

Age, provenance, and geochemical relationships amongst the Great Valley Group, Coast Range Ophiolite, and Franciscan subduction complex at Del Puerto Canyon, central California

Devon A. Orme^{1,†}, Natalee Weis¹, Kathleen D. Surpless², Andrew K. Laskowski¹, James L. Crowley³, Nicole Stine¹, Ian Colliver¹, Carly Ross¹, and Emma Sweet¹

¹Department of Earth Sciences, Montana State University, 226 Traphagen Hall, Bozeman, Montana 59717-3480, USA

²Department of Geosciences, Trinity University, One Trinity Place, San Antonio, Texas 78212, USA

³Department of Geoscience, Boise State University, Environmental Research Building 1160, MS 1535, Boise, Idaho 83725, USA

ABSTRACT

The Great Valley Forearc (GVF) basin of California, USA, preserves an extensive rock record of the Jurassic–Paleogene tectonic development of the California segment of the North American Cordillera. We present new U–Pb geochronology, zircon and whole-rock geochemistry, and petrographic analyses from the Great Valley Group (GVG), Franciscan subduction complex, and Coast Range Ophiolite (CRO) in the northern San Joaquin Valley to better understand the timing and location of initial forearc sedimentation, and how sediment routing systems may have evolved during Cretaceous time. Basal GVG strata of the Knoxville Formation were deposited ca. 145–140 Ma and are separated by an ~40 m.y. unconformity with overlying strata of the Upper Cretaceous Panoche Formation. Pre-Mesozoic zircon grains are present in both the Knoxville and Panoche formations, but are sparse (0%–7%) compared to other GVG sandstones. Zircon geochemistry records felsic igneous sources (Th/U 0.9–0.2) during both periods of deposition, and epsilon Hf signatures reveal a shift from juvenile to more evolved sources between Knoxville and Panoche deposition. Whole-rock geochemistry shows increasing compositional maturity from latest Jurassic crystallization of the CRO to Early and Late Cretaceous deposition of the GVG. Integrating these data, we present a tectonic model for the northern San Joaquin portion of the GVF basin from ca. 145 Ma to 80 Ma that documents the onset of basin deposition and details sediment pathways during the Early to earliest Late

Cretaceous. In addition, we discuss potential drivers for the ~40 m.y. unconformity within the San Joaquin Valley and implications of this work for global forearc basin processes.

INTRODUCTION

Forearc basins are a common component of subduction zones and form between the oceanic trench and volcanic arc (e.g., Dickinson, 1973, 1974; Karig, 1974; Seely et al., 1974; Dickinson, 1995; Noda, 2016). The sedimentary record of ancient forearc basins provides a record of the tectonics of extant subduction zones, including insight into spatial and temporal changes in topography and geochemistry of the upper plate (e.g., Ingersoll, 1982; Clift et al., 2000; DeGraaff-Surpless et al., 2002; Surpless et al., 2019; Kaporas et al., 2024). Forearc basin morphology often varies significantly along different parts of the same convergent margin, transitioning amongst end-member models for forearc basin classification, including whether the basin is landward of an accretionary or non-accretionary part of the margin (e.g., von Huene and Scholl, 1991; Stern, 2002; Clift and Vannucchi, 2004; Noda, 2016).

Forearc basin development depends on the mechanisms that generate and accommodate sediment. Periods of high flux magmatism result in crustal thickening and an increase in topographic elevation of the volcanic arc, which leads to enhanced erosion during and tens of millions of years after the increased magmatism (e.g., Lee et al., 2015; Karlstrom et al., 2018), resulting in increased rates of sediment accumulation within adjacent forearc, trench-slope, and trench basins (e.g., Noda, 2018; Orme and Graham, 2018). Without accommodation to pond sediment within a forearc basin, sediment from the upper plate will funnel directly to the

trench. Some working models for the generation of forearc basin accommodation include downward flexure from lithosphere load (e.g., Dickinson, 1995), trenchward growth and/or activation of back thrust structures within an accretionary prism due to increased basal friction (e.g., Scholl et al., 1980; Larroque et al., 1995; Dickinson, 1995; Noda, 2018), reversal of surface slope angle of the accretionary prism (i.e., negative alpha basin) as seen in numerical modeling (Tharp, 1985; Byrne et al., 1993; Fuller et al., 2006), and forearc extension (e.g., Trop et al., 2002; Stern, 2004; Romero et al., 2024).

The Great Valley forearc (GVF) basin in central California, USA, records sedimentation during latest Jurassic–earliest Cretaceous to Paleogene oceanic subduction along the North American margin and represents an opportunity to investigate the morphology of a long-lived ancient trench–forearc system. Located between the Franciscan subduction complex to the west and the Sierra Nevada and Klamath magmatic arcs to the east and north, respectively, these three lithotectonic units represent the major components of the ancient subduction zone (Fig. 1A; e.g., Cowan and Bruhn, 1992; Dickinson, 1995, 2004). Exposures of the early GVF basin stratigraphy, the Jurassic–Cretaceous Great Valley Group (GVG), are found within the central Coast Ranges, primarily in contact with the ancient accretionary wedge known as the Franciscan subduction complex. Often separating rocks of the GVF basin and the Franciscan subduction complex are remnants of middle–Late Jurassic ophiolitic material, collectively named the Coast Range Ophiolite (CRO). The CRO forms the basement to the GVF along its western margin. In the southern GVF, the San Joaquin Valley (SVJ), the Franciscan subduction complex, CRO, and GVG crop out along Del Puerto Canyon, Cali-

[†]devon.orme@montana.edu

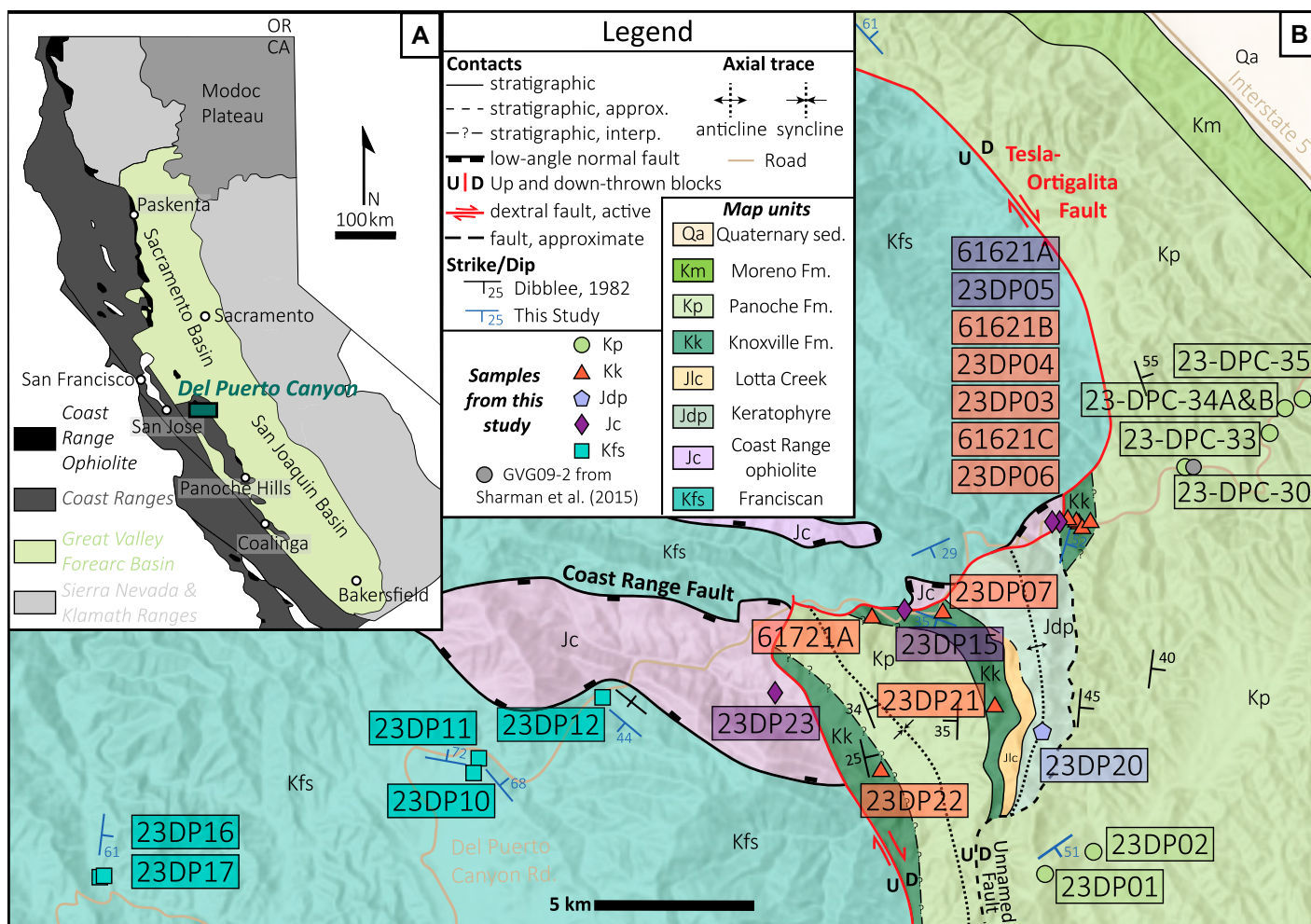


Figure 1. (A) Generalized geologic map of California highlighting the spatial extent of the Great Valley Forearc basin, modified from Orme and Graham (2018). (B) Revised geologic map of Del Puerto Canyon after Dibblee (1982), Dibblee and Minch (2007), Maddock (1964), Harms et al. (1992), and the results of this study. Stacked sample names are ordered from west to east, with the farthest west sample on top.

fornia, ~100 km and ~35 km southeast of the cities of San Francisco and San Jose, respectively (Fig. 1B). The exposure at Del Puerto Canyon provides a unique opportunity to study the age, geochemical, and provenance relations amongst the three lithotectonic units, which improves our understanding of the development of the GVF basin along the North American margin.

Studies of the oldest stratigraphy of the GVF (latest Jurassic–earliest Cretaceous) primarily focus on the western outcrop belt of the Sacramento basin, where the contact between the CRO and GVG is beautifully exposed and ample stratigraphic sections through Lower and Upper Cretaceous GVG are accessible for study (e.g., Lawton, 1956; Jones et al., 1969; Ingersoll, 1976, 1978, 1982; Romero et al., 2024). In contrast, outcrop exposures of the CRO–Lower Cretaceous GVG within the SJV are limited, obscured by later deformation that juxtaposes the Franciscan subduction complex against Cenomanian

(Upper Cretaceous) GVG strata (e.g., see summary by Moxon, 1988). Del Puerto Canyon is one of only a handful of locations within the SJV where the relationship between the CRO and GVG may be evaluated. At Del Puerto Canyon, the upper volcanics of the CRO immediately pre-date GVG sedimentation, which motivated us to investigate the role, if any, forearc basement may play in driving forearc basin initiation. In this study, we integrate zircon U-Pb geochronology, zircon trace element and Lu-Hf geochemistry, whole-rock geochemistry, and sandstone petrography to characterize the provenance and depositional ages of GVG strata and Franciscan subduction complex exposures in Del Puerto Canyon, California. We also use petrography and geochemistry to determine the composition of the underlying volcanics of the CRO to interrogate its tectonic origin and relationship to initial GVG sedimentation. Integration of these complementary methods represents a powerful

approach for basin analysis studies from which we interpret a tectonic model for the coevolution of these distinct lithotectonic units that aims to clarify the Jurassic–Cretaceous North American margin and the formation of the GVF basin. In turn, we compare our observations with those of another coeval ancient forearc basin to highlight the feedback between arc magmatism and subduction complex evolution in driving forearc basin subsidence.

GEOLOGIC BACKGROUND

Jurassic–Cretaceous California Margin

The preservation of the Franciscan subduction complex and segments of the CRO within the Coast Ranges, the GVF basin, and magmatic arcs of the Sierran and Klamath mountains provide a world-class natural laboratory for the study of ancient subduction dynamics.

The Coast Ranges are primarily composed of the Franciscan subduction complex, which is an assemblage of heterogeneous terranes characterized by varying metamorphic grades and lithologies that formed the accretionary wedge outboard of the GVF basin (Dickinson, 1981; Shervais et al., 2005; Hopson et al., 2008; Wakabayashi, 2011, 2015). Franciscan-related subduction is interpreted to be active as early as ca. 180 Ma (Mulcahy et al., 2018), with pulses of accretion between 148 Ma and 132 Ma (e.g., Dumitru et al., 2018; Wakabayashi and Dumitru, 2007; Snow et al., 2010) and intervening periods of subduction erosion until the Franciscan wedge became highly accretionary at ca. 123 Ma (Dumitru et al., 2010; Apen et al., 2021). The main stage of Franciscan exhumation was from ca. 100 Ma to 70 Ma with continuation into the Cenozoic (Tagami and Dumitru, 1996; Wakabayashi and Unruh, 1995). Inboard and within exposures of the Franciscan complex, the Coast Ranges include remnants of ophiolitic material that comprise the CRO and form the basement to the GVF along its eastern margin. The main stages of crystallization of ophiolitic remnants in California are between 168 Ma and 161 Ma (Hopson et al., 2008). Younger volcanic rocks of the CRO, found stratigraphically below or interbedded with the basal GVG, are ca. 152–144 Ma in age (Hopson et al., 1981, 2008). The tectonic setting in which the CRO remnants formed is debated and includes back-arc–inter-arc basin lithosphere (e.g., Moores, 1970; Schweickert and Cowan, 1975; Moores and Day, 1984; Ingersoll and Schweickert, 1986; Dickinson et al., 1996a, 1996b; Huot and Maury, 2002; Ingersoll, 2019, and references therein), paleo equatorial mid-oceanic rift spreading center (Berger and Winterer, 1974; Evarts, 1977; Evarts and Schiffman, 1983; Robertson, 1989; Hopson et al., 1991, 1996; Hull et al., 1993), and parautochthonous forearc lithosphere (e.g., Saleeby, 1982, 1996; Harper and Wright, 1984; Shervais and Kimbrough, 1985; Wyld and Wright, 1988; Shervais, 1990; Saleeby et al., 1992; Stern and Bloomer, 1992; Saleeby and Harper, 1993; Huot and Maury, 2002; Arkula et al., 2023).

The GVG consists of deep marine turbidites and shallow marine deposits, generally shoaling upward (e.g., Ojakangas, 1968; Ingersoll, 1978; Mansfield, 1979). The depositional age of the GVG was first determined through biostratigraphy, with the onset of sedimentation dated to the Late Jurassic (Jones et al., 1969; Imlay and Jones, 1970). The stratigraphically deepest GVG has also been dated using detrital zircon U–Pb geochronology, which yields variable maximum depositional ages (MDAs) of ca. 151–135 Ma, thus indicating that the beginning of the GVG deposition was diachronous during Late Jurassic

to Early Cretaceous time (Orme and Surpless, 2019; Romero et al., 2024). The provenance of the latest Jurassic–earliest Cretaceous GVG is mainly derived from Jurassic volcanics of the Sierran-Klamath magmatic arcs and North American continental interior, whereas the Late Cretaceous GVG is primarily sourced from deeply eroded portions of the proximal magmatic arc (e.g., DeGraaff-Surpless et al., 2002; Sharman et al., 2015; Dumitru et al., 2015; Orme and Surpless, 2019; Surpless et al., 2019; Romero et al., 2024). Volcanic sources for the GVG were first identified through petrographic analysis of sandstone, conglomerate clast counts, and clay mineralogy (e.g., Ojakangas, 1968; Ingersoll, 1979, 1983). The petrofacies in the GVG contain metamorphic, plutonic, and volcanic arc detritus, thus recording eroded magmatic arc as a source for GVG (Ojakangas, 1968; Dickinson and Rich, 1972; Ingersoll, 1978, 1983; Mansfield, 1979; Romero et al., 2024). Petrographic analysis shows that the lower GVG is dominated by volcanoclastic and sedimentoclastic sandstones, whereas the upper GVG is dominated by plutoclastic sandstones (Ingersoll, 1978). Detrital zircon studies that analyze the provenance of the GVG record a shift from generally multimodal age spectra to unimodal spectra in middle Cretaceous samples, interpreted to indicate changing source morphologies and sediment routing systems to the GVF (DeGraaff-Surpless et al., 2002; Dumitru et al., 2010; Sharman et al., 2015; Romero et al., 2024). Trace element geochemistry reveals basaltic and andesitic source composition for much of the Cretaceous strata within the northern GVF (the Sacramento basin) and the southern GVF (the San Joaquin basin), with significant sedimentary recycling (e.g., Cox et al., 1995; Surpless, 2015).

During the Early to Middle Jurassic Period, regions known today as the Sierra Foothills and portions of the Klamath Mountains accreted to the North American margin, after which the GVF developed (e.g., Schweickert and Cowan, 1975; Ingersoll and Schweickert, 1986; Ingersoll, 2000; Dickinson, 2008). Permo–Triassic to Cretaceous intra-arc strata of the Sierra Nevada reflect derivation from proximal arc activity, as well as from pre-Mesozoic sources from southwestern Laurentia (e.g., 1400–900 Ma, 500–300 Ma age modes; Attia et al., 2021). During Late Jurassic–Miocene Franciscan subduction and coeval deformation, calc-alkaline plutons of the Sierran batholith were emplaced in the North American upper crust (e.g., Schweickert et al., 1984; Barth et al., 2011; Cecil et al., 2012). There were magmatic high flux periods during Triassic (ca. 250–190 Ma), Middle and Late Jurassic (ca. 180–150 Ma), and mid-Cretaceous (ca. 125–75 Ma) time, with each episode involving varying depths

of pluton emplacement and magma addition rates (e.g., Paterson and Ducea, 2015).

Del Puerto Canyon, California

Del Puerto Canyon, located in the northern Diablo Range ~35 km southeast of San Jose, California, is part of the San Joaquin subbasin of the southern GVF basin (Fig. 1). Previous studies of Del Puerto Canyon have focused on detailed mapping and outcrop (e.g., Maddock, 1964; Dibblee, 1982; Dibblee and Minch, 2007) and petrologic descriptions of the CRO, with the latter exposed as an ~5.4-km-thick ophiolite sequence (e.g., Evarts, 1977, 1978; Evarts et al., 1992, 1999). Pioneering mapping of the region by Dibblee (1982) and Dibblee and Minch (2007) included defining the Del Puerto Keratophyre and the Lotta Creek Tuff as part of the CRO, and work by Maddock (1964) further divided the GVG into the Jurassic Knoxville Formation, Albion Shale, Upper Cretaceous Panoche Formation, and overlying Moreno Formation (van Vleck Anderson and Pack, 1915; Dibblee, 1982; Dibblee and Minch, 2007; Maddock, 1964; Bailey et al., 1970; Raymond, 1970, 1973; Figs. 1 and 2). We did not find the Albion Shale denoted by Maddock (1964) in our field area and therefore do not include it on our composite lithotectonic summary (Fig. 2; see Results section for further explanation).

