DOI:10.1130/B36076.1

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GSA Bulletin; Month/Month 2021; v. 133; no. X/X; p. 000–000; https://doi.org/10.1130/B36076.1; 11 figures; 3 tables; 1 supplemental file.
[1]Supplemental Material. **[[Please provide a brief description.]]**. Please visit https://doi.org/10.1130/GSAB.S.XXXX to access the supplemental material, and contact editing@geosociety.org with any questions.
SCIENCE EDITOR: ROB STRACHAN
ASSOCIATE EDITOR: JOHN JANSEN
MANUSCRIPT RECEIVED 10 FEBRUARY 2021
REVISED MANUSCRIPT RECEIVED 18 JUNE 2021
MANUSCRIPT ACCEPTED 26 OCTOBER 2021
Printed in the USA

Tectonic controls on Quaternary landscape evolution in the Ventura basin, southern California, USA, quantified using cosmogenic isotopes and topographic analyses

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ABSTRACT

The quantification of rates for the competing forces of tectonic uplift and erosion has important implications for understanding topographic evolution. Here, we quantify the complex interplay between tectonic uplift, topographic development, and erosion recorded in the hanging walls of several active reverse faults in the Ventura basin, southern California, USA. We use cosmogenic ^{26}Al/^{10}Be isochron burial dating and ^{10}Be surface exposure dating to construct a basin-wide geochronology, which includes burial dating of the Saugus Formation: an important, but poorly dated, regional Quaternary strain marker. Our ages for the top of the exposed Saugus Formation range from 0.36 $^{+0.18}/_{-0.22}$ Ma to 1.06 $^{+0.23}/_{-0.26}$ Ma, and our burial ages near the base of shallow marine deposits, which underlie the Saugus Formation, increase eastward from 0.60 $^{+0.05}/_{-0.06}$ Ma to 3.30 $^{+0.30}/_{-0.41}$ Ma. Our geochronology is used to calculate rapid long-term reverse fault slip rates of 8.6–12.7 mm yr^{-1} since ca. 1.0 Ma for the San Cayetano fault and 1.3–3.0 mm

yr^{-1} since ca. 1.0 Ma for the Oak Ridge fault, which are both broadly consistent with contemporary reverse slip rates derived from mechanical models driven by global positioning satellite (GPS) data. We also calculate terrestrial cosmogenic nuclide (TCN)-derived, catchment-averaged erosion rates that range from 0.05–1.14 mm yr^{-1} and discuss the applicability of TCN-derived, catchment-averaged erosion rates in rapidly uplifting, landslide-prone landscapes. We compare patterns in erosion rates and tectonic rates to fluvial response times and geomorphic landscape parameters to show that in young, rapidly uplifting mountain belts, catchments may attain a quasi-steady-state on timescales of $<10^5$ years even if catchment-averaged erosion rates are still adjusting to tectonic forcing.

1. INTRODUCTION

Landscape form is a product of the complex interplay between climate, tectonics, and the strength of the eroding rocks (e.g., Whipple and Tucker, 1999; Clark et al., 2004; Anders et al., 2005; Wobus et al., 2006; Yang et al., 2015). Climatic factors, such as precipitation and temperature, strongly influence many aspects of landscape evolution including erosion rates, channel morphology, and relief development (Portenga and Bierman, 2011; Champagnac et al., 2012; DeVecchio et al., 2012a; D'Arcy and Whittaker, 2014). Conversely, other studies have demonstrated that tectonic deformation is the primary control on erosion rates (Scherler et al., 2014; Bermúdez et al., 2013; Val et al., 2018; Roda-Boluda et al., 2019) and drives the development of topographic relief (Ellis and Barnes, 2015), which in turn can be modulated by lithological parameters like rock strength (Densmore et al., 2004; Duvall et al., 2004; Townsend et al., 2020). However, isolating tectonic signals from the landscape is often challenging because direct measurements of tectonic deformation (e.g., fault-slip rates) often have limited temporal and spatial resolution (Roberts and Michetti, 2004; Whittaker et al., 2008; Kirby and Whipple, 2012). A low-resolution record of tectonic rates can hinder our ability to analyze the contribution of tectonic forcing to erosion signals obtained from measures such as catchment incision data, channel morphology, or river longitudinal profile analyses (e.g., Whipple and Tucker, 2002; Whittaker et al., 2007; Stock et al., 2009; Cyr et al., 2010; Kirby and Whipple, 2012; Kent et al., 2017). Furthermore, isolating the tectonic contribution to landscape erosion can be complicated by spatial and temporal climate variability, which on a regional scale often exerts a stronger control on landscape morphology than tectonics (Champagnac et al., 2012; DeVecchio et al., 2012a; Bookhagen and Strecker, 2012). Therefore, recent studies aiming to extract tectonic and lithologic signals from the landscape have focused on the scale of individual basins, where climate can be considered spatially uniform (Godard et al., 2014; Ellis and Barnes, 2015; McCarthy et al., 2019).

A common starting point when extracting the tectonic contribution from landscapes is to compare tectonic uplift rates with catchment-averaged erosion rates and morphometric landscape parameters such as relief and/or channel steepness, the latter of which is typically obtained at far higher spatial density than the tectonic rate constraints (Vance et al., 2003; Densmore et al., 2009; Stock et al., 20009; Cyr et al., 2010; Miller et al., 2012; Wang et al., 2017). Therefore, the key requirement for any thorough assessment of landscape response to tectonic forcing is a high-resolution record of fault activity, e.g., tectonic rock uplift and/or fault slip rates, and an accurate quantification of erosion rates (Stock et al., 2009; Cyr et al., 2010; Roda-Boluda et al., 2019). Developments in Quaternary dating techniques over the past 20 years have provided geoscientists with the ability to quantify rock uplift rates via the precise dating of strain markers and quantification of erosion rates on various timescales between 10^2–10^6 years (Granger and Muzikar, 2001; Balco and Rovey, 2008; Balco et al., 2008; Granger et al., 2013). For example, improvements in the chemistry required for multiple cosmogenic isotope isochron burial dating (Corbett et al., 2016b) and accelerator mass spectrometry techniques (Rood et al., 2010; Wilcken et al., 2017, 2019) along with recent validation of the isochron burial dating method via a comparison with existing K–Ar and $^{40}Ar/^{39}Ar$ chronology (Zhao et al., 2016) provide the necessary tools to create a high-resolution geochronology for previously undatable Quaternary terrestrial sediments. Where data are available, such geochronology enables inferences about whether present-day relationships between rock uplift, erosion, and sedimentation are representative of a long-term signal that can be applied to model landscape evolution (Armitage et al., 2011; DeVecchio et al., 2012a; Corbett et al., 2016a) or to quantify seismic hazards (Kirby et al., 2008; Boulton and Whittaker, 2009; Whittaker and Walker, 2015; Bender et al., 2016).

In this paper, we address this important research challenge in the Ventura Basin, California, which contains several active reverse and thrust faults in close proximity to densely populated areas in the Santa Clara River Valley and the wider Los Angeles metropolitan area (Fig. 1). Due to its proximity to major population centers, calculating accurate fault displacement rates in the basin is critically important for seismic hazard assessment. We apply cosmogenic isotope isochron burial dating to quantify spatial variations in the age of the Saugus Formation, a regionally significant but poorly dated tectonic marker. We place the resulting geochronology in the context of previously published studies and apply the data to reevaluate slip rates. We also calculate terrestrial cosmogenic nuclide (TCN)-derived erosion rates and discuss the applicability of TCN-derived erosion rates in rapidly uplifting, landslide-prone areas. The displacement rates provide the tectonic context for a comparison with spatial patterns in relief, channel morphology, erosion rates, and fluvial response times to investigate the tectonic contribution to landscape evolution in the Ventura basin throughout the late Quaternary.

2. TECTONIC AND STRATIGRAPHIC SETTING

2.1 Late Cenozoic Geological Setting

In this work, we focus on the San Cayetano, Southern San Cayetano, and Ventura faults (Fig. 2), which are all thought to represent significant contemporary earthquake hazards (Field et al., 2014; Field et al., 2015) and are believed to have contributed significantly to the landscape evolution of the onshore Ventura basin (Rockwell et al., 1984; Azor et al., 2002; DeVecchio et al., 2012a). Furthermore, activity on the San Cayetano, Ventura, and Southern San Cayetano faults is thought to have initiated at different times within the same tectonic setting and a similar lithologic setting (Rockwell et al., 1984; Rockwell, 1988; Çemen, 1989; Hubbard et al., 2014; Hughes et al., 2018), which makes these faults an ideal location to examine the extent to which tectonic deformation has controlled patterns of erosion and relief development through time.

The Ventura basin is an east-west–trending, fault-bounded, sedimentary trough situated in the Western Transverse Ranges of southern California (Figs. 1–2). Within the basin, the marine Modelo Formation accumulated during the Miocene in a transtensional regime (Yeats et al., 1994) above an Oligocene-Eocene sedimentary succession (Figs. 2–3) (Bailey, 1947; Dibblee, 1950; Vedder et al., 1969). Subsequently, a switch to transpressional deformation occurred in the late Miocene to early Pliocene due to the formation of the “Big Bend” in the San Andreas Fault (e.g., Crowell, 1976; Wright, 1991). The current structural framework of the Ventura basin is largely a product of post-Miocene inversion of transtensional basins in a transpressional regime (Crowell, 1976; Yeats et al., 1994). Contemporaneous with transpressional deformation, up to 7 km of Plio-Pleistocene bathyal to shallow marine and alluvial deposits accumulated over older, deep marine strata (e.g., Yeats and Rockwell, 1991; Yeats et al., 1994), which have since been highly tilted and locally overturned (Campbell et al., 2014). Transpressional deformation is expressed by north-south–directed regional crustal shortening at rates of 7–10 mm yr^{-1} (Donnellan et al., 1993b; Marshall et al., 2013) and is accommodated by a series of east-west–striking reverse faults and associated folds (Fig. 2). Cenozoic bedrock in the basin is overlain by a series of late Pleistocene to Holocene strath terraces and alluvial fill terraces, which show varying degrees of tilting and deformation (Rockwell et al., 1988; DeVecchio et al., 2012a; Hughes et al., 2018).

The San Cayetano fault is a north-dipping reverse fault that trends east-west for ~40 km from Piru to the Upper Ojai Valley (Fig. 1; Rockwell, 1988). The fault is often separated into an eastern section and a western section (Figs. 1–2), which have different slip rates and geomorphic expression (Rockwell, 1988; Çemen, 1989; Dolan and Rockwell, 2001). The onset of surface uplift related to the San Cayetano fault is not well quantified, but estimates range from commencement at around 1.0 Ma (Çemen, 1989), to as long ago as ca. 3.3 Ma (Rockwell, 1982). Existing slip rates for the eastern San Cayetano fault have a broad range between 4.4 mm yr^{-1} and 10.4 mm yr^{-1} since 500 ka (Huftile and Yeats, 1996), and late Pleistocene to Holocene slip rates for the western San Cayetano fault range from 1.0 mm yr^{-1} to 5.0 mm yr^{-1} (Rockwell, 1988).

