Late Quaternary subsidence of Santa Catalina Island, Southern California Borderland, from submerged paleoshorelines

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Keywords
Marine Terraces
Last Glacial Maximum
Santa Catalina Island
California Channel Islands
Submerged marine terrace
Subsidence
Tectonics

Abstract

Submerged paleoshorelines surrounding Santa Catalina Island in the Southern California Continental Borderland require late Quaternary tectonic subsidence. Our geomorphic analysis of high-resolution bathymetry combined with seismic reflection profiles allows mapping of nine submerged wave-cut platforms around Catalina, preserved at 32 ─ 362 m depth. We identify the bathymetric expression of the Last Glacial Maximum (LGM) paleoshoreline at a depth of 131 m ± 1 m. The depth distribution of these submerged terraces correlates with sea-level lowstands on an ice-volume equivalent eustatic sea-level curve using approximately uniform Pleistocene subsidence rates. The most plausible correlation of terraces with sea-level still-stands is achieved using a time-integrated mean subsidence rate of 0.3 mm yr^{-1} over the last 355 ka, and a mean subsidence rate of 0.25 mm yr^{-1} over the last 1.15 Ma,

similar to decadal rates from GPS stations on the island. Catalina's terraces require radiometric dating or paleontological analysis to assign ages with greater confidence. However, the presence of terraces at depths of >350 m indicates that Catalina Island has subsided ~ 220 m since the formation of its deepest terraces. The use of submerged paleoshorelines to constrain late Quaternary slip rates is important for constructing tectonic evolution of continental margins, including the anastamosing San Andreas fault system in southern California, as well as for seismogenic tsunami hazard analysis in coastal communities.

1 Introduction

1.1 Geology

Santa Catalina Island (herein referred to as Catalina) is an exposed ridge crest running parallel to the coast of southern California (**Fig. 1**) and is the type locality for the Catalina Schist terrane of the Southern California Continental Borderland (SCCB). Catalina Island, one of Southern California's Channel Islands archipelago, consists of metamorphosed Farallon Plate, meta-sedimentary rocks, meta-volcanic and ultramafic rocks subducted in the late Mesozoic (Grove et al., 2007) **(Fig. 2).** This protolith was metamorphosed in blueschist-to-amphibolite facies during Farallon subduction (Grove et al., 2007), and unroofed during detachment of Late Cretaceous forearc strata and Jurassic basement from the western margin of the Peninsular Ranges Batholith (Atwater, 1970; Atwater and Molnar, 1973). Catalina Schist crops out on Catalina and on Palos Verdes peninsula (or Hills), and is inferred from boreholes to be the basement rock of much of the SCCB **(Fig. 1)** (Wright, 1991; Crouch and Suppe, 1993).

The igneous southern portion of Catalina Island was intruded by rocks of andesitic and dacitic composition during mid-Miocene (19 Ma) oblique rifting (Vedder et al., 1979; Legg et al., 2007). Miocene sedimentary rocks, including the San Onofre Breccia, form less than 5% of the subaerial geology of Catalina (**Fig. 2**) but also form part of Santa Cruz-Catalina Ridge north of the island. Catalina emerged from the Pacific following the deposition of late-Miocene deep marine fossiliferous sediments near Mount Banning (Smith, 1933) and Miocene to early-Pliocene shallow marine sediments in Cottonwood and Middle canyons **(Fig. 3a)**. These deposits are now approximately 140 m above modern sea level (Smith, 1933; Muhs et al., 2012). Catalina is currently affected by strike-slip faulting along the North American/Pacific plate boundary, translating ~44 mm/yr N44^{o}W relative to stable North America **(Table S1)**.

1.2 Physiography of Catalina

Active faulting has influenced the shape of the coastline and drainage patterns on Catalina Island. The Catalina Island structural block is bounded to the south and west by the Santa Cruz-Catalina fault (SCCF), and to the east by the San Pedro Basin (SPBF) (**Fig. 1**). The NE coast of Catalina between

Avalon and Two Harbors is defined by the Long Point Fault, mapped in this study by seismic and seismicity, that continues offshore for at least 8km **(Fig. 2)**. The SW-facing coastlines are defined by a stepover in the Santa Cruz-Catalina Fault that also marks the NE boundary of the Catalina Escarpment. The majority of large drainages on the SW side of the island converge toward Little Harbor where the Santa-Cruz-Catalina fault (SCCF) crosses the coastline. The SCCF was likely reactivated during Pliocene uplift of Catalina (Legg et al., 2007) and possibly also during Quaternary subsidence. Drainage capture where the SCCF crosses onshore suggests that it still controls topography. In contrast, drainages on the southernmost portion of Catalina Island are more likely to be influenced by sheeted mafic dikes that strike parallel to Silver Canyon. The north end of Catalina Ridge is adjacent to the southern portion of Santa Cruz Island, and is truncated by the Anacapa-Dume Fault (Chaytor et al., 2008).

Unlike neighboring Channel Islands, which bear the surficial features of their emergence from the Pacific Ocean, Catalina's rugged landscape contains sparse evidence of uplift (Lawson, 1893; Smith, 1897; Ritter, 1901) . However, water-lain andesites as well as mid-bathyal marine fossils of Pliocene age indicate that Catalina emerged from the Pacific no earlier than the Pliocene, and certainly following middle Miocene exhumation of the Catalina Schist from beneath the Peninsular Ranges Batholith (Grove et al., 2007). Any wave-cut platforms dating from this episode of emergence have apparently been dissected by ephemeral streams and canyons, and dismembered by landslides. Modern topography is influenced by the erodability of basement rocks, while the bathymetry is influenced by ocean currents and slope stability of sediments. Terrace slope stability is highest on terraces derived from granitic rocks, followed by those derived from lawsonite-blueschist. Epidote-albite and lawsonite-albite greenschist derived terraces are disrupted by frequent slides (**Fig. 2**). Landslides may add to seismogenic tsunami hazard posed by the island for nearshore communities (**Fig. S1**). Submarine landslides near Catalina occur on slopes <1° (see section 4.4.1).