There are two major fault systems in Del Puerto Canyon: the Coast Range fault and the Tesla-Ortogonalita fault (e.g., Vickery, 1925; Briggs, 1953; Bailey et al., 1970; Raymond, 1973; Jayko and Blake, 1989; Harms et al., 1992; Fig. 1). The Coast Range fault juxtaposes serpentinized harzburgite and zeolite facies of the CRO against jadeite-lawsonite-bearing metagraywacke of the Franciscan subduction complex (Jayko and Blake, 1989; Evarts, 1978). In the Diablo Range, map relations and kinematic analyses of the Coast Range fault are consistent with low-angle normal faulting and syn-subduction extensional offset of the CRO and subsidence of the GVF basin (Unruh, 2021; Wakabayashi, 2015; Unruh et al., 2007; Harms et al., 1992; Jayko and Blake, 1989). Extension is interpreted to have occurred as the Franciscan subduction complex was exhumed during latest Cretaceous time (Unruh, 2021; Unruh et al., 2007; Harms et al., 1992).

The Tesla-Ortogonalita fault is a high-angle (70° vertical dip) fault system with various strikes that form an almost continuous boundary between the Franciscan and GVG or younger Cenozoic basin fill from Del Puerto Canyon south to the Panoche Valley (Anderson and Pack, 1915; Cotton, 1972; Raymond, 1973; Anderson et al., 1982; Harms et al., 1992). The Tesla-Ortogonalita fault cuts beds

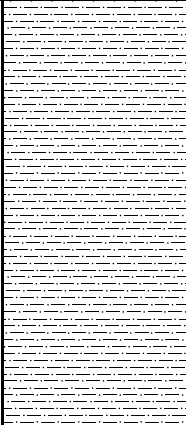
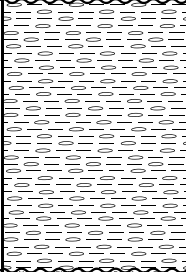
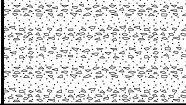
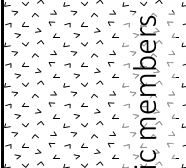
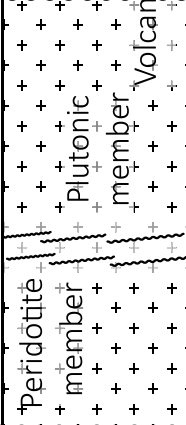
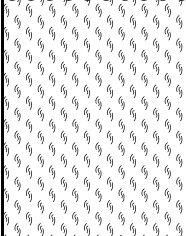
Thickness meters approx. scale	Lithology	Formation	Description	Depositional or Crystallization Age	Samples from this study *subsurface	U-Pb	Lu-Hf	Petrog.	GC
6,400		Panoche Formation (Kp)	Massive, fine-medium grained, concretionary sandstone with calcareous or argillaceous matrix. Soft, gray, silty shale interbedded with thin, fine grained layers of sandstone, and coarse conglomerate. Basal Adobe Flat Shale Member (Kpaf): Hard, dark gray to black silty shales. Gradational contact with the rest of the Panoche Formation. Concretions of gray, aphanitic limestone. Pea size conglomerate horizons mark unconformity between Kpaf and Jk. (Maddock, 1964; Anderson and Pack, 1915).	Turonian-Campanian 73-93 Ma (Bennison et al., 1991; Hosford Scheirer and Magoon, 2008)	23-DPC-35*	○			○
					23-DPC-34A*	○			
					23-DPC-34B*	○			
					23-DPC-33*	○	○		
					23-DPC-30*				○
					23DP02	○	○	○	○
1,500		Knoxville Formation (Kk)	Dark gray to black clay shale with tuffaceous and radiolarian debris. Limestone concretions abundant near basal contact. Base Mg Ox stained red yellow chert and sandstone along faulted base, but primary depositional relationship with CRO preserved (This Study)	Kimmeridgian-Berriasian 145 Ma (This Study)	23DP06	△	△	△	
					61621C	△		△	
					23DP03	△		△	
					23DP04G				△
					61621B	△		△	
					23DP07				△
					23DP21	△		△	
					61721A	△			
270		Lotta Creek Formation (Jlc)	Tuffaceous radiolarian chert with volcanic breccias, limited exposure of white albite tuffs lithics, and dark green shale. (Maddock, 1964).	Tithonian 149 Ma (Evarts et al. 1992; Maddock, 1964)					
450		Upper volcanics of Coast Range Ophiolite: Del Puerto Keratophyre	Intermediate volcanic rock. Groundmass of chlorite, albite, quartz, and magnetite, with phenocrysts of mostly albite and minor quartz. Bounded by faults. (Maddock, 1964).	Tithonian (?) (Maddock, 1964)	23DP20			◇	
5500 m		Coast Range Ophiolite (Jc)	Lower volcanic member: dike and sill complexes Plutonic member: mafic cumulates, basalt, plagiogranites overlain by hornblende rich gabbro cumulates and mafic to silicic intrusions. Serpentinite shear zone with mafic and ultramafic blocks Peridotite member: harzburgite, dunite and wehrlite with chromite, pyroxenite and piritic dikes (Evarts et al. 1992)	Oxfordian 158-161 Ma (Mattinson and Hopson, 2008; Lanphere, 1971)	61621A				◇
					23DP05			◇	◇
					23DP15			◇	
					23DP23			◇	
2,400- 6,000 not well documented in Del Puerto Canyon		Franciscan (Kfs)	Dark gray massive to thinly bedded metasandstone with interbedded siltstone and gray slatey shale. Ribbon chert horizons, melange, and rare block-in-matrix melange with high grade metamorphic blocks (Maddock, 1964; this study).	Aptian-Santonian (this study) Callovia to Berriasian 145-164 Ma (Maddock, 1964; Camp, 1942)	23DP12	□	□		
					23DP11	□			
					23DP10			□	
					23DP17	□		□	
					23DP16	□		□	

Figure 2. Tectonic stratigraphic column of Del Puerto Canyon, including sample names and analytical data type for this study. All thickness data are from the same reference as the description (van Vleck Anderson and Pack, 1915; Maddock, 1964; Bailey et al., 1970; Lanphere, 1971; Evarts, 1977; Dibblee, 1982; Dibblee and Minch, 2007). The upper age of the Panoche Formation at Del Puerto Canyon is reported to be as old as ca. 93 Ma from biostratigraphy (Bennison, 1991) but youngs within the subsurface to the south to ca. 83.5 Ma (Hosford Scheirer and Magoon, 2008). CRO—Coast Range Ophiolite; GC—geochemistry.

of the GVG including the Panoche Formation (Vickery, 1925; Huey, 1948; Raymond, 1973), and offsets segments of the Coast Range fault (Raymond, 1973; Fig. 1). The oldest beds the Tesla-Ortogonalita fault cuts are Lower Cretaceous (Albian) in age, so the fault must have been active after this time; in some areas, the fault truncates Miocene beds, indicating motion during the Neogene (Huey, 1948; Raymond, 1973). South of Del Puerto Canyon, the fault is referred to as the Ortogonalita fault zone and is characterized by right-lateral late Pleistocene–Holocene (active) motion and a series of pull-apart basins, including one that hosts San Luis Reservoir (van Vleck Anderson and Pack, 1915; Anderson et al., 1982; Petersen et al., 1996; Bryant and Cluett, 2000). At Del Puerto Canyon, motion along segments of the Tesla-Ortogonalita fault is complex, with reverse, normal, and dextral motion interpreted during various periods of this fault zone's history. Maddock (1964) and Harms et al. (1992) suggest the fault may have originated as a normal fault, associated with the Coast Range fault, which down-dropped the GVG relative to the Franciscan subduction complex as the latter was exhumed during latest Cretaceous time. Maddock (1964) and Raymond (1973) discuss how variations in fault dip, and the resulting interpretations of both thrust or normal-sense of motion, reflect the Tesla-Ortogonalita fault's complex relationship with the Coast Range fault and Cenozoic right-lateral deformation that post-dated initiation of the fault after Albian time.

The Franciscan subduction complex in Del Puerto Canyon defines the western portion of the field area and consists of mostly metasandstone and lesser components of greenstone, chert, and mélange (e.g., Raymond, 1973; Maddock, 1964; Dibblee, 1982; Dibblee and Minch, 2007). The metasandstone portion is dark gray to gray, massive to thinly bedded, fine-grained graywacke that is interbedded with siltstone or dark gray slaty shale (Maddock, 1964; Dibblee, 1982; Dibblee and Minch, 2007). Most metasandstones are jadeite-lawsonite-bearing and few high-grade blocks of amphibolite are present within mélange outcrops (Evarts et al., 1999; Bennison et al., 1991). Blocks of chert within the Franciscan are thinly bedded, brittle, red, white, green in color, and locally contorted (Dibblee, 1982). The greenstone (metabasalt) is black, massive, and locally brecciated, and it is often stained orange to red with iron and magnesium oxide (Maddock, 1964; Dibblee, 1982; Dibblee and Minch, 2007).

Biostratigraphy and preserved macrovertebrates suggest the Franciscan subduction complex in Del Puerto Canyon was accreted between ca. 152 Ma and 145 Ma (Camp, 1942; Crittenden, 1951; Hall, 1958; Maddock, 1964). These constraints on depositional age are older than those from detrital zircons extracted from Franciscan sandstones within the Diablo Range, north of Del Puerto Canyon, which have MDAs between ca. 115 Ma and 84 Ma (Dumitru et al., 2018; Apen et al., 2021; this study).

The CRO in Del Puerto Canyon is deformed and fragmented oceanic meta-igneous and igneous assemblages, which includes gabbroic dikes, calc-alkaline volcanics, chert, and serpentinite (Bailey et al., 1970). In contrast to the Franciscan subduction complex, the CRO is primarily zeolite grade, with the exception of higher-grade metamorphic facies lower in the ophiolite pseudostratigraphy, interpreted as evidence for sea-floor hydrothermal metamorphism (Evarts and Schiffman, 1983). The crystallization ages from the CRO are between 161 Ma and 155 Ma, recorded by U–Pb in zircon and $^{40}\text{Ar}/^{39}\text{Ar}$ in hornblende found in plagiogranite veins in quartz diorite and tonalite (Lanphere, 1971; Hopson et al., 1981; Evarts et al., 1992; Mattinson and Hopson, 2008). A series of dikes that cut the cumulates and a rhyolite at the top of the volcanic portion of the ophiolitic succession yields dates between ca. 150 Ma and 149 Ma (Evarts et al., 1992). The Del Puerto Keratophyre is a mappable unit, interpreted as the upper volcanic section of the CRO as first defined by Dibblee (1982) and described by Evarts et al. (1992), and consists of intermediate volcanic rock that has a groundmass of chlorite, albite, quartz, and magnetite, with phenocrysts including albite and minor quartz (Dibblee, 1982; Dibblee and Minch, 2007; Maddock, 1964). The Del Puerto Keratophyre is bounded by the Tesla-Ortogonalita fault to the north and an unnamed fault to the east and is deformed into an anticline (Maddock, 1964; Raymond, 1973; Dibblee, 1982; Dibblee and Minch, 2007).

The Lotta Creek Formation is stratigraphically above the Del Puerto Keratophyre and crops out south of the main canyon (Figs. 1B and 2). The volcanic formation is composed of tuffaceous radiolarian chert with volcanic breccia, limited exposure of white crystalline tuff (albite), white tuffaceous sandstone, lithics, and dark green shale (Maddock, 1964; Raymond,

1970, 1973; Dibblee, 1982; Dibblee and Minch, 2007). An andesite boulder within the formation dates to ca. 150 Ma ($^{40}\text{Ar}/^{39}\text{Ar}$ on hornblende: Evarts et al., 1992).

The Knoxville Formation conformably and gradationally overlies the Lotta Creek Formation and contains dark gray, black, and purple clay shales with gray limestone concretions near the basal contact containing fossils and thinly bedded sandstone layers interbedded with the shales (Figs. 1B and 2; Crittenden, 1951; Hall, 1958; Maddock, 1964). In contrast to the Franciscan subduction complex, the Knoxville Formation shows minimal diagenesis or metamorphism, with chlorite supporting up to zeolite grade in places. The Knoxville Formation is classically assigned a latest Jurassic depositional age and is the most basal member of the GVG within the SJV (e.g., Crittenden, 1951; Hall, 1958; Maddock, 1964). At Del Puerto Canyon, the Knoxville Formation, previously referred to as the Hawk Shale at this locale, contains Late Jurassic (Tithonian) to Early Cretaceous (Valanginian) age *Buchia* species (Schilling, 1962).

The Panoche Formation unconformably overlies the Knoxville Formation and makes up most of the GVG strata mapped in Del Puerto Canyon (Figs. 1B and 2; van Vleck Anderson and Pack, 1915; Maddock, 1964; Dibblee, 1982; Dibblee and Minch, 2007). A thin, pea-size conglomerate horizon marks the unconformity between Knoxville and Panoche (van Vleck Anderson and Pack, 1915; Maddock, 1964). The lower Turonian Adobe Flat Shale Member forms the base of the Panoche Formation and is a hard, dark gray to black silty shale that grades into overlying sandstones (Maddock, 1964). By contrast, the bulk facies found within the Panoche Formation consist of massive, fine- to medium-grained sandstone with argillaceous or calcareous matrix interbedded with soft, brown to gray, silty shale (van Vleck Anderson and Pack, 1915; Maddock, 1964). The Panoche Formation has been dated using fossils and U–Pb MDAs from detrital zircons to be ca. 93–87 Ma (Maddock, 1964; Sharman et al., 2015; this study).

The Moreno Formation is present in the northeast corner of the field area and is composed of brown clay shale that has limestone concretions and interbedded massive gray sandstone (Fig. 1B; van Vleck Anderson and Pack, 1915; Maddock, 1964; Raymond, 1970; Bishop, 1970). It overlies the Panoche Formation with grada-

tional depositional contact, and only the lower portion of the formation is visible in this field area (Maddock, 1964). The presence of foraminifera, ammonites, and reptile remains indicates marine deposition of the Moreno Formation in the latest Late Cretaceous (Maddock, 1964).

METHODS

Field Geology and Core Repository

At each field site, we took GPS coordinates, examined contacts, conducted a detailed lithofacies description, and measured strike and dip of bedding. We were unable to complete measured stratigraphic sections or detailed mapping beyond single field stations due to limited access to private property. At many field stations, we collected samples for detrital U-Pb geochronology ($N = 13$), Lu-Hf geochemistry ($N = 4$), sandstone petrography ($N = 8$), igneous and/or metamorphic petrology ($N = 7$), and whole-rock geochemistry ($N = 5$; Figs. 1B and 2). On Figures 2–4, our samples are presented in their west to east context as (1) the gross structure of the study area preserves the original architecture of the subduction zone margin, moving from the trench (Franciscan) to the upper plate (CRO and GVG), and (2) the age of the units gets younger moving east from the Jurassic CRO through Early to Late Cretaceous GVG strata, reflecting the overall eastward dip of stratigraphy. We note that the Franciscan subduction complex in the western portion of the field area is younger than the CRO, but structurally below it.

Given our limited access to the field area mapped as the Panoche Formation, we collected subsurface sandstone samples for U-Pb geochronology ($N = 4$), Lu-Hf ($N = 1$), and whole-rock geochemistry ($N = 2$) from the California Core Repository in Bakersfield, California (Fig. 2). Open access drilling reports and well logs with gamma and sonic ray curves were used to identify sandstone intervals in core sections. To ensure enough zircon yield for detrital analyses, each sample is a composite from core chippings from -91 m and -293 m below the Kelly Bushing datum (Supplemental Material File S1¹;

Fig. 1B). All the chipping samples we collected were consistently medium-grained sandstone and did not exhibit any changes in bulk mineralogy with depth.

Geochronology

U-Pb geochronology was completed for detrital zircons from 18 sandstone samples to determine provenance and maximum depositional ages of sedimentary rocks (Williams, 1997; Gehrels, 2011). Samples include nine medium-grained sandstones from mapped outcrops or subsurface wells of the Panoche Formation, one medium-grained sandstone from mapped Knoxville Formation, four medium-grained sandstones at the easternmost contact with the CRO that are mapped as Knoxville Formation, Albion Shale, or lower Turonian Adobe Flat Shale member of the Panoche Formation (Maddock, 1964; Bishop, 1970; Dibblee, 1982; Dibblee and Minch, 2007), and four medium- to coarse-grained sandstones from the Franciscan subduction complex (Fig. 1B). Mineral separation for zircon was completed using standard heavy mineral separation techniques (e.g., Gehrels et al., 2008). We used backscattered electron images to randomly select detrital zircon grains ($n = 315$) without significant inclusions or cracks.

We primarily analyzed zircons with laser ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) at the University of Arizona LaserChron Center, Tucson, Arizona, USA, using a Thermo Element2 single collector ICP-MS. U-Pb analyses were conducted with a 20 μ m spot diameter, resulting in a 12 μ m pit depth on each evaluated grain. U-Pb data were filtered to exclude ages with high common Pb, $>5\%$ reverse discordance, $>10\%$ uncertainty, or $>20\%$ discordance. The reported ages are based on the $^{206}\text{Pb}/^{238}\text{U}$ ages for grains $<ca. 1.0$ Ga and on the $^{206}\text{Pb}/^{207}\text{Pb}$ ages for grains $>ca. 1.0$ Ga (Gehrels et al., 2008; Gehrels, 2011). Th/U ratios measured during this analysis were used for zircon trace element geochemical analysis. All data were reduced using AgeCalcML, a MATLAB-based graphical user interface for reducing, visualizing, and reporting data collected by LA-ICP-MS (Sundell and Saylor, 2021). U-Pb data are reported in Supplemental Material File S1.