The Southern San Cayetano fault is a low-angle thrust fault that was recently identified immediately east of the Ventura fault in the footwall of the San Cayetano fault (Hubbard et al., 2014; Hughes et al., 2018; Hughes et al., 2020) and is thought to have been active since ca. 58 ka with a slip rate of 1.0–1.8 mm yr^{-1} (Hughes et al., 2018).

The Ventura fault is a north-dipping reverse fault that is mapped onshore for 17 km from Saticoy (Fig. 1) to the coast and continues westward offshore as the Pitas Point fault (Fig. 1) (Sarna-Wojcicki et al., 1976; Sarna-Wojcicki and Yerkes, 1982; Hubbard et al., 2014). Surface uplift related to the Ventura fault is thought to have begun with the commencement of uplift of the Ventura Avenue anticline (Fig. 2) at around 200–250 ka (Sarna-Wojcicki et al., 1976; Sarna-Wojcicki and Yerkes, 1982; Rockwell et al., 1988), and the Holocene slip rate for the Ventura fault is calculated at 4.4–6.9 mm yr^{-1} (Hubbard et al., 2014).

2.2 The Saugus Formation

The Saugus Formation is an important but poorly dated strain marker in the Ventura basin (Fig. 2) and comprises a deformed sequence of terrestrial fluvial and alluvial sediments of Pleistocene age (Levi and Yeats, 1993). The Saugus Formation is often described as the youngest of the deformed bedrock strata and is argued to record the Pleistocene onset of activity on several major active faults and related drainage reorganizations (Levi and Yeats, 1993; DeVecchio et al., 2012a, 2012b). However, while the Saugus Formation is thought to be time-transgressive, few direct ages (i.e., ages calculated from samples taken in situ rather than inferred by correlation with ex-situ chronostratigraphic markers) for the Saugus Formation exist (Wehmiller et al., 1978; DeVecchio et al., 2012a). As a result, there are insufficient data points to directly quantify the spatial variability in the age of either the top or base of the exposed Saugus Formation across the entire basin. Despite uncertainties in the chronology, slip rates for multiple faults in the Ventura basin are calculated using the Saugus Formation as a strain marker. Specifically, slip rates for important active faults, such as the Oak Ridge (Yeats, 1988; Yeats et al., 1994), San Cayetano (Huftile and Yeats, 1995, 1996), and Santa Susana faults (Fig. 1; Huftile and Yeats, 1996; Levy et al., 2021) are based on projecting ages of a time-transgressive unit over tens of kilometers, which results in large inherent uncertainties in the fault slip rates. Therefore, to track patterns of late Quaternary deformation across the Ventura basin, a robust chronology of direct ages across the entire Saugus Formation is required.

There are different interpretations for which lithological units or depositional environments should be included in the definition of the Saugus Formation. The convention adopted here is to apply the term Saugus Formation only to terrestrial deposits and use separate terms for shallow marine and brackish deposits that underlie the Saugus Formation throughout the Ventura basin (Figs. 2–3; Winterer and Durham, 1962; Campbell et al., 2014; Swanson and Irvine, 2015). In the western basin, shallow marine deposits are assigned to the Las Posas Formation, whereas in the central basin the underlying shallow marine deposits comprise the Grimes Canyon deltaic facies (Figs. 2–3; Campbell et al., 2014). In the eastern basin, the Sunshine Ranch Member is a transitional unit that consists of a mixture of both shallow marine and brackish water deposits (Figs. 2–3; Winterer and Durham, 1962; Campbell et al., 2014).

Previous studies of the terrestrial Saugus Formation in the east Ventura basin suggest an age range of 0.6–2.3 Ma based on an assumption of constant sedimentation rates projected above and below the 0.76 Ma Bishop ash (Levi and Yeats, 1993). The Saugus Formation was also assigned to the Matuyama reversed magnetic chron based on a paleomagnetic transect through the Saugus Formation outcrop in the eastern Ventura basin (Levi and Yeats, 1993). The minimum age of the Saugus Formation in the western Ventura basin is constrained by an age of 205 ± 25 ka for the underlying Las Posas Formations at Ventura, which is based on amino acid racemization on *Macoma* mollusk shells (Wehmiller et al., 1978). However, the racemization age has been questioned by various authors and could be an underestimate, because both the kinetic and thermal models employed in the calculation assumed the same thermal history for the bedrock units as the overlying marine terrace despite an obvious angular unconformity being present between the two units (Yeats, 1988; Wehmiller, 1992; Huftile and Yeats, 1995). More recent work used a combination of optically stimulated luminescence (OSL) and cosmogenic nuclide dating to calculate a lower age for the Camarillo member of the Saugus Formation in the southern Ventura basin of 125 ka and an upper age of 60–25 ka, which is significantly younger than previously thought (DeVecchio et al., 2012a; DeVecchio et al., 2012b). To improve the existing chronology of the Saugus Formation, we focus on characterizing the age of the youngest exposed terrestrial Saugus Formation deposits and the base of underlying shallow marine or brackish deposits in the hanging walls of major reverse faults from east to west across the Ventura basin.

3. METHODS

In this study, we produce complementary geochronological and landscape morphology data sets that we employ to quantify fault activity for three active reverse and thrust faults and to assess how the interaction and evolution of these faults has exerted a first-order control on the Quaternary landscape evolution of the Ventura basin.

3.1 Cosmogenic Dating

We employed three different cosmogenic isotope dating techniques (isochron burial dating, ^{10}Be surface exposure dating, and ^{10}Be-derived catchment-averaged erosion rates) to characterize various tectonic and erosion parameters throughout the Ventura basin. We used isochron burial dating to quantify the age of the Saugus Formation from east to west across the basin and to calculate displacement rates on 10^6 year timescales. We calculated a ^{10}Be surface exposure age for an extensive uplifted alluvial fan surface above Bear Canyon, which was previously termed the "Bear Canyon surface" (Fig. 2B) and dated to 80–100 ka by soil correlation with a similar terrace on the Ventura River (Rockwell, 1988). A date for the Bear Canyon surface is useful because there are no sediments younger than Miocene preserved in the hanging wall of the western San Cayetano fault (Fig. 2), and an age of ca. 80–100 ka would mean that the Bear Canyon surface could potentially provide a measure of 10^5-year rock uplift or incision rates for the western San Cayetano fault. The goal of the ^{10}Be-derived catchment-averaged erosion rate sampling was to track fault-parallel patterns in erosion rates and compare results with fault displacement rates, geomorphic parameters, and lithological distribution.

3.1.1 Isochron Burial Dating

Isochron burial dating is a key tool for dating Quaternary sediments and has been applied to a wide range of terrestrial and marine deposits (Balco and Rovey, 2008; Erlanger et al., 2012; Balco et al., 2013; Çiner et al., 2015; Bender et al., 2016). The method assumes that at the time of deposition, the slope of a linear regression on a plot of ^{26}Al concentration versus ^{10}Be concentration measured from a specific stratigraphic unit reflects the surface production ratio of ^{26}Al and ^{10}Be. As the unit is buried by deposition, the overlying deposits begin to shield the underlying unit from exposure to cosmic rays and the concentration of ^{10}Be and ^{26}Al starts to decrease via radioactive decay. As a result, the slope of the regression on the ^{26}Al/^{10}Be concentration plot evolves because of the differing half-lives of the two nuclides so that the difference between the slope of the isochron and the surface production ratio is indicative of the burial age of the sample (Balco and Rovey, 2008; Balco et al., 2013). Any post-burial nuclide production will uniformly increase the concentrations of ^{10}Be and ^{26}Al in all samples taken from the same depth horizon and will also alter the intersection of the line with the *y* axis but not affect the slope (Balco and Rovey, 2008). Hence, post-burial production can be treated as a constant, defined by the intersect of the linear regression with the *y*-axis. The burial time, t_b, of the deposit is obtained using the following equation (Balco and Rovey, 2008):

$$t_b = \frac{-\ln\left(\frac{R_m}{R_{in}}\right)}{\lambda_{26} - \lambda_{10}} \qquad (1)$$

where R_{in} is the ^{26}Al/^{10}Be ratio at the time of deposition, R_m is the ^{26}Al/^{10}Be ratio measured in the samples, and λ_{26} and λ_{10} are the decay constants for ^{26}Al and ^{10}Be, respectively, derived from the half-lives of the two isotopes.

We collected 64 individual samples from key stratigraphic horizons at eight specific sampling localities across the Ventura basin (Fig. 2). The stratigraphic and geographic sample locations are shown in Figure 2 and described in the Supplemental Material[1]. We selected sampling localities from locations that we interpreted to represent either the top of the exposed Saugus Formation or the base of various shallow marine units that underlie the Saugus Formation throughout the Ventura basin (e.g., Las Posas Formation, Grimes Canyon deltaic facies, and Sunshine Ranch Member) (Figs. 2–3). Here, the top of the Saugus Formation is referred to as the top of the "exposed" Saugus Formation because we have no quantification of the amount of material eroded from above the sample location. Details for laboratory extraction methods for ^{10}Be and ^{26}Al are included in section 3.1.4 below.

Outliers can occur in isochron burial dating data sets and may indicate that important assumptions associated with this method are invalidated. For example, the assumption that sediment has experienced the same exposure-burial history or an assumption of insignificant post-depositional production. On one hand, clasts with complex burial histories (i.e., multiple periods of exposure and burial) may be initially exposed and then buried beneath the attenuation depth of cosmic rays (e.g., in fluvial terraces, basins, or aeolian dunes) and subsequently be exhumed, eroded, and redeposited at the sample location. In this case, the clast would have experienced radioactive decay during the previous burial episode, which would result in a decrease in the ^{26}Al/^{10}Be ratio of the clast relative to the surface production ratio. Such a clast would plot as an outlier below an isochron line relative to a suite of clasts that only had one

DOI:10.1130/B36076.1

common exposure-burial history. On the other hand, clasts that have been reworked since original deposition may experience post-burial nuclide production, which can, especially for low-concentration clasts in rapidly eroding landscapes, overprint the signal of post-burial decay and result in an effectively reset ^{26}Al/^{10}Be ratio that is indistinguishable from the surface production ratio. A sample who's ^{26}Al/^{10}Be ratio has been reset by post-depositional production will plot on the surface production line. Sampling strategies, such as sampling from the same depth horizon to increase the likelihood of a shared post-burial history, can minimize the occurrence of outliers, but outliers may still occur. In our analysis, we systematically treat any sample with either a ^{26}Al or a ^{10}Be concentration more than three sigma away from the most likely isochron line as an outlier and do not include these samples in the regression or uncertainty analyses for the slope and intercept values.