1.3 Tectonics and vertical motion of the SCCB

The vertical tectonic motion of the Catalina block has been a subject of debate for over a century, as is summarized well in other papers (Davis, 2004; Schumann et al., 2012). The conspicuous absence of emergent marine terraces on Catalina and the presence of broad submarine shelves deeper than -130 m circumscribing Catalina suggest Quaternary subsidence (Ritter, 1901; Emery, 1958) (**Fig. 2**). Smith (1933) reported a variety of "Quaternary emergent terraces" on Catalina Island and argued for Quaternary uplift, though subsequent investigators interpret these as fluvial terraces (USDA, 2008; Schumann, 2012) or ascribe other origins. Smith's identification of marine surfaces was further complicated by the presence of Native Americans on Catalina, who transported significant quantities of shells and wave rounded stones to upslope localities (Glassow, 1980; Davis, 1985), and all "uplifted marine terraces" reported by Smith have been dismissed by later workers (Davis, 2004). New arguments for Quaternary emergence based on geomorphology have been proposed based on knickpoints in Catalina's drainages (Schumann et al., 2012). However, most knickpoints in the Schumann study are located at major thrust faults, at lithologic contacts, or where landslides deflect drainages, supporting the alternate hypothesis that the "knickpoints" are controlled by local conditions, not by uplift (see section 4.2). Models of continuous Quaternary uplift provide no satisfactory explanation for the lack of uplifted terraces or for the presence of Catalina's submerged shelves that were first mapped by Emery (1958).

Fossil-bearing uplifted marine terraces on neighboring Channel Islands in the SCCB (eg. Muhs, 1983; Pinter et al., 2001) have been used to date Quaternary vertical motion following the methodology of LaJoie (1986), which correlates radiometrically dated terraces at known elevations to sea-level highstands assuming linear uplift rates. Uplift rates on neighboring landforms such as Palos Verdes Peninsula, the San Joaquin Hills and most of the Channel Islands **(Fig. 4b)** have been constrained using radiometric dating (Bryan et al., 1987; Grant et al., 1999). Vertical motion estimates for the SCCB are based mostly on uplifted terraces with very few quantitative investigations of submerged terraces.

The use of submergent terraces in constraining vertical motion is well established (e.g. Steinen et al., 1973; Chiocci et al., 1996, 1997; Rohling et al., 1998; Passaro et al., 2011). These investigations typically combine bathymetry, multi-channel seismic and ideally some form of age control, e.g. radiometric dating of shells recovered from terraces during submersible dives (Chaytor et al., 2008). Fossils from submerged marine terraces of Pilgrim Banks, an isolated bathymetric high at ~130m depth, and around Santa Cruz Island have been used to constrain vertical motions (Chaytor et al., 2008). A particular target is the LGM paleoshoreline (Chaytor et al., 2008), the topographic expression of global sea-level lowstand between 19 and 23 ka (Marine Isotope Stage 2, or MIS 2) (Yokoyama et al., 2000). Global sea level was ~120-130m below modern sea level during the LGM (Fleming et al., 1998). LGM aged fossils recovered from ~130m depth from Pilgrim Banks show that sea level during the LGM did reach at least -120 m, even allowing for rapid (0.5 mm/yr) subsidence since the LGM. Other authors who neglect this fossil evidence have argued that the LGM lowstand level could have been as high as -90 m in southern California after consideration of glacio-isostatic adjustments (Muhs et al., 2012).

Uplifted terraces on San Clemente Island to the south yield an average uplift rate of 0.2 mm/year since 125 ka (Muhs, 1983), and extend up to ~550 m above sea level. San Clemente Island, which hosts over a dozen uplifted marine terraces, is the textbook example of emergent marine terraces, yet it has a submerged terrace at -120 to -130 m which matches regional observations (Chaytor et al., 2008) of maximum sea level drop during LGM **(Fig. 3).** Because sea-level over the last 1 Ma has not been significantly lower than -130 m for a sustained length of time (Lisiecki & Raymo, 2005) we do not expect to see any deeper terrace on an uplifting island. The existence of the one submerged terrace and many uplifted terraces shows that San Clemente Island has been uplifting for at least 1 Ma. Schumann et al. (2012) identify four additional submerged terraces on San Clemente Island, but these are clearly shown with newer bathymetry to be artifacts of gridding an aliased dataset **(Fig. S2), (**see Passaro et al. (2011), their section 2)**.**

In contrast, Catalina has no emergent terraces but instead a stair-stepped series of flat to very gently dipping submarine surfaces surrounding the island that we interpret as equivalent submergent terraces. We suggest that Catalina has been subsiding for the last 355 ka (the lowstand at MIS 10), and possibly for 1.15 Ma (lowstand at MIS 34) or more, based on our analysis of submerged terrace remnants around Catalina at depths of 32-362 m below sea-level. Using multiple generations of seismic and bathymetric data collected around the island, we have interpreted paleo-sea level during formation of the sequence boundaries prior to the LGM, and we correlate these features with an ice-volume equivalent eustatic sea level curve to establish a likely chronology for the subsidence of Catalina. Our preferred chronology of the terraces implies a mean subsidence rate of 0.25 mm yr^{-1} for the last 1.15Ma, which we suggest is driven by the position of Catalina Island relative to local restraining and releasing fault segments of the San Andreas System. Using sequence-stratigraphic relationships we show that Catalina's sequence of terraces requires successive sea-level lowstands of variable depth on a subsiding margin, or multiple progressively shallower lowstands on a tectonically stable margin. Volumetric constraints on the amount of global ice volume preclude the latter hypothesis of progressively shallower lowstands. In what follows, we describe our bathymetric and seismic datasets around Catalina, and present and justify our interpretation of nine submerged terraces.

