Pleistocene vertical motions of the Costa Rican outer forearc from subducting topography and a migrating fracture zone triple junction

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ABSTRACT

Understanding the links between subducting slabs and upper-plate deformation is a longstanding goal in the field of tectonics. New 3D seismic sequence stratigraphy, mapped within the Costa Rica Seismogenesis Project (CRISP) seismic-reflection volume offshore southern Costa Rica, spatio-temporally constrains several Pleistocene outer forearc processes and provides clearer connections to subducting plate dynamics. Three significant and/or slope erosional events at ca. 2.5–2.3 Ma, 1.95–1.78 Ma, and 1.78–1.19 Ma, each with notable differences in spatial extent, volume removed, and subsequent margin response, caused abrupt shifts in sedimentation patterns and rates. These shifts, coupled with observed deformation, suggest three primary mechanisms for Pleistocene shelf and slope vertical motions: (1) regional subsaerial erosion and rapid subsidence linked to the southeastward Panama Fracture Zone triple-junction migration, with associated abrupt bathymetric variations and plate kinematic changes; (2) transient, kilometer-scale uplift and subsidence due to inferred subducting plate topography; and (3) progressive outer wedge shortening accommodated by landward- and seaward-dipping thrust faults and fold development due to the impinging Cocos Ridge. Furthermore, we find that the present-day wedge geometry (to within ~3 km along strike) has been maintained through the Pleistocene, in contrast to modeled landward margin retreat. We also observe that deformation, i.e., extension and shortening, is decoupled from net margin subsidence. Our findings do not require basal erosion, and they suggest that the vertical motions of the Costa Rican outer forearc are not the result of a particular continuous process, but rather are a summation of plate to plate changes (e.g., passage of a fracture zone triple junction) and episodic events (e.g., subducting plate topography).

INTRODUCTION

The Costa Rica convergent margin (CM) is thought to be an end-member-type erosive margin (von Huene and Scholl, 1991; Ranero and von Huene, 2000; Clift and Vannucchi, 2004), where basal erosion (von Huene et al., 2004) is driving widespread Costa Rican outer forearc subsidence (Hinz et al., 1996; Vannucchi et al., 2013). However, a recently collected 3D seismic-reflection volume offshore southern Costa Rica (CRISP seismic experiment; Kluesner et al., 2013; Bangs et al., 2014) reveals structures that conflict with the generic erosional model, showing a margin wedge that is composed of a layered fabric that is pervasively folded and thickened by thrusting. Subsequent Integrated Ocean Drilling Program (IODP) drilling successfully penetrated the margin wedge and recovered clastic sediments (Vannucchi et al., 2011; Harris et al., 2013a), consistent with the layered fabric seen in 3D seismic imaging. These results have led some to propose a temporary phase of recent accretion (Bangs et al., 2016) or a new type of CM, a depositionary margin, where extreme, episodic basal erosion removes forearc basement that then drives subsidence and rapid terrigenous sedimentation (Vannucchi et al., 2016).

Reconciling Pleistocene outer forearc subsidence with shortening and thickening offshore Costa Rica requires a detailed investigation into the timing and amounts of outer forearc vertical motions and shortening. For example, work offshore Honshu found a spatiotemporal link between outer forearc subsidence, backarc spreading, and plate kinematic changes (Regalla et al., 2013). Regalla et al. (2013) proposed that changes in the shallow-slab geometry could drive outer forearc subsidence. Shallow-slab geometry changes can result from two possible mechanisms: (1) an increase in slab effective elastic thickness or (2) a change in slab buoyancy. Both mechanisms can be driven by a change in the convergence rate; specifically, an increase in the convergence rate can either increase the radius of slab curvature or drive negative slab buoyancy, both potentially producing downward flexing of the slab under the outer forearc (Furlong et al., 1982; Buijten et al., 2001; Regalla et al., 2013). Furthermore, although subducting topography can remove and/or tectonically erode the frontal prism and lower slope (Lallemand and Le Pichon, 1987; Bal lace et al., 1989; von Huene and Scholl, 1991; Sak et al., 2004), the effects of subducting topography farther down dip are less known. Analog sandbox models suggest lower slope removal, middle and upper slope uplift, scaled to the size of the subducting feature, and ensuing rapid subsidence (Domínguez et al., 1998). These findings and models provide alternative mechanisms to basal erosion for outer forearc subsidence.
Here we report on a detailed 3D mapping of unconformities and a subdivision of Pleistocene strata into depositional sequences on the southern Costa Rica subduction margin, utilizing the newly collected 3D seismic-reflection volume (CRISP) and drilling results from IODP Expeditions 334 and 344. The integration of seismic and well data allows us to temporally constrain sedimentation, vertical motions, and shortening (e.g., Ladd and Schroeder, 1985). We show evidence for three main mechanisms controlling Pleistocene vertical motions: (1) regional subaerial erosion and rapid subsidence linked to the southeastward Panama Fracture Zone triple-junction migration, with associated abrupt bathymetric variations and plate convergence changes; (2) transient, kilometer-scale uplift and subsidence due to inferred subducting plate topography; and (3) outer wedge shortening accommodated by landward- and seaward-dipping thrust faults and fold development due to the impinging Cocos Ridge. Our results underscore the importance of 3D seismic imaging and integration of well data in the documentation and decoupling of complex convergent margin histories.

**TECTONIC SETTING**

Subduction processes offshore Costa Rica likely began during the Late Cretaceous, when andesitic volcanism initiated near the present-day volcanic arc (Lundberg, 1991). The Caribbean plate is thought to have formed either “in situ” (Frisch et al., 1992) or to the west and then subsequently emplaced piecemeal beginning in the Late Cretaceous and extending through much of the Cenozoic (Burke, 1988; Pindell and Barrett, 1990). The Cocos and Nazca plates are thought to have their origins from the splitting of the Farallon plate ca. 27 Ma, based on early magnetic and bathymetric studies (Hey, 1977; Lonsdale and Klitgord, 1978), although later work revised this event to ca. 23 Ma (Lonsdale, 2005). The present-day Cocos plate is subducting under the Caribbean plate with a convergence direction approximately normal to the Middle America Trench (MAT) at ~70–90 km/m.y. (depending on which velocity model is used) and dipping ~19° (offshore Osa Peninsula), while the Nazca plate is moving 37–48 km/m.y. to the ENE relative to the Caribbean plate and is subducting more steeply at ~33° (Fig. 1; DeMets et al., 1994; DeMets et al., 2001; Morell et al., 2008; DeMets et al., 2010; Argus et al., 2011; Kobayashi et al., 2014; Morell, 2015). The Caribbean plate can be subdivided further with its Panama–southern Costa Rica portion referred to as the Panama microplate (Marshall et al., 2000), which moves relative to the Caribbean plate at 25 km/m.y. to the NE (Kobayashi et al., 2014). From this, Morell (2015) calculated a Cocos-Panama convergence rate of ~70 km/m.y. and a Nazca-Panama convergence rate of ~20 km/m.y.