When calculating slope and intercept values and associated uncertainties from measured isochrons, we adopted a Bayesian approach (Muzikar, 2011; Bender et al., 2016). Our Bayesian analysis incorporates a reference surface production ^{26}Al/^{10}Be ratio (R_{sp}, where $R_{sp} = R_{in}$) of 6.75 $\pm$ 0.5 (1σ) in the sediment at the time of deposition (based on Nishiizumi et al. [1989], normalized to standard values presented in Nishiizumi et al. [2007]). To validate our choice of R_{sp}, we collected a suite of sediment samples of different grain sizes from the modern San Gabriel River, which has its source in the San Gabriel Mountains (Fig. 1), which themselves are the ancestral source of most Saugus Formation deposits. Samples collected from the modern San Gabriel River produced an isochron with a slope of 6.97 $^{+0.31}/_{-0.28}$ (uncertainty in slope is 95% confidence interval), which overlaps with the canonical reference surface production ratio ^{26}Al/^{10}Be ratio of 6.75 $\pm$ 0.5 (1σ; Fig. 4H). Additionally, while it has been demonstrated that R_{sp} can vary with latitude and/or altitude (Argento et al., 2013; Lifton et al., 2014; Corbett et al., 2017), our use of R_{sp} = 6.75 at our sites is consistent with R_{sp} measured at a similar latitude to our study area elsewhere in the western USA at Promontory Point in Utah (Lifton et al., 2015). Furthermore, we performed a sensitivity analysis and demonstrated that our choice of initial surface production ratio, i.e., not only across the full 1s range of canonical surface production ratio values but also using our representative results from the modern San Gabriel River, has an insignificant effect on our interpretations of the burial ages within the reported uncertainties of our Bayesian methods (Table S1; see footnote 1). Full details of the Bayesian data reduction method and additional information on laboratory analysis for all cosmogenic dating methods, burial dating background, and burial dating sampling strategy are included in the Supplemental Material, and full burial dating samples details are summarized in Table S2 (see footnote 1).

3.1.2 ^{10}Be Surface Exposure Ages

We sampled 10 boulders for laboratory analysis and selected boulders standing >1 m above the Bear Canyon surface to minimize the chance that boulders had been exhumed or rotated since deposition. While care was taken to select boulders that had experienced minimal weathering, no accurate quantification of the amount of erosion on the boulder surface is available. Consequently, we assumed zero erosion in our age calculations. We also made no attempt to model inheritance for the boulders (see discussion in section 4.2). Details for each boulder sample are included in Table S3 (see footnote 1), and further details on sampling strategy and data reduction for the boulder samples are described in the Supplemental Material.

3.1.3 ^{10}Be-Derived, Catchment-Averaged Erosion Rates

We extracted 25 catchments of >~4.0 km^2 from U.S. Geological Survey 10 m National Elevation Data set digital elevation models (DEM) using TopoToolbox (Schwanghart and Scherler, 2014). Catchments were selected to cover the entire length of the hanging walls of the

San Cayetano, Southern San Cayetano, and Ventura faults. We based fault traces on existing geologic mapping of the study area (e.g., Dibblee, 1987; Dibblee and Ehrenspeck, 1988; Dibblee, 1990a, 1990b; Dibblee and Ehrenspeck, 1992a, 1992b, 1992c, 1992d; Tan et al., 2004; Campbell et al., 2014; Hughes et al., 2018) and the Southern California Earthquake Center 3-D Community Fault Model (CFM), version 5.2 (Plesch et al., 2007; Nicholson et al., 2017).

In total, we collected 18 samples comprised of ~2 kg of sand-sized sediment from bars and active channels at the mouths of fault-bounded catchments just upstream from a fault. We calculated catchment-averaged erosion rates from ^{10}Be concentrations using the CRONUS online calculator version 3 (Balco et al., 2008). A description of sample parameters and inputs for the CRONUS calculator are included in Table S4 (see footnote 1).

Landslides are present throughout the study area and can be a major contributor to bias in catchment-averaged erosion rate calculations (Niemi et al., 2005; Densmore et al., 2009; Yanites et al., 2009; Roda-Boluda et al., 2019). We used a preliminary version of the California Landslide Inventory Database (CaLSI) (available at http://maps.conservation.ca.gov; accessed August 2021) in conjunction with landslides included on geological and landslide maps from the study area (e.g., Tan et al., 2004) to avoid sampling immediately downstream from major mapped landslides and incorrectly modeling erosion rates. We also checked mapped landslides in close proximity to potential sample locations in the field, on Google Earth™, and where available with a high-resolution DEM derived from lidar data with 5 m horizontal accuracy and 0.45 m vertical accuracy that covers part of the study area (Airborne1, 2005).

3.1.4 Laboratory Analysis

We undertook quartz separation and chemistry for the erosion rate samples, the isochron burial dating samples, and the surface exposure age samples in laboratories at the Scottish Universities Environmental Research Centre (SUERC); the University of Vermont, Burlington; and in the CosmIC Laboratory at Imperial College London. We sieved bulk sediment samples to isolate the 250–500 μm fraction and isolated and purified ~10–30 g of quartz from the samples following the methodology of Kohl and Nishiizumi (1992). We undertook all Be and Al isolation following the method of Corbett et al. (2016b), which is described in the Supplemental Material. We measured ratios of ^{10}Be/^{9}Be for the exposure age samples and ^{10}Be/^{9}Be and ^{26}Al/^{27}Al for the burial dating samples by accelerator mass spectrometry (AMS) at the Centre for Accelerator Science at the Australian Nuclear Science and Technology Organization (ANSTO) using the 6 MV Sirius tandem accelerator (Wilcken et al., 2017, 2019) and at SUERC (Xu et al., 2015). We measured AMS ratios for the erosion rate samples at the Center for Accelerator Mass Spectrometry at Lawrence Livermore National Laboratory (Rood et al., 2010). Details of the measurement standards, beam currents, process blanks, and backgrounds for each AMS run are included in the Supplemental Material.

3.4 Fault Displacement Rates

In this study, fault slip is defined as the amount of displacement along the fault plane (dip-slip displacement), and we calculated the vertical component of fault slip (fault throw) by multiplying dip-slip displacement by the sine of fault dip. Uplift that is a product of both fault throw and folding is termed “rock uplift.”

We extracted dip-slip offsets and associated uncertainties from cross sections contained in Huftile and Yeats (1996) (Fig. S1; see footnote 1). For the eastern San Cayetano fault, we converted slip rates to throw rates to facilitate a direct comparison with erosion rates using a fault dip of 50° taken from CFM (Plesch et al., 2007; Nicholson et al., 2017). To revise the Holocene throw rate and slip rate for the western San Cayetano fault, we used published analyses of a fault scarp in a ~7.3 ka alluvial fan at the mouth of Bear Canyon (Fig. 2; Rockwell, 1988; Hughes et al., 2018). We calculated incision into the older, uplifted Bear Canyon surface in the hanging wall of the western San Cayetano fault (Fig. 2B) by subtracting channel incision into the surface from the maximum elevation. We determined maximum relief using maximum and minimum elevation values extracted from a swath profile parallel to the stream. To calculate an incision rate, we combined the value for maximum incision into the Bear Canyon surface with the TCN exposure age from boulders on the uplifted Bear Canyon surface, which we used as a proxy for the minimum Late Pleistocene rock uplift rate.

3.5 Landscape Analysis

We quantified mean catchment slope, maximum catchment relief, and catchment-averaged normalized channel steepness indices (k_{sn}) to provide geomorphic context to the catchment-averaged erosion rates. We measured catchment relief as the maximum difference in elevation within the catchment measured upstream from the fault to the drainage divide, and we measured mean catchment slopes using the Topographic Analysis Kit (TAK) for TopoToolbox (Schwanghart and Scherler, 2014; Forte and Whipple, 2019).

Above a threshold slope, mean hillslope gradient becomes decoupled from erosion rates and mean gradient is no longer an effective measure of erosion or tectonic processes (Burbank et al., 1996; Montgomery and Brandon, 2002; Ouimet et al., 2009). In contrast, k_{sn} fundamentally reflects how steep a river is for a given drainage area, and numerous studies have shown k_{sn} to be sensitive to both uplift and bedrock erodibility (Kirby and Whipple, 2001; Kirby et al., 2003; Cyr et al., 2010; DiBiase et al., 2010; DiBiase and Whipple, 2011; D'Arcy and Whittaker, 2014). Therefore, to supplement data on catchment maximum relief and catchment mean slope, we calculated catchment-averaged k_{sn} values for catchments using TAK for TopoToolbox (Schwanghart and Scherler, 2014; Forte and Whipple, 2019). In our calculations we employed a reference concavity of 0.50 based on regressions of slope versus upstream drainage area (Fig. S2; see footnote 1). Further details on the background and applications of k_{sn} and our justification for the reference concavity of 0.5 are included in the Supplemental Material.

4. RESULTS

4.1 Isochron Burial Ages

4.1.1 Western Ventura Basin

The isochron burial ages for the base of the Las Posas Formation and the top of the exposed Saugus Formation in the hanging wall of the Ventura fault are 0.60 $^{+0.05}/_{-0.06}$ Ma (Fig. 4A) and 0.36 $^{+0.18}/_{-0.22}$ Ma (Fig. 4B), respectively. All burial ages are mode and have 95% confidence limits throughout. We do not know how much material has been eroded above the sample location for the top of the exposed Saugus Formation, so the age for the top is a maximum age. The relatively large uncertainties associated with the upper age result from insufficient burial time for the $^{26}Al/^{10}Be$ ratio to deviate from the surface production ratio and low nuclide concentrations (Table S2). As a result, the top of the exposed Saugus Formation at Ventura is approaching the minimum age limit for these sediments using the $^{26}Al/^{10}Be$ nuclide pair. The burial age of 0.60 $^{+0.05}/_{-0.06}$ Ma for the base of the Las Posas Formation at Ventura overlaps with the 0.63 Ma age of the Lava Creek B ash (Matthews et al., 2015), which is mapped 60 m above the base of the Las Posas Formation in the hanging wall of the Ventura fault (Fig. 5; Sarna-Wojcicki et al., 1987).

4.1.2 Central Ventura Basin

In the hanging wall of the Oak Ridge fault, the isochron burial age from near the base of the Grimes Canyon deltaic facies, which underlies the Saugus and Las Posas Formations, is 1.06

±0.12 Ma (Fig. 4E). The Grimes Canyon deltaic facies interfingers with upper Pico Formation on the north flank of Oak Ridge (Fig. 6A), where the Bailey ash is mapped within the Pico Formation, approximately 3 km to the west of the sample location for the Grimes Canyon deltaic facies (Fig. 6A; Campbell et al., 2014). The Bailey ash is dated at 1.2 ± 0.3 Ma using fission track methods (Izett et al., 1974; Boellstorff and Steineck, 1975) and map relations indicate that our sample location for the Grimes Canyon deltaic facies may be stratigraphically slightly above the Bailey ash (Fig. 6A), which is consistent with our burial age.