2 Methods

2.1 Seismic and Bathymetric Data

We used multiple generations of seismic data, including 1970s USGS boomer and airgun single-channel data, 2008-2009 California State University, Long Beach 16-channel sparker data, as well as 2014 Stanford University 35-channel boomer data **(Table S2)** to interpret sequence stratigraphy, the extent of sedimentation and bedrock geometry**.** To correlate seismic data with bathymetry we used a digital-elevation model (DEM) at 2m lateral resolution produced by California State University, Monterey Bay (CSUMB) to generate several slope-enhanced shaded relief raster images **(Figs. 2, 3, S2)** that we imported into SMT Kingdom Suite™.

2.2 Interpretation Methods and Correlation

Marine terraces mark surfaces formed during sea-level still-stands and therefore provide constraints on Quaternary paleo-sea level if the sequence boundaries can be dated (LaJoie, 1986; Osterberg, 2006). Marine terraces surrounding Catalina Island can be confidently identified using combined seismic and bathymetric data. Terraces T1 through T9 were mapped at increasing depths and distances from Catalina, using our bathymetric DEM and paying special attention to identifying the intact outer edge of each terrace tread (**Fig. 2, Table 1**). We selected portions of each surface that are well-developed north of the island based on our DEM and minimally dissected or covered by new sediment based on our seismic profiles. We then sampled them using the zonal statistics tool in ArcGIS. The mean elevations (column 3, Table 3) are estimated from bathymetry, and correspond closely to peaks in the elevation histogram (column 5, table 3)**(Fig. 6c)** (Passaro et al., 2011). Where high-resolution seismic and bathymetry data exist on upper near-horizontal terraces, we find there is good agreement between bathymetrically-derived terrace depths and those identified using seismic data. The histogram unequivocally demonstrates the existence of the upper terraces around Santa Catalina **(Fig. 6c).**

Bathymetry provides a modern depth to each subplanar surface, but leaves paleo sea-level poorly constrained. Seismic data provide evidence that each terrace observed in the bathymetry correlates with a sequence boundary and that these planar surfaces are indeed marine terraces covered by younger sediments. Bathymetry provides a poor approximation of the terrace back-edge elevations because new sediment and landslides usually overlie the abandoned terrace back edge. Seismic data provide constraints on the elevation of terrace back edges, representing the inland extent of marine terrace erosion. Using several sets of seismic data, we measured the mean depth of clearly identifiable sequence boundaries using water acoustic wavespeed of 1500 m/s and a sediment acoustic wavespeed of 1900 m/s. For example, we examined digitized paper records of high-resolution single-channel Uniboom seismic data acquired across several terraces southeast of Avalon (**Fig. 5**)**.** From 3 – 3.5 km from the start of seismic line 65 **(Fig. 5)**, beneath the sea-floor at 0.125s, a horizontal reflection at 0.145 s clearly truncates reflections dipping seaward at 1 – 5°. We identify this clear erosional unconformity as a wave-cut platform, now buried by 0.02 s travel-time or 19 m thickness of younger sediments. Submerged terrace T2 is the seafloor underlain by these younger sediments.

The observation of the unconformity on the seismic data, not possible with bathymetric data alone, provides conclusive evidence that submerged terrace T2 is associated with the unconformity and was thus formed as a result of wave erosion during a sealevel stand, even though it is not the actual wave-cut platform. Two other sequence boundaries (for terraces T5 and T6) were similarly interpreted on profile 65. These and other terraces were further mapped using modern digitally-recorded sparker or boomer data (**Figs. 5, 6**). Sparker data **(Figs. S1, S3)**, despite having lower resolution than boomer data, can eliminate explanations for the planar surfaces other than as wave-cut platforms, such as asperities in the exposed bedrock or landslide deposits. 35-channel 1.5 k-J uniboom data collected by us in 2014 have been processed and prepared for preliminary interpretation **(Fig. 6)**. These ultra-high resolution data have a vertical resolution of ~0.5m and allow detailed mapping of individual sequence boundaries and erosional incisions. On seismic line 2205 **(Fig. 6)** beneath the seafloor at 0.25 seconds between 4.5-4.8 km, steeply dipping downlap sequences were truncated and overlain by sediment transported from higher elevations. This sequence boundary, which underlies T4 has bathymetric expression **(Fig. 6b)** despite being covered by a minimum of 0.01s (9.5 m) of sediment; because it is so narrow it appears only weakly on the histogram (**Fig. 6c**). The depths derived from seismic lines north of the island agree well with terrace depths derived from bathymetric data. South of the island, terrace depths vary due to tilting of Catalina toward the mainland (see section 4.3.1).