The differing Cocos and Nazca relative plate motions are accommodated by the north-striking Panama Fracture Zone (PFZ), a distributed right-lateral transform plate boundary with sharp bathymetric steps (Fig. 1; von Huene et al., 2000). The PFZ transition separates large crustal thickness differences (Sallarès, 2003) and drives lateral offsets of the MAT (Moore and Shipley, 1999), the Cocos Ridge and PFZ and the slight clockwise rotation of the Cocos Ridge axis relative to either of the relative plate motion models. EPR—East Pacific Rise.
The PFZ has been migrating southeastward along the MAT at ~30–55 km/m.y. since the Middle Pliocene (Silver et al., 1990; McIntosh et al., 1993; Morell et al., 2008; Morell, 2015). This southeastward migration has been linked to inner forearc shortening (Morell et al., 2008, 2013), outer forearc shortening (McIntosh et al., 1993; Morell et al., 2011), outer forearc uplift (Morell et al., 2011), outer forearc subsidence (Corrigan et al., 1990), and possibly to the ~2-km-high, ~100-km-wide, and ~150-km-long Coiba Ridge, forming as a result of compression (MacMillan et al., 2004).

The Cocos-Nazca spreading center ridge (CNS) generated new crust through three sequential ridge orientations, all while interacting with the existing Galapagos hot spot, which emplaced thick sequences of volcanic rocks on the north and south sides of the ridges (Meschede et al., 1998). This process resulted in the formation of the NE-trending Cocos Ridge (Fig. 1) and the E-trending Carnegie Ridge (located on the Nazca plate) (Barckhausen et al., 2001). The Cocos Ridge is a ~2-km-high, ~250-km-wide, and >20-km-thick NE-trending ridge that extends from the CNS to the MAT and is truncated on its eastern side at the PFZ (Sallarès, 2003), exhibited by an ~2 km bathymetric scarp (Fig. 1). The Cocos Ridge broadly deflects the MAT landward over ~350 km along strike. The Cocos Ridge is oriented ~10° clockwise from the Cocos-Caribbean relative plate motion vector (~N24°E; Nuvel-1A model), resulting in a slow NW migration of the Cocos Ridge across the MAT with continued subduction (Barckhausen et al., 2001). Timing of the initiation of subduction of the Cocos Ridge is debated, with ages ranging from ca. 0.5 to 8 Ma (Gardner et al., 1992; Abratis and Worner, 2001); however, recent studies have narrowed in on ca. 2–3 Ma (Morell et al., 2012; Morell, 2015). The Cocos Ridge exhibits strong controls on upper-plate deformation processes, including landward deflection of the MAT and increased horizontal velocities (LaFemina et al., 2009), outer forearc uplift (Corrigan et al., 1990; Gardner et al., 1992; Sak et al., 2004; Morell et al., 2011), inner forearc shortening (Fisher et al., 2004; Sitchler et al., 2007), and backarc shortening (Collins et al., 1995).

West of the Galapagos hot spot, oceanic crust generated along CNS results in a dense array of seamounts (~40% of surface area), while Cocos plate crust generated farther west and north along the East Pacific Rise results in a smoother topography (Fig. 1; Ranero and von Huene, 2000). The zone of seamounts consists of discrete, conical seamounts up to ~2 km high and ~20 km diameter and broad, flat seamounts and plateaus up to ~2 km high and ~40 km in width (e.g., the Quepos Plateau) (Fig. 1; von Huene et al., 1995, 2000). The zone of seamounts is estimated to have begun subduction ca. 3–4 Ma (Morell, 2015). Cocos plate variations outboard of the MAT (i.e., changes in bathymetric roughness and thickness) parallel the along-strike changes of the outer forearc (e.g., changes in margin wedge thickness and width, slope continuity and trench geometry; Fig. 1) (Hinz et al., 1996; Fisher et al., 1998; von Huene et al., 2000; Sak et al., 2009). These variations also parallel tectonic changes (e.g., coastal uplift, fault kinematics, and arc migration) within the onshore inner forearc and volcanic arc (Gardner et al., 1992; Marshall and Anderson, 1985; Fisher et al., 1998; Marshall et al., 2000; Gardner et al., 2001; Marshall et al., 2003; Fisher et al., 2004; Sak et al., 2004).

A regionally extensive seismic discontinuity, known offshore the Nicoya Peninsula as the Base of Slope Sediment (BOSS) reflection (Vannucchi et al., 2001), extends across the entire outer forearc offshore Costa Rica (Hinz et al., 1996) and is correlative to the onshore Mal Pais unconformity (Vannucchi et al., 2001). This regional unconformity demarcates a surface along which widespread regional erosion occurred, although not occurring at the same moment in time. The unconformity offshore the Nicoya Peninsula separates an Early-Middle Miocene margin wedge from an overlying Middle Miocene to present sedimentary apron (Kimura et al., 1997), while offshore the Osa Peninsula, the unconformity separates an Early Pleistocene to Pliocene margin wedge from Early Pleistocene to present slope and shelf sediments (Vannucchi et al., 2011; Harris et al., 2013a). The overlying sediments typically contain a fining-upward section, recording deepening water depths (Kimura et al., 1997; Vannucchi et al., 2011; Harris et al., 2013a). To the SE of the CRISP volume, ~15 km, Vannucchi et al. (2013) found Pliocene deep-water sediments overlain by nearshore sediment facies and benthic foraminifera, recording rapid uplift from ~1 km below sea level (bsl) to sea level. The nearshore sediments are then overlain by a fining-upward section, recording deepening water depths (Vannucchi et al., 2013). This regional unconformity demonstrates widespread regional surface erosion followed by subsidence.

**PREVIOUS SEISMIC-REFLECTION STUDIES AND IODP DRILLING**

Multichannel seismic-reflection data offshore Costa Rica were first collected ~35 years ago (Crowe and Buffler, 1983), followed by Sea Beam multibeam sonar and high-resolution, water-gun seismic reflection (Shipley and Moore, 1986) and limited deep tow seismics (Moore and Shipley, 1988). Many different models were suggested for the internal structure of the forearc margin, including (1) the margin is a sedimentary body with young offscraping (Crowe and Buffler, 1983; Shipley and Moore, 1986); (2) the deformed wedge is much older than the undeformed veneer of sediments and therefore is non-accretionary (Crowe and Buffler, 1983); or (3) the deformed wedge is a result of growing thrust duplexes with the undeformed veneer of sediments protected by roof thrusts (Silver et al., 1985). Two-dimensional and 3D seismic-reflection data were later collected offshore the Nicoya Peninsula (Stoffa et al., 1991), and these data revealed evidence of possible offscraping at the trench, duplexing and out-of-sequence faulting, large variability in fault spacing and reflector geometry, and young, active faults cutting the seafloor (Shipley et al., 1992; McIntosh and Silver, 1996).