Two sandstones (61721A and 23DP22) analyzed for U-Pb geochronology with LA-ICP-MS were further interrogated with chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-ID-TIMS). CA-ID-TIMS provides a more precise age than LA-ICP-MS, therefore allowing closer inspection of outlier grains, such as a young tail of single ages that do not overlap the bulk of analyses. CA-ID-TIMS was performed at Boise State

University, Boise, Idaho, USA, following the protocols modified after Mattinson (2005). Cathodoluminescence (CL) images were obtained with a Hitachi TM4000Plus scanning electron microscope (Supplemental Material File S2). Grains selected for TIMS analysis based on CL images and LA-ICP-MS data were removed from the mounts and annealed at 900 °C for 60 h in a muffle furnace. Following chemical abrasion at 190 °C for 12 h, zircon was spiked with the Boise State University mixed ^{233}U – ^{235}U – ^{205}Pb tracer solution (BSU-1B) and dissolved and rinsed in hydrofluoric acid. Uranium and Pb were separated from the zircon matrix using an hydrochloric acid–based anion-exchange chromatographic procedure (Krogh, 1973), eluted together and dried with 2 μ l of 0.05 N H_3PO_4 . Pb and U were loaded on a single outgassed Re filament in 5 μ l of a silica-gel/phosphoric acid mixture (Gerstenberger and Haase, 1997), and U and Pb isotopic measurements made on a GV IsoProbe-T multicollector thermal ionization mass spectrometer equipped with an ion-counting Daly detector. Pb isotopes were measured by peak-jumping all isotopes on the Daly detector for 160–200 cycles. CA-ID-TIMS U-Pb dates and uncertainties (two-sigma) were calculated using the algorithms of Schmitz and Schoene (2007), U decay constants recommended by Jaffey et al. (1971), and $^{238}\text{U}/^{235}\text{U}$ of 137.818 (Hiess et al., 2012). We corrected $^{206}\text{Pb}/^{238}\text{U}$ ratios and dates for initial ^{230}Th disequilibrium using $D_{\text{Th/U}} = 0.20 \pm 0.05$ (1s) and the algorithms of Crowley et al. (2007). All common Pb in analyses was attributed to laboratory blank and subtracted based on the measured laboratory Pb isotopic composition and associated uncertainty. Analytical data can be found in Supplemental Material File S3.

Zircon U-Pb Age Spectra

The complete distribution of detrital zircon ages within a sample can help characterize source terranes and provide information on the tectonic setting of strata (e.g., Cawood et al., 2012; Gehrels, 2011, 2014). AgeCalcML was used to construct a series of vertically stacked histograms and kernel density estimates to visually compare age spectra from multiple samples (Fig. 3; Gehrels et al., 2008; Horstwood et al., 2016; Sundell et al., 2021). Kernel density estimates display the continuous probability of the age distribution of a sample, but do not incorporate uncertainties since the Gaussian kernel bandwidth can be set by the user to optimize data visualization based on the distribution of ages (Vermeesch, 2012; Sundell et al., 2021). In AgeCalcML (Sundell

¹Supplemental Material. Supplemental File S1: Sample information, zircon U-Pb (LA-ICP-MS) isotopic data, maximum depositional ages, sandstone modal analyses, zircon Lu-Hf geochemistry, whole-rock geochemistry, and detrital zircon U-Pb compilation. Supplemental File S2: Cathodoluminescence images of zircon grains analyzed by LA-ICP-MS and CA-ID-TIMS for U-Pb with corresponding dates. Supplemental File S3: CA-ID-TIMS U-Pb isotopic data. Please visit <https://doi.org/10.1130/GSAB.S.28820891> to access the supplemental material; contact editing@geosociety.org with any questions.

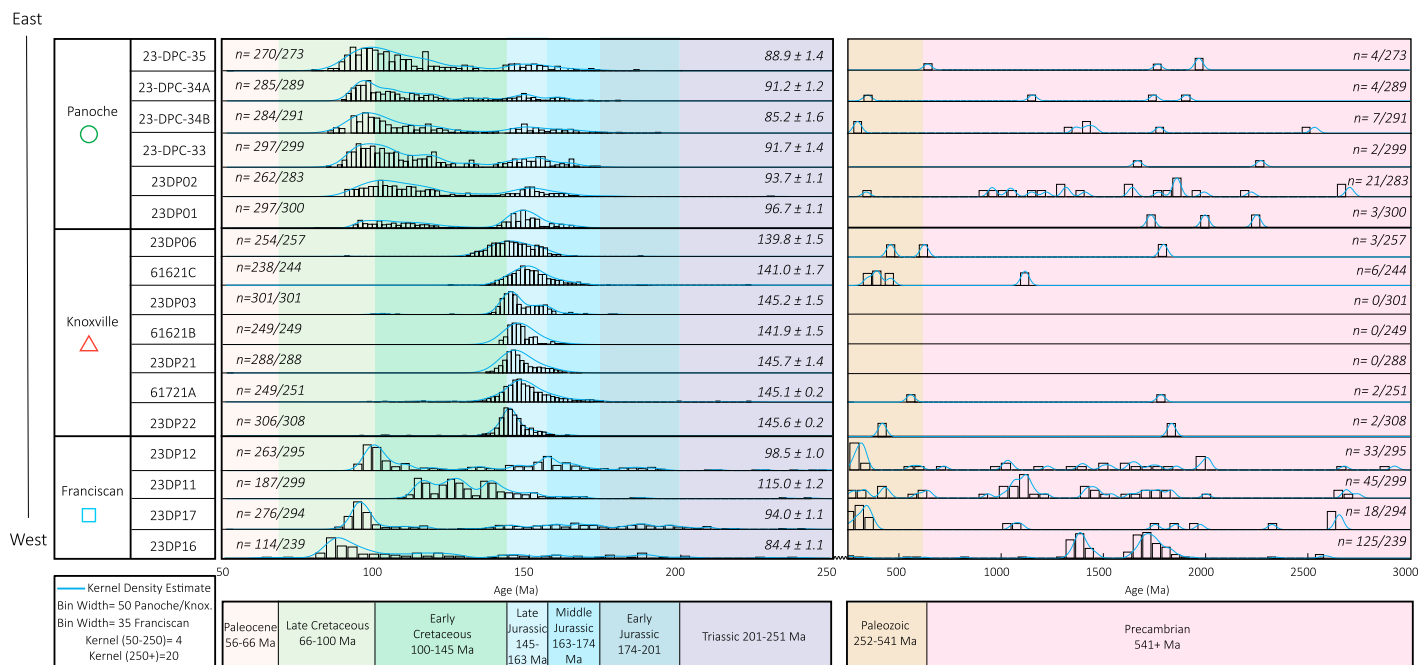


Figure 3. Detrital zircon kernel density estimates (KDEs) of 17 sandstones from Del Puerto Canyon stacked from east (top) to west and grouped by geologic formation. Mesozoic KDEs (250–50 Ma) are normalized to each other by formation and youngest statistical population maximum depositional age (MDA) of each sample reported (see Table 1 and Fig. 4 for additional MDA details). Following a break in scale, Paleozoic and Precambrian (3000–251 Ma) KDEs are normalized to each other and highlight the lack of these age grains within Great Valley Group sandstones at Del Puerto Canyon.

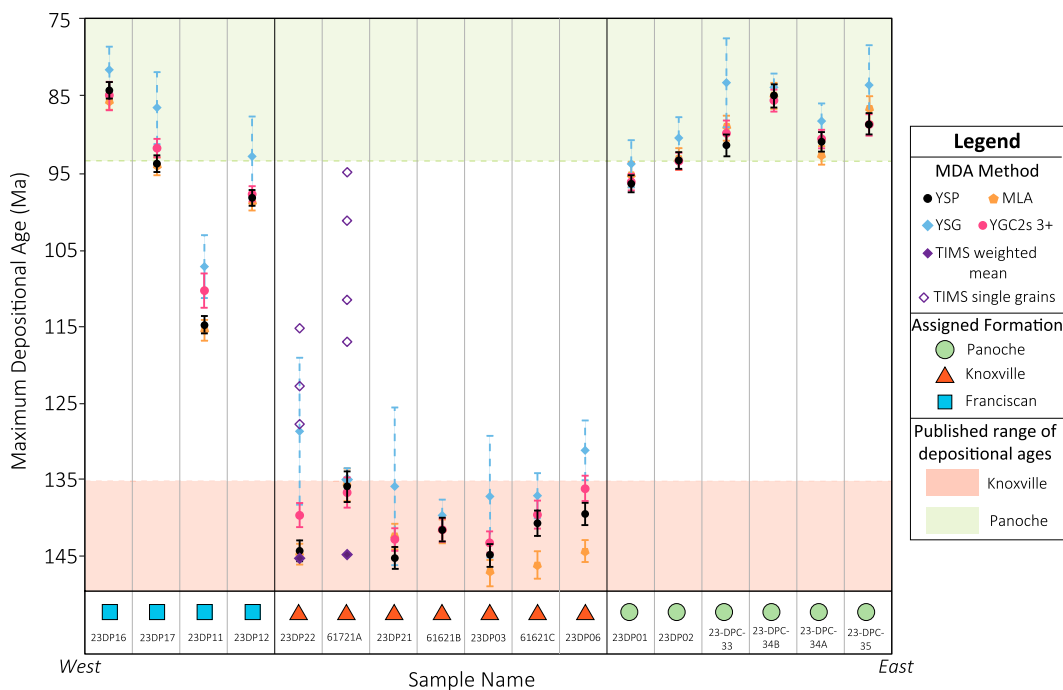


Figure 4. Maximum depositional ages for all 17 samples shown in Figure 3, calculated using methods described in the text including youngest statistical population (YSP), maximum likelihood age (MLA), youngest single grain (YSG), youngest cluster of three or more grains with overlapping error of 2-sigma (YGC2s 3+), thermal ionization mass spectrometry (TIMS) weighted mean. MLA and weighted means calculated with IsoPlotR (Vermeesch, 2018); all other methods calculated with DZmda (Sundell et al. 2024). Chemical abrasion–isotope dilution–TIMS grains shown as open purple diamonds did not overlap with uncertainties with other dates and therefore were not used for the weighted mean calculations. (See Discussion section for more details.)

et al., 2021), we set bin width for all GVG ages to 50 m.y. and a bin width of 35 m.y. for the Franciscan samples. We set a kernel band-

width for 250–50 Ma grains to 4 m.y., and kernel bandwidth for grains older than 250 Ma to 20 m.y.

Multidimensional scaling (MDS) was implemented to compare detrital zircon populations from the Panoche Formation at Del Puerto

Canyon to other known localities in the central-northern SJV using the MATLAB program DZmds from Saylor et al. (2018). Cross-correlation amongst samples ($N = 18$) allows for the visualization of the relative dissimilarity between detrital populations by converting intersample dissimilarity into disparity via linear transformation. In turn, the disparities are rearranged in Cartesian space while minimizing the misfit between disparity and Cartesian distance. We used the cross-correlation metric because it is more sensitive to the number and relative proportion of ages within a detrital age distribution, as compared to other statistical tools such as the Kolmogorov-Smirnov test (Vermeesch, 2013; Saylor et al., 2018). The grouping of samples in MDS space suggests similar age populations and is used to inform provenance interpretations amongst different areas (e.g., Quinn, et al., 2018; Ronemus et al., 2021; Surpless et al., 2024).

Maximum Depositional Ages

The youngest component of a sedimentary rock can be used to constrain its maximum depositional age (MDA), which may approximate the true depositional age (TDA) in convergent tectonic settings with active arcs and young zircon grains (e.g., Dickinson and Gehrels, 2009; Cawood et al., 2012; Spencer et al., 2016; Coutts et al., 2019; Herriott et al., 2019; Huang et al., 2022). There are a range of statistical methods used to determine the MDA, including the youngest single zircon grain (YSG), youngest cluster of at least three grains whose errors overlap within two sigma uncertainty (YGC2 σ 3+), and youngest statistical population (YSP; e.g., Dickinson and Gehrels, 2009; Coutts et al., 2019). YSP is a weighted average of the youngest group of two or more grains that have a mean square weighted deviation (MSWD) of ~ 1.00 (Coutts et al., 2019). We use the MATLAB-based program DZmda to calculate YSG, YSP, and YGC2 σ 3+ (Sundell et al., 2024). In addition, we consider the maximum likelihood age (MLA), which calculates an MDA using all grains from a sample, thus avoiding incorrectly identifying the youngest component of a sample (Galbraith, 1988; Galbraith and Laslett, 1993; Vermeesch, 2018, 2021). We calculate the MLA using IsoplotR Online (Vermeesch, 2018).

We report the age of the youngest single grain from each sample, with no filtering. However, prior to calculating YSP, YGC2 σ 3+, and MLA, we examined the youngest series of dates and did not include any grains that did not overlap with another grain within 2-sigma error in MDA calculations. We report a range of MDA calculations to highlight differences in the resulting MDA based on each method (Fig. 4), but we use

the YSP date as the best estimate TDA in Del Puerto Canyon, following Herriott et al. (2019). YSG and YGC2 σ 3+ often record MDAs that are too young and can routinely underestimate MDAs, whereas YSP calculations are more conservative and are less likely to capture erroneously young grains or tails that could have been affected by Pb loss (Herriott et al., 2019). For two samples with a scattering of young dates that do not overlap the youngest age population, we use weighted mean calculations of the youngest subset of grains that were analyzed using CA-ID-TIMS. Reasons for not including several CA-ID-TIMS dates in the MDA calculation are discussed within the Results section.

Lu-Hf Geochemistry

Lutetium-hafnium (Lu-Hf) isotopic ratios, reported as epsilon hafnium (ϵ Hf) values, can be combined with U-Pb age data to give insight on the type of magma from which zircon crystallized (Vervoort, 2014). The ϵ Hf values are categorized as evolved (negative ϵ Hf) or juvenile (positive ϵ Hf) in comparison to CHUR (chondritic uniform reservoir). Evolved signatures imply the zircon was crystallized from evolved arc magmas, often intruding older continental material, whereas juvenile signatures support crystallization from undifferentiated, more mafic magma sources (e.g., Hofmann, 1997; Vervoort, 2014). We analyzed five samples for Lu-Hf after U-Pb analysis at the Arizona LaserChron Center, using the practices outlined by Gehrels and Pecha (2014). We chose grains from the two primary zircon age populations found in GVG strata (160–140 Ma and 100–90 Ma), and we selected grains that would fit a 40-micron laser spot in addition to the 15-micron U-Pb analysis spot. After data reduction in AgeCalcML (Sundell et al., 2021), ϵ Hf values are plotted as individual points overlain on heat maps to distinguish between different units in the field area.

Whole-Rock Trace Element Geochemistry

Seven samples from Del Puerto Canyon were analyzed for trace element concentrations using ICP-MS and loss on ignition at Washington State University, Pullman, Washington, USA, following routine methods of Knaack et al. (1994) and Johnson et al. (1999). Samples were collected from specific units as mapped by Dibblee (1982), Dibblee and Minch (2007), and Maddock (1964), with two mudstones within the Knoxville Formation, three mudstones within the Panoche Formation, and two andesite samples from the uppermost CRO (Fig. 2). Analyzing the trace element geochemistry of the sedimentary samples can distinguish mafic

components, source region, and degree of weathering and recycling of sediments (McLennan et al., 1993; Fralick, 2003; Surpless, 2015). The trace element geochemistry of the igneous samples is used to gain insight into the type of magma from which the rocks crystallized, and therefore infer their tectonic setting (e.g., Hofmann, 1997). Mantle convection leads to a globally heterogeneous mix of magma composition; as a melt separates from the mantle, elements that are incompatible (e.g., U, Th, K) migrate to other regions (Hofmann, 1997). Because of their varying tectonic settings and therefore mantle interactions, mid-ocean-ridge basalt (MORB), ocean-island basalt (OIB), and igneous units that have interacted with continental crust or evolved mantle will have varying chemical signatures, both isotopically and within their trace element compositions (e.g., Hofmann and Magaritz, 1977; Sun and McDonough, 1989).

In sedimentary geochemistry, trace elements help determine provenance because they are generally excluded from seawater and have low post-depositional mobility (McLennan et al., 1993). In fine-grained sedimentary rocks, such as shale and mudstone, immobile trace elements such as Th, Sc, and La help determine provenance and composition (Armstrong-Altrin and Verma, 2005; Ryan and Williams, 2007). In a La-Th-Sc ternary diagram, juvenile, MORB, and crustal components plot toward the Sc corner of the diagram, whereas more evolved North American Shale Composite and upper continental crustal components plot closer toward the La corner of the plot in a linear trend (Bhatia and Crook, 1986; McLennan et al., 1990, 1993; Surpless, 2015). In addition, a Zr/Sc versus Th/Sc plot is used to compare our samples to what we would expect from MORB, andesite, and upper crust (granodiorite); Zr/Sc ratio increases due to increased sedimentary recycling, and Th/Sc increases due to increased magmatic differentiation.

Petrography

Fifteen rock samples from Del Puerto Canyon were collected, cut, and stained for Ca-plagioclase and K-feldspar as standard petrographic thin sections. Eight are sandstones from GVG strata, and the remainder are igneous, metamorphic, and sandstone samples from the CRO and Franciscan subduction complex (Figs. 1 and 2). The composition and provenance of the GVG samples were analyzed by counting 300 framework grains according to the Gazzi-Dickinson point-counting method (Gazzi, 1966; Dickinson, 1985), with only samples composed of $<15\%$ matrix analyzed. In addition to the 300 framework grains, we recorded the composition of the

matrix and cement and took note of any sedimentary structures and textures to gain insight into the depositional and post-depositional environment of the sedimentary samples. Point-counting data, including grain types and modal parameters for all 24 samples, are available in Supplemental Material File S1.

The mineralogy of the CRO volcanic and Franciscan sandstone samples was also analyzed by documenting the primary groundmass and minerals present. We recorded any igneous textures and metamorphic fabrics for these samples to determine the origin and extent of metamorphism and deformation, respectively.

RESULTS

U-Pb Age Spectra

Samples collected from the GVG predominantly consist of Mesozoic grains, with <2% pre-Mesozoic grains, and fall into two groups (Fig. 3). Seven samples (23DP22, 61721A, 23DP21, 61621B, 23DP03, 61621C, 23DP06) consist of a unimodal age distribution centered at ca. 145 Ma with zero to six pre-Mesozoic grains in each sample (Fig. 3, middle panel). Sample 23DP21 was collected within strata mapped by Dibblee (1982), Dibblee and Minch (2007), and Maddock (1964) as the Knoxville Formation, 23DP22 and 61721A were collected from mapped Panoche Formation, and samples 61621B, 23DP03, 61521C, and 23DP06 were collected progressively east from the contact with the Del Puerto Keratophyre (Fig. 1B). Six samples from mapped or subsurface Panoche Formation (23DP01, 23DP02, 23-DPC-33, 23-DPC-34B, 23-DPC-34A, 23-DPC-35) consistently show a bimodal age spectrum with modes at ca. 93–89 Ma and ca. 150 Ma (Fig. 3, top panel); these samples contain two to 21 pre-Mesozoic grains per sample (Fig. 3). The four sandstones collected from the Franciscan subduction complex (23DP16, 23DP17, 23DP11,

23DP12) show a multimodal age spectrum with variable Mesozoic age modes between 250 Ma and 80 Ma (Fig. 3) and 6%–52% pre-Mesozoic grains. Pre-Mesozoic grains for three of the four (23DP11, 23DP12, 23DP17) Franciscan samples show modes of various magnitudes, ranging between ca. 450–250 Ma, ca. 1200–1000 Ma, ca. 1600–1400 Ma, ca. 1800–1600 Ma, and a few >2000 Ma grains. Sample 23DP16 differs from other Franciscan samples in the field area, as it contains age populations between 1470–1330 Ma and ca. 1800–1600 Ma and lacks the ca. 450–250 Ma grains.