2.3 Chronology

Terrace numbers T1-T9 were assigned in order of increasing depth, not increasing age, to laterally continuous sub-planar surfaces that circumscribe the island. Our terrace depth data, combined from bathymetry and seismic data, were plotted along the vertical axis of the sea-level curve at time t=0 **(Fig. 7)**. We used Lisiecki and Raymo's (2005) Pliocene-Pleistocene stack of 57 globally distributed benthic δ^{18}O records to generate a sea-level curve, and have assumed a linear relationship between δ^{18}O and sea level, so that

$$depth = \frac{\delta^{18}\mathrm{O}(t) - \delta^{18}\mathrm{O}(t_0)}{-9.173\times10^{-3}}, \qquad (1)$$

where the denominator was chosen so that the amplitude of sea-level fluctuations matches geologic observations. The range of irregularly sampled δ^{18}O values converted to depths provides relative paleo-sea level depth-ranges during high and low sea-level stands (Osterberg, 2006). We smoothed the irregularly time-sampled output of equation 1 with a moving average window of 6 samples (6 samples ≈ 2 ka), effectively bandpassing the $\delta^{18}O$ time series to better fit global observations of Holocene/Quaternary sea-level stands (Lambeck et al., 2002; Hanebuth et al., 2009). The discrepancy between dated coral terraces and the $\delta^{18}O$ curve from benthic foraminifera is minimal at high and lowstands, and the relative magnitudes and timing of stands are preserved (Chappell and Shackelton, 1986). Our 6-sample smoothing increased ages of lowstands by approximately 2% which will decrease our estimates of subsidence rates by up to 2%. Smoothing also may reduce the amplitude of sea-level fluctuations which could inflate our estimated subsidence rates by a few %.

In order to assign ages we next match our terrace depths with our paleo sea-level curve. Sea-level lowstands during the late Quaternary epoch have a frequency of ~110 ka. The product of this frequency and the subsidence rate gives us the expected vertical distance between successive submerged terraces below global sea level minima. T1-T9 have an average separation of 40 m **(Table 1).** This spatial frequency of terraces limits the subsidence rate to <~0.4 mm/yr, assuming one major terrace is cut per lowstand. This constraint on the subsidence rate rules out the possibility that T4 (now at 165 m depth) was cut during the LGM because this scenario implies vertical subsidence of ~1 to 2 mm/yr with respect to Pilgrim Banks (now at ~-130 m), and subsidence 1-2 mm/yr would produce terraces ~110m apart on Catalina. The limited subsidence rate, <0.4 mm/yr, also rules out T2 (-109 m) as having been cut during the LGM elevation at -130 m because this correlation would require rapid uplift of ~1.0 mm/yr. This rapid uplift scenario would also expose any terraces cut during the Last Interglacial (LIG), a series of globally observed sea level highstands dated ~80-125 ka at a maximum elevation of ~6 m above mean sea level. Within the LIG, the MIS 5 terrace is a radiometrically dated paleosurface, as wide as 1 km from shelf-edge to back-edge, observed on uplifting Channel Islands including San Clemente (Muhs, 1983). 1.0 mm/yr of uplift would place the MIS 5 terraces at +80-125 m today. We observe no subaerial terraces on Catalina, hence we infer Catalina's LIG terrace has submerged since it was cut. Any LIG terrace cut on Catalina requires ~20 m of subsidence between MIS 5 and MIS 2 to accommodate post-LGM uplift yet leave no trace of an uplifted LIG terrace, requiring rapid changes in both the magnitude and direction of the velocity of the Catalina block.

Since neither T2 or T4 can reasonably represent the LGM terrace, the remaining possibility is T3. This assignment allows us to explain the distribution of all of Catalina's terraces at the correct vertical spatial frequency while honoring sequence stratigraphic relationships. The average terrace separation implies a maximum subsidence rate ~0.4 mm/yr since T9 was cut, and the absence of a LIG MIS 5 subaerial terrace (cut at +6m) implies a subsidence rate > 0.06 mm/yr since ~100 ka. In figure 10 we link the modern T3 depth to the LGM lowstand and T1 to the LIG and assume the subsidence rate defined by our submerged terrace sequence is approximately uniform. We then search for approximately parallel (uniform subsidence rate) lines linking all terraces T2 and T4-T9 to other significant lowstands on Figure 7. We also assume that terraces represent the youngest, deepest lowstand on any parallel uplift trajectory (T6 represents the 355 ka lowstand, not the older and shallower 450 ka and 560 ka lowstands. This logic and these minimal assumptions provide the unique chronology of **Figure 7** and **Table 2**, extending back to ~1150 ka for our deepest terrace, T9. It is possible that additional unconformities may be present in the seismic data, which would allow Catalina to have subsided for a longer time at a slower pace, although it would be difficult to create a better correlation of terrace depths and dated stillstands than already present in **Fig. 7**.

After correlating erosional sequence boundaries with sea-level lowstands, we solved for time-variant uplift rates using

$$h_{T_n} = z_n + \sum_{k=1}^{n} m_n \Delta t \quad (2)$$

$$\Delta t = t_n - t_{n-1} \quad (3)$$

where h_{Tn} is the elevation at the time t_n during which sequence boundary of terrace *n* was cut, z_n is the current depth of the sequence boundary *n*, m_n is the subsidence rate in m/kyr over the n^{th} interval, and Δt is the interval time in thousands of years between successive sequence boundaries (**Table 2, Fig. 7).** The deepest terraces are not exposed on the southwest side of the island, so we have sampled depths exclusively from the north side and North Point.

3 Results

3.1 Terrace Data

T1 - The shallowest terrace outcrops only on west-facing shores at a depth of ~ -32m **(Figs. 2, 3)**. Near Whale Rock **(Fig. 3a)**, this terrace is composed of low-intermediate grade metamorphic detritus and much of the outer edge of the terrace had been dismembered by landslides. Closer to Little Harbor the terrace extends 1km laterally from shore. T1 is not present where several streams meet the sea at Little Harbor. Modern sediment is deposited on a shelf that is ~ 22 m below present storm-weather wave base at 10-15m depth.