Drilling efforts offshore Costa Rica continued concomitantly with these early bathymetric and reflection studies, with ODP Legs 170 and 205 (Kimura et al., 1997; Morris et al., 2003). Leg 170 proved especially valuable, revealing that all incoming strata offshore Nicoya were subducted (Kimura et al., 1997; Silver, 2000), in contrast to previous accretionary margin models (Shipley et al., 1990, 1992).
Further geophysical investigations included broad 2D multichannel seismic-reflection surveys (e.g., SO-76, SO-81, and BGR99) that covered large portions of the Costa Rica convergent margin (Hinz et al., 1996; von Huene et al., 2000; Ranero et al., 2008). These studies revealed information suggesting abundant fluid-flow along the shallow portions of the plate boundary interface within the subduction zone (Ranero et al., 2008), subduction of sparse trench sediments (Ranero and von Huene, 2000), a transitional backstop (von Huene et al., 2000), and spatially episodic accretionary and erosional processes (von Huene et al., 2000).

Recent drilling efforts have focused offshore the Osa Peninsula with IODP Expeditions 334 and 344 (Vannucchi et al., 2011; Harris et al., 2012). IODP Expedition 334 drilled wells along 2D multichannel seismic-reflection lines to the southeast of the CRISP 3D volume, while IODP Expedition 344 drilled wells within the 3D volume (Fig. 1), with the exception of a re-entry and deepening of well U1380 from Expedition 334.

IODP Expedition 334 efforts to drill through the sedimentary cover and into the margin wedge within the slope failed; however, Site U1379 (located on the outer shelf in ~125 m water depth; Fig. 1) drilled to ~960 m below seafloor (mbsf), penetrating margin wedge material at ~880 mbsf, documenting the entire shelf sedimentary cover sequence (Vannucchi et al., 2011). Site U1379 recovered margin wedge material consisting of Early Pleistocene (ca. 2.5 Ma) middle bathyal paleodepth rocks (800–1200 m below sea level [mbsl]), exhibited by interpreted forearc basin sediment facies and benthic foraminifer assemblages, overlain by nearshore sediment facies (Vannucchi et al., 2013). Per Vannucchi et al. (2013), rapid uplift, followed by rapid subsidence occurred during Early Pleistocene (ca. 2.5–2.0 Ma), subsequent to moderate uplift from 1.5 Ma to the present (Vannucchi et al., 2013). Sites U1378 and U1380 are located on the upper slope in ~500 m water depths and were drilled to depths of ~500 mbsf. They encountered ca. 1.5 Ma rocks and sediment facies and benthic foraminifera assemblage changes that show abyssal paleodepth rocks (>2000 mbsl), overlain by sequentially shallower paleodepth rocks to the present (Vannucchi et al., 2013). Sites U1378, U1379, and U1380 all encountered generally coarsening sediments down well, with sandy intervals becoming more frequent and more massive with depth (Vannucchi et al., 2011).

Expedition 344 revisited site U1380 from Expedition 334 and extended it to the margin wedge, and drilled two wells within the CRISP 3D seismic-reflection volume, Site U1414, located seaward of the trench and on the down-going Cocos plate, and Site U1413, located within the upper slope (Fig. 1; Harris et al., 2013a). Site U1414 drilled through the overlying hemipelagic and pelagic sediments and into basalt flows at ~375 mbsf to well bottom at ~465 mbsf (Harris et al., 2013b), documenting the shallowest material being carried into the subduction system. Site U1413 drilled to ~584 mbsf and encountered terrigenous sediments of sedimentary and magmatic origins with grain sizes that generally increase at greater depths, punctuated by turbidite sands, tephras, general mass transport deposits, and major benthic foraminifera assemblage changes (Harris et al., 2013d). U1380 was extended to ~800 mbsf, encountering terrestrial sourced sediments of similar structure and composition to slope sediments from U1413. U1380 also penetrated ~250 m into the Pliocene margin wedge and encountered sediments similar to margin wedge sediments from Expedition 334 U1379, namely clays and silts with occasional thin sand layers (Harris et al., 2013c).

**METHODS**

In 2011, we collected an 11 × 55 km² 3D seismic-reflection data volume offshore the Osa Peninsula (Fig. 1) to study the subduction thrust zone and processes governing seismogenesis of an erosive convergent margin (Kluessner et al., 2013; Bangs et al., 2014). Seismic reflections were generated using two 27-gun arrays with a separation of 75 m and a 3200 L displacement operating in flip-flop mode. Data were recorded using four 6-km-long streamers spaced 150 m apart, resulting in a bin size of 12.5 × 18.75 m with ~60 fold (Bangs et al., 2014). Subsequent processing of the data removed multiples and suppressed noise using standard seismic processing workflows (Yilmaz, 2001). Post-stack-time migration was then performed. These data were used to generate a 3D velocity model that was utilized in a full pre-stack depth migration (Bangs et al., 2014). The data set images the subducting Cocos plate and overlying Caribbean plate down to depths >10 km.

We utilize seismic sequence stratigraphy techniques (Vail and Mitchum, 1977) for stratigraphic interpretation of the young, up to 2.5-km-thick, stack of reflections that drape the margin wedge. As such, strata are subdivided into depositional sequences, which are bounded by subaerial unconformities and their correlative marine conformities (Figs. 2 and 3). We pick the marine correlative conformity in our study as the seafloor at the onset of relative base-level fall (Kolla et al., 1995). We describe stratial stacking patterns as one or a combination of the following: upstepping, forestreeing, backstepping, and downstepping (Fig. 4). In shallow-water environments, these stacking patterns express three types of shoreline shifts: forced regression (foresteeding and downstepping), normal regression (foresteeding and upstepping), and transgression (backstepping) (Fig. 4; Catuneanu et al., 2011).

Synchronous reflections are thought to represent either stratal surfaces, which follow geologic time lines, or discontinuities, such as unconformities, which signify geologic time gaps (Vail et al., 1977). These stratal or unconformable reflections have low diachronity in the dip direction of the slope but can be more diachronous along strike (e.g., the Miocene to Pleistocene regional shelf and slope unconformity that extends across Costa Rica) (Catuneanu, 2002). Diachronous reflections, such as the methane-hydrate–related, bottom-simulating reflector, are often fluid-rich interfaces, or can be representative of diagnostic changes, such as the transition from Opal A to Opal CT (Vail et al., 1977). Fluids, especially gas, can attenuate the acoustic signal, and fluid pathways can cause zones of chaotic, low-energy reflections (Leseth et al., 2000). Within the CRISP 3D survey region, there are abundant indicators for fluid-rich zones, fluid migration, and seafloor seepage across the slope and shelf (Kluessner et al., 2013), and they limit our ability to map depositional sequences across portions of the data.
The application of seismic stratigraphy methods to deep-water settings (e.g., slope) is challenging, especially along the Costa Rican convergent margin, which has high orthogonal convergence rates of 70–90 km/m.y. There is the possibility of physical disconnections between shallow-water portions of depositional sequences and their deep-water equivalents, usually due to slope instability and mass transport deposits, which can lead to lateral instead of vertical stacking patterns (Catuneanu et al., 2011). Furthermore, relative base-level changes are only seen indirectly in deep-water settings, although notable exceptions occur.