Maximum Depositional Ages

The seven sandstones with a unimodal age spectrum yield YSP MDAs of ca. 139–145 Ma, with other MDA methods generally overlapping this range (Fig. 4). Initial LA-ICP-MS analyses on samples 61721A and 23DP22 yielded young tails of scattered dates ($n = 7$ each) that do not overlap the unimodal age spectra centered at ca. 145 Ma. Therefore, we performed CA-ID-TIMS on a subset of these grains to test for the accuracy of LA-ICP-MS dates and the potential effects of Pb loss leading to erroneously young dates. Sample 61721A yields six CA-ID-TIMS dates: 145.14 ± 0.12 Ma, 145.10 ± 0.32 Ma, 117.32 ± 0.11 Ma, 111.86 ± 0.08 Ma, 101.55 ± 0.11 Ma, and 95.18 ± 0.15 Ma (Table 1; Fig. 4). The two oldest dates (ca. 145 Ma) are significantly older than their LA-ICP-MS dates of ca. 125 Ma and ca. 124 Ma, respectively, suggesting Pb loss may have affected LA-ICP-MS dates (Table 1). In contrast, three other dates overlap LA-ICP-MS dates, and one is a bit younger. We choose to use the oldest two dates that overlap within 2-sigma error to calculate a weighted mean of 145.1 ± 0.22 Ma, which we report as the MDA for this sample. We do not choose to include the other CA-ID-TIMS dates in any MDA calculation as they do not overlap within error of each other, although they

are concordant. Sample 61721A was collected within strata mapped as the Panoche Formation, which is consistent with the youngest single CA-ID-TIMS date and scattering of Aptian–Albian ages. However, the sedimentary facies from which the sample was collected are dark gray to black shales, with limestone concretions and thin (<5 cm) sandstone beds, which are typical of the Knoxville Formation. Therefore, we assign this sample to the Knoxville Formation due to its proximity (<300 m) to thick, tabular massive sandstones characteristic of the Panoche Formation, and suspect contamination from detritus eroding from upslope to explain the youngest grains. Sample 23DP22 was also collected within strata mapped as the Panoche Formation, but within an outcrop of predominantly black shale. This sample yields six CA-ID-TIMS dates: 146.19 ± 0.22 Ma, 145.90 ± 0.19 Ma, 145.30 ± 0.11 Ma, 128.09 ± 0.09 Ma, 123.11 ± 0.17 Ma, and 115.56 ± 0.31 Ma. Three dates (ca. 146–145 Ma) are significantly older than their LA-ICP-MS dates of ca. 133 Ma, ca. 102 Ma, and ca. 94 Ma (Table 1). The remaining three grains show overlap between CA-ID-TIMS and LA-ICP-MS dates (Table 1). From these six dates, the oldest three overlap within 2-sigma error and yield a weighted mean of 145.60 ± 0.17 Ma, which we report as the MDA for this sample. The youngest TIMS dates are excluded from the weighted mean MDA calculation because the dates are more dispersed, with 5–10 m.y. between them and lack of correlation with ages of strata in the field area. We attribute three young grains from sample 23DP22 to be sourced from contamination from the upslope Panoche Formation during sample collection because there is no evidence in the CL images or Th/U ratios (>0.3) for metamorphic overgrowth or hydrothermal alteration leading to Pb loss. Shale sedimentary facies, sandstone petrography, and the lack of a distinct Late Cretaceous age range do not support the sample being from the Panoche Formation. In addition,

TABLE 1. CA-ID-TIMS COMPARISON AND CLASSIFICATION

Sample name	Zircon grain number	LA-ICP-MS $^{206}\text{Pb}/^{238}\text{U}$ date (Ma)	$\pm$ (2-sigma)	CA-ID-TIMS $^{206}\text{Pb}/^{238}\text{U}$ date (Ma)	$\pm$ (1-sigma)	Resulting unit assignment based on TIMS alone	Final unit assignment
61721A	251	125.89	1.45	145.14	0.12	Knoxville Formation	Knoxville Formation
61721A	75	124.47	2.73	145.10	0.32	Knoxville Formation	
61721A	96	115.64	1.89	117.32	0.11	Albian Shale	
61721A	158	110.93	1.30	111.86	0.08	Albian Shale	
61721A	12	103.85	1.58	101.55	0.11	Albian Shale	
61721A	271	103.11	1.39	95.18	0.15	Albian Shale	Knoxville Formation
23DP22	227	133.90	9.20	146.19	0.22	Knoxville Formation	
23DP22	109	102.60	3.50	145.90	0.19	Knoxville Formation	
23DP22	188	94.70	3.20	145.30	0.11	Knoxville Formation	
23DP22	266	129.00	9.60	128.09	0.09	Unknown ?	
23DP22	193	121.00	5.70	123.11	0.17	Unknown ?	Albian Shale
23DP22	180	111.30	4.70	115.56	0.31	Albian Shale	

Note: Laser ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) and chemical abrasion–isotope dilution–thermal ionization mass spectrometry (CA-ID-TIMS) U-Pb ages from grains within samples 23-DP-22 and 61721A are reported and used to classify each sample by age into the Knoxville Formation (central California, USA). Bold text are grains used in maximum depositional age calculation.

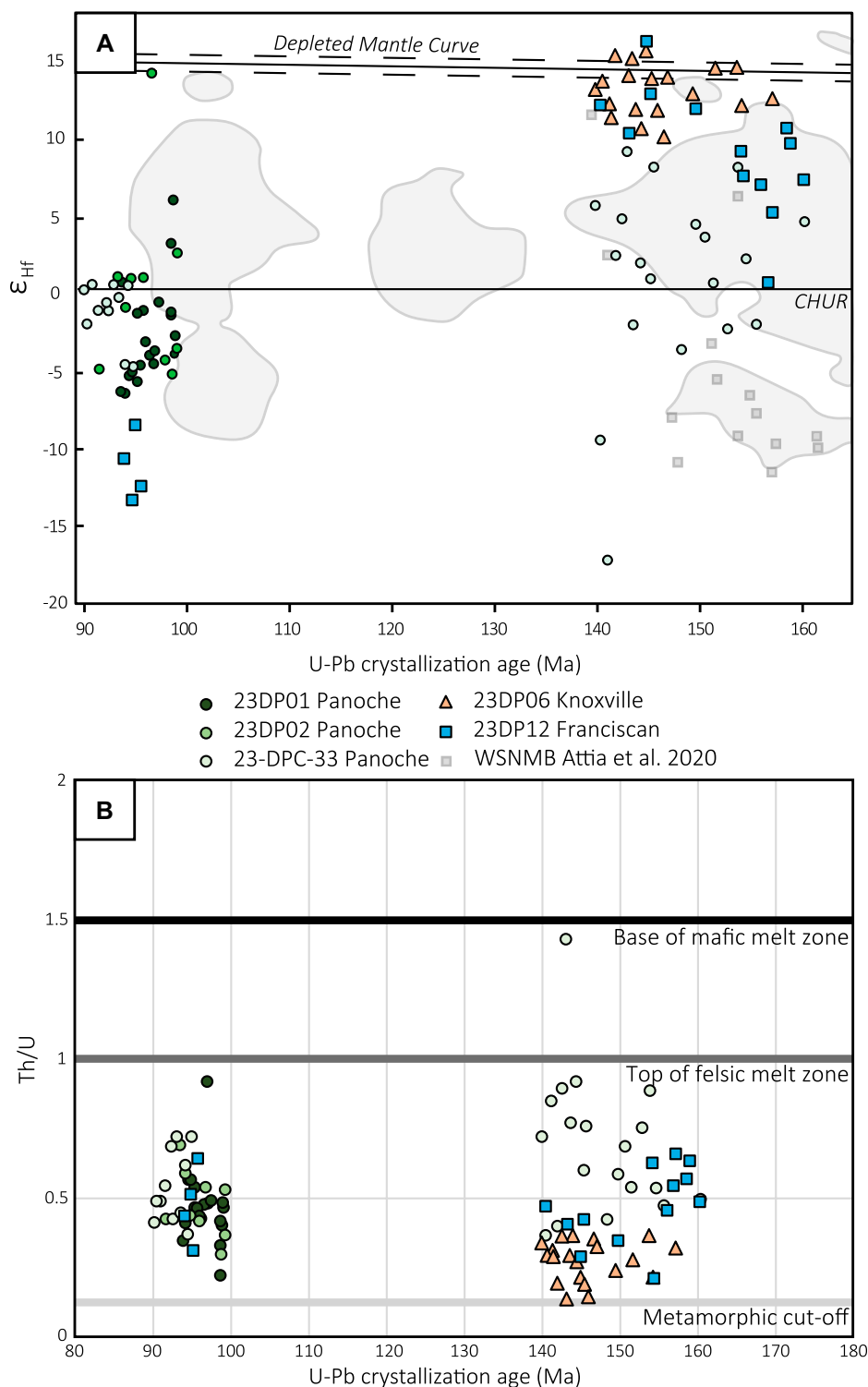


Figure 5. (A) Epsilon hafnium (ϵ_{Hf}) vs. U-Pb for samples from Del Puerto Canyon in age ranges of 100–90 Ma and 160–140 Ma. CHUR—chondrite uniform reservoir. Gray polygons indicate the ϵ_{Hf} curve at 95% confidence of Sierra Nevada arc from Attia et al. (2020). Gray squares represent Western Sierra Nevada Metamorphic Belt (WSNMB) samples from Attia et al. (2020), highlighting the wide range from -11 to $+12$ ϵ_{Hf} values. These are the first Lu-Hf analyses from the Great Valley Forearc basin. Highly juvenile values (positive ϵ_{Hf}) for the Knoxville Formation are consistent with whole-rock geochemistry from coeval strata of the Stony Creek Formation in the Sacramento basin, but differ from Lower Cretaceous strata in the San Joaquin basin farther south that show evolved signatures (see Fig. 9; Surpless, 2015). (B) U-Pb crystallization age vs. Th/U ratios of zircon grains from (A). The lines indicating the base and top of the mafic melt zone and the metamorphic cut off are from Teipel et al. (2004) and Linnemann et al. (2011). See Supplemental Material File S1 for plot of all grains U-Pb vs. Th/U (see text footnote 1).

Zircon Lu-Hf and Th/U Geochemistry

Five sandstones from Del Puerto Canyon from two age groups (160–140 Ma and 100–90 Ma) were analyzed for Lu-Hf geochemistry (Fig. 2). These populations were chosen to reflect the general depositional age groupings of the Knoxville (latest Jurassic–earliest Cretaceous) and Panoche (earliest Late Cretaceous) Formations. We analyzed a total of 96 zircons for Lu-Hf, with 53% ($n = 51$) of the grains crystallized from 160 Ma to 140 Ma, and 47% ($n = 45$) from 100 Ma to 90 Ma. The ϵ_{Hf} values for the older zircons range from -17 to $+16$ (Fig. 5A), with the majority (88%) plotting above the CHUR reference line. Within the 160–140 Ma zircons, the Knoxville Formation sample, 23DP06, has the smallest ϵ_{Hf} range: $+10$ to $+15$ (Fig. 5A). Grains from the Panoche sample, 23-DPC-33, range from -17 to $+9$, with 68% of grains plotting above the CHUR reference line in older zircons (Fig. 5A). Older grains from the Franciscan subduction complex sample, 23DP12, range from $+1$ to $+16$, primarily overlapping with zircons of the Knoxville Formation. Late Cretaceous zircons from the Panoche Formation and Franciscan are more evolved with ϵ_{Hf} values ranging from -13 to $+15$. The 100–90 Ma grains from Panoche samples are clustered around the CHUR reference line, with 90% of grains ranging from -6 to $+1$ (Fig. 5A). Zircon

visible modern mud drapes from upslope blanket much of this outcrop, supporting contamination as a source for these anomalous grains.

Panoche sandstone samples yield YSP MDAs of ca. 97–85 Ma, and all other MDA methods overlap the YSP within error (Fig. 4). Three sandstones (23DP16, 23DP17, 23DP12) from the

Franciscan subduction complex yield YSP MDAs of ca. 98–84 Ma, with other MDA methods overlapping within error (Fig. 4). In contrast, sample 23DP11 yields a YSP MDA of 115.0 ± 1.2 , which overlaps with the MLA of 115.8 ± 1.4 Ma, but slightly older than other methods that yield MDAs of ca. 110–107 Ma (Fig. 4).

grains from the Franciscan subduction complex sample, 23DP12, are increasingly negative (−8 to −13) as compared to zircons from the coeval Panoche Formation. The Th/U concentrations of the subset of Mesozoic zircons analyzed for Lu–Hf geochemistry were also investigated to aid with provenance interpretation (Fig. 5B). These samples consist of three Panoche samples (23DP01, 23DP02, and 23-DPC-33), one Knoxville sample (23DP06), and one Franciscan subduction complex sample (23DP12). All samples, regardless of age or formation, plot within the felsic melt zone, with one outlier grain that lies within the transition zone (Fig. 5B). Zircons from the Panoche Formation have the widest spread of Th/U ratios, ranging from ~0.2 to 0.9, excluding an outlier with a ratio of 1.43. Zircon grains from the Knoxville Formation are densely clustered and Th/U ratios range from ~0.1 to 0.4, and the Franciscan subduction complex sample grains range from ~0.2 to 0.7 (Fig. 5B).

Comparing U–Pb and Th/U of all ~5000 zircons analyzed by LA-ICP-MS indicates no change in Th/U concentrations between the pre-Mesozoic and Mesozoic grains (Supplemental Material File S1); the entire data set is not shown in Figure 5B for the purposes of clarity.

Whole-Rock Trace Element Geochemistry

Within rare earth elements (REEs) shown in a chondrite-normalized spider diagram, all samples show a negative trend and a significant negative Eu anomaly, including those from the volcanics of the upper CRO (Fig. 6A). From La to Sm, Panoche samples yield the highest concentrations, with the Knoxville close below, and the CRO samples have the least concentrated values (Fig. 6A). On a La–Th–Sc ternary diagram, all samples plot between La and Sc, and far from Th (Fig. 6B). CRO samples plot between MORB and continental arc references,

Knoxville samples right above the continental arc signature, and Panoche samples plot between upper continental crust and North American Shale Composite reference points (Fig. 6B), reflecting more evolved geochemistry through time. Samples from the Panoche Formation are consistent with Upper Cretaceous SJV mudstones from Surpliss (2015), but the Knoxville samples plot above the trend for Lower Cretaceous samples from the same study.

A comparison of Zr/Sc versus Th/Sc shows the CRO samples close to the andesite reference (Fig. 6B). The Knoxville samples plot above andesite, and the Panoche samples show a trend toward the granodiorite reference (Fig. 6C). As in the La–Th–Sc ternary diagram, samples from the Panoche Formation plot with Upper Cretaceous samples from Surpliss (2015), with the Knoxville just below this trend, and the CRO samples plot with Lower Cretaceous samples from the same study.

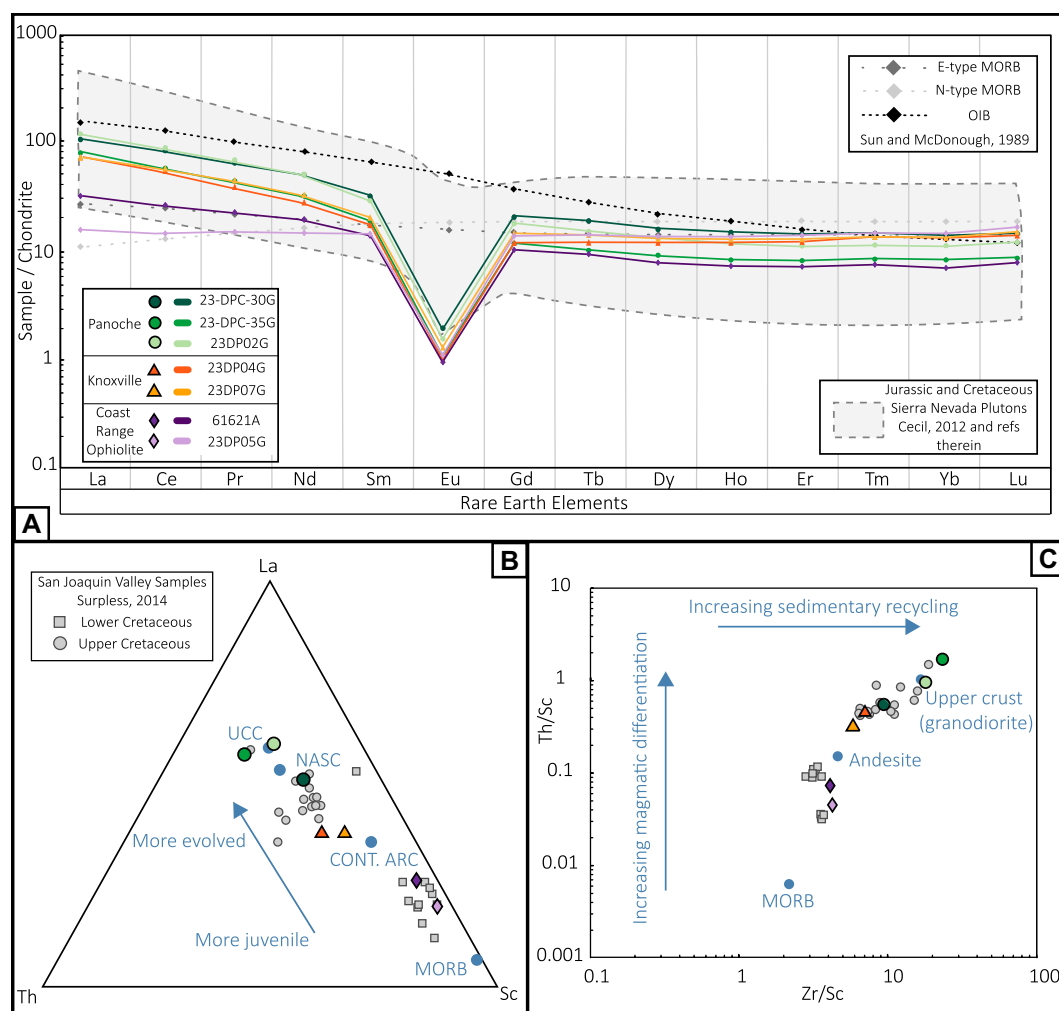


Figure 6. Geochemistry of Del Puerto Great Valley Group (GVG) strata and volcanics from the upper Coast Range Ophiolite (CRO). (A) Spider diagram of rare earth elements (REE) normalized to chondritic reservoir (Sun and McDonough, 1989), which highlights the similarity amongst volcanics of the upper CRO and GVG. Notably, all samples show a strong negative Eu anomaly compared to mid-ocean-ridge basalt (MORB) and ocean-island basalt (OIB) reference lines reflecting the input of continental crustal material. The gray envelope with dotted border overlying the REE concentrations represents compiled Jurassic and Cretaceous aged Sierra Nevada plutons from Cecil et al. (2012) and references therein. (B) La–Th–Sc ternary plot with potential sources from Taylor and McLennan (1985) and McLennan et al. (1993). (C) La/Sc vs. Zr/Sc plot, after McLennan et al. (1990); values for upper crust granodiorite from Taylor and McLennan (1985) and MORB from Sun and McDonough (1989). Data from

Upper and Lower Cretaceous strata from Surpliss (2015) shown in gray in (B) and (C). The legend in (A) is applied to all panels. CONT.—continental; NASC—North American Shale Composite; UCC—upper continental crust.