T2 - The most prominent sub-planar and nearly horizontal terrace exists at ~ 90 m depth. Where T2 is imaged by seismic data **(Fig. 5)**, ~ 19 m of sediment overly the planar sequence boundary that represents the wave-cut surface responsible for T2's bathymetric expression. On west facing shores, T2 contains some rocky outcrops (e.g. Farnsworth Bank, **Fig. 3a**) rise above T2 with as much as 30 m bathymetric relief. On windward (south and west-facing) shores T2 is over 4 km from outer edge to back edge, while on the leeward side of Catalina, T2 is less than 0.5 km from outer edge to back edge.

T3 - T3 is preserved on windward shores of the Catalina Escarpment, where erosion would have been intense during MIS 2 (LGM), and current drives sediment away from T3. Interpretation of T3 is complicated by the presence of a sequence boundary associated with an older terrace also at ~-130m **(Fig. 8a).**

T4 - T4 is a narrow shelf underlying T3 at ~ -165m, and is expressed in bathymetry on west facing shores and NE of the northern tip of the island **(Fig. 6)**.

T5 - The T5 surface is apparent on bathymetric slope maps **(Fig. 6b)** and histograms **(Fig. 6c)**. The T5 surface corresponds to a buried sequence boundary. Northern T5 contains a small area of rocky shoals with a maximum of 10m surface relief and dips to the north $0.7° \pm 0.2°$. The T5 sequence boundary is nicely imaged in seismic south of Avalon **(Fig. 5).**

T6 - This is the second most prominent terrace on the island, apparent in bathymetry surrounding the entire island, except at the Catalina Escarpment. The sequence boundary associated with T6 is overlain by a sediment prism ~53 m thick over the back edge, denoted by the red circle labeled T6-228 m **(Fig.**

6a), which tapers to <1m thickness at the shelf outer edge. T6 dips to the north at 1.5° ± 0.2° and is clearly shallower at the SE tip of the island (**Fig. 5a, inset**) than at the NW tip **(Fig. 6a)**.

T7 -T7 is associated with the largest sequence boundary contained in sediments near Catalina and marks a transition upwards to younger higher-energy deposition, as evidenced by the change in steepness of reflectors at this interface. The T7 sequence boundary in line 2205 **(Fig. 6a)** has an apparent north dip of 1.7° (258m at the back edge to ~320m at the seafloor). T7 is absent on other portions of the island so we do not know if the increased dip of 1.7° is because of tilting of Catalina.

T8 & T9 - These surfaces are visible in the bathymetry north of Two Harbors **(Figs. 2, 3)** and correspond to wave-planed basement at the depths listed in column 6 of **Table 1 (Fig. S4)**. Elevations for T8 and T9 come from wave-planed basement in seismic line Stanford-2202.

4 Discussion

4.1 Submerged terraces

Wave-cut terraces are a reliable proxy for paleo-sea-level and provide us a datum for understanding vertical tectonic motions of the Channel Islands, and in turn the anastomosing strike-slip faults that dissect the borderland as a part of the greater San Andreas Fault System. Understanding of lowstand deposits, although represented in the literature (e.g. Steinen et al., 1973: Chiocci et al,. 1996, 1997: Rohling et al., 1998: Passaro et al., 2011), has proceeded at a much slower pace than that of highstands (e.g. Lajoie, 1986) in part because of the expense and difficulty of surveying and coring submerged features. Terraces in subsiding clastic margins are rarely preserved, and remain distinct surficial features on the seafloor only when subsidence produces more accommodation space than sediment can occupy. This is the case for Catalina where sedimentation rates are low and subsidence is relatively rapid, so that Catalina's terraces are a retrogradational parasequence set. Subsidence aids in the cutting of broad terraces since frictional resistance to wave-cutting is diminished as freshly cut surfaces subside below storm-weather wave base. Terraces cut by deeper and longer-duration sea-level lowstands are more likely to be preserved in subsiding sedimentary sequences whereas those caused by lower-amplitude (<80 m) sea-level fluctuations are likely to be eroded or overlain by subsequent cycles of transgression and regression (Bradley and Griggs, 1976). Low-amplitude lowstands (eg. 450 ka, 550 ka) likely to originally have produce terraces were therefore excluded from our analysis as likely eroded by younger deeper lowstands (e.g. 355 ka).

In clastic emergent margins sea-level highstands and stillstands produce surfaces used in correlation with the sea-level curve (Lajoie, 1986). Conversely, in subsiding margins, low stands and still stands are equally likely to be responsible for carving the morphology we observe in the bathymetry. We suggest that highstands are unlikely to be preserved as submerged terraces because the periodicity and amplitude of Milankovitch cyclicity causes highstand terraces to be eroded, buried and /or reoccupied by another sea-level stand before they subside out of reach of wave action. For example, it is likely that if Catalina continues subsiding and sea level fluctuations continue to match Quaternary behavior, T1 will be obliterated/obscured/reoccupied before reaching the "safe zone," below 130m, in which a terrace can be preserved indefinitely. However, the T5 sequence boundary **(Fig. 6a)** truncates sediments deposited at a sea-level highstand. If the lowstand that cut T5 was at the same elevation or higher than the original highstand shelf, the highstand terrace may have been preserved. Although none of Catalina's submerged terraces are strictly highstand terraces, it is possible that submerged highstand terraces may exist elsewhere.