In spite of these challenges, deep-water sequences can include several important shallow-water equivalent surfaces, including those that form at the points of maximum shoreline regression and transgression (Catuneanu et al., 2011). These deep-water equivalent surfaces are mappable in the 3D volume.

We focused our mapping efforts on unconformities and their correlative conformities, referred to as events (e.g., L1, L2, M1, M2, etc.; Figs. 2 and 3), and in mapping them in high resolution (every 5–10 crosslines and 5–10 inlines). Mapping was performed within OpendTect 6.0.5 seismic interpretation.
Software. This effort resulted in well-constrained gridded horizons that show significant detail (e.g., Fig. 5), such as folding and thrusting (Figs. 5 and 6) and paleodrainage networks (Figs. 5 and 7). Once horizons were mapped, a gridding algorithm was utilized to interpolate between picks and generate 3D surfaces. The gridding algorithm was augmented by a calculated dip-steered volume, which is a 3D volume of local dip and azimuth data for each seismic sample that structurally guides interpolation. Gridded horizons were then examined, exposing mapping errors or inconsistencies, which were then reassessed iteratively. Once satisfactory, we applied a median filter with a 2 × 2 step-out (inline and crossline) to the gridded horizons to subdue acquisition footprint noise and other gridding artifacts. We then calculated vertical thicknesses between horizons (isopachs) and projected those thicknesses onto the younger horizon (e.g., Fig. 7). Isopach results were utilized to estimate sedimentation rates during depositional intervals. Note that isopachs are vertical thicknesses; thus, if strata are dipping or offset, isopach results will show apparent thicknesses, which are greater than real thicknesses (e.g., some isopach results along the limbs of thrusts A–D; Fig. 7). Thus, thickness values used for sedimentation rates are not always maximum observed thicknesses but are those more closely interpreted to be real thicknesses.

We integrate drilling results from IODP Expedition 344 to the 3D pre-stack depth migrated volume. Depth imaging allows correlation to U1414 and U1413 magnetostratigraphy, lithostratigraphy, and recovered benthic foraminifera assemblages. Incomplete biostratigraphic data limit some magnetic polarity interval correlation; however, several polarity intervals are well correlated (e.g., the Olduvai subchron to the normal polarity section of 480–520 mbsf in U1413) (Harris et al., 2013d).

RESULTS

Shelf and Slope Sediment Stratigraphy

We separate the ~0.7–2.5-km-thick shelf and slope sediment stratigraphy into three stratigraphic domains: Lower Strata, Middle Strata, and Upper Strata (Figs. 2 and 3). These stratigraphic domains are separated by three large erosional events: Lower Unconformity (L1), Middle Unconformity (M1), and Upper Unconformity (U1). Lower, Middle, and Upper Strata are bounded by these three large erosional events and the seafloor (i.e., L1 and M1 bound Lower Strata, M1 and U1 bound Middle Strata, and U1 and the seafloor bound Upper Strata). Depositional sequences within each stratigraphic domain are named sequentially upward, with each depositional sequence number bearing the same number as its underlying sequence boundary number; e.g., L1 (Lower Unconformity) and its overlying Ls1 (Lower Sequence 1), L2 (Lower Unconformity), and Ls2 (Lower Sequence 2), M1 and Ms1, M2 and Ms2, etc. (Figs. 2 and 3).
L1 Unconformity

L1 (Fig. 5) is a regional erosional event that shows a high acoustic impedance contrast, producing a strong, positive polarity reflection that separates the margin wedge from the overlying shelf and slope sediments. L1 is mappable to the landward extent of the CRISP volume and trenchward to the middle slope, a length of ~35–42 km and an area of ~430 km², at which point L1 terminates against underlying reflections or is lost due to poor reflectivity between steeply dipping forelimbs of anticlinal thrusts. L1 is unconformable across the shelf, upper slope, and partially the middle slope, overlying discontinuous margin wedge reflections and truncating landward- and seaward-dipping margin wedge reflections (e.g., Figs. 2 and 3). Its correlative conformable reflection is observed variably down slope. L1 depths range from ~0.85–2.9 km below sea level (bsl; Fig. 5). IODP drilling results constrain L1 to the Early Pleistocene, ca. 2.5 Ma (Vannucchi et al., 2011; Harris et al., 2013a).

L1 displays a remarkable topographic contrast between its slope and shelf portions, where it is densely deformed across the slope and is undeformed and channelized across the shelf (Fig. 5). This contrasting L1 topography marks the inner wedge (limited contractile deformation) and the outer wedge (folded...
and thrusted; Fig. 5), analogous to inner and outer wedges seen along other margins (Wang and Hu, 2006). The transition between these wedges is singular and steeply dipping to vertical within the central to eastern portion and distributed (two thrusts) and shallowly landward dipping to the west (Figs. 6–7). This transition is coincident with the present-day shelf break along the western portion of CRISP but deviates from the landward-deflected portion of the shelf break, i.e., the scalloped or arcuate shelf bite mark, by up to ~4 km, across the eastern portion of CRISP (Fig. 5).

Two major channel systems are observed across the L1 shelf (Figs. 5, 7, and 8). The main channel system, with its tributaries, encompasses >80 km², generally trends southwest, and extends >8 km in length adjacent to an old domal uplift, which we refer to as the “eastern dome” (Figs. 5 and 6). This main channel system cuts to ~500 m depth, relative to adjacent, average L1 depths (Figs. 5 and 7). The other L1 channel system, which occupies the NW corner of the volume and is only partially revealed, encompasses a minimum area of >16 km² and cuts to ~400 m depth (labeled ch2; Figs. 5 and 7). These channel systems feature branching, rugose tributaries with high relief at acute to nearly orthogonal confluence angles (Fig. 5), in contrast to present-day slope channels with straighter, anastomosing profiles and lower relief (Kluesner et al., 2013). Furthermore, well U1379 from IODP Expedition 334 recovered near-shore sands overlying L1 (Vannucchi et al., 2013). Thus, the observed channel systems likely formed due to fluvial incision and subaerial erosion of a formerly subaerial landscape.

**Lower Strata Depositional Sequences**

Immediately overlying the L1 event, Ls1 onlapping reflectors are observed ~10–15 km down slope from the present-day shelf break (Fig. 2). These Ls1 onlapping clinoforms backstep to, then against and over the inner/outer wedge transition, transgressing the shelf (Figs. 2 and 6) but do not transgress the

![Figure 6. Inlines 2085 (top) and 2410 (lower) of the inner/outer (i/o) wedge transition. Thick pink lines denote major erosional events L1 and M1. Dashed yellow lines are prominent thrusts. Solid white line is the L2 unconformity. Dashed white lines are younger lower strata unconformities. Solid black arrows are prominent onlapping clinoforms, and open black arrows are downlapping clinoforms. Two-sided arrows show normal faults, and one-sided arrows denote depths at which those normal faults terminate. Lower Strata Sequences 1 and 2 (Ls1 and Ls2) are labeled. Note the consistent thickness of Ls1 across the i/o transition at inline 2085 and the thinning of Ls1 up the eastern dome (projected eastern dome slope is shown with thick dashed pink [L1] and white [L2] lines). Note the thinning Ls2 across the i/o transition within both inlines. Also note the small offsets (<100 m) of normal faults and the depths that they cut to (shallower than L1 and mostly L2).](https://pubs.geoscienceworld.org/gsa/geosphere/article-pdf/14/2/510/4101799/510.pdf)
“eastern dome” (Figs. 6 and 7). Ls1 shelf clinoforms transgress landward of CRISP and then upstep until an abrupt seaward shift, L2 (Fig. 2). Ls1 reaches thicknesses of ~850 m within the slope and ~500 m within the shelf (where L1 channels are filled), which are the largest thicknesses of any Pleistocene depositional sequence and mark a transgressive succession >36 km in horizontal length (Fig. 7; greater than the landward extent of CRISP imaging). Ls1 reflectors demonstrate that the “eastern dome” predates earliest Pleistocene sedimentation and that the inner/outer wedge transition established itself soon after L1, along the eastern portion, and soon after L2, along the western portion (Figs. 2, 6, and 7).