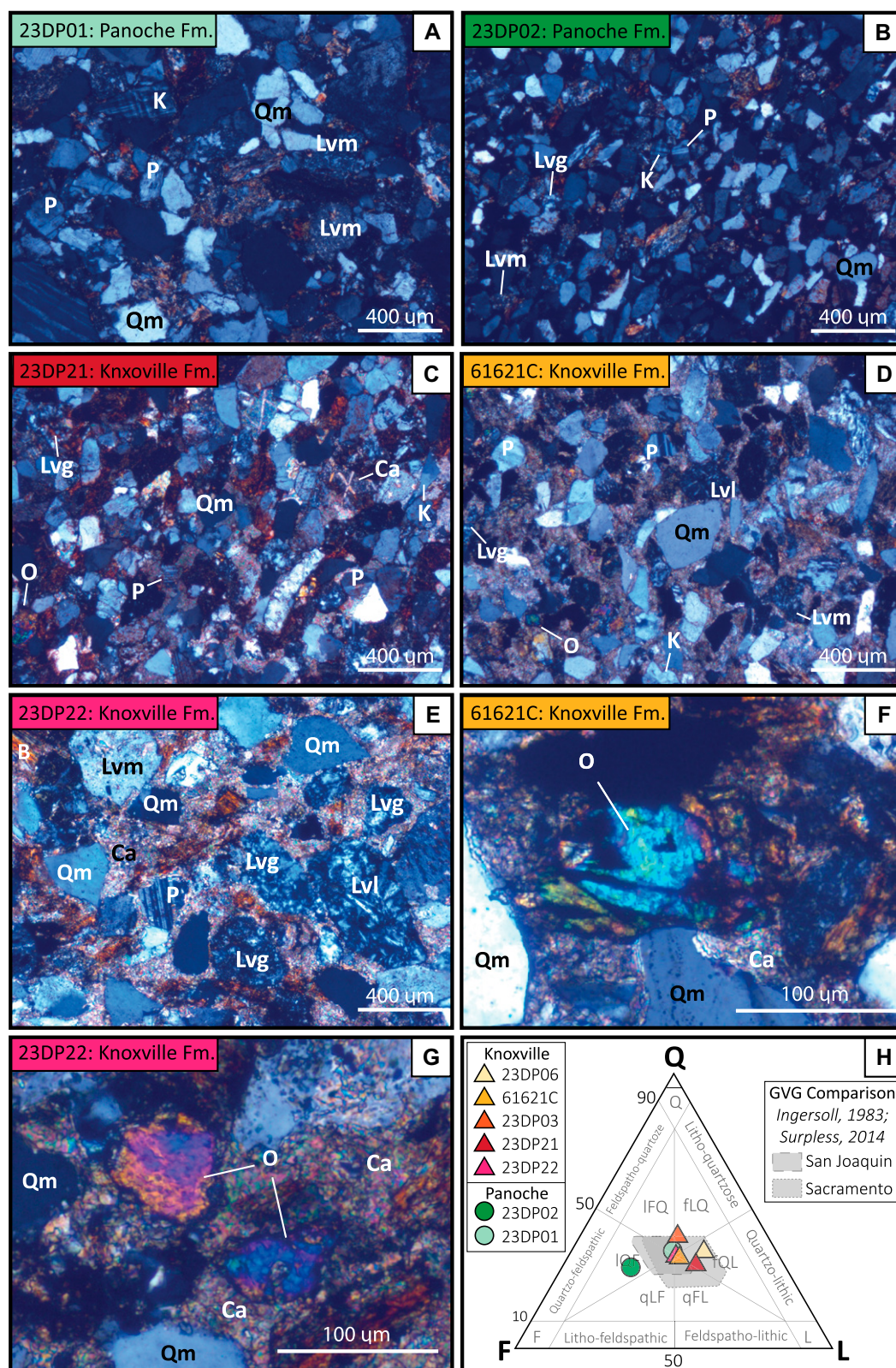


Figure 7. (A–G) Characteristic thin section images from Great Valley Group (GVG) strata in Del Puerto Canyon. Magnification for panels A–E is 4×, and the scale bar is 400 microns in length. The magnification of panels F and G is 10×, and the scale bars are 100 microns long. Images all taken under cross polarized light (XPL). B—biotite; Ca—calcite; K—potassium feldspar; Lvg—volcanic, felsitic, granular; Lvl—volcanic, lathwork; Lvm—volcanic, microlitic; O—olivine; P—plagioclase feldspar; Qm—monocrystalline quartz. (H) Ternary diagram adapted from Garzanti (2019) with abundance of total quartz (Q), total feldspar (F), and total lithics (L). Gray references fields of San Joaquin Valley and Sacramento basin from Surpless (2015) adapted from Ingersoll (1983).

Petrography

The average composition of eight samples of GVG strata from Del Puerto Canyon is 33%

quartz, 32% feldspars, and 35% lithics (Fig. 7; Supplemental Material File S1). Ninety-five percent of the quartz is monocrystalline (Qm) and the remainder is polycrystalline (Qp).

Among the feldspar grains, 50% are plagioclase feldspar (P) and 50% are potassium feldspar (K). Lithics are primarily volcanic, with lathwork (17%), granulitic (55%), and

microlitic (28%) textures. Notably, Knoxville samples have a total of 161 detrital olivine (O) present, and olivine comprises 3%–14% of the framework grains of each sample (Fig. 7; Sup-

plemental Material File S1). Both Knoxville and Panoche samples have accessory detrital minerals including muscovite, biotite, calcite, and chlorite. All GVG samples lack evidence

of deformation, contain minor plagioclase replacement by calcite, and generally contain cement and matrix composed of micas, iron oxides, and calcite (Figs. 7A–7G).

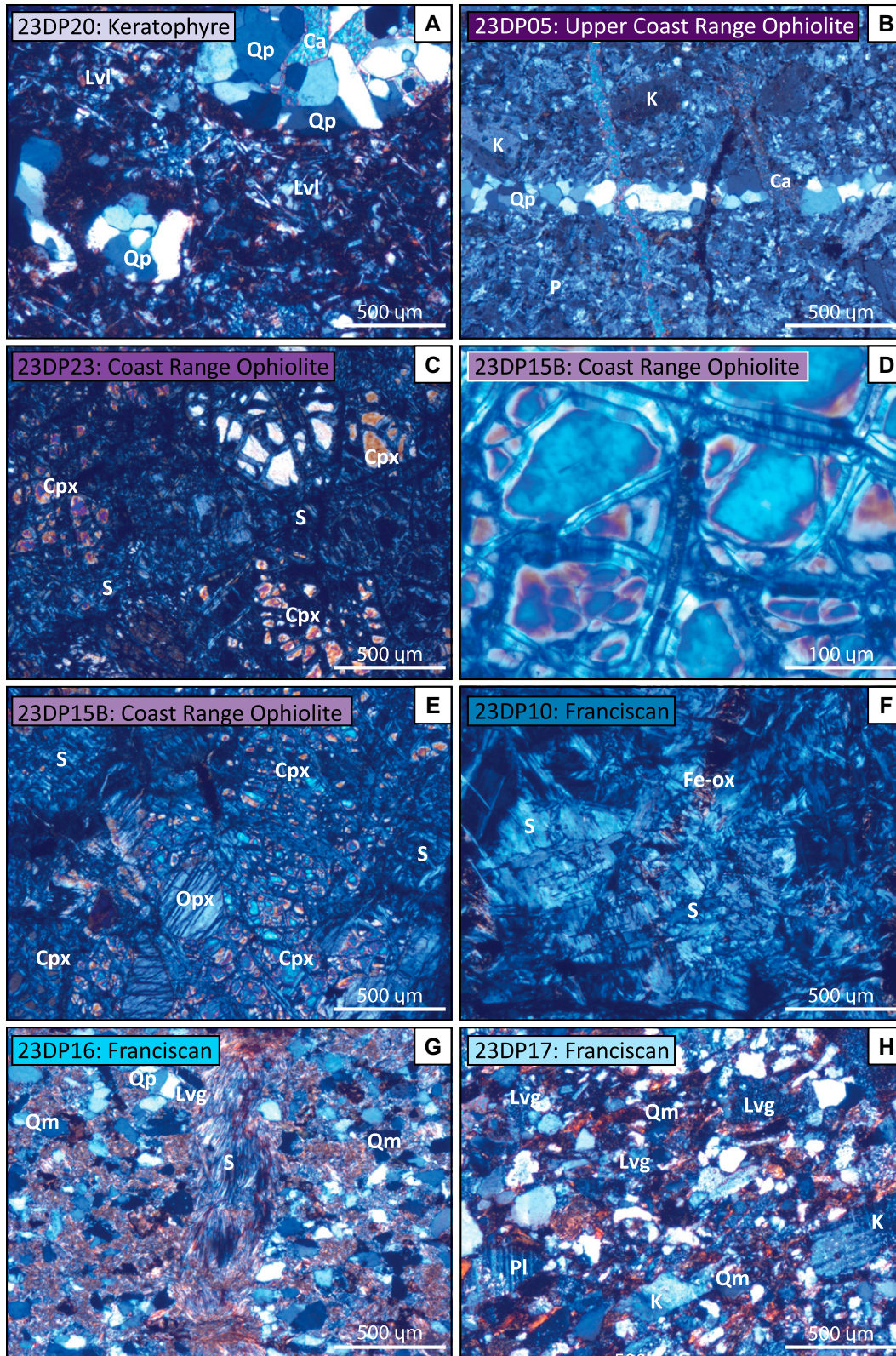


Figure 8. (A–E) Characteristic thin section images from Coast Range Ophiolite (CRO) in Del Puerto Canyon. (F–H) Characteristic thin section images from Franciscan subduction complex in Del Puerto Canyon. Magnification for samples is 4×. (D) was taken at 20× magnification. Sample 23DP15B was collected from a highly serpentinized (D) outcrop of CRO, but in thin section shows the presence of relict harzburgite minerals such as clino- and orthopyroxene and olivine that has altered to serpentinite (E). (A) and (B) show the volcanic textures of the keratophyre and CRO, respectively, including the presence of polycrystalline quartz and diagenetic calcite. Images were all taken under cross polarized light (XPL). Ca—calcite; Cpx—clinopyroxene; Fe-ox—iron oxide; K—potassium feldspar; Lvg—volcanic, felsitic, granular; Lvl—volcanic, lathwork; Opx—orthopyroxene; P—plagioclase feldspar; PI—plagioclase; Qp—polycrystalline quartz; Qm—monocrystalline quartz; S—serpentinite.

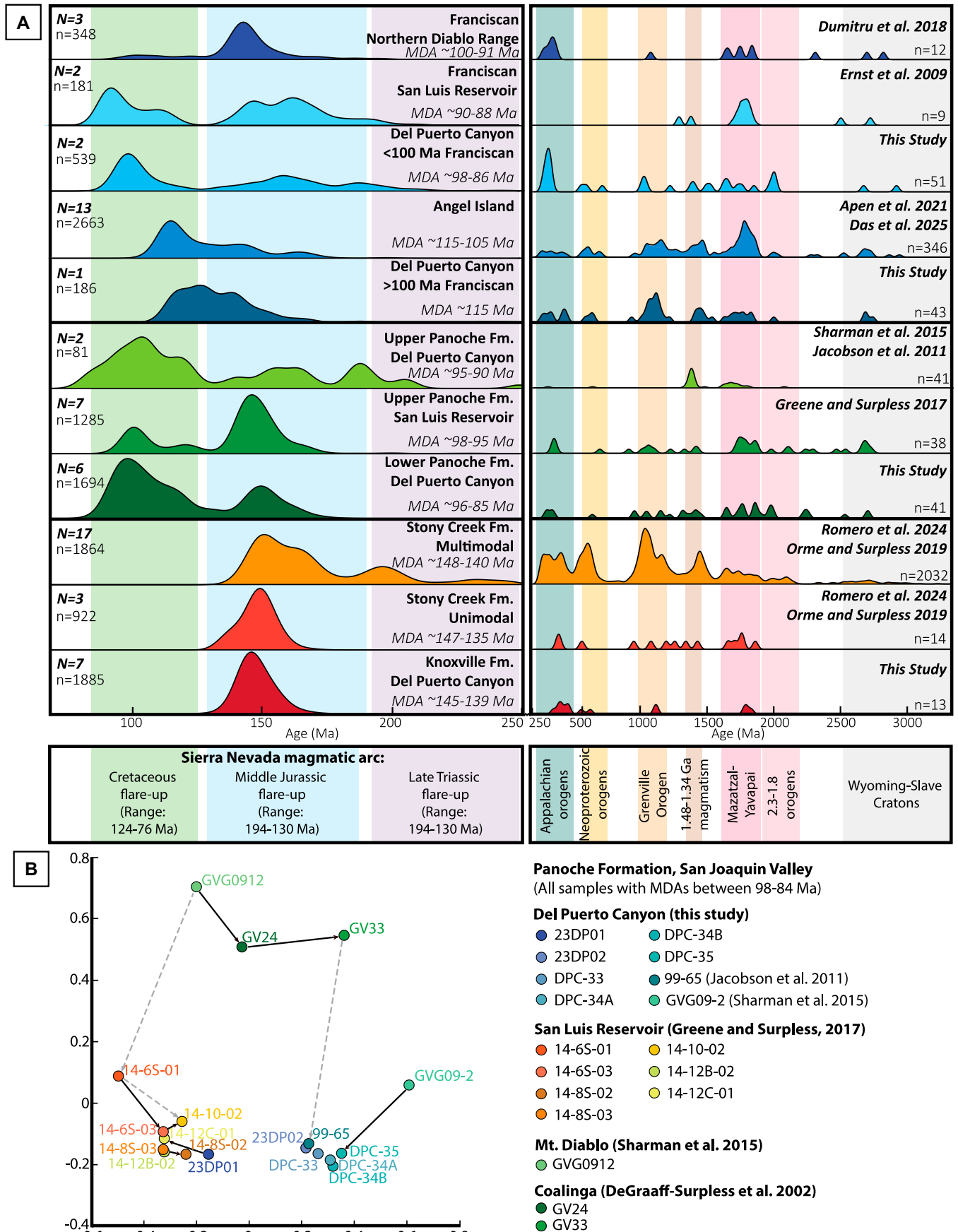


Figure 9. (A) Detrital zircon kernel density estimates (KDEs) compilation of Franciscan subduction complex (blue), Panoche Formation (green), and age equivalent Knoxville (San Joaquin Valley, SJV) and Stony Creek (Sacramento basin) Formations of the Great Valley Forearc. KDEs are grouped by formation, and within each formation, by age, with the oldest on the bottom of each formation. Mesozoic KDEs (250–50 Ma) are normalized to each other and have a kernel of 4. Following a break in scale, Paleozoic and Precambrian (3000–251 Ma) KDEs are normalized to each other and have a kernel of 20. Samples used are represented by “N=”; total grains for each KDE are shown as “n=”. (B) Two-dimensional multidimensional scaling plot of the Panoche Formation at Del Puerto Canyon (blue symbols) and other localities where the Panoche Formation has detrital zircon data from samples with maximum depositional ages (MDAs) between 98 Ma and 84 Ma. This analysis highlights the provenance differences amongst sandstones deposited coevally within the SJV. The plot was generated using DZmds (Saylor et al., 2018) using the cross-correlation statistic due to large variation in the number of grains amongst samples. Where distances between samples are great, black arrows point to the nearest neighbor and dashed gray lines point to the second nearest neighbor.

The composition and texture of the Del Puerto Keratophyre and CRO andesites are shown in representative thin section photos (Figs. 8A–8E). Del Puerto Keratophyre (23DP20) shows polycrystalline quartz with lathwork volcanics and minor calcite (Ca; Fig. 8A). From the CRO, sample 23DP05 is mostly lathwork volcanics and plagioclase feldspar, and it contains veins of calcite and polycrystalline quartz (Fig. 8B). Sample 23DP23 is composed of clinopyroxene, serpentine, lizardite, and olivine, and it contains microfractures within the grains (Fig. 8C). Sample 23DP15B shows lizardite and corona texture with mostly clinopyroxene (Cpx), orthopyroxene (Opx), and serpentine (S; Figs. 8D and 8E).

The variation in texture and composition of Franciscan subduction complex samples in Del Puerto Canyon is shown in representative thin section photos (Figs. 8F–8H). Sample 23DP10 is a serpentinite clast within siliciclastic mélange composed entirely of serpentinite with interpenetrating texture and minor iron oxide ribbons (Fig. 6F). By contrast, samples 23DP16 and 23DP17 show framework grains composed of monocrystalline quartz, polycrystalline quartz, plagioclase feldspar, potassium feldspar, and accessory detrital serpentinite ($n = 7$; 1.2%; Figs. 8G and 8H). The texture detrital serpentinite is fibrous, with elongate blades and occasional interpenetrating fabric.