4.2 Comparative morphology

The development of topography on Catalina Island is markedly different than that of San Clemente Island, which hosts marine terraces overlying basalt up to 550 m above sea level (**Fig. 3**). If the uplift rate for San Clemente of 0.2mm/yr since 125 ka (Muhs, 1983) applies to the entire uplift history of the island, then San Clemente's highest and likely oldest terrace should have been cut at ~2.7 Ma. Since San Clemente's emergence from the Pacific, its terraces have only begun to be dissected by the canyons that have propagated upslope from the coast, the longest of which is ~8km. Flights of non-dissected marine terraces are preserved between canyons. In contrast, terraces which formed during Catalina's emergence have been completely removed by erosion, and a mature topography with dendritic drainages covers the island. No original topography resembling San Clemente's wave-planed paleo-topography exists today, and canyons as long as 12 km (Middle Canyon) extend up to the highest ridgeline, suggesting Catalina emerged (possibly quite rapidly) during the late Miocene-early Pliocene, and has since been resurfaced by fluvial processes. Suggestions by Schumann (2012) that Catalina has been resurfaced during the Quaternary in an episode of rapid uplift are inconsistent with the rates of canyon propagation (3 km/Ma) yielded from San Clemente; had Catalina uplifted during the late Quaternary, the resulting knickpoints should be no further than 1 km inland from the coast.

4.3 Catalina-specific terrace information, Last Glacial Maximum, regional considerations

Submerged terrace morphology on the south and west-facing shores of the island is dominated by erosion from large waves from the open Pacific, while the NE side of the island experiences calm seas year round with the exception of the occasional Santa Ana winds (Shepard, 1937). The NW tip of the island is impacted by the California Current from the northwest, which drives longshore sediment transport. South-facing shores are subjected to large, long-period south swells from tropical systems and southern hemisphere storm activity. Catalina terraces exposed to this current are the largest terraces in lateral extent. The submerged terraces of Catalina, and all other Channel Islands exhibit a bimodality between the windward (NW to SW) and leeward (E), with terraces on the E side of the island being narrower than on the SW (Emery, 1958).

The lateral extent of the 90 m T2 surface varies significantly around the island, and may be influenced by highstands prior to 65ka (**Fig. 7,** dashed line). Near Farnsworth Bank where T2 is largest (up to 5 km from outer edge to back edge) only 4 terraces are visible in the bathymetry at 30, 90, 125, and 170 m depths. In comparison, the terraces north of Catalina Island are narrower (up to 2 km from outer edge to back) and 9 are observed rather than 4. It is likely that the broader T2 surface near Farnsworth Bank has been reoccupied by several sea level stands, while the more rapidly or uniformly subsiding northern end of the island has recorded several discrete terrace-cutting episodes. We acknowledge that the lateral extents of the terraces are governed not only by the duration of sea-level stands during a cutting episode, but by the frequency of strong storms, the shape of the coastline, local uplift or subsidence, and currents which drive long-shore drift of sediment. The interplay of these causes remains poorly understood, yet they are critical for using subsided terraces to understand both tectonics and paleoclimates.

The locations where an LGM paleoshoreline have been identified (Pilgrim Banks and Catalina) are both likely to be subsiding, as suggested by a succession of marine terraces on Pilgrim Banks similar to Catalina's, deeper than 130 m depth. It is therefore likely that LGM sea level in the SCCB was a few meters higher than -130 m during the LGM. Using the 0.25 mm/yr vertical rate derived from our subsidence analysis, it is possible that Catalina's LGM terrace is ~5 m below its original incisional depth of ~126 m. During post-glacial sea level rise, T2 was likely briefly reoccupied, then abandoned after the Older Dryas (~14ka).

4.4 Active tectonics

4.4.1 Tilting of Catalina

T5 and T6 are important strain markers for the motion of the Catalina block. T5 dips approximately 0.7 ± 0.2° NNE and T6 dips 1.5 ± 0.3° NNE indicating progressive rotational subsidence (**Fig. 8**). Our best-fitting chronology suggests T5 records tilt accumulated after 278 ka and T6 tilt after 355 ka. Seismic reflectors beneath T6 steepen progressively to a maximum dip of ~1.9°, the dip of the basement between 0.5 and 1.75 km on seismic line 2055. Catalina's tilting toward the mainland was first noted by Smith (1897) based on ridgeline morphology. This tilting destabilizes slopes on the NE side of the island, threatening a landslide in the direction of Los Angeles and Orange Counties. Catalina has produced landslides in excess of 1.3 km^3 with downslope motion of over 5 km (**Fig. S1**) (Legg & Francis, 2011). Numerical models of tsunami runup resulting from the 1.5-1.75 km^3 Goleta Slide, north of Santa Cruz Island, suggest up to 10 m runup along a ~30 km stretch of coastline (Lee et al., 2009). Models of seismogenic tsunamis induced by slip on faults on or near Catalina predict a maximum coastal run-up of 1.5-2.2 m in near-shore communities (Legg et al., 2004). Landslides are a likely greater tsunamogenic hazard than fault motion alone. Legg and Kamerling (2003) describe large basement-involved submarine landslides in the SCCB and recognize that detachment fault surfaces provide slip surfaces for such large-scale slides. Submarine landslides occur at slopes of less than 1° and the morphology of the terraces suggests that Catalina has already tilted over 1.5°. Submarine landslides appear more prevalent in terraces composed of low-metamorphic-grade detritus than in those derived from granitic rocks (**Fig. 2**), suggesting that parent bedrock lithology is a first-order control on slope stability of Catalina's submerged terraces.

4.4.2 Vertical tectonic motion of the Channel Islands

We have evidence for vertical motion on three different time scales: from the sedimentary record during the Miocene-Pliocene, from terrace depths in the late Quaternary, and from GPS records for the last decade. The sedimentary record on Catalina suggests it was exhumed during the mid to late-Miocene and emerged from the ocean during the Pliocene (Vedder, 1979). Cessation of uplift along the Santa-Cruz-Catalina fault occurred sometime between the youngest early Pliocene marine strata and the oldest mid-Quaternary subsided terraces, and may be contemporaneous with the inception of a continuous San Pedro Basin -San Diego Trough Fault (SPB-SDTF).