Generally, the first half of Lower Strata depositional sequences, Ls1–Ls6, feature repetitive successions of backstepping clinoforms that build to, against, and over the inner/outer wedge transition (except Ls2), until abrupt seaward, or regressive, shifts (Fig. 2). These early backstepping clinoforms and their equivalent down-slope reflections are truncated by the younger M1 slope erosional event, although we see a portion of forestopping and upstepping clinoforms of Ls4–Ls6 (Fig. 2). The regressive shifts are interpreted as maximum regressive surfaces at the onset of transgression (i.e., most seaward onlapping reflector) and are measured between this seaward point and their preceding most landward point (often extends landward of survey, thus giving minimum value). Notable regressive shifts are L2, with an ~30 km regressive shift, L4, with an ~25 km regressive shift, and L7, which broadly truncates underlying Ls6 reflectors. The subsequent depositional sequences, Ls7–Ls10, onlap L7 sequentially across the shelf, with coeval forestopping and upstepping clinoforms across the outer shelf and uppermost upper slope (down-slope continuity is obscured due to truncation by M1 unconformity; Fig. 2). Generally, Lower Strata features a succession of backstepping clinoforms and equivalent down slope and/or shelf forestopping and upstepping clinoforms that are separated by marked regressive shifts. It is suggested by the coherent vertical stacking patterns, good lateral continuity, and large landward transgressions and seaward shifts that Lower Strata backstepping reflectors represent coastal onlap.

Lower Strata stack in patterns consistent with sequence stratigraphy models. At times, rapid Early Pleistocene tectonic subsidence did not outpace sedimentation (e.g., forestopping and upstepping clinoforms of Ls8), and at times...
During IODP drilling, ~80 m of Lower Strata were encountered at site U1413 and were composed of fine to medium sandstone to conglomerate from ~500 mbsf to well bottom (Fig. 9). Magnetostratigraphy constrains Lower Strata to the Olduvai subchron and its reversal and older (>1.78 Ma; Harris et al., 2013d). This, coupled with magnetostratigraphy results from wells U1379 (shelf) and U1380 (middle slope) ~15 km to the SE, which date the regional unconformity at 2.2 Ma, constrain Lower Strata to the Early Pleistocene (Harris et al., 2013c; Vannucchi et al., 2013).

Lower Strata stacking patterns are important for several reasons, including: (1) they demonstrate vertical stacking patterns and good along-strike continuity, relative to Middle and Upper Strata; (2) the inner/outer wedge transition established itself right after L1 and L2 and continued activity during Lower Strata deposition; (3) the "eastern dome" is a long-lived area of uplift that may represent a pre-L1 shelf break; and (4) even though the slope outpaced the shelf, subsidence was fairly coherent (shelf to at least middle slope), and broad (seen by large horizontal regressions and transgressions).

**M1 Unconformity**

Deposition of Lower Strata was interrupted by the slope M1 erosional event. M1 is a variably reflective event (positive and negative polarity or absent reflectivity) that truncates Lower Strata and margin wedge material (lower slope) across the entire slope (Figs. 2, 3, and 10) and is itself truncated locally updip by the U1 event (Fig. 10). Its downcutting portion is generally bounded on its landward side by the inner/outer wedge transition, and its correlative conformity extends landward across the shelf and is gently dipping landward, paralleling shelf strata (Figs. 2 and 10). M1 truncates upper-slope Lower Strata...
down to Ls4 (Fig. 2), constraining the thickness of material removed to ~900 m. Its upper slope is characterized by linear, sinuous, anastomosing channels and levees that track parallel to the slope and deeply channelize its surface (Figs. 3 and 10). Its upper-slope portion is dipping seaward ~8°–10° (Figs. 2 and 6), more than double that of the present-day slope (~3°). M1 shallows to sub-parallel the underlying Lower Strata across the middle slope (close to horizontal when not offset) and then cuts down section across the lower slope until it is truncated by the seafloor (Fig. 10). M1 is mappable down to within ~5–10 km of the trench (is locally obscured due to the BSR), truncating older lower-slope, landward-dipping thrusted sections (Fig. 10). M1 does not truncate tightly imbricated reflections of the frontal prism. Generally, its SE middle and upper-slope extent is more intensely deformed by folding (Fig. 10).

M1 was intersected by well U1413, at ~500 mbsf, constraining the event to the Olduvai subchron (1.95–1.78 Ma) (Fig. 9) (Harris et al., 2013d). No lithology change marks the event; however, a significant benthic foraminifera assemblage change at ~504 mbsf is encountered, with the appearance of Brizalina cf. dilatata below M1 (Harris et al., 2013d).

It is not known whether any portion of the M1 unconformity was formed due to subaerial erosion. It is possible that some portion of the upper slope was subaerial but was then overprinted by intense channelization along a
steeply dipping coastal gradient undergoing rapid subsidence. Regardless, the pervasive channelization, steep down cutting, and total slope surface erosion (excepting frontal prism) suggest a major slope collapse that removed a thick section of Lower Strata and margin wedge material.

Middle Strata Depositional Sequences

Sedimentation following M1 began within the middle to upper slope, where the M1 slope shallows from 8° to 10° to subhorizontal. Ms1 onlapping reflectors backstep up M1 to within ~6 km of the inner/outer wedge transition while coeval forestepping clinofoms extend ~8 km down slope to the base of a middle-slope anticlinal thrust (labeled F; Fig. 2), before a small abrupt seaward shift (M2; Fig. 2). The following Ms2–Ms4 backstep and upstep across M1 landward and coeval clinofoms forestep down slope, with Ms3 extending across anticlinal thrust F (Fig. 2). These early Middle Strata sequences generally stack vertically and are separated by small seaward shifts. Ms6 reflectors backstep to within ~1 km of the inner/outer wedge transition at which point they transgress the transition and forestepping clinofoms and then build over the transition (Fig. 2). The remaining sequences, Ms7 and Ms9–Ms12, feature successions of upstepping, levied channels across the upper slope (Fig. 3) and coeval upstepping and forestepping clinofoms across the middle and lower slope, although upstepping levies, or aggrading channels, are present locally throughout Middle Strata across the entire slope (Figs. 3 and 11). Ms8 consists of backstepping clinofoms well seaward of its earlier sequence counterparts (Ms1–Ms7), backstepping landward until the M9 unconformity (Fig. 2). Some lateral and/or seaward shifts are prominent, including M5, which truncates underlying Ms3 and Ms4 reflectors broadly, M8, which truncates Ms6 and Ms7 down slope up to ~15 km from the inner/outer wedge transition, and M9 and M10 (Fig. 2). The upstepping portions of all Middle Strata sequences within the upper slope (proximal to the steeply dipping portion of the M1 surface) consist of channels, levies, and channel fill (Fig. 3).