Samples from the top of the exposed Saugus Formation near the axis of the Long Canyon syncline gave a burial age of 1.06 $^{+0.23}/_{-0.26}$ Ma (Fig. 4C). We also dated sediments from near the core of the Happy Camp Syncline (Fig. 6) that are mapped as either within the terrestrial Saugus Formation (Dibblee and Ehrenspeck, 1992a) or tentatively within the Grimes Canyon deltaic facies (Fig. 6A; Campbell et al., 2014), which gave a burial age of 0.98 $^{+0.20}/_{-0.28}$ Ma (Fig. 4D). Both samples were collected near the core of the respective synclines and should approximate the youngest section of the exposed sediments (Fig. 6B). The good agreement between the two ages indicates a possible age range for the top of the exposed Saugus Formation in the hanging wall of the Oak Ridge fault of 0.7–1.29 Ma. This age range overlaps with the age range from near the base of the underlying Grimes Canyon deltaic facies of 0.94–1.18 Ma despite several hundred vertical meters of sedimentary deposits separating the two ages (Fig. 6B). However, the total age range for the top of the exposed Saugus Formation of 0.78–1.29 Ma overlaps with an independent age of 0.78–0.85 Ma, which was calculated by comparing a mammalian fossil assemblage with magnetostratigraphic data near Moorpark (Fig. 5; Wagner et al., 2007). Therefore, we suggest that the true age from near the base of the Grimes Canyon deltaic facies is probably towards the 1.18 Ma upper bound that is consistent with the ~1.2 Ma age of the Bailey ash. Likewise, the true age for the top of the exposed Saugus Formation is probably nearer to the 0.70 Ma lower bound that is consistent with the 0.78–0.85 Ma fossil age (Wagner et al., 2007).

A sample from the top of the exposed Saugus Formation in the hanging wall of the Southern San Cayetano fault did not return a burial age. Several of the samples plot on the line of the surface production ratio, which indicates that they may be too young for $^{26}Al/^{10}Be$ isochron burial dating (Fig. S3; see footnote 1).

4.1.3 Eastern Ventura Basin

In the eastern Ventura basin, samples from strata that are tentatively mapped as the brackish water Sunshine Ranch Member of the Saugus Formation (Fig. 3) gave a burial age of 3.30 $^{+0.30}/_{-0.41}$ Ma (Fig. 4F). A recent $^{87}Sr/^{86}Sr$ age of 5.1–4.3 Ma was calculated in shallow marine deposits from the upper Pico Formation, which is located stratigraphically just below our burial sample location (Figs. 5A and S4; see footnote 1; Buczek et al., 2020). However, the total possible age range suggested for the $^{87}Sr/^{86}Sr$ samples was 5.3–3.6 Ma, and the $^{87}Sr/^{86}Sr$ ages may be an overestimate because the $^{87}Sr/^{86}Sr$ ratio in the samples may have been altered by contact with pore water during diagenesis (Buczek et al., 2020). Therefore, the isochron burial age may be a more reliable indication of the true age of the Saugus/Pico boundary in the east Ventura basin.

For sediments mapped near the contact between the undifferentiated Saugus Formation and an upper member of the Saugus Formation (Campbell et al., 2014), we calculated a burial age of 0.83 $^{+0.36}/_{-0.41}$ Ma (Fig. 4G). A previous estimate for the age of this contact is 0.6–0.7 Ma based on magnetostratigraphy and extrapolation of sedimentation rates from the 0.76 Ma Bishop ash, which is located stratigraphically just below the contact (Fig. S4; Levi and Yeats, 1993).

Although the burial age is located stratigraphically just above the Bishop Ash (Fig. S4), the burial age is consistent with the age of the ash within the 95% confidence limit (Fig. 5).

4.2 Exposure Age of the Bear Canyon Surface

When boulder height above the uplifted Bear Canyon surface is plotted against the ^{10}Be surface exposure age, the ages of the three tallest boulders overlap within the uncertainties, but boulder ages decrease systematically as a function of decreasing height for the seven boulders below 2.4 m height (Fig. 7B, Table S3). Based on these observations, our preferred exposure age is 121.2 ± 11.6 ka (N = 3, most likely value based on the three oldest boulders; 2σ uncertainty; Fig. 7). This age assumes that the remaining boulders below 2.4 m tall have been exhumed since deposition by surface lowering, which has gradually exposed boulders that were either buried or partially buried at the time of deposition (see discussion in Supplemental Material).

Our boulder ages assume zero inheritance. However, a large inherited nuclide concentration increases the apparent exposure age of the boulder surface. We note that some of the ^{10}Be-derived erosion rates have been perturbed by deep-seated local landslides (see section 5.2), which range from 1 m to 5 m deep within the study area (Harp and Jibson, 1995; Townsend et al., 2020). Moreover, inheritance values estimated from cosmogenic isotope depth profiles calculated using sand-sized particles in the Ventura basin are all low, which implies that past erosion rates were high and rapid exhumation from depth for sand-sized particles in the study area (DeVecchio et al., 2012a; Hughes et al., 2018). While neither of these factors provide precise quantification of the inheritance in boulders, they at least are compatible with a model of frequent erosion involving rapid exhumation of deeply sourced boulders, which is consistent with low inheritance. Conversely, while there appears to be a linear correlation between boulder height and boulder age for boulders below 2.4 m in height, there is some variation in age for specific heights (Fig. 7B). For example, there are two boulders with heights of 2.2 m with respective ages of ca. 102 ka and 78 ka (Fig. 7B). This deviation may indicate some variation in inheritance or even some differential exhumation between boulders of similar heights. Consequently, our surface exposure age may be a maximum.

Our surface exposure age from the Bear Canyon surface is a marked improvement on existing age constraints. The previous age of 80–100 ka is not an in situ age for the surface but is based on the correlation of soil profiles with what was argued to be a similar aged terrace on the Ventura River (Rockwell, 1988). The 80–100 ka Ventura River terrace was itself dated by extrapolating a slip rate of 0.37 ± 0.02 mm yr^{-1} for the Arroyo Parida/Santa Ana fault since ca. 38 ka using the relative height of the terrace above the modern day Ventura River (Rockwell, 1982). Given these assumptions and the fact that the boulder age is directly taken from the uplifted Bear Canyon alluvial fan surface, we prefer the boulder age of 121.2 ± 11.6 ka for this fan surface. The ca. 121 ka age is also consistent with a regional aggradation event within the Ventura basin that is thought to have initiated around 125 ka (DeVecchio et al., 2012a).

4.3 Displacement Rates

We recalculated existing displacement rates for the San Cayetano fault by incorporating the results of the cosmogenic nuclide dating presented above. Details of marker horizons and ages used to make the slip rate calculations are shown in Table 1.

4.3.1 Eastern San Cayetano Fault

There are no Saugus Formation deposits in the hanging wall of the San Cayetano fault to use in slip rate calculations, but we dated the Saugus Formation and the Grimes Canyon deltaic facies in the hanging wall of the Oak Ridge fault to the south (Fig. 2). The end of Pico Formation deposition, defined as a transition from fine-grained marine sediments to fossiliferous sandstone, is thought to have been isochronous across the Santa Clara River Valley between Oak Ridge and what is now the hanging wall of the San Cayetano fault (Huftile and Yeats 1995, 1996). Furthermore, the Grimes Canyon deltaic facies either caps or interfingers with the Upper Pico Formation in the hanging wall of the Oak Ridge fault (Fig. 6). Therefore, the burial age near the base of the Grimes Canyon deltaic facies can be used to estimate the end of Pico Formation deposition in the central Ventura basin. Dip-slip separation since the end of deposition of the Pico Formation across the eastern San Cayetano fault was previously measured as 10.1–11.8 km (Table 1) by assuming that the current thickness of Saugus and Pico Formation deposits mapped from well data in the footwall was once present in the hanging wall and projecting the thickness of the footwall strata across the fault (Huftile and Yeats, 1996). Accordingly, we divided the 10.1–11.8 km of dip-slip offset for the Pico Formation by the age range of 0.94–1.18 Ma from near the base of the Grimes Canyon deltaic facies to calculate an average minimum slip rate for the eastern San Cayetano fault of 8.6–12.7 mm yr^{-1}. By multiplying the dip-slip amount by the sine of a fault dip of 50°, we convert total dop-slip to fault throw and obtain a fault throw rate of 6.6–9.7 mm yr^{-1}. The upper bound for all rates stated here uses the maximum possible offset from the relevant reference (Table 1) and the youngest age value from the uncertainty associated with the appropriate age (Fig. 4). The lower bound uses the minimum possible offset from the relevant reference (Table 1) and the oldest age value from the uncertainty associated with the appropriate age (Fig. 4).

4.3.2 Western San Cayetano Fault

For the western San Cayetano fault, we divided 8.0–10.0 m vertical offset across a Holocene alluvial fan at the mouth of Bear Canyon (Rockwell, 1988) by a depth-profile age of 5.6–9.1 ka for the alluvial fan (Hughes et al., 2018) to give a Holocene throw rate of 0.8–1.8 mm yr^{-1}. We converted the Holocene throw rate to a slip rate of 1.2–2.5 mm yr^{-1} by assuming a fault dip of 45° (Rockwell, 1988). The incision rate for the uplifted Bear Canyon surface is 1.3–1.7 mm yr^{-1}. This rate is based on maximum incision into the surface of 166–184 m (we assign an arbitrary ± 5% uncertainty to incision calculated using a swath profile) divided by the 111–132 ka boulder age range of the Bear Canyon surface (Fig. 7). The overlap between the Holocene throw rate and the late Pleistocene incision rates indicates that displacement rates on the western San Cayetano fault have not varied significantly over the last ca. 120 ka.

The contemporary reverse slip rate across the entire San Cayetano fault measured from mechanical models driven by GPS data is 5.4 ± 1.7 mm yr^{-1} (Marshall et al., 2017). However, the GPS-derived values do not differentiate between the eastern San Cayetano fault and the western San Cayetano fault. Assuming the that the slip rate for the western San Cayetano fault has not increased by three to fourfold in the last ca. 9 ka, then the high GPS-derived slip rate for the San Cayetano fault must reflect strain localization on the eastern San Cayetano fault (Hughes et al., 2020).