Post-rift thermal contraction is responsible for much of the Early-Mid Neogene subsidence of southern California (Turcotte and McAdoo, 1979), including the Channel Islands. The long-wavelength topography, becoming gradually deeper between Los Angeles and the edge of the continental shelf to the west, and the numerous submerged islands in the borderland that are deeper than -130m (e.g. Emery Knoll), indicate a regional subsidence since mid Miocene time (Emery, 1958). Today's residual thermal contraction subsidence signal should be <0.05 mm/yr. Additional subsidence is the result of regional subsidence and localized zones of transtension or transpression.

Scripps Orbit and Permanent Array Center (SOPAC) GPS data suggest that Catalina is subsiding relative to the other Channel Islands (**Table S1, Fig. 4**). Vertical GPS timeseries from the Channel Islands are strikingly well correlated, mostly due to reference-frame errors and unmodeled atmospheric effects. Vertical GPS timeseries vary depending on the method used when computing solutions, and consensus on the most accurate method for computing vertical timeseries has not been reached. However, differencing between vertical timeseries records over long occupation intervals reveals a coherent subsidence signal for Catalina Island relative to all other Channel Islands **(Fig. 4b)**. Glacio-isostatic adjustment (GIA), i.e. vertical crustal response to shifting water mass due to glaciation and deglaciation events, may impart a subsidence signal on far-field localities like Los Angeles causing coastal subsidence as well (Muhs et al., 2012). However GIA and regional subsidence should act similarly on all the Channel Islands and can not account for the difference in vertical motion among different islands, including Catalina's or Pilgrim Banks' Quaternary subsidence relative to other Channel Islands (Chaytor et al., 2008).

Time-variant differential motions of faults related to the anastomosing San Andreas Fault System account for the yo-yo tectonics observed in the SCCB. The uplift of Catalina in the Pliocene to ~200m above its current elevation occurred largely along the Catalina Escarpment which links the SC-CF with the SDTF (Legg, 2007). The SC-CF may have been reactivated along vertical transpressional faults not imaged by seismic, assisting in the uplift of the island, and again during Catalina's subsidence. Post-Pliocene subsidence may be in part due to transfer of motion from the SDTF-SC-CF restraining bend to the SPB-SDTF (Francis and Legg, 2011; Ryan et al., 2012). The increasing dips of Catalina's terraces indicate that some subsidence has been accommodated by oblique dextral slip on the SPB-SDTF and that the remaining subsidence of the island has been relatively uniform, either accommodated by slip along bounding faults or as the entire region subsides.

5 Conclusions

Catalina's submerged terraces correspond to erosional wave-cut platforms formed during sea-level lowstands. Our analysis of submarine bathymetry and high-resolution seismic data for Catalina Island requires subsidence since 355 ka, and possibly much longer, contradicting recent interpretations of Quaternary uplift (Schumann et al., 2012). The lack of subaerial marine terraces on Catalina Island is due to erosion of the original wave-cut terraces during Quaternary subsidence driven by tectonics. Our study suggests a subsidence rate of 0.32 mm/yr for the last 355 ka, and a mean subsidence rate of 0.25mm/yr for the last 1.15 Ma, although possibly slower over a longer time interval. Catalina is subsiding tectonically, possibly due to a transfer of motion from the Santa Cruz-Catalina fault to the San Pedro Basin-San Diego Trough fault. Differenced SOPAC GPS vertical time series records reveal a subsidence signal relative to other Channel Islands, agreeing in direction of vertical motion, although the magnitude of this subsidence is poorly constrained by GPS. The northern portion of Catalina is sinking faster than the south, destabilizing slopes and contributing to seismogenic tsunami hazard in Orange County and southern Los Angeles County. The broader significance of this paper is to affirm the importance of submarine geomorphology, alongside the better studied sub-aerial counterpart, in understanding regional tectonics.

Acknowledgments

Bathymetric data used in this study were acquired, processed and distributed by Cal State Monterey Bay's Seafloor Mapping Lab. Seismic processing used OpenCPS tools, donated by Open Geophysical. Kingdom Suite software was available thanks to a generous grant from IHS to California State University, Long Beach and Stanford. Vertical GPS data is from the Garner GPS archive, hosted by the Scripps Orbit and Permanent Array Center (SOPAC). Horizontal GPS data is based on data provided by the UNAVCO Facility with support from the National Science Foundation NSF and NASA under NSF Cooperative Agreement No. EAR-073515. CSULB seismic surveys and analysis were supported by a grant from the American Chemical Society Petroleum Research Fund. Stanford seismic surveys were supported by Conoco-Phillips Inc. and equipment loans from Geometrics, San Jose, CA, and Subsea Systems, Ventura, CA. Tim McHargue and Paul Segall at Stanford University assisted with sequence stratigraphy and GPS, respectively. CMC is supported by an NSF Graduate Fellowship. Conni De Masi Castillo assisted with field work.

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Table Captions

Table 1 : Terrace elevation data from bathymetry and seismic data.

