fits into the temporal gap between the end of Klamath-style  
 493 subduction and the onset of accretionary tectonics in the Franciscan, with underplat-  
 494 ing of the lower CMS starting before the earliest coherent underplating in the Francis-  
 495 can and then continuing during underplating of the South Fork Mountain Schist.

### 496 6.2 The CMS as a record of earliest Franciscan underplating

497 Reconstructions of primary subduction margins from the Klamath terranes are par-  
 498 ticularly challenging given the complex structural relationships of multiple overprinting  
 499 accretion events. The aggregated Klamath terranes preserve evidence of multiple fring-  
 500 ing subduction margins that were subsequently stacked along the North American mar-  
 501 gin. Imbricated arc and basinal terranes that sandwich rare subduction complexes (e.g.,  
 502 Snoko & Barnes, 2006, and references therein) collectively suggest the existence of four  
 503 fringing subduction margins affiliated with, from east to west: 1) an unidentified con-  
 504 tinental North American arc, 2) the Redding arc and related terranes (Trinity and Yreka),  
 505 3) the North Fork arc, and 4) the Rattlesnake Creek-Western Hayfork (RSC-WHF) arc



**Figure 7.** a) Schematic cross sections showing evolution of the Klamath margin. The Klamaths preserve evidence of a single subduction margin outboard of the Rattlesnake Creek and Western Klamath terranes from the Late Triassic to the Early Cretaceous. Colors and abbreviations follow legend in Fig. 1. b) Cross-section along X-X' in Fig. 1 showing the correlation that is most consistent with the data presented in this study. c) Timeline of Klamath and Franciscan tectonics including geochronologic constraints from this study. All lower CMS ages are presented in this study. Others are sourced as follows: <sup>1</sup>Dumitru et al. (2010); <sup>2</sup>Saleeby and Harper (1993); <sup>3</sup>Snoke and Barnes (2006); <sup>4</sup>for the upper CMS: Hacker et al. (1995); Saleeby and Harper (1993), for the Franciscan: Dumitru et al. (2010); Morissani (2006). The dashed line marks onset of accretionary subduction in the Franciscan (c.f., Dumitru et al., 2010). d) Geothermal evolution of the Franciscan margin through time, constrained by Franciscan P-T estimates from: <sup>1</sup>Cooper et al. (2011); <sup>2</sup>Page et al. (2007); <sup>3</sup>Massonne (1995); <sup>4</sup>Krogh Ravna and Terry (2004); <sup>5</sup>Krogh et al. (1994); <sup>6</sup>Tsujimori et al. (2006); <sup>7</sup>Wakabayashi (1990); <sup>8</sup>Ukar and Cloos (2016); <sup>9</sup>Dumitru et al. (2010); <sup>10</sup>Schmidt and Platt (2020). All blocks from the same location are plotted with ages from Cooper et al. (2011). Where ages are not available, dashed symbols are plotted against an average age for block metamorphism from Cooper et al. (2011). SFMS: South Fork Mountain Schist, VS: Valentine Springs, YB: Yolla Bolly.

506 (Saleeby, 1990; Wright & Wyld, 1994; Snoke & Barnes, 2006). Based on our data from  
 507 the lower CMS, coupled with regional relationships, we hypothesize that the RSC-WHF  
 508 arc was built above a proto-Franciscan margin, based on the following lines of evidence:

- 509 1. The Franciscan Eastern Belt is in thrust contact with the Western Klamath ter-  
 510 rane (Fig. 1a). Although the Eastern Belt post-dates the Western Klamath ter-  
 511 rane, the Eastern Belt represents underplated material subducted on the down-  
 512 going Farallon slab along the Franciscan margin (e.g., Dumitru et al., 2010). The  
 513 Western Klamath terrane comprises arc- and forearc-related units (e.g., Snoke &  
 514 Barnes, 2006). The simplest interpretation would place a subduction margin be-  
 515 tween these two units, which would have been the Franciscan margin at the time  
 516 of Eastern Belt subduction ( $\leq 123$  Ma).
- 517 2. The remains of a different trench do not exist between the RSC-WHF arc and the  
 518 Franciscan Eastern belt. If Franciscan subduction initiated along a new trench out-  
 519 board of the RSC-WHF margin, juxtaposing the Franciscan Eastern Belt (down-  
 520 going slab of this Franciscan margin) against the Western Klamath terrane (which  
 521 is in the overriding plate of the RSC-WHF margin) would require removal of: 1)  
 522 any subduction complex material affiliated with the RSC-WHF margin, 2) the in-  
 523 tervening downgoing plate material between the RSC-WHF margin and the Fran-  
 524 ciscan margin, and 3) backarc and forearc deposits, as well as arc rocks, affiliated  
 525 with the Franciscan margin. We do not see evidence of these units in the Klamaths.
- 526 3. The RSC is an ophiolitic mélangé overlain by volcanic cover that formed the base-  
 527 ment for the WHF arc (Wright & Wyld, 1994; Frost et al., 2006). RSC-WHF vol-  
 528 canics and plutons date from the late Triassic to ca. 170 Ma (Snoke & Barnes, 2006)  
 529 (Fig. 7a). After imbrication of the Klamath terranes, the Wilson Point thrust be-  
 530 tween the Eastern and Western Hayfork terranes is cut by a suite of Middle to Late  
 531 Jurassic plutons (168-152 Ma) that indicate construction of a new arc on top of  
 532 the juxtaposed North Fork, Eastern and Western Hayfork, and RSC terranes (Fig.  
 533 1 and 7a) (Saleeby, 1990; Wright & Wyld, 1994; Snoke & Barnes, 2006). These  
 534 plutons are typically interpreted to represent continued volcanism in the RSC-WHF  
 535 arc, where suturing of the upper plate terranes has no effect on continued subduc-  
 536 tion along the RSC-WHF margin, except to change the configuration of the over-  
 537 riding plate (e.g., Wright & Wyld, 1994).  
 538 Subsequent forearc spreading forming the Josephine Ophiolite (ca. 165-161 Ma)  
 539 and associated basinal deposits (158-151 Ma) (Saleeby & Harper, 1993; Hacker  
 540 et al., 1995; Snoke & Barnes, 2006; Yule et al., 2006; Surpless et al., 2023), and  
 541 forearc basin closure (ca. 153-150 Ma) (Snoke & Barnes, 2006, and references therein)  
 542 shuffled the relative location of the trench with respect to the RSC-WHF arc, but  
 543 did not change subduction along the margin (Fig. 7a). Ongoing volcanism in the  
 544 Rogue-Chetco arc (Yule et al., 2006; Garcia, 1982) suggests ongoing subduction  
 545 along the RSC-WHF margin through the Late Jurassic.

546 Given these different lines of evidence, we propose that Late Triassic onset of vol-  
 547 canism in the RSC-WHF arc is indicative of proto-Franciscan subduction initiation, which  
 548 occurred earlier in the north, outboard of the Klamaths, than in the south. The mar-  
 549 gin between the Western Klamath terrane and the Franciscan Eastern Belt was the Fran-  
 550 ciscan subduction margin by the Early Cretaceous. There is no evidence that a secondary  
 551 margin existed between the Franciscan and the rest of the Klamaths, suggesting that the  
 552 Franciscan margin is the same as the RSC-WHF margin, thus designating the RSC-WHF  
 553 margin the proto-Franciscan margin. Progressive arc development from subduction ini-  
 554 tiation in the RSC to the Middle Jurassic arc that overprints the sutured Klamath ter-  
 555 ranes to the Rogue-Chetco arc all suggest continued subduction along this proto-Franciscan  
 556 margin from the Late Triassic to the Middle Jurassic.

557 If the lower CMS subducted along this proto-Franciscan margin, it represents large  
 558 scale coherent underplating in the early period of non-accretionary subduction. This pe-  
 559 riod of subduction along the Franciscan margin (ca. 170-123 Ma) was previously assumed  
 560 to only be preserved in high grade blocks and isolated slabs in the younger, lower grade  
 561 melange of the Franciscan Central Belt. The onset of accretionary subduction was co-  
 562 incident with wide-scale underplating of the Franciscan Eastern Belt, increased sediment  
 563 supply due to the rise of a laterally-continuous Cordilleran arc and Great Valley fore-  
 564 arc and development of the accretionary wedge that is preserved in the Franciscan Com-  
 565 plex (e.g., Dumitru et al., 2010; Orme & Surpless, 2019). Geochemical evidence from the  
 566 lower CMS suggests incorporation of ultramafic and mafic lenses sourced from tectonic  
 567 erosion of the overriding plate (Tewksbury-Christle et al., 2021), consistent with the early,  
 568 non-accretionary Franciscan subduction margin. For example, the Scraggy Mountain epi-  
 569 dote blueschist lens has an arc-affinity geochemical signature (Tewksbury-Christle et al.,  
 570 2021) and a crystallization age comparable to the Western Hayfork arc, which occupied  
 571 the hanging wall during subduction and is a possible source for the protolith of this blueschist  
 572 lens.