Generally, the earliest Middle Strata sequences, Ms1–Ms4, consist of successions of channelized backstepping and upstepping reflectors that transgress up the M1 event, with coeval forestepping and sometimes upstepping

Figure 10. The M1 horizons with an oblique perspective view looking SE. Meters below sea level (mbsl) contours and colors (same as Fig. 5) are overlain. Other labels are same as Figure 5 (including dashed black line as inner/outer wedge transition and dashed white line as shelf break). Lower left inset of semblance coefficient values (measure of similarity between traces) calculated with dip steering (structurally guided) over the outer shelf. Blacker values denote lower semblance (greater dissimilarity). Semblance values are used here to highlight normal fault networks cutting the M1 horizon. Lower-right subset is of channels cutting down the upper-slope portion of M1, highlighted by shading and by using an upslope perspective view (looking landwards). Note thrust folds G and H and the orientations of labeled anticlinal folds (WSW trending). C.P. – Cocos plate.

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Cobb Mountain reversal (Harris et al., 2013d). Shows that the bulk of Middle Strata sediments were deposited during the Strata unconformities, M5 and M8, to ca. 1.5 Ma and ca. 1.35 Ma. Thus, U1413 assuming a constant rate of sedimentation, we constrain two important Middle of the Matuyama chron (ca. 1.78–1.19 Ma and ca. 1.95–1.78 Ma; Fig. 9). As-straining Middle Strata to the Cobb Mountain reversal and Olduvai subchrons are thought to vary between 0 and 450 m (Fig. 8).

Middle Strata depositional sequences show poor continuity and high vari-ability in thickness along strike (Figs. 11 and 12), compared to the relatively continuous and consistent Lower Strata. Ms5–Ms12 sequences thin and termi-nate against slope anticlinal thrusts (Figs. 11). Middle Strata thicknesses within the slope reach up to ~1000 m, while shelf thicknesses are obscured due to data loss but reach up to ~1000 m, while shelf thicknesses are obscured due to data loss but extend up to >150 m thicknesses, and its correlative shelf conformity is obscured by data loss (Fig. 13). The U1 event delineates an unconformity that broadly truncates Middle Strata across the middle and upper slope and Lower Strata across the outermost outer shelf and down the lower slope (via channelization at great water depths). Its most landward erosional extent is coincident with the inner/outer wedge transition to the NW, and to the SE, is coincident with the “eastern dome” (long-lived Pleistocene domal uplift; Fig. 10). Subhorizon-tal erosion extends down to at least ~1.1 km bsl and channelized erosion down >1.9 km bsl, to within ~5 km of the trench (Figs. 13 and 14).

U1 3D geometry is broad and subhorizontal, paralleling the present-day seafloor. Its surface is relatively featureless, with subtle, short channels that run generally parallel to the slope. Small offset (<25 m), orthogonal, NW- and NE-striking normal faults cut the U1 event across the upper slope (Fig. 13). Its eastern upper-slope portion is deformed by landward- and seaward-dip-ping anticlinal thrusts (Figs. 13 and 15). Its southeastern middle-slope por-tion is overprinted by a major paleocanyon system that reaches ~350 m in relative depth (measured from bounding-wall heights) and extends >7 km in length (Figs. 13–15). The canyon head is ~600 m wide and significantly widens down slope to beyond the width of the CRISP survey (Fig. 15). The canyon is bounded by relatively steep walls (~13°; Fig. 14).

Considering these observations, it is likely that the U1 unconformity was formed by subaerial erosion, via wave-plain erosion or wave scour, extending across an area that now reaches down to ~1.1 km bsl. The paleocanyon system overprints the broad and flat wave-plain erosion and seems to have formed at a transient break, potentially a coastal break or a paleoshelf break, following subsidence, analogous to many small canyon or channel systems cutting into shelf breaks along Costa Rica (Fig. 1; von Huene et al., 2000).

Upper Strata Depositional Sequences

Upper Strata reflectors begin onlapping ~6–7 km down slope from the inner/outer wedge transition, between landward- and seaward-dipping anti-clinal thrusts labeled A and B (Figs. 3, 13, 15–17). These reflectors, labeled Us1, partially fill the synclinal depression until an abrupt seaward shift, designated U2 (Fig. 17). Subsequent reflectors, labeled Us2, continue to onlap local syn- clinal basins within the central to eastern portion of the volume (mostly be-tween anticlinal thrusts A–D) and on both the landward and seaward sides of the inner/outer wedge transition uplift (Figs. 2, 3, 13, 15, and 17). Onlapping reflectors then abruptly shift to ~3–4 km seaward of the inner/outer wedge transition and backstep over the crest of the anticlinal thrust A (Us3) (Fig. 17). An abrupt seaward shift truncates Us3 (U4), cutting Us3 into a lobate shape.
Overlying reflectors then stack vertically to the seafloor and generally upstep across the slope (Us4; Figs. 3 and 17).

Upper Strata sediments seem to be constrained to the slope, although data loss in the shallow portions of the shelf obscure any correlative conformities. Upper Strata are much thinner than Lower and Middle Strata, ranging from 0 to 500 m, with a median thickness of 94 m (Figs. 8, 12, and 15). Thicknesses away from anticlinal thrusts A–E, the inner/outer wedge transition, and the paleo canyon are limited to <150 m (Fig. 15). The thickest subbasin (up to 498 m) developed between a pair of landward- and seaward-dipping anticlinal thrusts A and B, with sediment thicknesses increasing to the east (Fig. 15).

Site U1413 drilled through ~180 m of Upper Strata, constraining these sediments to the Cobb Mountain reversal of the Matuyama chron, 1.78–1.19 Ma and younger (Fig. 9) (Harris et al., 2013d).

**Sedimentation Rates**

Based on IODP magnetostratigraphy and biostratigraphy (Vannucchi et al., 2011, 2013; Harris et al., 2013d) and CRISP sequence stratigraphy results, we broadly constrain sedimentation rates for each stratal domain.

**Lower Strata**

The L1 event, ca. 2.5–2.3 Ma (Vannucchi et al., 2013; Harris et al., 2013c), and M1 event, ca. 1.95–1.78 Ma (Harris et al., 2013d), constrain the duration of Lower Strata sedimentation to 0.35–0.78 m.y. This time frame, and the calculated isopach thicknesses for the shelf (up to ~1.3 km) and slope (up to ~1.6 km), constrain Lower Strata sedimentation rates for the shelf to ~1.7–3.7 km/m.y. and slope to ~2.1–4.6 km/m.y. These sedimentation rates are the highest for the margin during the Pleistocene. Furthermore, both shelf and slope rates are high and demonstrate that subsidence was broad and rapid across both domains.