Table 2: Subsidence rate calculations. The three values in bold (top down) in each time column are the time in ka, depth (at that time), and the subsidence rate between today and the time in row two of each column. The time and depth coordinates of lowstands selected from the sea level curve **(Fig. 7).**

Figure Captions

Figure 1: Bathymetry of the Southern California Continental Borderland. Contour interval = 250 m. Major faults and fault zones are shown with thick lines: A-DF—Anacapa-Dume fault; FFZ—Ferrelo fault zone; MCF— Malibu Coast fault; NIFZ— Newport-Inglewood fault zone; PVF—Palos Verdes fault; RCF—Rose Canyon fault; SCCF—Santa-Cruz-Catalina Fault; SCF—San Clemente fault; SCIF— Santa Cruz Island fault; SMF—Santa Monica fault; SPB-SDTF—San Pedro Basin - San Diego Trough Fault. After Chaytor et al. (2008).

Figure 2: Geology, bathymetry, and submerged terrace outer edges of Santa Catalina Island. Geology adapted from Vedder (1979) and Grove et al. (2008). Bathymetry: slope-enhanced shaded relief map, data from Cal State University, Monterey Bay, Seafloor Mapping project (CSUMB) and National Geophysical Data Center. Undissected outer edges of submerged marine terraces are mapped in black lines. Terraces were mapped using a combination of 2D seismic, slope-enhanced shaded relief maps, and low-sun angle hillshade.

Figure 3: Topography and Bathymetry of (a) Catalina and (b) San Clemente Islands. Catalina topography: hillshade azimuth 315^{o} and altitude of 45^{o}. San Clemente topography: slope map selected to show stair-step terraces onshore. San Clemente bathymetry: Slope-enhanced shaded relief map of NGDC data gridded using MB-System. Catalina bathymetry: slope-enhanced relief map. Data: Cal State University, Monterey Bay, Seafloor Mapping project (CSUMB) and National Geophysical Data Center. Underlying bathymetry (both images): grayscale shaded relief map, data: Southern California Coastal

Ocean Observing System (SCCOOS) 1/2 second bathymetric grid of Southern California. Topography (both images): USGS seamless server.

Figure 4: A: UNAVCO GPS velocities **(Table S1)**, relative to Palos Verdes relative to SNARF. Faults after Chaytor et al. (2008). B: Projection of Channel Islands looking shoreward along N40E. Red arrows and numbers are rates of uplift relative to Catalina calculated by differencing GPS vertical timeseries over long occupation intervals. Black arrows indicate vertical motion rates calculated from marine terrace data.

Figure 5: A: Seismic line KELEZ-65 **(Table S1)**. Vertical resolution: ~2m. Inset contains interpretation of sequence boundaries and "bright" reflectors. **b.** Cutaway view of Line 65. Cartoon representation of modern bathymetry and underlying sequence stratigraphy. At cutaway 1 all sediment above wave cut platform T5 has been removed. At cutaway T6 we removed sediment overlying platform 3.

Figure 6: **a.** Seismic line 2205, constant-offset (40 m) section data collected in 2014 **(Table S2)**. 1.5 kJ boomer, vertical resolution ~0.5m. Terrace back-edges circled in red. Terrace treads shown with brackets over terraces. Inset: interpretation of major sequence boundaries in line 2205, note different spatial/temporal scale on bottom and right side of figure. Red lines are sequence boundaries responsible for terraces observed in the bathymetry. **b.** Slope map and bathymetric contours, north tip of Catalina Island. Contour interval 25 m with 100 m index contours. Topography from Los Angeles Region Imagery Acquisition Consortium, bathymetry from CSUMB Seafloor Mapping Lab. Underlying bathymetry: grayscale shaded relief map, data: Southern California Coastal Ocean Observing System (SCCOOS) 1/2 second bathymetric grid of Southern California. **c.** Histogram of bathymetry data using 1m vertical bin size. Inset shows location of DEM: Points from 0 to -500m are captured in the histogram.

Figure 7: Best-fitting correlation of sequence boundaries with sea level curve. Blue line is sea level curve generated from Lisiecki and Raymo's stack of 57 globally distributed benthic $\delta^{18}O$ records, bold black line is sea level curve normalized to a window of six samples. Each terrace's corresponding sequence boundary is plotted on the Y-axis of the graph at t=0. Lines of positive slope represent subsidence rate.

Figure 8: 3D view of Catalina looking east at North Point with 5x vertical exaggeration. Topography is in brown and higher elevations have "mist." Bathymetry: 3D perspective view of DEM overlain with slope map. Brighter colors are steeper slopes.

Supplementary Material Captions

Figure S1: Seismic line CSULB-183 **(Table S2)**. 1.3 km^3 landslide west of Catalina Island. covering the scarp and toe of a 1.3 km^3 submarine landslide near Catalina Escarpment.

Figure S2: Left: Gridded bathymetric DEM used in Schumann (2012). Artifacts interpreted by Schumann as marine terraces are artifacts of gridding an aliased dataset. See Passaro et al. (2011) for a description of these artifacts. Dashed lines are 100ft contours which corespond to the false terraces. Right: Bathymetric grid of submerged features off south tip of San Clemente Island. 10m grid produced in MB-System from multibeam bathymetry collected on R/V McDonald.

Figure S3: Seismic line CSULB-187 **(Table S2)**. Migrated 16-channel seismic data. 2 kJ sparker, vertical resolution: ~15m. Data from north of Avalon, Long Point Fault shown in black. Only major sequence boundaries are effectively interpreted with these sparker data.

Figure S4: a. Seismic line 2202, constant-offset (40 m) section of 35-channel data collected in 2014 **(Table S2)**. 1.5 kJ boomer, vertical resolution ~0.5m. Inflections in acoustic basement are possibly wave-cut notches.

Table S1 : GPS data for Channel Islands and Palos Verdes Peninsula. Data and errors provided by UNAVCO (2013). SNARF: Stable North American Reference Frame

Table S2 : Seismic data used in this study.