DISCUSSION

Controls on Depositional Age

Sample 23DP21, collected from strata mapped as Knoxville on all previously published maps and along the western limb of the Del Puerto Keratophyre anticline, stratigraphically above the Lotta Creek Tuff, contains exclusively Mesozoic grains with MDA of 145.56 ± 1.52 Ma (Figs. 1, 3, and 4). Similarly, five samples (61621B, 23DP04, 23DP03, 61621C, 23DP06) collected on the east limb of the anticline, at the contact with the CRO, yield unimodal modes with MDAs ranging from ca. 146 Ma to 139 Ma (Figs. 1, 3, and 4). These sandstones were col-

lected within ~300 m of stratigraphy comprising fissile black shales interbedded with fine-grained sandstone intervals and scattered limestone concretions. Previous mapping assigns these rocks to the Panoche Formation (Dibblee, 1982; Dibblee and Minch, 2007) or an Albian Shale (Bennison et al., 1991). However, due to similarities with Knoxville sedimentary descriptions, unimodal age spectra, and ca. 145 Ma MDAs, we reassign these to the Knoxville Formation (Figs. 1–4). We use the YSP method as our preferred MDA as it is the most conservative and utilizes MSWD values to determine if grains are part of a single population. Alternatively, if one considers other MDA methods such as the youngest group of three or more grains overlapping within 2-sigma error (YGC2s 3+; Fig. 4), the uppermost sample from the Knoxville Formation, east of the contact with the CRO, could be as young as ca. 136 Ma (23DP06). This age is consistent with previous biostratigraphy that found fossils assigned as Valanginian in age (Schilling, 1962).

Similarly, along Del Puerto Canyon Road on the western side of the Del Puerto Keratophyre anticline, sample 61721A was collected from black shales and fine-grained sandstones interbedded with limestone boulder size concretions (Figs. 2 and 3). Given the similar lithofacies, the unimodal Mesozoic age distribution, and an MDA of 145.1 ± 0.11 Ma for 61721A, we reassign sample 61721A to the Knoxville Formation (Figs. 1–4). Furthermore, we interpret these MDAs as approximating TDA and not significantly younger than ca. 140 Ma, as their ages broadly agree with biostratigraphic records that interpret Tithonian through Valanginian age (ca. 151–135 Ma) in the region (Schilling, 1962). In addition, Valanginian–Hauterivian (ca. 140–130 Ma) strata were deposited atop the basal Knoxville Formation just south of our field area at Pacheco Pass and Coalinga (e.g., Moxon, 1988). Although there is a period of low flux Sierran-Klamath arc magmatism after ca. 140 Ma until ca. 124 Ma, there is still volcanic activity during this interval, as reflected in the detrital zircon record of more complete

sections of the GVF basin that were deposited during this interval (e.g., DeGraaff-Surpless et al., 2002; Orme and Surpless, 2019; Romero et al., 2024; Paterson and Ducea, 2015, and references therein). Therefore, we do not interpret these MDAs as significantly younger than their TDA owing to any lack of zircon availability, but acknowledge these dates are MDAs, not absolute ages from volcanic tuffs.

The five sub-surface Panoche samples (23-DPC-30, 23-DPC-33, 23-DPC-34A, 23-DPC-34B, 23-DPC-35) and two surficial Panoche samples (23DP01, 23DP02) yield MDAs ranging from ca. 97 Ma to 85 Ma, consistent with each other and with published depositional ages. The MDAs from our samples generally young up section (west to east) or remain the same (Fig. 4) and are younger than two Campanian sandstones from the Panoche Formation farther east (Fig. 9; Jacobson et al., 2011; Sharman et al., 2015). The ~40 m.y. duration between deposition of Lower Cretaceous Knoxville strata and Upper Cretaceous Panoche strata highlights a previously documented unconformity (Moxon, 1988), and its implications are discussed further within the Tectonic Interpretations sections (Figs. 3 and 4).

Four sandstone samples from the Franciscan subduction complex (23DP12, 23DP11, 23DP17, 23DP16) yield MDAs that range from ca. 115 Ma to 84 Ma, which, apart from the oldest ca. 115 Ma MDA (23DP11), temporally overlap with deposition of the Panoche Formation. Notably, these MDAs are much younger than previous constraints from biostratigraphy that suggest accretion of the Franciscan preserved at Del Puerto Canyon occurred between ca. 152 Ma and 145 Ma (Camp, 1942; Crittenden, 1951; Hall, 1958; Maddock, 1964). Detrital zircon age distributions from metasandstones at neighboring localities in the northern Diablo Range report MDAs between ca. 115 Ma and 86 Ma (Dumitru et al., 2018), with Apen et al. (2021) and Das et al. (2025) assigning rocks with MDAs between 115–110 Ma and 100–92 Ma to the Angel Island accreted units in the Tiburon Peninsula ~100 km north of Del Puerto Canyon; restoring Cenozoic strike-slip deforma-

tion places these units ~30–50 km south of the Diablo Range (Apen *et al.*, 2021). Two of our samples (23DP12 and 23DP17) have MDAs and detrital zircon age spectra (discussed within the Provenance section) consistent with rocks of the Northern Diablo Range and Franciscan meta-graywackes at San Luis Reservoir (Ernst *et al.* 2009), whereas sample 23DP11 yields an MDA of ca. 115 Ma and age spectra consistent with metasediments of Angel Island accreted units (Das *et al.*, 2025; Apen *et al.*, 2021; Fig. 9). Sample 23DP16 yields an MDA (ca. 84 Ma) younger than previously reported MDAs of the Franciscan subduction complex exposed within the northern Diablo Range.

Revised Mapping

Compiling previous work by Dibblee (1982), Dibblee and Minch (2007), Maddock (1964), Raymond (1973), and Harms *et al.* (1992) with our field observations and the California State Geologic Map (Jennings *et al.*, 1977; Wills *et al.*, 2015), we present a revised map in Figure 1. We find that the Lotta Creek Formation does not reach the present location of Del Puerto Canyon Road and moved the contact farther south to reflect this. In addition, we find that GVG strata at the easternmost contact with the CRO has lithofacies, sandstone compositions, and geochronologic ages consistent with the Knoxville Formation, and reestablish the presence of this unit at this contact (see Provenance section for details). We incorporated the Coast Range fault and Tesla-Ortiguera fault mapped by Anderson and Pack (1915), Dibblee (1982), and Raymond (1973) on our map and the addition of one unnamed fault in the southeast portion of the map that was originally mapped by Dibblee (1982) and Dibblee and Minch (2007), with a portion remapped by Harms *et al.* (1992; Fig. 1). Although active motion along the Tesla-Ortiguera fault is primarily dextral, kinematic analyses and cross-cutting relationships reveal a normal sense of motion as old as post-Albian. Therefore, on our new map, we include up (U) and down (D) indicators in black text to indicate this older history. In addition, based on kinematic analyses of Harms *et al.* (1992), the Coast Range fault is interpreted as a series of detachment faults, with top-to-the-east displacement of the Franciscan, and we include the detachment symbol to reflect these interpretations.

Provenance

Knoxville Formation

The detrital olivine present in all of our Knoxville samples ($n = 161$) indicates that at the time of Knoxville deposition, the CRO was exposed

and had not undergone hydrothermal alteration to serpentinite (Rai and Matthew, 2021; Figs. 5F, 5G, and 10A). Similar petrographic and age relationships are observed in the northern GVF where the most basal GVG, the Stony Creek Formation, and an underlying breccia interval that separates the GVG from the CRO, contain ophiolitic and serpentinite detritus (Ingersoll, 1983; Romero *et al.*, 2024). Notably, differences in the degree of alteration, as reflected by the presence of olivine versus serpentinite, suggest variability in the timing of the serpentinization process of the CRO between the Sacramento and SJV basins and a lack of alteration during sedimentary burial within the Del Puerto region. The Knoxville samples' unimodal detrital zircon U-Pb age spectra (Fig. 3) and relative reduction in volcanic lithics as compared to the majority of the oldest GVG in both the Sacramento basin and SJV (Fig. 5) suggests that there was a source of broadly uniform age during deposition. This source may have been Jurassic (ca. 150–145 Ma) plutons within the central or southern Sierra Nevada magmatic arc, including potential Jurassic sources within the Western Sierra Nevada Metamorphic Belt (WSNMB; Ingersoll, 2000; Cecil *et al.*, 2012; Schweickert, 2015) or the underlying CRO (e.g., Ingersoll, 1982; Romero *et al.*, 2024). A unimodal Mesozoic age mode contrasts with much of the basal GVG in the northern GVF, the Stony Creek Formation, which commonly shows >55% pre-Mesozoic grains (Fig. 9; Orme and Surpless, 2019; Romero *et al.*, 2024). Nevertheless, a few samples from the Stony Creek Formation within the Sacramento basin also show a similar unimodal peak at ca. 150–145 Ma and no pre-Mesozoic grains (DeGraaff-Surpless *et al.*, 2002; Surpless *et al.*, 2006; Orme and Surpless, 2019); these samples are also interpreted to reflect contributions from Jurassic sources within the proximal undissected magmatic arc and WSNMB (Ingersoll, 1982; DeGraaff-Surpless *et al.*, 2002; Romero *et al.*, 2024).

Four of our Knoxville Formation samples contain pre-Mesozoic grains found within basement terranes of North America and are petrographically similar to the basal GVG samples from the Sacramento basin that contain abundant pre-Mesozoic detrital zircon grains (Dickinson and Rich, 1972; Ingersoll, 1983; Surpless, 2015; Romero *et al.*, 2024). We interpret that deposition of the Knoxville Formation, the oldest GVG strata in the SJV, occurred proximal to North America, consistent with previous studies of the oldest GVG in the Sacramento basin (DeGraaff-Surpless *et al.*, 2002; Orme and Surpless, 2019; Romero *et al.*, 2024). In addition, our zircon and whole-rock geochemical results link the Knoxville Formation to the Sierra Nevada continental

arc during Knoxville deposition, consistent with geochemistry of coeval GVG strata (Surpless, 2015) and records from the Sierra Nevada magmatic arc and the western Sierra Nevada magmatic belt (Attia *et al.*, 2020, 2021; Figs. 8 and 9). This is most apparent in La-Th-Sc and Zr/Sc versus Th/Sc plots, in which Knoxville samples plot between continental arc and upper continental crust signatures (McLennan *et al.*, 1993), and in between andesite and upper crust signatures (Taylor and McLennan, 1985). Rare earth element spectra of central and southern Jurassic Sierra Nevada plutons display concave to flat trends, with several outliers showing negative Eu anomalies due to high SiO₂ intrusions (Cecil *et al.*, 2012, and references therein), which aligns with Jurassic strata from the Knoxville Formation in Del Puerto Canyon (Fig. 8A).

Zircon ϵ_{Hf} values from Knoxville overlap with data points from the Sierran arc but show a shift to slightly more juvenile values as compared to the 95% confidence interval from the arc (Fig. 7A; Attia *et al.*, 2020). These values are consistent with a more juvenile source such as the CRO or accreted island-arc material of the WSNMB. Zircon ϵ_{Hf} data from the WSNMB are limited ($n = 14$), and range from -13 to $+13$, with 11 of 14 analyses plotting below CHUR (Fig. 7A). Given that we only analyzed one sample from the Knoxville for Lu-Hf, and that whole-rock and zircon U-Pb geochemical data support a link with a felsic continental arc, we interpret the CRO and WSNMB as secondary sources to Jurassic plutons of the Sierra Nevada. Compared to other Lower Cretaceous strata within the SJV (Surpless, 2015), our Knoxville samples plot closer to the continental arc and within the lower part of the upper granodiorite crust of these discrimination plots (Fig. 9). One explanation for this difference may be that Del Puerto is located farther north than the Lower Cretaceous SJV strata studied by Surpless (2015) and that the proposed southern sources, including major input from the WSNMB, did not reach this far north.

Panoche Formation

Samples from the Panoche Formation within our study area have similar age spectra, with Middle–Late Jurassic (ca. 170–145 Ma) and Early–Late Cretaceous (ca. 140–90 Ma) zircon populations (Fig. 3) and similar petrologic compositions (Fig. 5). The abundant Cretaceous zircons coincide with the well-documented mid-Cretaceous high-flux magmatism of the Sierra Nevada, which comprises ~50% of all exposed plutons within the Sierran arc today (e.g., Chen and Moore, 1982; Paterson and Ducea, 2015). These include proximal plutons of the Tuolumne Intrusive Suite, adjacent to the dextral Bench

Canyon and Quartz Mountain shear zones, both of which are interpreted to be active during Campanian Panoche Deposition (e.g., Coleman et al., 2004; McNulty, 1995; Tobisch et al., 1995; Tikoff and Greene, 1997; Greene and Surpless, 2017; Tomek et al., 2024). By contrast, Jurassic sources are less abundant and spatially scattered within the modern outcrop of the Sierra Nevada. The presence of the same Late Jurassic age mode (ca. 150–145 Ma) and similarities between the petrography of both Knoxville and Panoche Formations suggests that during Panoche deposition, Jurassic sources that fed the Knoxville were still feeding the GVF (e.g., Dumitru et al., 2015; Greene and Surpless, 2017). The age range of Jurassic grains (ca. 170–146 Ma) present in our samples is found within the Jurassic Sierra Nevada arc and WSNMB, and also are abundant in various parts of the CRO, including at Del Puerto Canyon, where crystallization ages range from ca. 161 Ma to 155 Ma (Lanphere, 1971; Hopson et al., 1981; Evarts et al., 1992; Mattinson and Hopson, 2008).

Within the SJV, including the exposures of the Panoche Formation at San Luis Reservoir ~45 km southwest of Del Puerto Canyon, paleocurrent analyses indicate a predominantly west or southwest-directed current (Ingersoll, 1979; Cherven, 1983; Bennison, 1991; Bennison et al., 1991; Greene and Surpless, 2017). At San Luis Reservoir, Greene and Surpless (2017) document both felsic and mafic plutonic rocks (granitoids to gabbro) and metavolcanic and/or metavolcaniclastic rocks within conglomerate intervals, and interpret the latter as derived from the CRO and/or the WSNMB. Our petrographic results from the Panoche Formation overlap with those of Greene and Surpless (2017) and the SJV Grabast petrofacies of Mansfield (1979) and Ingersoll (1983) and include similar diagenetic features such as presence of calcite cement, altered volcanic grains, and dissolved feldspar grain boundaries. However, we do not observe a high abundance of metamorphic grains, as Ingersoll (1983) observed. The abundance of potassium feldspar to total feldspar in all of our samples is consistently 50%, unlike the 0%–50% increase from Jurassic to Cretaceous found by Ingersoll (1983).

Panoche samples are geochemically more evolved than Knoxville samples (Fig. 8), but both formations' geochemical signals tie their provenance to the Sierra Nevada arc, as seen in the overlap of Del Puerto GVG strata ϵ_{Hf} with the Sierran arc signature (Fig. 7A; Attia et al., 2020), felsic zircon Th/U ratios (Fig. 7B), and the REE trends of central and southern Cretaceous Sierra Nevada plutons (Fig. 8; Cecil et al., 2012). Integrating the petrographic, geochronologic, and geochemical data with the paleo-

current data from previous studies, we favor proximal Sierra Nevada Cretaceous and Jurassic plutons as the primary source of sediment for the oldest Panoche sandstones at Del Puerto Canyon, with little input from the WSNMB. As this differs from the evidence that supports the WSNMB as a source for the Panoche Formation ~45 km south at San Luis Reservoir, we hypothesize separate sediment dispersal systems feeding these two depositional centers from slightly varying sources. More input from the WSNMB to the San Luis Reservoir area is supported by the distribution of Jurassic (43%) and Cretaceous (54%) age populations, as compared to the abundance of Late Cretaceous age zircons in the Panoche Formation at Del Puerto Canyon (74% Cretaceous, 23% Jurassic). This difference is highlighted in our MDS analysis that shows a distinct grouping of samples from the two areas (Fig. 9B). Further MDS comparison of detrital zircon age distributions from coeval strata deposited at Mt. Diablo and Coalinga show strong dissimilarity with Del Puerto Canyon and San Luis Reservoir, as these localities are dominated by middle Cretaceous (139–101 Ma) zircons (Fig. 9B). These two areas preserve strata of mid-Cretaceous age that are absent in Del Puerto Canyon and San Luis Reservoir.

East and upsection of this study's field area, a sandstone from the Panoche Formation contains Proterozoic grains likely derived from a ca. 1380 Ma pluton within the Lemhi subbasin of Idaho (Sharman et al., 2015; Dumitru et al., 2016). These grains are hypothesized to be transported to the GVF basin and Franciscan subduction complex between ca. 85 Ma and 80 Ma via the Kione River (Dumitru et al., 2016). Our samples are down-section in both outcrop and the sub-surface, and yield MDAs older than two Campanian sandstones from the Panoche Formation analyzed by Sharman et al. (2015) and Jacobson et al. (2011; Fig. 9). Given the lack of the ca. 1380 Ma peak in our samples and their lower stratigraphic position, we propose that the lower Panoche Formation did not receive detritus from the Kione River system, consistent with its arrival after 85 Ma.

Franciscan Subduction Complex

Franciscan subduction complex sandstones share similar framework grains with the Panoche Formation in Del Puerto Canyon (Figs. 5 and 6) but also contain accessory detrital serpentinite (~1%; Fig. 6), not observed in our Panoche sandstones. The fibrous blades and interpenetrating texture are evidence for recrystallization processes prior to deposition, in addition to sandstone diagenesis that is observed in the chloritic matrix (e.g., Perrotta et al. 2024). Detrital serpentinite in sandstones from the Franciscan

is rare, with estimates between 0.5 and 5% per sample (J. Wakabayashi, 2025, personal commun.). However, clastic serpentinite matrix surrounding high-grade blocks is abundant within the Franciscan (e.g., Wakabayashi, 2015) and is a potential source of this detritus, in addition to serpentinite of the CRO.