4.3.3 Oak Ridge Fault

While not the primary focus of this study, the burial ages presented here also facilitate reevaluation of slip rates of the Oak Ridge fault (Fig. 1). Dip-slip separation of 1.7–2.1 km for the Oak Ridge fault was calculated in Huftile and Yeats (1996) by restoring a balanced cross section so the top of currently exposed Saugus Formation in the hanging wall of the Oak Ridge fault onlaps onto the southern limb of the Oak Ridge anticline (Fig. S1). Therefore, we divide 1.7–2.1 km of dip-slip separation by the age range of 0.7–1.29 Ma for the top of the exposed Saugus Formation in the hanging wall to calculate a slip rate of 1.3–3.0 mm yr^{-1} for the Oak Ridge fault. This rate is slightly slower than the previous estimate of 3.7–4.5 mm yr^{-1} (Huftile

DOI:10.1130/B36076.1

and Yeats, 1996), although our slip rate is a minimum because we have not quantified the amount of material eroded from the top of the exposed Saugus Formation. Nevertheless, our slip rate shows good agreement with a contemporary reverse slip rate of 2.6 ± 0.5 mm yr^{-1} based on mechanical models derived from GPS data (Marshall et al., 2017). All displacement rates calculated here along with additional displacement rates, which are not reevaluated in this study but are used in our analysis, are included in Table 1.

4.4 Erosion Rates and Landscape Analysis Results

4.4.1 ^{10}Be Erosion Rates

Catchment-averaged erosion rates in the 18 catchments sampled range between 0.05 ± 0.01 mm yr^{-1} and 2.30 ± 0.39 mm yr^{-1} (all uncertainties associated with erosion rates are 1 s; Fig. 8). Erosion rates in the hanging wall of the Ventura fault are consistent along strike and range from 0.17 ± 0.02 mm yr^{-1} to 0.51 ± 0.06 mm yr^{-1}. There is no discernible pattern along strike for erosion rates for catchments in the hanging wall of the Southern San Cayetano fault that range from 0.20 ± 0.02 mm yr^{-1} to 1.14 ± 0.18 mm yr^{-1}. In the hanging wall of the western San Cayetano fault, erosion rates are between 0.05 ± 0.01 mm yr^{-1} and 0.33 ± 0.04 mm yr^{-1} with the notable exception of catchment 5, which has a high rate of 2.30 ± 0.39 mm yr^{-1}. Erosion rates in the hanging wall of the eastern San Cayetano fault are consistent along strike and generally higher than the western San Cayetano fault with rates between 1.04 ± 0.21 mm yr^{-1} and 1.14 ± 0.18 mm yr^{-1} (Fig. 8).

4.4.2 Geomorphic Parameters

Throughout the study area, there is little correlation between TCN-derived erosion rates and landscape metrics such as catchment mean slope, catchment-averaged k_{sn}, or catchment drainage area (Fig. 9). The extent that tectonic parameters scale with landscape metrics and catchment-averaged erosion rates in the Ventura basin is explored in Figure 10. **[[Is this text inserted correctly?]]** In general, relief, slope, and catchment-averaged k_{sn} values are greater in the hanging wall of the San Cayetano fault than in the hanging wall of the Ventura fault (Fig. 10). In the hanging wall of the San Cayetano fault, relief is bimodal with a peak of ~1750 m along the western San Cayetano fault and a peak of >1250 m along the eastern San Cayetano fault (Fig. 10A). Catchment-averaged k_{sn} is broadly similar to the pattern of relief along strike with maximum values of 268 m and 174 m for the western San Cayetano fault and eastern San Cayetano fault, respectively (Fig. 10A). In the hanging wall of the San Cayetano fault, catchment slopes ranges from 28° to 34° and reach a maximum value of 34° in the center of the fault (Fig. 10A). Both throw rates and erosion rates increase eastward from the hanging wall of the western San Cayetano fault to the hanging wall of the eastern San Cayetano fault (Fig. 10B). However, for the eastern San Cayetano fault, throw rates are greater than erosion rates by a factor of three to four and, for the western San Cayetano fault, throw rates are greater than erosion rates by a factor of seven to 10 (Fig. 10B).

In the hanging wall of the Ventura fault, relief ranges from ~300 m to ~800 m and relief increases eastward to a maximum of ~1350 m in the hanging wall of the Southern San Cayetano fault (Fig. 10C). For the Ventura fault, catchment averaged k_{sn} values are in the narrow range of 44–70 m but increase eastward in the hanging wall of the Southern San Cayetano fault to a maximum value of 184 m (Fig. 10C). Catchment mean slopes gradually increase eastwards from a minimum of 20° in the hanging wall of the Ventura fault to a maximum of 30° in the hanging wall of the Southern San Cayetano fault. Erosion rates are lower than fault throw rates for the Ventura fault by a factor of 10 (Fig. 10D). All geomorphic parameters are summarized in Table 2.

DOI:10.1130/B36076.1

5. DISCUSSION AND IMPLICATIONS

In the following discussion, we justify the reasons for removing outliers from the isochron burial ages and discuss possible alternative interpretations of these ages. We then discuss the extent to which TCN-derived, catchment-averaged erosion rates in the Ventura basin can represent catchment-averaged rates given abundant mapped landslides. Having established which erosion rates represent a catchment-averaged rate, we calculate fluvial response times and compare the catchment-averaged erosion rates with fault displacement rates and morphometric landscape parameters to quantify the landscape response to Quaternary tectonics.

5.1 Outliers and Alternative Isochron Burial Ages

Outliers are possible in isochron burial dating data sets because of the inherent geological complexities and assumptions associated with this method. Therefore, it is important to explore the consequences of identifying and excluding outliers from the burial ages and to develop a reasonably objective criteria for justifying that certain samples are outliers. Clasts that have a lower ^{26}Al/^{10}Be ratio relative to other clasts that share a common exposure-burial history are usually assumed to have undergone a complex burial history involving a previous burial event for a duration similar to or greater than the respective isotope half-lives (Balco and Rovey, 2008). In general, our cobble and the pebble samples have ^{26}Al/^{10}Be ratios that are consistent with a shared exposure-burial history, a similar initial production ratio at the time of deposition, and a shared post-burial history (Fig. 4). However, in four isochron data sets, the sample of amalgamated sand is identified and omitted as an outlier because it has ^{26}Al concentrations $<<3\sigma$ and ^{10}Be concentrations $>3\sigma$ away from the most likely regression line based on the samples from the coarser clasts that are consistent with a common single exposure-burial history (See section 3.1.2 for outlier criteria; Fig. 4). These four isochrons are from the terrestrial Saugus Formation samples, which include the two data sets from the top of the exposed Saugus Formation in the hanging wall of the Oak Ridge fault and the two samples from the east Ventura basin (Fig. 4; Table S6).

One explanation for the anomalously low ^{26}Al/^{10}Be ratio in the terrestrial sand samples is that it results from deposition of aeolian sand within the terrestrial Saugus Formation. Aeolian sand is known across the globe to have a lower ^{26}Al/^{10}Be ratio than the surface production ratio because of long-term storage in sand dunes (Klein et al., 1986; Vermeesch et al., 2010; Davis et al., 2012), and modern dunes fields in our study area are documented along the coast at Ventura (Muhs et al., 2009). Furthermore, the coastal plain and associated dune fields are thought to have been much larger when the continental shelf was exposed during Pleistocene periods of low sea-level (Muhs et al., 2009). Prevailing onshore winds could have redistributed low ^{26}Al/^{10}Be ratio sand from the large coastal plain into the ancestral Santa Clara River and provided an influx of sand with an ^{26}Al/^{10}Be ratio lower than the surface production ratio. This scenario would result in a lower ^{26}Al/^{10}Be ratio of the sand sample relative to the pebbles and cobbles in our isochrons, and, in turn, our identification and omission of the sand sample as an outlier. However, it is unclear why the sand samples from the shallow marine Las Posas Formation or the Grimes Canyon deltaic facies do not have similarly low ^{26}Al/^{10}Be ratios, because they presumably share sediment provenance with the terrestrial samples. Most importantly, regardless of the reason for the anomalously low ^{26}Al/^{10}Be ratio in some of the sand samples, we suggest that our omission of the sand samples as outliers is not only both reasonable and justified, but also, when the sand samples are removed from the regressions as outliers, the preferred burial ages are more consistent with independent age constraints (Table S6). A detail comparison of alternative ages is provided in the Supplemental Material.

5.2 Landslide Influences on Cosmogenic Nuclide Concentrations

5.2.1 Landslides and Isochron Burial Dating

Regarding our isochron burial ages, another possible explanation for the anomalously low $^{26}Al/^{10}Be$ ratios for certain samples (see Section 5.1 above) is that these ratios reflect that the different sediment grain sizes were sourced from different depths. Several studies document differences in cosmogenic nuclide concentrations for sand versus gravel and pebbles, and often attribute the difference in nuclide concentrations to an increase in grain size with depth, where shallow-source fine sediment experiences more nuclide production than coarse sediment (e.g., Aguilar et al., 2014; Carretier et al., 2015). Moreover, in rapidly eroding landscapes, the $^{26}Al/^{10}Be$ production ratio varies from ~6.75–7.25 between 0 m and 2 m depth to ~7.25–7.75 between 2 m and 5 m depth. The increase in $^{26}Al/^{10}Be$ production ration with depth occurs because nuclide production below depths of 2–3 m becomes dominated by muons, which results in a higher $^{26}Al/^{10}Be$ production ratio than production by spallation (Braucher et al., 2013; Akçar et al., 2017; Zhang et al., 2021). If the cobbles and pebbles are generally sourced from depths below the spallation-dominated production zone (~2 m) during landslides, then the low $^{26}Al/^{10}Be$ ratio of the sand compared to the cobbles could be a product of different initial ratios at the time of deposition due to nuclide production via muons in the deeply sourced cobbles and pebbles versus spallation in the shallowly sourced sand. In this case, the assumption of deposition with an initial $^{26}Al/^{10}Be$ ratio of 6.75 for all grain sizes would be invalid, and the burial ages based on the cobbles and pebbles would be an underestimate. However, the sediment samples from the modern San Gabriel River are each from a different grainsize, including amalgamated cobbles, amalgamated pebbles, and amalgamated sand, which define an isochron with a slope and $^{26}Al/^{10}Be$ production ratio of 6.97 $^{+0.31}/_{-0.28}$ (Fig. 4G). This initial production ratio, which comes from a representative modern catchment, is consistent with the range of production ratios of 6.75 but consistent with the range of production ratios of 6.75–7.25 for material sourced from the upper 2 m in rapidly eroding landscapes (Akçar et al., 2017). Therefore, we suggest the anomalous sand ratios probably do not result from different source depths for the different grain sizes.