**Middle Strata**

Middle Strata shelf and slope sedimentation rates are less constrained, due to the poor temporal constraint of the U1 event (1.78–1.19 Ma). We estimate that U1 occurred toward the tail end of the Cobb Mountain reversal by assuming a constant sedimentation rate, approximating the age to 1.3 Ma (Fig. 9), constraining the duration of Middle Strata sedimentation to 0.48–0.65 m.y. With that esti-
Figure 13. The U1 horizon with an oblique perspective view looking SE. Inline 2630 is shown for reference. Colors and labeled contours on U1 denote meters below sea level (mbsl); blues to purples are greater depths; and greens to yellows to reds are progressively shallower depths. Anticlinal thrusts are labeled with white letters. Thick white dashed line across the middle slope denotes the down-slope extent of wave-plain erosion. Inset is of the western upper slope and displays normal faults (oblique to perpendicular to the slope) and channels (parallel to the slope) by semblance coefficient values. Blacker values denote lower semblance (greater dissimilarity). Note the subtle, parallel shadows with varying orientations, linear to arcuate ridges that strike oblique to perpendicular to the slope and are highlighted by shading. These are small offset normal faults cutting the U1 horizon (some labeled with n.f.). Note the paleocanyon extending down slope from the middle to lower slopes.

Figure 14. Crossline 3892 showing the initiation of anticlinal thrust F after M8 and G likely before L1, L1, M1, and U1 erosional events are shown with solid white lines. The M8 unconformity is shown with a dashed white line. Solid black arrows show onlapping reflectors (uncertainty is shown with question mark). One-sided arrows denote base of some normal faults, in this case likely associated with folding. Anticlinal thrust G likely initiated before L1 and continued activity through Lower Strata and early Middle Strata. Note the paleocanyon (down-slope extent of U1) cutting Middle Strata.
mation, and with observed maximum Middle Strata shelf and slope thicknesses of ~450 m and ~1000 m, we estimate a shelf rate up to 0.69–0.94 km/m.y. and slope rate up to ~1.5–2.1 km/m.y. The decline from Lower to Middle Strata in the shelf and slope sedimentation rates is approximately tenfold to half.

**Upper Strata**

Because the U1 unconformity is only imaged across the slope, we can only constrain Upper Strata (ca. 1.3 Ma–present) sedimentation rates for the slope. Upper Strata thicknesses reach up to ~500 m (between anticlinal thrusts A and B), and with an estimated U1 age of 1.3 Ma, the sedimentation rate is up to ~0.38 km/m.y. However, away from young anticlinal thrusts and the paleocanyon system, thicknesses are generally <150 m, thus sedimentation rates are generally <0.12 km/m.y., an order of magnitude less than Middle or Lower Strata. Thus, slope sedimentation rates tapered off quickly following U1, with sedimentation only occurring in the spaces created by active faulting and folding continued (i.e., eastern portion of the upper-slope and down-slope portions of middle and lower slopes; Fig. 15).

**Faulting**

**Thrust Faulting**

The subduction thrust appears to serve as a source for a series of imbricate, landward-dipping thrust faults that extend from the toe to the inner/outer wedge transition (Fig. 16). Thrust faults are observed to have a greater density and to accommodate greater offset nearer to the toe, or deformation front, and toward the SE portion of the volume, i.e., along strike, within the middle and...
upper slope (Fig. 16). Often paired with these landward-dipping thrust faults are seaward-dipping thrust faults (e.g., anticlinal thrusts A–D), which accommodate greater offset toward the inner/outer wedge transition and to the SE portion of the volume (Figs. 16–18). Deformation accommodated by thrust faulting has resulted in inclined folding, local uplift and subsidence, with generally tighter anticlinal and broader synclinal hanging-wall thrusts (Fig. 17). Middle- and upper-slope fold axes E–H are WSW trending, while fold axes A–D are W trending (Figs. 5 and 10). Both fold trends are oblique to the NW-trending trench axis, lower-slope ridges (seafloor expression of frontal prism imbricated and thrusted slices), and inner/outer wedge transition (Kluesner et al., 2013).

The inner/outer wedge transition is coincident with the shelf break along the western portion of CRISP, and it deviates from the indented or deflected portion of the shelf break along the central and eastern portions (Fig. 5, 7, 8, 10, 12–13, 15, 19). The outer wedge, generally coincident with the slope, is characterized by a series of imbricate thrust faults and paired conjugate faults, while the landward portion, delineated by the shelf, is mostly absent shortening in Pleistocene strata (Fig. 5). Interestingly, prominent Pleistocene subsidence is observed across both the inner and outer wedges (Figs. 2 and 19).

The timing of middle- and upper-slope thrusting and folding (hence local uplift and subsidence) postdates early, rapid, broad subsidence and seems to have occurred in pulses of activity. The earliest Pleistocene shortening that is concomitant with earliest Lower Strata deposition is the inner/outer wedge transition, where there is observable thinning of Ls1 across the structure.

Figure 16. Inlines 2090 (top) and 2625 (bottom) are shown to demonstrate increased shortening to the SE, nearer to the impinging Cocos Ridge. L1, M1, and U1 unconformities are shown with white dashed lines. Thrust faults are shown with dashed yellow lines and are labeled. The plate boundary (décollement) is labeled pb. Trench is on the right side; shelf is to the left. Shortening is calculated relative to L1 change in length. Note the decrease in thrust spacing, increase in fault dip, the steepness, singularity, and amount of offset of inner/outer wedge transition toward the SE (inline 2625), and the formation of seaward-dipping thrusts to the SE (inline 2625).
within the eastern portion of the volume (Figs. 6 and 7). The central and western portions of the inner/outer wedge began shortening soon after (e.g., Ls2 thinning; Figs. 6 and 7). There was a small degree of middle- and upper-slope shortening down slope during Lower Strata deposition, as seen by several slope basin inversions (e.g., Ls2 and Ls4 in Fig. 17). Middle-slope thrust F likely initiated during Ls1, because there is observable Ls1 thinning over its crest, and the overlying Ls2 and Ls3 forestep its inclined back limb (Figs. 2, 7, and 8). Middle-slope thrust D initiated during the later stages of Lower Strata, as early Middle Strata (Ms1–Ms3) sequences onlap and prograde over its back- and forelimbs (Fig. 18). Thrusts A–C postdate early Middle Strata (Ms1–Ms4) and seem to have initiated after the M5 unconformity (ca. 1.5 Ma; Figs. 11 and 18), and thrust E, after M8 (ca. 1.35 Ma; Fig. 18). Interestingly, thrusts A–D (which are two paired landward- and seaward-dipping anticlinal thrusts), all seem to have had a coeval pulse of activity following M5 (ca. 1.5 Ma), and thrusts E and F, following M8 (ca. 1.35 Ma; Figs. 11, 14, and 16–18). Anticlinal thrusts A–F all bound the thickest Middle Strata basin, and thrusts D, F, and G bound an adjacent, thick, down-slope Middle Strata basin, all WSW trending (Fig. 12). Thrusts G and H were active during Middle Strata and were likely active during Lower Strata deposition, because Lower Strata taper out across their crests and limbs. Continued shortening during Upper Strata deposition was first accommodated by thrusts A and B, marked by Us1 (Fig. 17). Subsequent shortening was accommodated by thrusts A–E and the inner/outer wedge transition (Figs. 13, 15, 17, and 18). Thrusts A and B bound the thickest Upper Strata basin (Fig. 15). Thrusts G and H are not well constrained during Upper Strata due to BSR interference, although it is likely they were and are active.