573 Constraints on the early thermal structure of the Franciscan margin are difficult  
 574 due to limited preservation and overprinting retrogression. The lower CMS preserves a  
 575 coherently underplated record with limited retrogression that can be used to better con-  
 576 strain the early Franciscan conditions. Pressure-temperature (P-T) estimates from high  
 577 grade blocks and slabs in the Central Belt suggest a wide range of geotherms (Cooper  
 578 et al., 2011; Page et al., 2007; Massonne, 1995; Krogh Ravna & Terry, 2004; Krogh et  
 579 al., 1994; Tsujimori et al., 2006; Wakabayashi, 1990; Ukar & Cloos, 2016). Geothermal  
 580 gradients are better constrained in the coherently underplated Franciscan Eastern Belt  
 581 (Fig. 7d) (Dumitru et al., 2010; Schmidt & Platt, 2020). We calculated the geotherm  
 582 using the graphite crystallinity temperatures presented in this paper and previous pres-  
 583 sure constraints (Tewksbury-Christle et al., 2021). The lower CMS fits broadly into the  
 584 estimated geotherms and the cooling trend in the Franciscan Eastern Belt units. Fur-  
 585 ther characterization of the lower CMS metamorphic history could provide further con-  
 586 straints and insights on early Franciscan subduction conditions.

### 587 **6.3 Implications for mass and volatile recycling at erosive subduction** 588 **margins**

589 Our improved timing constraints have further implications for recycling of mass  
 590 and volatiles to Earth's deep interior. Underplating in modern tectonically erosive mar-  
 591 gins is not currently accounted for in global mass and volatile budgets (e.g., Clift et al.,  
 592 2009; Clift, 2017). Tewksbury-Christle et al. (2021) estimated that the lower CMS rep-  
 593 represents underplating of 10-60% of the incoming sediment despite subducting along a tec-  
 594 tonically erosive margin. Uncertainty in underplating timing and duration, which affects  
 595 the estimated sediment supply as well as the assumed plate velocities, drives the signif-  
 596 icant uncertainty in this estimate. With the new ages presented here, the onset of meta-  
 597 morphism near the top (ca. 133 Ma) and at the bottom (ca. 128 Ma) of the structural  
 598 package brackets underplating duration for approximately 9 km of the lower CMS struc-  
 599 tural thickness. Using the cross sectional area and sediment supply rates calculated in  
 600 Tewksbury-Christle et al. (2021) and Farallon plate velocities from Engebretson et al.  
 601 (1985), in conjunction with our improved duration estimate, the lower CMS represents  
 602 ca. 75% of the incoming sediment. This estimate is consistent with upper bounds of pre-  
 603 vious estimates of underplating percentages from Hikurangi (60%, calculated from Clift  
 604 et al., 2009; Bassett et al., 2010), and the Andes (80%, Clift & Hartley, 2007). Signif-  
 605 icant preservation predicted for the lower CMS and other modern erosive margins sug-  
 606 gests that underplating in erosive margins is an important factor to incorporate into mass  
 607 and volatile budgets.

## 7 Conclusions

The western margin of North America records protracted underplating in the Klamath Mountains and the Franciscan Complex. Subduction during assembly of the Klamaths was characterized by fringing island arcs and terrane accretion and transitioned to Andean-style subduction resulting in the formation of the Franciscan Complex. This transition between the end of Klamath terrane accretion and the onset of coherent underplating in the Franciscan is poorly preserved due to a lack of accretion. The CMS is a subduction complex exhumed through a window in the overlying older Klamath terranes that records down-dip coherent underplating and up-dip tectonic erosion. Previously published age constraints on the upper CMS, in addition to regional structural relationships, suggest that the upper CMS can be correlated with other Klamath terranes (e.g., Klein, 1975; Saleeby & Harper, 1993). Previously published age constraints for the epidote-blueschist facies lower CMS, however, overlap with both the youngest Klamaths and oldest Franciscan.

We constrained lower CMS deposition/crystallization and emplacement using multiple geochronometers. Our geochronologic results constrain lower CMS deposition and subduction/underplating to 153-135 Ma and 119-133 Ma, respectively. These ages preclude direct correlation to both the youngest Klamath terranes (>150 Ma) and to the oldest, coherent Franciscan (underplated ca. 123 Ma). The lower CMS subducted after assembly of the other Klamath terranes, and subduction and emplacement began before and overlapped with underplating in the Franciscan. P-T conditions in the lower CMS are consistent with other constraints from the early Franciscan history. Because the earliest Franciscan history is poorly preserved, efforts to reconstruct the thermal structure of the subduction margin rely on highly retrogressed isolated blocks and slabs. The lower CMS is a coherently underplated record of this transitional time frame with limited retrogression, and further characterization of CMS P-T conditions can provide further insights into evolution of the western North American margin.

## Open Research Section

For the purposes of peer review, all data collected by the authors have been uploaded as excel files and indexed in the supplementary information. If this manuscript is accepted, the authors will submit the data to the ETH Research Collection, a publicly-accessible repository for research data and publications. These files will be publicly available and indexed with a DOI.

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