Further, our sensitivity analysis, which was performed to explore the change in burial age for surface production ratios between 6.25 and 7.28 (i.e., the full range of values defined by the uncertainties in both the canonical $^{26}Al/^{10}Be$ production ratio and our representative estimate from the modern San Gabriel River), produced a range of burial ages that are generally within the 95% confidence limits of our preferred burial ages (Table S1). On balance, if landslides up to 5 m deep are a major contributor to the erosion signal in the Ventura basin, then landslides could affect the $^{26}Al/^{10}Be$ ratio and the burial ages; however, observed landslide depths in our study area are generally <4 m (Harp and Jibson, 1995; Townsend et al., 2020). With landslide depths <4 m deep, our sensitivity analysis indicates that our burial ages are robust regardless of the specific depth the material is sourced from because any potential change in the surface production ratio with depth will not alter our burial ages outside of the preferred 95% confidence limits.

5.2.2 Landslides and 10Be Erosion Rates

In landscapes where numerous landslides are present, the assumption that all sediment has undergone a steady transition through the penetration zone of cosmic rays may not be applicable (Niemi et al., 2005; Densmore et al., 2009; Yanites et al., 2009; Roda-Boluda et al., 2019). The assumption of a steady transition may be invalid because nuclide production at depths below the penetration depth of cosmic ray neutrons will be dominated by muon production, which occurs at much slower rates than production by neutron spallation near the surface (Braucher et al., 2013). Consequently, erosion rates calculated using standard methods in areas of deep-seated landslides could overestimate the catchment-averaged erosion rate if a recent landslide has mobilized low-concentration sediment sourced from beneath the penetration depth of cosmic ray neutrons (Niemi et al., 2005; Yanites et al., 2009; West et al., 2014). However, in areas where frequent but shallow landslides are the dominant erosion signal from the landscape, TCN-derived erosion rates can be a reliable indicator of catchment-averaged erosion rates (Roda-Boluda et al., 2019).

Assuming that landslides dominate the erosion signal, two landscape-derived ratios can help assess whether TCN-derived erosion rates represent a reliable catchment-averaged, long-term rate. Firstly, for a given erosion rate, reliability decreases as the mean landslide depth (h_{ls}) approaches or exceeds the mean attenuation path length (z^*; typically, 80 cm). With decreasing h_{ls}/z^*, landslides will become more frequent, and the erosion will more closely resemble a continuous process that erodes material from within the production zone of cosmogenic nuclides rather than a sporadic series of large events (Yanites et al., 2009). Modeled earthquake-induced landslide depths in the study area resulting from the 1994 Northridge earthquake vary by lithology with mean values of 0.8 m or 1.3 m for the Saugus Formation (different values correspond to whether a Culmann or a Newark model for depth is applied; see Townsend et al., 2020), 1.7 m or 1.1 m for the Pico Formation, and 3.8 m or 2.1 m for the Modello Formation (Townsend et al., 2020). The relatively shallow earthquake-triggered landslide depths and associated relatively low h_{ls}/z^* in the Saugus and Pico Formations ($h_{ls}/z^* = 1$ or 1.6 and 2.1 or 1.4, respectively) indicate that catchments in the hanging wall of the Ventura fault and the western section of the Southern San Cayetano fault, which contain predominantly Saugus and/or Pico Formations (Fig. 2A), may experience frequent, shallow landslides where sediment approximates a steady transition through the production zone. Conversely, the Modello Formation is the predominant lithology in the hanging wall of the eastern San Cayetano fault (Fig. 2A), which results in deeper modeled landslides and a higher h_{ls}/z^* of 4.8 or 2.6. Therefore, the reliability of the erosion rates in areas with deeper landslides depends on the ability of the fluvial system to mix more deeply sourced sediment from larger, less frequent landslides (Niemi et al., 2005; Yanites et al., 2009; Roda-Boluda et al., 2019).

Information on the ability of the fluvial system to mix landslide-derived sediment is contained in a second landscape-derived ratio: the ratio of the residence time of fluvial sediment in the catchment (m_t) to the average landslide repeat time for a given point in the catchment (μ) (Yanites et al., 2009). Assuming that landslides provide most of the sediment to the system, m_t/μ can be used as a metric to assess the extent to which the fluvial sediment sourced from landslides is spatially and temporally mixed. In other words, for catchments with deeper landslides, the reliability of TCN-derived erosion rates relies on the ability of the fluvial system to mix material from landslides that occurred at different times. TCN-derived erosion rates are likely to be representative of the basin-averaged, long-term rate if m_t/μ is $>10^{-2}$ or, in simple terms, that landslide material originates from >1% of the catchment area (Yanites et al., 2009). The landslides mapped in Figure 8B are from an amalgamation of different sources that contain both dormant and active landslides and were mapped for different purposes (e.g., seismic hazards, geological mapping, landslide susceptibility, etc.). Moreover, we have no data on the precise age of the landslides and what percentage of the landslides occurred within the period covered by the catchment averaged erosion rates (<~4500 years; Table S4). However, landslides triggered by the 1994 Northridge earthquake were previously mapped for the eastern San Cayetano fault,

DOI:10.1130/B36076.1

including within catchments 9 and 10 (Harp and Jibson, 1995; Townsend et al., 2020). The percentage of total surface area covered by landslides triggered by the Northridge earthquake in catchments 9 and 10 are 2% and 5%, respectively (Fig. S5; see footnote 1). While the results from two catchments may not be applicable to the entire study area, the percentage of catchment area covered by landslides in the hanging wall of the eastern San Cayetano fault is above the theoretical 1 % threshold from the 1994 Northridge earthquakes alone. This is consistent with the suggestion that earthquake-triggered landslides provide sufficient material to the fluvial system to become embedded in the erosion rate and record reliable catchment-averaged rates.

A third consideration is that above a threshold slope, mean hillslope gradient becomes decoupled from erosion rates and mean gradient is no longer an effective measure of erosion or tectonic processes (Burbank et al., 1996; Montgomery and Brandon, 2002; Ouimet et al., 2009). We observed no correlation between slope and erosion rate in the study area (Fig. 9C). While the lack of correlation does not indicate a threshold behavior, it does indicate that erosion rates are decoupled from slope, which implies that the landscape response will be driven by increased landslide frequency rather than increased hillslope gradients (Montgomery and Brandon, 2002). Additionally, we measured a low ^{10}Be concentration (Table S4) and an anomalously high erosion rate of 2.30 ± 0.39 mm yr^{-1} in catchment 5 and, to a lesser extent, the rate of 1.14 ± 0.18 mm yr^{-1} in catchment 14, both of which contrast with erosion rates from the surrounding catchments that are generally <0.30 mm yr^{-1} (Fig. 8B). We suggest that the sample from catchment 5, and possibly also catchment 14, may comprise significant sediment from an individual deep-seated landslide near the sample location and that the erosion rates may not reflect a catchment-averaged rate (Niemi et al., 2005; Yanites et al., 2009; West et al., 2014). As a result, we exclude the erosion rate from catchment 5 in the comparison with tectonic rates, catchment-averaged relief, slope, and ksn (Figs. 9–10).

While we suggest that, in general, the TCN-derived erosion rates in the Ventura basin are reliable catchment averaged rates, some of our erosion rates appear to embed the signal of deep-seated, locally sourced, recent landslides. We recorded low ^{10}Be concentrations and high resulting uncertainties associated with erosion rates of 6.43 ± 3.55 mm yr^{-1} in catchment 5 and 3.94 ± 3.55 mm yr^{-1} in catchment 8 (Table 2). We suggest that the samples from catchments 5 and 8 contain sediment from individual locally sourced, deep-seated landslides near the sample location and that the erosion rates in these catchments do not reflect a catchment-averaged rate (Niemi et al., 2005; Yanites et al., 2009; West et al., 2014). As a result, we do not include the erosion rates from catchments 5 and 8 in the comparison with tectonic rates, catchment-averaged relief, slope, and k_{sn} (Figs. 9–10).

We note an anomalously low erosion rate of 0.05 mm yr^{-1} in catchment 3. Just upstream of our sample location in catchment 3, the creek is littered with very large boulders that are sometimes in excess of 2 m tall (Fig. S6; see footnote 1). Large boulders are also common around the sample locations in catchments 1 and 4 and must have also been deposited over the past ca. 120 ka based on the presence of the >2 m tall boulders recorded on the Bear Canyon surface, which is located above catchment 4 (Fig. 7A). The presence of such large boulders indicates that large landslides or debris flows feed large boulders into the channel in the hanging wall of the western San Cayetano fault, which may suppress channel incision and decrease erosion rates measured from channel sands (Bennett et al., 2016).

5.3 Landscape Response to Fault Slip

The reevaluated fault displacement rates presented in section 4.3 provide the tectonic framework to investigate the landscape response to tectonic forcing via a comparison with erosion rates and landscape parameters. The erosion rates in the hanging wall of the Southern San Cayetano fault of 0.20–1.14 mm yr^{-1} are difficult to interpret because the uplift rate is variable along strike (Hughes et al., 2018) and catchments along the eastern section of the Southern San Cayetano fault are cut by the western San Cayetano fault, which implies that the uplift rate may vary within the catchments. Therefore, we focus our analysis on the San Cayetano and Ventura faults.

5.3.1. Landscape Response Times

There is an apparent discrepancy between the TCN-derived erosion rates for catchments in the hanging wall of the Ventura fault and erosion inferred from structural uplift. The Saugus Formation was potentially uplifted via folding to a maximum of 2.7 km across the Ventura Avenue anticline since the end of Saugus Formation deposition, which we suggest here has a maximum age of ca. 360 ka (Fig. 4B). The amount of uplift was calculated by projecting the tilt of bedding values on the fold limb to the fold hinge (Rockwell et al., 1988). If true, this would imply that a maximum of ~2.2 km of structural relief generated by folding of the Saugus Formation could have been eroded from the hanging wall of the Ventura fault, which requires extremely high long-term erosion rates of ~5–6 mm yr^{-1} based on the maximum current topographic relief of 485 m on the Ventura Avenue anticline. The long-term erosion rate is similar to the rock uplift rate of 3.6–4.9 mm yr^{-1} since 15.9 ± 0.2 ka (Hubbard et al., 2014) but is much higher than millennial-scale erosion rates of 0.17–0.33 mm yr^{-1} (Fig. 10D).

The disparity between high long-term and low short-term erosion rates could simply be a product of integration time. The TCN erosion rates from the hanging wall of the Ventura fault apply to the age range of 1500–4800 years (Table S4). This timescale is similar to the recurrence interval for large-magnitude (>Mw 7.5) multi-fault earthquakes that include the Ventura fault of 1000–2000 years (Rockwell et al., 2016). If the TCN-derived erosion rates in the hanging wall of the Ventura fault are only integrated over one to two earthquakes then the TCN-derived millennial-scale erosion rates may not necessarily show a simple relationship with erosion rates inferred from rock uplift measured over several seismic cycles, especially in an area dominated by landslides (Densmore et al., 2009). However, regardless of the integration time for the TCN-derived erosion rates, the similarity of the long-term erosion rate and the late Pleistocene rock uplift rate, coupled with the absence of tectonic knickpoints in streams in the hanging wall of the Ventura fault (Fig. 11B), could indicate that the streams may be in a quasi-steady state, which requires rapid landscape response times if uplift commenced at a maximum of ca. 360 ka.