Apart from sample 23DP11 (MDA ca. 115 Ma), the MDAs of the Franciscan subduction complex sandstone coincide with the Panoche MDAs (Fig. 4), but age spectra contain more pre-Mesozoic grains (~6%–52%) than the Panoche or Knoxville samples (Fig. 3). These data are consistent with previous studies of the Franciscan subduction complex that show an abundance of syn-volcanic and variation in the abundance of pre-Mesozoic age grains (Apen et al., 2021; Dumitru et al., 2015; Snow et al., 2010; Ernst et al., 2009) that are interpreted to represent clastic sedimentation from wide catchments across the North American margin (e.g., Dumitru et al., 2018; Wakabayashi, 2015). Pre-Mesozoic grains (ca. 450–250 Ma, ca. 1200–1000 Ma, ca. 1600–1400 Ma, ca. 1800–1600 Ma, and a few >2000 Ma grains) are found in samples with MDAs between ca. 115 Ma and 94 Ma. These age populations comprise 6%–15% of these samples' age spectra, and are consistent with original derivation from the Appalachian orogen, Grenville Province, anorogenic granitic plutons, Yavapai-Mazatzal Provenance, and older Paleoproterozoic North American basement (e.g., Dickinson et al., 2012; Gehrels and Pecha, 2014). We interpret the relative abundance of pre-Mesozoic age grains within the Franciscan compared with the Panoche Formation to reflect the dominance of proximal Sierran sources for the latter and input from farther traveled axial systems that likely were active within the trench. The age spectrum of 23DP16 is distinct from our other Franciscan subduction complex samples, but resembles the Campanian Panoche sample of Sharman et al. (2015). Sample 23DP16 has an MDA of ca. 84 Ma and it contains 52% pre-Mesozoic grains that primarily cluster around ca. 1470–1330 Ma and ca. 1800–1600 Ma (Figs. 3 and 10). Notably, this sample is strikingly similar to a coeval Franciscan metagraywacke from the nearby San Luis Reservoir region that also shows the presence of these age peaks, which its older counterparts in the region lack (Fig. 9A; Ernst et al. 2009). Following the pioneering observations of Dumitru et al. (2016), we interpret 23DP16 to record the arrival of the Kione River system, sourced from central Idaho to the Franciscan subduction complex after 85 Ma, where this signature is evident in the one Panoche sample of Sharman et al. (2015).

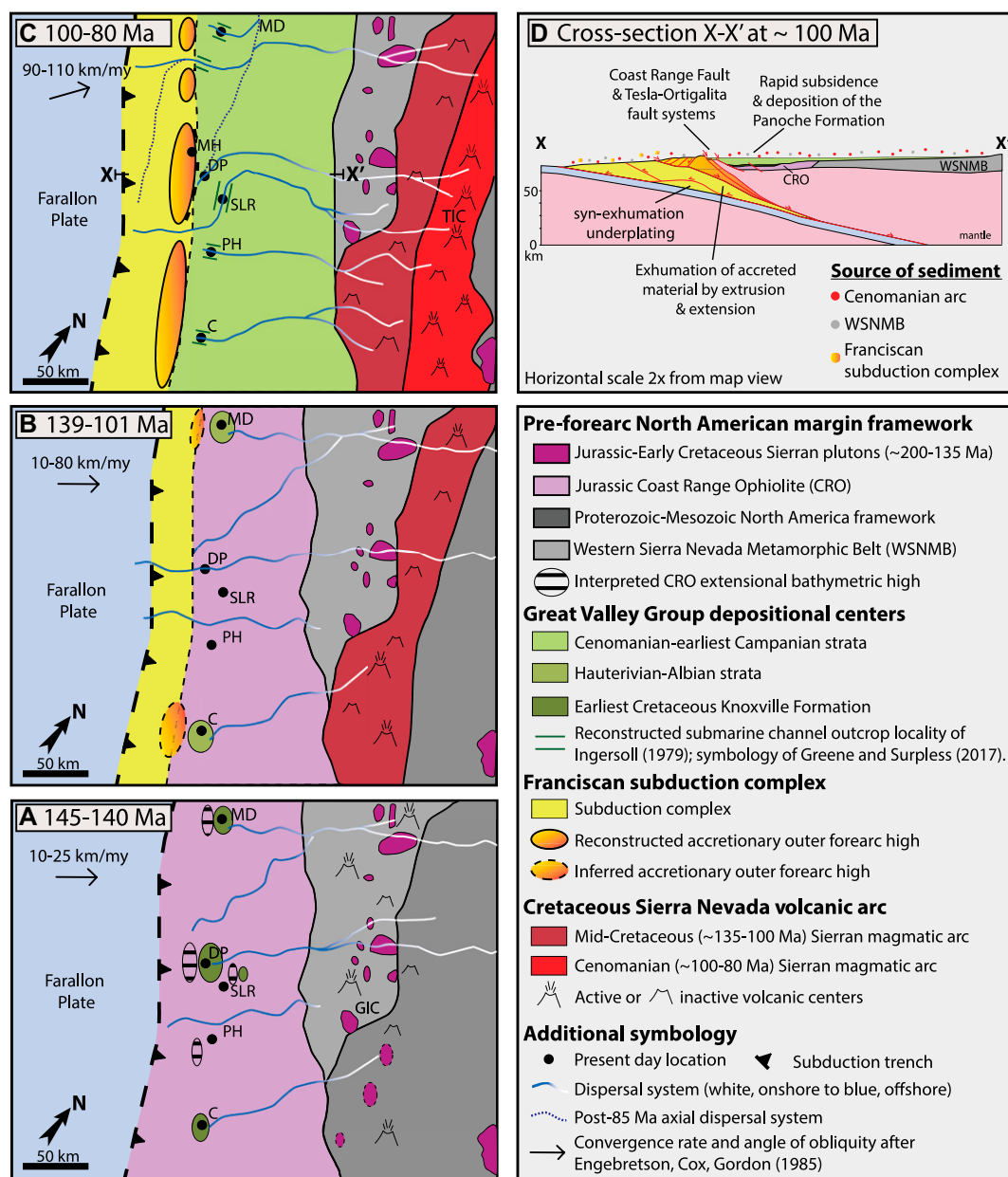


Figure 10. Tectonic model of the Great Valley Forearc (GVF) centered around Del Puerto Canyon (DP, latitude of ~37°N) in the San Joaquin Valley during Early to earliest Late Cretaceous. Legend applies to all panels. Location of the Western Sierra Nevada Metamorphic Belt (WSNMB) and Cretaceous Sierra Nevada volcanic systems from Greene and Surpless (2017). C—Coalinga; GIC—Guadalupe Igneous Complex; MD—Mt. Diablo; MH—Mount Hamilton; PH—Panoche Hills; SLR—San Luis Reservoir; TIC—Tuolumne Igneous Complex. (A) Earliest Cretaceous (145–140 Ma): During onset of the GVF, sediment ponded in isolated basins (green polygons) atop the Coast Range Ophiolite (CRO) and its submarine topography. Provenance was primarily from Jurassic sources of the CRO and plutons of the WSNMB and Sierra Nevada, with minimal input from the Proterozoic–Mesozoic North America framework. (B) Middle Early Cretaceous (139–101 Ma): Depiction of sediment bypass to the trench to explain little to no preservation of sediments of this age within the San Joaquin Valley at the latitude of Del Puerto Canyon while the Sacramento basin accumulated sediments. The middle-Cretaceous magmatic arc developed east of the WSNMB and experienced low-flux events as compared to the Jurassic and late Cretaceous arcs. Earliest records of the Franciscan at these latitudes suggest accretion of sediments accelerated between 123 Ma and 115 Ma (e.g., Dumitru et al. 2018; this study). (C) Earliest Late Cretaceous (100–80 Ma): Forearc basin widens as the Franciscan subduction complex grows westward and generates outer forearc topographic highs behind which to trap sediment. Differences in provenance between localities supports variable sediment pathways to forearc depositional centers, with Del Puerto Canyon primarily sourced from the Campanian volcanic arc, whereas San Luis Reservoir, only 40 km to the south, received more detritus from the WSNMB (see Fig. 9; multidimensional scaling). By contrast, samples from the Panoche Formation at Mt. Diablo to the north and Coalinga to the south are primarily sourced from the middle Cretaceous (139–100 Ma) volcanic arc. We interpret axial dispersal systems within the forearc and trench by 85 Ma, which transported Proterozoic age zircons from the interior of North America to the margin (Dumitru et al., 2018; this study). Axial transport is documented within coeval strata at San Luis Reservoir (Greene and Surpless, 2017). (D) Schematic cross section adapted from Wakabayashi (2015) and Greene and Surpless (2017) highlighting the role of exhumation and extension across the Franciscan–GVF interface to drive basin subsidence atop the CRO.

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Detrital zircon age distributions from neighboring localities of the Franciscan in the northern Diablo Range have MDAs between ca. 104 and 93 Ma, with prominent age modes at ca.

160–140 Ma and ca. 120–115 Ma (Das et al., 2025; Dumitru et al., 2018; Ernst et al., 2009) and abundant pre-Mesozoic grains that lack the ca. 1380 Ma Kione signal. Our three Franciscan

samples with MDAs between ca. 115 Ma and 94 Ma have age spectra consistent with these prior samples and support correlations with the Angel Island or Alcaraz accreted terranes. Our

sample with a ca. 115 Ma MDA shows a strong similarity with sandstones collected from the Franciscan subduction complex at Angel Island with MDAs spanning ca. 115–105 Ma (Das et al. 2025; Apen et al., 2021). By contrast, our samples with MDAs <100 Ma more strongly match age peaks from coeval Franciscan sandstones at San Luis Reservoir to the south (Ernst et al. 2009) and those from across the Northern Diablo Range, including the Alcatraz terrane (Dumitru et al. 2018; Fig. 9A). Notably, our pre-Mesozoic age spectra are almost identical to those from San Luis Reservoir, but the latter lacks the distinct Appalachian-derived signal (ca. 450–250 Ma; Fig. 9A), which is found in other Northern Diablo Franciscan sandstones; we suspect the lack of this Paleozoic age range reflects the low-*n* (<100/sample) and is not reflective of process. Th/U of zircons from sample 23DP12 plot solidly in the felsic melt zone and overlap with Panoche and Knoxville grains (Fig. 7B). Interestingly, ϵ Hf data of 23DP12 show overlap of Jurassic age grains with GVG strata and the Sierran arc signature but show more evolved signatures for four Cretaceous grains relative to the GVG (Fig. 7A; Attia et al., 2020). Additional zircon geochemical data (Th/U and ϵ Hf) from the Franciscan subduction complex is needed to fully investigate this interpretation. This general pull-down in ϵ Hf space, although only evident from four grains, is consistent with the Hf isotopic signatures of North American basement (e.g., Gehrels and Pecha, 2014) and does not require input from the Kione River system, which we interpret to arrive ca. 85 Ma, after deposition of these samples and at the time of deposition of sample 23DP16.

Tectonic Interpretations of the Geologic History in Del Puerto Canyon

Late Jurassic (ca. 161–149 Ma):

Emplacement of the Coast Range Ophiolite

At Del Puerto Canyon, the crystallization age of the CRO spans from ca. 161 Ma for the plutonic facies to as young as ca. 149 Ma for the uppermost volcanics (Lanphere, 1971; Hopson et al., 1981; Evarts et al., 1992; Mattinson and Hopson, 2008). This age range is slightly younger than the majority of CRO ophiolite fragments, which have crystallization ages of ca. 164–168 Ma, although there are several exceptions (see summary in Arkula et al., 2023). Cross-cutting dikes of the uppermost CRO and the overlying Lotta Creek Tuff are dated to ca. 149 Ma (Evarts et al., 1992), indicating an $\sim$ 3 m.y. unconformity between the uppermost volcanics of the CRO and the Knoxville Formation (149–146 Ma). The geochemical signatures of the Del Puerto Canyon CRO

volcanic samples (61621A and 23DP05G) plot between MORB and continental arc on a Th-La-Sc ternary diagram, and closer to andesite than MORB on a Zr/Sc versus Th/Sc plot (Fig. 8). In addition, our CRO volcanic samples generally follow the same REE trends as Jurassic and Cretaceous plutons from northern, central, and southern Sierra Nevada (Fig. 8A; Cecil et al., 2012). These REE spectra are more geochemically evolved than previous CRO analyses that lack the large negative Eu anomaly (Wakabayashi et al., 2010; Shervais and Choi, 2012; Giaramita et al., 1998). These previous studies focused on the basaltic and plutonic assemblages of the CRO. By contrast, our samples are calc-alkaline andesites collected within 3 m of the contact with the Knoxville Formation, which was deposited $\sim$ 3 m.y. after the youngest dated volcanics (149–146 Ma). Thus, we interpret the negative-Eu anomaly of the uppermost CRO volcanics at Del Puerto Canyon to be the product of continental crustal contamination. Our CRO volcanic samples plot amongst Lower Cretaceous strata from farther south within the SJV (Surpless, 2015; Figs. 8B and 8C). This striking similarity suggests a shared magmatic origin. Surpless (2015) proposed the foothills of the southwestern Sierra Nevada western metamorphic belt (i.e., WSNMB) as a potential source for their SJV sandstones at San Luis Reservoir due to their abundance of metavolcanics and zircon-poor (i.e., more mafic) rocks. Our geochemistry suggests the CRO may have also been a source for the Lower Cretaceous strata studied by Surpless (2015; discussed in the following paragraph), in addition to metavolcanic sources within the WSNMB and plutons of the Sierran magmatic arc.

Our petrographic and geochemical results do not directly constrain the tectonic origins of the lower (peridotite member) and middle (plutonic-lower volcanic members) portions of the CRO, which at Del Puerto Canyon are suggested to be an island-arc or suprasubduction zone setting based on previous geochemical analyses (Evarts, 1977, 1978; Evarts et al., 1999). However, our petrography and geochemistry of the uppermost volcanics, collected at the contact with the Knoxville Formation, are most consistent with input from continental crust. Therefore, we speculate that these volcanics erupted in a forearc position, within the upper plate. This interpretation is consistent with the pronounced negative Eu anomaly and similarity with other GVG mudstones analyzed farther south within the SJV basin (Surpless, 2015). In addition, this is also consistent with the timing of arc-arc collision of the Sierra Nevada western metamorphic belt with North America (Nevadan Orogeny), which occurred by ca. 165 Ma at the latitude of

the central Sierra Nevada (Edelman and Sharp, 1989; Attia et al., 2021). The timing of the suturing of the western metamorphic belt therefore precedes formation of oceanic crust of the Del Puerto CRO and is indirect evidence that these younger ophiolitic segments formed in a forearc position.

Earliest Cretaceous (ca. 145–140 Ma):

Initiation of the Forearc Basin

The Knoxville Formation records the earliest GVG sedimentation atop the CRO, $\sim$ 3 m.y. (149–146 Ma) after the CRO crystallized and overlying volcanics erupted. The Knoxville Formation was sourced from the CRO and the proximal Sierra Nevada magmatic arc and western foothills (Fig. 10A). The Knoxville Formation appears to be more extensively preserved at Del Puerto Canyon than in other regions of the SJV yet is significantly thinner than the latest Jurassic–earliest Cretaceous Stony Creek Formation, which forms the basal strata of the GVG in the northern GVF (Moxon, 1988). Given minimal accretionary processes at the Franciscan trench during Knoxville deposition (Dumitru et al., 2010; Mulcahy et al., 2018), the SJV part of the GVF may have been a sloped forearc with a minor trench slope break to trap sediment behind it (Dickinson and Seely, 1979; Dickinson, 1995; Fig. 10A).

Within the Sacramento basin, extension of the CRO prior to and coeval with deposition of the basal GVG is invoked to explain the generation of accommodation during the earliest stages of basin deposition (Constenius et al., 2000; Romero et al., 2024). Part of the evidence for extension comes from seismic data that shows high bathymetric relief ($\sim$ 6 km) of the CRO-GVG contact and attenuation of the CRO along strike of the outcrop belt within the northern Sacramento basin (Constenius et al., 2000). Original thicknesses of the CRO in the SJV are challenging to reconstruct owing to deformation along the Coast Range and Tesla-Ortigalita fault systems, and minimal preservation of the GVG-CRO contact in outcrop. However, it is interesting that at Del Puerto Canyon, the CRO preserves an almost complete ophiolite sequence (5.4 km thick; Hopson et al., 1981; Evarts et al., 1992), but $\sim$ 40 km south, just north of Pacheco Pass, the CRO is only 0.91 km thick (Schilling, 1962; Moxon, 1988). We speculate that these differences in thickness of the CRO may reflect extensionally driven thinning of the CRO prior to and concurrent with GVG deposition.

Thermal relaxation (i.e., cooling) of a newly formed ophiolite combined with extension-related generation of bathymetric relief of the CRO may have been the driving mechanisms for accommodation and accumulation of GVG

sediments at the onset of forearc basin development within the SJV. The ca. 149 Ma dikes that cross-cut older, ca. 160 Ma ultramafics and plagiogranites of the CRO at Del Puerto Canyon support the idea that the CRO was in a state of extension ~ 3 m.y. prior to GVG deposition. In addition, the highly negative Eu anomaly present in the uppermost volcanics of the CRO suggest they erupted proximal to a continental arc, a process we speculate may have been accommodated by extension within the forearc region. The arc-arc collision of the Nevadan Orogeny may have generated “tectonic highlands” (Ingersoll, 1982, p. 462) along the suture zone and variable paleobathymetric features within the older, pre-165 Ma parts of the CRO, for example, in the northern Sacramento basin (Ingersoll, 1982; Romero et al., 2024). However, at Del Puerto Canyon, crystallization of the CRO post-dates the arc-arc suturing process. Given the ~ 5.4 -km-thick ophiolite sequence preserved at Del Puerto Canyon, lack of documented deformation of the CRO in the subsurface to date, and very thin to missing ophiolitic material to the south within the SJV, we favor a model whereby extension of the forearc basement drove the generation of sea floor topography within which to trap sediment in the newly forming forearc basin. This is consistent with the proposed timing of extension in the northern Great Valley forearc during the onset of GVG sedimentation (Romero et al., 2024).

Middle Cretaceous (ca. 139–101 Ma):

Absence of Preserved Sediment

Within the SJV at Del Puerto Canyon, there is no stratigraphic evidence of late Early Cretaceous forearc basin deposition, with an ~ 40 m.y. unconformity between the youngest Knoxville Formation (ca. 139 Ma, Valanginian YSP MDA) and onset of Panoche Formation deposition (ca. 97 Ma, Cenomanian YSP MDA). This is consistent with previous biostratigraphy of the region, and applying different MDA methods does not change the duration of this unconformity. By contrast, Lower Cretaceous (Valanginian–Hauterivian) strata are preserved in the Mt. Diablo and Coalinga areas of the SJV, ~ 70 km north and ~ 170 km south, respectively, of Del Puerto Canyon (Moxon, 1988). At these localities, Albian or Cenomanian strata of the Panoche Formation overlie Lower Cretaceous strata, resulting in an unconformity within the SJV of between 18 m.y. (Coalinga) and 40 m.y. (Del Puerto Canyon). This lack of rock record contrasts with ample Early Cretaceous strata in the Sacramento basin, including deposition of the Platina and Lodoga Formations (ca. 132–100 Ma; e.g., Ingersoll, 1979, 1983) during this interval.

Here, we discuss two plausible explanations for the timing and duration of the ~ 40 –18 m.y. unconformity in the northern SJV, centered at the latitude of Del Puerto Canyon.