Similarly, streams in the hanging wall of the San Cayetano fault may also be in a quasi-steady state. The initiation of surface uplift from the San Cayetano fault is not well quantified, with estimates ranging from 1 Ma (Çemen, 1989) to 3.3 Ma (Rockwell, 1982). Apatite cooling ages indicate long-term exhumation rates from the hanging wall of the San Cayetano fault of 1.2–1.6 mm yr^{-1} since 3 Ma (Townsend et al., 2021). For the eastern San Cayetano fault, erosion rates of ~1 mm yr^{-1} are seven to 10 times lower than the throw rate of 6.6–9.7 mm yr^{-1} and, by a factor of approximately three and, for the western San Cayetano fault, erosion rates between 0.05 mm yr^{-1} and 0.33 mm yr^{-1} are also significantly lower than Holocene fault throw rates of 0.8–1.8 mm yr^{-1} (Fig. 10B). Therefore, while the magnitude of erosion rates is slightly less than the fault throw rates, the erosion rates agree with the long-term exhumation rate, and the pattern of an eastward increase in erosion rates mirrors the eastward increase in fault throw rates (Fig. 10B). The stream profiles are difficult to interpret because abundant landslides and small-scale variations in lithology create numerous small perturbances to the profiles (Fig. 11A).

Nevertheless, while there are some minor knickpoints that cannot be attributed to lithology or landslides, there appear to be no major non-lithologic knickpoints (Fig. 11).

Some simple calculations of fluvial response times help constrain the extent to which streams in the study area have adjusted to tectonic forcing. Assuming a detachment-limited stream power erosion law, the speed, v, at which a knickpoint would propagate upstream from a starting point such as a fault through an entire catchment can be expressed as:

$$v = KA^{m}S^{n-1} \qquad (2),$$

where K is the bedrock erodibility with units of m^{1-2m} yr^{-1} (e.g., Whittaker and Boulton, 2012; Royden and Perron, 2013; Zondervan et al., 2020). Using estimates of bedrock erodibility derived from average channel gradients in our study catchments and our tectonic/cosmogenic constraints on fluvial erosion and uplift rates, we model v as a function of declining drainage area upstream for each catchment for a base case where $m = 0.50$ and $n = 1$. We use this to constrain to the first order how long a knickpoint would take to get within 0.5 km of the headwaters of each channel (see Supplemental Material for detailed calculations of K and v).

For the total range of uplift rates of 3.0–25.2 mm yr^{-1} for the Ventura fault (Hubbard et al., 2014), we calculate K values of 2.95×10^{-5} m yr^{-1} to 2.48×10^{-4} m yr^{-1} (minimum and maximum values; Table 3). Using the range of uplift rates for the San Cayetano fault calculated here (Table 1), we calculate average K values of $2.58–3.80 \times 10^{-5}$ m yr^{-1} for the eastern San Cayetano fault and $2.09–4.18 \times 10^{-6}$ m yr^{-1} for the western San Cayetano fault (Table 3). These relatively high K estimates are comparable to those calculated within similar stratigraphy in the Santa Ynez Mountains (Fig. 1) just west of the study area (Duvall et al., 2004). These calculations yield fluvial response times of $0.10–1.70 \times 10^{5}$ years for the Ventura fault, $0.6–2.2 \times 10^{5}$ years for the eastern San Cayetano fault, and $6.4–23.0 \times 10^{5}$ years for the western San Cayetano fault (Table 3).

5.3.2 The Ventura Fault

With a maximum uplift age of ca. 360 ka, the short response times ($10^{4–5}$ years), in principle, allow sufficient time for a knickpoint to propagate through the entire catchment so that the streams appear to be in a quasi-steady state. However, several observations indicate that topographic steady-state may not have been achieved for the Ventura Avenue anticline. Late Pleistocene uplift rates of 3.6–4.9 mm yr^{-1} are in sharp contrast to the TCN-derived erosion rates from the Ventura Avenue anticline, which range from 0.17 mm yr^{-1} to 0.33 mm yr^{-1} (Fig. 10A). Moreover, the stream paths of catchments 19 and 24, which drain the south flank of Sulfur Mountain, have been redirected by younger catchments in response to the uplift of the Ventura Avenue anticline since at most 0.36 Ma (Fig. 8A). Given ongoing growth of the Ventura Avenue anticline (McAuliffe et al., 2015), this drainage reorganization is also presumably still ongoing, which is inconsistent with the attainment of topographic steady state.

The absence of tectonic knickpoints in potentially transient streams in the hanging wall of the Ventura fault may suggest that the streams eroding the Ventura anticline do not in fact lie at the detachment-limited end member and show some diffusive (sediment flux-dependent) behavior. In response to a change in uplift regime, for channels limited by their ability to transport sediment, tectonic knickpoints do not tend to develop because the stream undergoes progressive steepening throughout the entire stream length (Whipple and Tucker, 2002). In this case, erosion rates would be controlled by the sediment transport capacity of the streams rather than the channel incision rate, which could explain the contrast between low millennial-scale, TCN-derived erosion rates of 0.17–0.33 mm yr^{-1} and high long-term uplift rates >3.6 mm yr^{-1}. However, the correspondence between the inferred long-term erosion rates of 5–6 mm yr^{-1} and

DOI:10.1130/B36076.1

the late Pleistocene uplift rates of 3.6–4.9 mm yr^{-1} indicate that this disparity may not have existed in the past.

Hybrid behavior has been suggested for streams with similar lithology to the study area in the Santa Ynez Mountains, which lie just west of the study area (Duvall et al., 2004). Hybrid transport-limited, bedrock-incising channels can occur with decreasing uplift rates or in areas with rapid uplift rates and where abundant hillslope-derived sediment floods the bedrock channel (Whipple and Tucker, 2002; Gasparini et al., 2006; Kent et al., 2021). Either of these criteria could be applied to the Ventura fault. Landslides are abundant in the hanging wall of the Ventura fault, and although the uplift rate of 3.6–4.9 mm yr^{-1} since 15.9 ± 0.2 ka is high, it decreased from 8.6–25.2 mm yr^{-1} in the early stages of uplift (Rockwell et al., 1988; Hubbard et al., 2014). For hybrid streams, a decrease in rock uplift rate will bring about transport-limited behavior because changes in sediment supply lag behind the channel response (Whipple and Tucker, 2002; Gasparini et al., 2006). Therefore, the initial rapid uplift stage of the Ventura Avenue anticline could have been accompanied by a period of rapid detachment-limited erosion and channel downcutting at high rates of ~5–6 mm yr^{-1} that were possibly facilitated by the low strength of the uplifting, poorly lithified Plio-Pleistocene sediments (Townsend et al., 2020). With uplift rates >3 mm yr^{-1}, any knickpoints could propagate through the entire stream in ~10^5 years (Table 3). Then, as the uplift rate decreased over time, a transition to transport-limited behavior led to the current decoupling of TCN-derived erosion rates from uplift rates.

5.3.3 The San Cayetano Fault

There are pronounced east-west gradients in tectonic and erosion metrics along strike in the hanging wall of the San Cayetano fault (Fig. 10A). Erosion rates and landscape response times mirror the pattern of fault throw rates, with higher erosion rates and faster landscape response times for the eastern San Cayetano fault than for the western San Cayetano fault (Fig. 10B and Table 3). Relief and k_{sn} are correlated throughout the study area (Fig. 9A) similar to the results of studies from a variety of tectonic and lithologic settings (D'Arcy and Whittaker, 2014). However, contrary to the eastward increase in erosion rates and fault throw rates, both relief and k_{sn} decrease eastward in the hanging wall of the San Cayetano fault (Fig. 10B). Rock strength is relatively uniform along strike in the hanging wall of the San Cayetano fault (Townsend et al., 2021). With uniform rock strength and climate, the key variable controlling the eastward increase in both the erosion rates and the landscape response times in the hanging wall of the San Cayetano fault must be the eastward increase in uplift rate (Fig. 10B). In contrast, relief and k_{sn} appear to be decoupled from tectonic forcing and erosion rate as previously documented elsewhere in low rock-strength mountain belts (Densmore et al., 2004, 2007; Barnes et al., 2011).

The fluvial response times for the San Cayetano fault suggest a quasi-steady state. For the eastern San Cayetano fault, the streams appear to be well adjusted with no major tectonic knickpoints (Fig. 11A). This observation is supported by the short fluvial response times of 1–2 × 10^5 years, which appear to be fast enough for any previous tectonic knickpoints to have propagated through the entire length of the stream. Conversely, with slower response times of 0.6–2.3 × 10^6 years for the western San Cayetano fault, the streams may not have had sufficient sufficient time to adjust if surface uplift commenced at ca. 1.0–3.3 Ma (Cemen, 1989; Rockwell, 1992). However, the knickpoint celerity estimates assume a constant K through time. During the early stages of uplift, the eroding rocks would have comprised a thick succession of Plio-Pleistocene sediments with lower rock-strength than the Miocene–Eocene succession that is currently mapped in the hanging wall of the San Cayetano fault (Townsend et al., 2021). Therefore, the past erodibility of the San Cayetano fault may have been higher than the

DOI:10.1130/B36076.1

erodibility indicated by contemporary *K* values, and potentially similar to the erodibility of current stratigraphy mapped in the hanging wall of the Ventura fault where the outcrop is dominated by Plio-Pleistocene sediments (Fig. 2). With a previously higher *K* value, the response times may have been fast enough to allow sufficient time for any knickpoints to propagate through the entire stream length in both the eastern San Cayetano fault and the western San Cayetano fault. Further, a temporal change in K could also cause knickpoints to develop in streams, which may be an alternative explanation for the minor "tectonic" knickpoints shown in Figure 11.