First, the unconformity coincides with reduced magmatism within the Sierra Nevada and little accretion at the trench at the latitude of the SJV (e.g., Paterson and Ducea, 2015; Dumitru et al., 2018; Mulcahy et al., 2018; Apen et al., 2021). Without an actively growing accretionary wedge, the forearc lacked an outer forearc high, behind which to trap sediments. Therefore, the sloped forearc may have transitioned from underfilled to filled after localized deposition of the Knoxville Formation, leaving no accommodation for sediments to accumulate and causing sediment to bypass to the trench or nearby basins in the GVF (Fig. 10B). By contrast, deposition of middle Cretaceous sediments at Mt. Diablo and Coalinga may reflect the initiation of an outer forearc high that formed by topographic growth of the Franciscan subduction complex (Fig. 10B).

A reduction in erosion rates from along convergent margins correlates with magmatic lulls, during which the rate of igneous pluton emplacement and topographic gradients reduce (Val et al., 2018); this reduction in sediment is consistent with limited deposition within the forearc and accumulation of sediments within the trench. Within the Sacramento basin in the northern GVF, Lower Cretaceous sediments accumulated at low sediment flux (e.g., Moxon, 1988; Orme and Graham, 2018); volumetric sediment accumulation rates within the Sacramento basin were less than $2000 \text{ km}^3/\text{m.y.}$ before ca. 100 Ma, but as the magmatism peak approached ca. 90 Ma, sediment accumulation increased ~ 7 -fold to over $15,000 \text{ km}^3/\text{m.y.}$ by ca. 84 Ma (Orme and Graham, 2018). During this same Early Cretaceous period, the Franciscan recorded accretion of the South Fork Mountain Schist outboard of the Sacramento basin and the Yolla Bolly terranes outboard of the Sacramento basin and northern Diablo Range of the SJV basin (Fig. 10B; e.g., Worrall, 1981; Dumitru et al., 2018; Apen et al., 2021). In addition, successions of Upper Cretaceous rocks deposited after this unconformity are thickest in the SJV compared to the Sacramento basin. This spatial shift in deposition is observed in both subsurface and in outcrop studies and supported by observed southward down-lap geometries (Moxon, 1988; Orme and Graham, 2018).

Together, these observations suggest that the forearc had segmented depositional centers, where sediments ponded within discrete forearc basins in some areas and directly fed to the trench basins in others (Fig. 10B). This segmented forearc basin morphology is common

in the recent ancient to modern forearc records along the Java-Sumatra, Aleutian, and Japanese island arc systems (e.g., Schlüter et al., 2002; Yang and Gao, 2020; Okamura and Shishikura, 2020; Utsunomiya et al., 2023). Similar morphologies explain the stratigraphic architecture of Upper Cretaceous strata of the northern GVF (Williams and Graham, 2013).

An alternative hypothesis to explain the development of the unconformity is that sediments of Lower Cretaceous age were deposited in the SJV and subsequently eroded. Greene and Surpless (2017) discuss evidence from Tagami and Dumitru (1996) for rapid exhumation of the Franciscan wedge near Mt. Hamilton, ~ 30 km west of our study area, during ca. 100–70 Ma, as a potential mechanism to erode Lower Cretaceous strata. Importantly, this model is invoked to explain the CRO as a potential source of the Panoche Formation, given the abundance of metavolcanics within clast counts at San Luis Reservoir (Greene and Surpless, 2017). Our sandstone petrographic work does not support the CRO as a source for the lower Panoche Formation at Del Puerto Canyon, and instead, we propose slight provenance differences between the lower pre-Campanian Panoche at Del Puerto Canyon and Campanian strata of the Panoche Formation upsection at Del Puerto Canyon (Sharman et al., 2015) and at San Luis Reservoir (Greene and Surpless, 2017). Application of low-temperature thermochronology to strata preserved across the Lower–Upper Cretaceous contact could shed light on any record of basin uplift and erosion not evident in the preserved stratigraphic record. Therefore, without evidence for deposition and subsequent erosion of strata between ca. 140 Ma and ca. 100 Ma, we favor a model that invokes sediment bypass at this latitude during this time interval (Fig. 10B). Coeval sediment accumulation within the Sacramento basin while parts of the San Joaquin basin primarily experienced sediment bypass is consistent with modern forearc systems that highlight spatial variability of depositional centers (e.g., Aleutian, Kamchatka; e.g., Williams and Graham, 2013).

Late Cretaceous (ca. 100–80 Ma): Accretion at the Trench and Renewed Forearc Basin Deposition

Deposition of the Panoche Formation occurred at a time of (1) increased plate convergence and obliquity (e.g., Engbreton et al., 1985; Liu et al. 2008), (2) high-flux arc magmatism and intra-arc shear zones (e.g., Paterson and Ducea, 2015; Glazner, 1991; Tobisch et al. 1995; Tikoff and Greene, 1997), and (3) wedge accretion and exhumation of high-grade blocks within the Franciscan subduction complex at the latitude of the northern SJV (Fig. 11; e.g.,

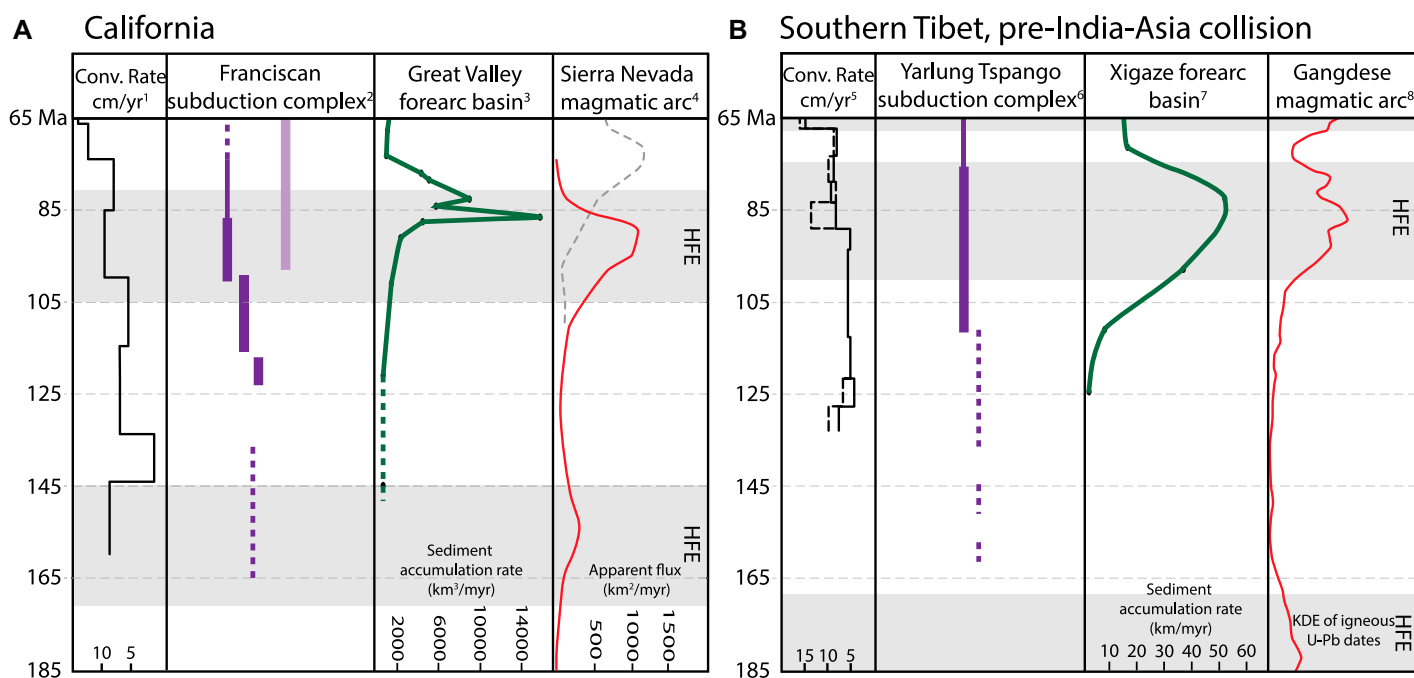


Figure 11. Comparison of Cretaceous sediment flux within the Great Valley (California) and Xigaze (southern Tibet) forearc basins to plate convergence rates, periods of high rates of accretion (dark purple), and exhumation (light purple) of outboard subduction complexes, and records of adjacent magmatic arcs. Both margins experienced an increase in plate convergence during high flux magmatic events (HFE). In both forearc basins, the peak in Cretaceous sedimentation rates occur ~5–10 m.y. after the apex of high flux magmatism and high rates of accretion and/or exhumation of the subduction complex. References: 1—Farallon–North America convergence rate from Engbreton et al. (1985). 2—Compilation from DeCelles and Graham (2015). Dashed purple line indicates periods of high-grade metamorphic block generation. 3—Volumetric sediment accumulation rates from Orme and Graham (2018), Sacramento basin; San Joaquin basin-wide sedimentation accumulation curve for this time interval not plausible owing to lack of boreholes penetrating Cretaceous strata >75 Ma (Hosford Scheirer, 2013). 4—Apparent magmatic flux rate reproduced from Orme and Graham (2018) after Paterson and Ducea (2015; Sierra Nevada) and Gaschnig et al. (2017; Idaho Batholith). 5—India–Eurasia convergence rate at the eastern syntax (dashed) and the western syntax (solid) from van Hinsbergen et al. (2011). 6—Periods of accretion within subduction complex from Metcalf and Kapp (2019). 7—Sediment accumulation curve constructed from compilation in Orme et al. (2015). 8—Igneous U–Pb dates shown as a kernel density estimate (KDE) from Ma et al. (2022, 2023).

Tagami and Dumitru, 1996; Dumitru et al., 2018; Apen et al., 2021; Das et al., 2025). Rapid subsidence and sediment accumulation occurred during the Cenomanian across the Great Valley forearc (e.g., Ingersoll, 1978, 1982; Moxon and Graham, 1987; Orme and Graham, 2018), and Moxon (1990) and Greene and Surpless (2017) correlated this to an increase in the rate of convergence and obliquity between the Farallon and North American plates (Page and Engbreton, 1984). Similarly, changes in plate obliquity have been called upon to explain Cretaceous high-flux magmatism and associated changes in igneous fabrics (e.g., Tomek et al., 2024) and an overall shift to oblique transpression along the margin (e.g., Liu et al., 2008). East of Del Puerto Canyon at Eyer Mountain and Mt. Hamilton, the zircon fission track system records cooling of the Franciscan subduction complex below 210 °C at ca. 100 Ma and to ~90 °C by 70 Ma, interpreted as major exhumation of the subduction complex during this time (Tagami and Dumitru, 1996).

This exhumation supports the establishment of an outer forearc high, or ridged forearc (Dickinson, 1995), behind which sediment shed from the Sierran arcs accumulated with the GVF basin (Fig. 10C). As Tagami and Dumitru (1996) discuss, exhumation of the Franciscan subduction complex at this time was not caused by a collision event due to the “lack of island arc or other large crustal fragments” (p. 11,361) within Franciscan detritus. Rather, these authors and others (Platt, 1986; Jayko et al., 1987; Wakabayashi and Unruh, 1995; Wakabayashi, 2015) invoke accelerated accretion of the frontal Franciscan wedge and extensional thinning along the eastern margin of the wedge (i.e., western forearc) to drive exhumation of the subduction complex. In turn, the interpreted extension within the forearc may have been the driver of observed accelerated basin subsidence and sediment accumulation. At Del Puerto Canyon, the Tesla–Ortogonalita fault was active after Albian time (e.g., Raymond, 1973), and kinematic indicators suggest a

normal sense of motion during latest Cretaceous. Integrating these observations, we propose the Tesla–Ortogonalita and Coast Range faults formed a western forearc basin bounding detachment system that accommodated an increase in basin subsidence during deposition of the Panoche Formation (Fig. 10D).

Based primarily on paleomagnetic reconstructions, several models for the Late Cretaceous evolution of the North America Cordillera have hypothesized collision and translation of the Insular and Intermontane superterrane, now located within the Alaska and Canadian segments of the Cordillera, with the western margin of Laurentia (e.g., Maxson and Tikoff, 1996; Cowan et al., 1997; Tikoff et al., 2023). The recent “hit-and-run” model (Tikoff et al., 2023) proposes collision of the Insular terrane with the Franciscan–Great Valley–Sierra Nevada between 100 Ma and 85 Ma, followed by translation of the Insular terrane after 85 Ma. The proposed timing of collision and translation

coincides with deposition of the Panoche Formation in the northern (96–84 Ma; this study) and central (85–73 Ma; Hosford Scheirer, 2013) SJV and deposition and exhumation of the Franciscan subduction complex (post–100 Ma; Tagami and Dumitru, 1996; Unruh et al., 2007). Tikoff et al. (2023) cite increased rates of deposition in the GVF and Franciscan exhumation as two lines of evidence for the proposed collision and northward translation of the Insular terrane. However, exhumation of the Franciscan subduction complex by collisional tectonics is inconsistent with geologic evidence for extension and GVF basin subsidence at this time, as evidenced by the Coast Range and Tesla-Ortgalita detachment faults at Del Puerto Canyon and GVG deposition, respectively (Fig. 10D; Harms et al., 1992; Raymond, 1973). In addition, the proposed collision should affect the source of detritus being deposited at the trench (i.e., non-North America sources), but detrital zircon and geochemical signatures of the Franciscan show a shared provenance with the GVG that contain sediments derived from North America (e.g., this study; Ghatak et al. 2013; Sharman et al., 2015; Dumitru et al., 2018; Apen et al., 2021). Furthermore, we argue that increased rates of subsidence within the GVF were a natural basin response to changes within the magmatic arc and exhumation of the subduction complex related to oceanic subduction beneath North America.

Our study documents the response of a forearc basin to increased rates of sediment production and erosion during high-flux magmatism and increased sedimentation within the trench. To illustrate this, we compare the GVF to the Xigaze forearc basin in southern Tibet, which developed along the southern margin of Asia during Cretaceous time (Fig. 11; e.g., Einsele et al., 1994; Orme et al., 2015; Wang et al., 2017). In both forearc basins, peak sedimentation rates followed the apex of high flux magmatic addition within ~5–10 m.y. and a coeval increase of accretion and exhumation along their neighboring subduction complexes (Fig. 11). Increased sediment accumulation in forearc basins is consistent with numerical models that show rates of erosion remain high following the peak in high-flux magmatism (e.g., Lee et al. 2015). In addition, an increase in the flux of sediment entering the trench leads to high rates of accretion of neighboring subduction complexes (e.g., Wakabayashi, 2015). In turn, this leads to increased topographic growth of an outer forearc high behind which to trap sediments, one of the primary mechanisms of forearc basin development (e.g., Dickinson, 1995). Periods of exhumation of the subduction complex are accommodated by extension, which is well documented by the Coast Range fault in the GVF-Franciscan system

(e.g., Harms et al., 1992; Wakabayashi, 2015). Extension will further drive forearc basin subsidence, in addition to the increased sediment load from adjacent high-flux magmatism. We argue that the Late Cretaceous GVF record, both at Del Puerto Canyon and basin-wide, reflects a forearc basin response to these convergent margin processes and does not constitute geologic evidence for terrane collision or translation.

CONCLUSIONS

The Late Jurassic through early Late Cretaceous geologic history of Del Puerto Canyon in central California highlights the integrated history of the Franciscan subduction complex, CRO, GVG, and Sierra Nevada magmatic arc. The geochemistry of the uppermost volcanics of the CRO at this locality suggests they erupted in an upper plate, forearc position. The most basal GVG strata in Del Puerto Canyon, the Knoxville Formation, were deposited from ca. 145 Ma to 139 Ma, ~2–5 m.y. after emplacement of the upper volcanics of the CRO as its underlying basement. The Knoxville Formation was derived from the proximal felsic to intermediate Jurassic–earliest Cretaceous plutons of the Sierra Nevada magmatic arc, with minor components of mafic minerals sourced from exposed CRO. The basin was likely a sloped forearc, with little accretionary prism outboard and may have become an overfilled basin and locus of sediment bypass shortly after deposition of the Knoxville Formation atop the CRO, resulting in an ~40 m.y. unconformity with the Late Cretaceous Panoche Formation. We favor a model of sediment bypass to explain the unconformity, because the time interval (ca. 140–100 Ma) coincides with little evidence for accretion or metamorphism within the Franciscan trench outboard of the SJV basin at the latitude of Del Puerto Canyon and occurs during a magmatic lull in the Sierra Nevada arc, which is characterized by lower erosion rates across the GVF.

At Del Puerto Canyon, the Late Cretaceous Panoche Formation and Franciscan subduction complex record a return to sedimentation and rapid subsidence and accretion along the trench, respectively. The increase in rapid sediment accumulation within the GVF basin during this time likely reflects the growth of the accretionary prism acting as an outer forearc high, behind which to trap sediments, preventing their escape into the trench, and accelerated basin subsidence resulting from extension along the western margin of the basin. In turn, these processes are consistent with and reflect the return to high flux magmatism within the Sierra Nevada, owing to increased rates of oblique convergence (Engelbreton et al., 1985). After ca. 85 Ma, the mul-

timodal age spectra of the Panoche Formation, including a ca. 1380 Ma peak, indicates sediment sources from the interior of North America (i.e., Kione River from Idaho; Dumitru et al., 2016), in addition to the Jurassic sources present during initial deposition of the Knoxville Formation within the GVF basin. This study integrates multiple data sets across different lithotectonic units preserved within an ancient forearc to better understand processes active during forearc basin initiation and rapid basin subsidence.

ACKNOWLEDGMENTS

Funding was provided by the National Science Foundation (NSF) grant EAR-CAREER-1942460 to D.A. Orme. We are grateful to the landowners in Del Puerto Canyon for graciously allowing us to visit their properties and to collect samples from their land. Thank you to the California Well Sample Repository, specifically Charles James, for sample collection assistance, the University of Arizona LaserChron Center and staff for analytical support (NSF EAR-2050246), ZirChron for assistance with mineral separation, Washington State University for geochemical analyses, and Spectrum Petrographics for thin sections. We are thankful for Kurt Sundell for his assistance with Lu-Hf data reduction and help with MATLAB programs, and for Zachary Morrow, Natali Kragh, and Rebekah Kennedy for discussions regarding petrography and geochemistry. Thank you, John Wakabayashi, an anonymous reviewer, *GSA Bulletin* Associate Editor M. Elliot Smith, and *GSA Bulletin* Science Editor Troy Rasbury for thoughtful reviews of this work.

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SCIENCE EDITOR: TROY RASBURY
ASSOCIATE EDITOR: M. ELLIOT SMITH

MANUSCRIPT RECEIVED 29 OCTOBER 2024
REVISED MANUSCRIPT RECEIVED 17 DECEMBER 2024
MANUSCRIPT ACCEPTED 17 APRIL 2025