6. CONCLUSIONS

We compiled a basin-wide geochronology and used the geochronology to calculate displacement rates in the Ventura basin, southern California, USA. We used the resulting displacement rates as the tectonic template to compare with terrestrial cosmogenic nuclide (TCN)-derived catchment-averaged erosion rates and morphometric landscape parameters within the Ventura basin. Our main findings have important implications for regional earthquake hazards and interpreting landscape response in rapidly uplifting young mountain belts:

• We used cosmogenic $^{26}Al/^{10}Be$ isochron burial dating to derive a basin-wide geochronology for a key regional strain marker: the Saugus Formation. Our ages for the top of the exposed Saugus Formation from west to east across the basin are 0.36 $^{+0.18}/_{-0.22}$ Ma, 1.06 $^{+0.23}/_{-0.26}$ Ma, 0.98 $^{+0.20}/_{-0.28}$ Ma, and 0.83 $^{+0.36}/_{-0.41}$ Ma. The burial ages near the base of shallow marine deposits, which underlie the Saugus Formation throughout the basin, are 0.60 $^{+0.05}/_{-0.06}$ Ma, 1.06 ±0.12 Ma, and 3.30 $^{+0.30}/_{-0.42}$ Ma for the western, central, and eastern Ventura basin, respectively. The burial ages for the Saugus Formation are generally consistent with independent ages and provide a robust geochronological framework to improve our understanding of the uplift and erosion history and of seismic hazards within the Ventura basin.

• Consistent with previous work, late Pleistocene throw rates for the western section of the San Cayetano fault of 0.9–1.8 mm yr^{-1} since 7.3 ka and 1.3–1.7 mm yr^{-1} since ca. 121 ka are lower than a rapid long-term fault throw rate of 6.6–9.7 mm yr^{-1} since ca. 1.0 Ma for the eastern section of the San Cayetano fault. Based on the burial ages, slip rates for the Oak Ridge fault are 1.3–3.0 mm yr^{-1} since ca. 1 Ma and agree with contemporary estimates of reverse fault slip rates derived from mechanical models driven by GPS data (Marshall et al., 2017).

• Our work demonstrates the applicability of TCN-derived, catchment-averaged erosion rates in rapidly uplifting, landslide-prone landscapes. For catchments that we interpret to represent a long-term, catchment-averaged rate, erosion rates range from 0.05–1.14 mm yr^{-1}. By analyzing the ratio of modeled landslide depths to the attenuation length of cosmic rays and examining landslide density within the catchments, we show that, in general, TCN-derived catchment-averaged erosion rates in the Ventura basin record the catchment-wide erosion signal of frequent, shallow landslides.

• A comparison of tectonic rock uplift and fault throw rates with TCN-derived erosion rates, fluvial response times, and geomorphic landscape parameters shows that in young mountain belts with high uplift rates and weak rocks, rapid fluvial response times can cause stream profiles to attain a quasi-steady-state on timescales of ~10^5 years. However, millennial-scale, TCN-derived erosion rates will lag behind fault throw or rock uplift rates if decreasing uplift rates cause a transition from detachment-limited to sediment flux-dependent stream behavior. If uplift rates remain fairly stable, then TCN-derived erosion rates may become partially coupled to tectonic slip but will not necessarily correlate with geomorphic parameters such as relief or channel steepness.

ACKNOWLEDGMENTS

The authors thank Ruth Davy, Alex Seal, and Charlie Singer for assistance during field work and sample collection and Jeff Knott for insightful and productive discussions on the age of the Saugus Formation. Well data in Figure 6 were taken from the online database provided by the California Department of Conservation, Division of Oil, Gas, and Geothermal Resources website, which is freely available at: https://gis-california.opendata.arcgis.com/datasets/cadoc::well-finder (accessed August 2021). Work was supported by funding from the Southern California Earthquake Center (SCEC), which is funded by National Science Foundation Cooperative Agreement EAR-1033462 and U.S. Geological Survey Cooperative Agreement G12AC20038. This work was funded by award numbers 13105 and 15100 to D.H. Rood and D.E. DeVecchio, 17184 to D.H. Rood, 16049 to T.K. Rockwell, and 17024 to T.K. Rockwell. This is SCEC contribution number #11502. The authors also acknowledge financial support from the Australian Government for the Centre for Accelerator Science at the Australian Nuclear Science and Technology Organization (ANSTO) through the National Collaborative Research Infrastructure Strategy (NCRIS). This work was completed thanks to ANSTO award number 10125 to D.H. Rood. Additional funding was supplied by National Science Foundation EAR award number 1735676 to P.R. Berman supporting L.B. Corbett, a Royal Geological Society postgraduate research grant to A. Hughes, an Imperial College Janet Watson bursary to A. Hughes, and two postgraduate research grants from the British Society for Geomorphology to A. Hughes.

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Figure 1. The tectonic setting of the Western Transverse Ranges with faults from the Southern California Community Fault Model (Plesch et al., 2007; Nicholson et al., 2017). Thin white lines are rivers. Focal mechanism solution is for the Mw 6.7 1994 Northridge earthquake (Huftile and Yeats, 1996). The line A-A′ refers to cross section in Figure S1 (see footnote 1). SCRV—Santa Clara River valley; WSCF—western section of the San Cayetano fault; ESCF—eastern section of the San Cayetano fault; SSCF—Southern San Cayetano fault; LA—Los Angeles; LFS—Lion fault set (Sisar, Big Canyon, and Lion Canyon faults); APF—Arroyo-Parida fault; SAF—San Andreas fault.

Figure 2. (A) Geological map of the Ventura basin shows locations of isochron burial dating samples with shaded stars. The geological units to the east are based on the mapping of Campbell et al. (2014), and the western half of the map is based on the T.W. Dibblee, Jr., geological maps (see reference list). (B) Location and extent of the Bear Canyon surface. (C) Mean annual precipitation for the period 1971–2000 for catchments in the hanging wall of the Ventura, Southern San Cayetano, and San Cayetano faults adapted from Stillwater Sciences (2011). AP-SAF—Arroyo Parida/Santa Ana fault; DVF—Del Valle fault; SSCF—Southern San Cayetano fault; SSMF—south Sulfur Mountain fault; BC—Bear Canyon; OC—Orcutt Canyon.

Figure 3. Idealized stratigraphic section across the Ventura basin, which is amended from Swanson and Irvine (2015). The letters after the colon denote the lithological units in Figure 2A, and the letters in brackets denote the primary lithology within the unit.

DOI:10.1130/B36076.1

Figure 4. ^{26}Al-^{10}Be isochrons which were constructed using a Bayesian linear regression (Bender et al., 2016). The pale gray to dark gray lines are the range of results from 1,000,000 trial runs that potentially fit the isotope data with light gray representing lower likelihood and dark gray representing higher likelihood. Black crosses represent 1σ uncertainties in ^{26}Al or ^{10}Be concentrations for each sample included in the regression, and pale gray or dotted crosses represent outliers omitted from the isochron analysis (either ^{10}Be or ^{26}Al concentration >3σ from the mode line of best fit). Slope, intercept, and ages are most likely (modal) values, and uncertainties are based on the 95% confidence limits of the data. GCDF—Grimes Canyon Deltaic facies; SRM—Sunshine Ranch member of the Saugus Formation.

Figure 5. (A) Geological map that shows the location of key chronostratigraphic markers within the study area. (B) Comparison of isochron burial ages for the Saugus Formation and underlying shallow marine deposits with existing geochronology for the Ventura basin. *$^{87}Sr/^{86}/Sr$ age (Buczek et al., 2020) is thought to be an overestimate. AAR—amino acid racemization; OSL—optically stimulated luminescence; SSCF—Southern San Cayetano fault.

Figure 6. (A) Geological map shows the location of isochron burial samples in the Long Canyon syncline. Mapped geological units and the location of the Bailey ash (dashed black line) are from Campbell et al. (2014). The line of cross section X-X′ refers to the section in part B. (B) Cross section across the Oak Ridge anticline and the Long Canyon syncline, which is based on well data obtained from the California Department of Conservation (www.maps.conservation.ca.gov) and surface mapping of Campbell et al. (2014). Details for wells used to construct the cross section are included in Table S5 (see footnote 1).

Figure 7. A summary of ^{10}Be exposure ages from boulders on the Bear Canyon surface. The location of the Bear Canyon surface is shown in Figure 2B. (A) Photo of the boulders on the Bear Canyon surface. The largest boulder visible here is ~2 m tall. (B) Plot of boulder height against boulder age, which shows a systematic decrease in boulder age for boulders less than ~2 m in height. This plot demonstrates that boulders <2 m tall were likely exhumed from below the surface by erosion of surrounding material and do not represent the true exposure age of the surface. Uncertainties on ages are 1 σ and the measurement uncertainties for boulder height are approximated at 10%. (C) Probability density function and associated ages incorporating all 10 samples. Uncertainties are 2σ. (D) Probability density function incorporating the oldest three samples from the surface, which are used to derive the preferred age of the surface. Note that these ages assume no erosion of the boulder surface and no inheritance.

Figure 8. Map of erosion rates and channel steepness in the study area. (A) Catchments are shaded by catchment-averaged normalized channel steepness indices (k_{sn}) with lighter color representing lower k_{sn}. Streams are thick, black lines within catchments. The numbers at the catchment outlets are the catchment identifiers referred to in the text and are included in Tables 2, 4, and S4 (see footnote 1). (B) Erosion rates and landslides in the study area. Mapped landslides are taken from the California Landslide Inventory database. Numbers in bold are average erosion rates of catchments in mm yr^{-1}. Rates in italics may have been artificially increased by landslides in proximity to the sample location (see text). VF—Ventura fault; VAA—Ventura Avenue anticline; WSCF—western San Cayetano fault; ESCF—eastern San Cayetano fault; SSCF—Southern San Cayetano fault.

Figure 9. Graphs comparing erosion rates and geomorphic parameters throughout the study area. (A) A plot of k_{sn} as a function of relief. (B) A plot of upstream drainage area versus erosion rate. (C) Erosion rate as a function of k_{sn}. (D) A plot of mean catchment slopes versus catchment-averaged erosion rates. Error bars on erosion rates are 1 s and errors on k_{sn} are one standard error.

Figure 10. Fault-parallel plots that show erosion and tectonic metrics along-strike. (A) Plot of maximum catchment relief, mean catchment slope, and catchment-averaged normalized channel steepness (k_{sn}) in the hanging wall of the San Cayetano fault. (B) Catchment-averaged erosion rates and fault throw rates in the hanging wall of the San Cayetano fault. Small black dots connected by the dotted line are stratigraphic separation values taken from Rockwell (1988) and Çemen (1989), which represent a proxy for fault throw along strike. (C) Plot of maximum catchment relief, mean catchment slope, and k_{sn} in the hanging wall of the Ventura fault and the Southern San Cayetano fault (SSCF). (D) Catchment-averaged erosion rates and vertical deformation rates (uplift or throw rate) in the hanging wall of the Ventura fault and the Southern San Cayetano fault. Appropriate references for vertical deformation are included in Table 1, and data for plots are included in Table 2. Error bars on erosion rates are 1σ, and errors on k_{sn} are one standard error.

Figure 11. River long profiles in the Ventura basin. (A) Streams in the hanging wall of the San Cayetano fault. Gray streams are the eastern section of the San Cayetano fault and black streams are the western section of the San Cayetano fault. (B) Streams in the hanging wall of the Ventura fault and along the western end of the Southern San Cayetano fault.