**Normal Faulting**

Moderately to steeply dipping discontinuities that accommodate normal motion are observed across the slope and shelf (Figs. 6, 7, 10, 11, 13, 14, 17, and 18). These normal faults cut pervasively through all Pleistocene stratal domains and appear as distinct groups or networks (Figs. 10, 13, 17, and 18), suggesting that extension is episodic and occurs locally. These fault networks consist of intersecting arrays or tightly spaced parallel segments of varying geometries (Figs. 10 and 13), suggesting that stresses spatially and temporally vary.

Normal faults can be spatially associated with the hinges of anticlinal folds (Figs. 11, 17, and 18) but are also prevalent across the shelf (Fig. 10), upper slope (Figs. 10 and 13), and possibly even the M1 erosional event. Normal fault activity and its control on sediment accumulation is minimal, with offsets <100 m (Figs. 6, 11, 14, and 18), demonstrating that normal faults contribute little to vertical motions. Growth strata associated with normal faulting are rare, suggesting that these faults are not long lived. Normal faults are shallow and do not generally reach into the margin wedge or even rarely into Ls1 (Figs. 6, 11, 14, 17, and 18).
Figure 18. Four inlines shown across the slope to demonstrate the initiation of anticlinal thrusting. Solid white lines show L1, M1, and U1 erosional events. Dashed white lines denote important unconformities (M5 and M8). Solid black arrows denote onlapping clinoforms. One-sided arrows denote the base of prominent normal faults. Inline 2480 (upper left) shows the initiation of thrusts C and D following the M5 sequence boundary. Inline 2555 (upper right) shows the initiation of D after M1 and then another pulse of shortening following M5, continuing through Middle Strata. Thrust G initiation is less clear, but it seems pre-dated L1 have been active through both Lower and Middle Strata. Inline 2370 (lower left) shows the initiation of E following M8. Inline 2280 (lower right) shows the initiation of F after M1 and continued during Middle Strata deposition.
INTERPRETATION

Early Pleistocene Outer Forearc Erosion and Subsidence

Panama Fracture Zone Subduction and Plate Kinematic Changes

Recovered margin wedge sediments from U1379 and U1380 demonstrate that before the L1 erosional event, the region was a marine basin filling with terrigenous sediments. Fossils and lithofacies from Vannucchi et al. (2013) suggest middle bathyal water depths (800–1200 m) for the present-day outermost shelf. Thus, to account for rapid uplift and broad extent of Early Pleistocene outer forearc (shelf and slope) subaerial erosion recorded by L1, we cite the spatio-temporal link of the southeastward PFZ triple-junction migration across the CRISP portion of the forearc ca. 2–4 Ma (age depends on the projected paleo-orientation of the PFZ and on which velocity model is used; Fig. 1; DeMets et al., 1990; DeMets et al., 1994, 2010; Kobayashi et al., 2014; Morell, 2016). How much sediment was removed is not constrained, but the broad extent of surface erosion (across the shelf and down to the middle slope) and deeply incised fluvial drainages (up to ~500 m) across the shelf suggest significant thinning of the margin wedge (Fig. 20). There are several present-day analogue L1 erosional environments, including the Burica Peninsula, inboard of the PFZ, and Coiba and Jicarita islands, inboard of the Coiba Fracture Zone. Furthermore, Morell et al. (2011) found a temporal link between marine terrace uplift across the Burica Peninsula and PFZ SE migration. We suggest that at CRISP, the SE migration of the irregular and steep PFZ, or a related precursor to the PFZ, drove rapid, transient uplift and subaerial erosion at CRISP beginning ca. 2.5 Ma (Fig. 20). This subaerial erosion thinned the margin wedge, and eroded sediments likely ended up in the trench, due to the transient steepness of the slope, where they were carried and/or subducted to the SE along with the PFZ.

To account for rapid and broad Early Pleistocene subsidence following uplift, we point to the continued SE migration and passage of the PFZ triple-junction and associated bathymetric, crustal thickness and plate kinematic changes. The short time frame (~0.5 m.y.) associated with robust subsidence (~1.6 km within the slope and ~1.3 km within the shelf) resulted in the highest sedimentation rates (shelf ~1.7–3.7 km/m.y. and slope ~2.1–4.6 km/m.y.) during any Pleistocene time frame, and thus makes it likely that the passage of...
the steep topography of the PFZ was an important driver of rapid subsidence (since the margin was eroded across its surface, thinned broadly, and subsided quickly). However, the threefold increase in relative convergence and decrease in crustal thickness, and the potential downward flexing of the subducting plate under the outer forearc (Furlong et al., 1982; Buiter et al., 2001), likely contributed to broad and continued subsidence as well (Lower Strata are laterally continuous and include large transgressive periods, >35 km, Ls1, and regressive shifts, >30 km, L2). To what degree each mechanism drove subsidence is not constrained, but the bulk of the observed net margin Pleistocene subsidence occurred during this Early Pleistocene ~0.5 m.y. window. Furthermore, the maintenance of the present-day position of the inner/outer wedge transition through the Pleistocene (Figs. 2, 5–8, 10, 12–13, 15, 19, and 20), including during rapid Early Pleistocene subsidence (Figs. 5–8), shows that the margin did not retreat landward during subsidence, as modeled by basal erosion and suggests a link with plate boundary dynamics that do not require upper-plate net material loss.

The spatiotemporal link of the PFZ triple-junction migration and coincident outer forearc subsidence is also found offshore the Nicoya Peninsula (McIntosh et al., 1993). The same triple-junction migration–outer forearc subsidence link is observed along other convergent margins, including Japan, Izu-Bonin, Mariana, and Tonga, although backarc extension is also observed (Regalla et al., 2013).

Pleistocene Slope Erosional Events

Rapid Early Pleistocene subsidence is overprinted by two younger slope erosional events, the M1 and U1 events. These two large slope erosional events truncated and removed thick sections of strata, have erosional spatial footprints that extend across the entire slope, and they shifted or altered the existing states of sedimentation (Figs. 2, 10, and 13). Each event differs in character, thickness of sediments removed, and the amount of ensuing slope subsidence, but both M1 and U1 unconformities require anomalous mechanisms, because they both fall outside the window of relative base-level fall and marine unconformities within the sequence stratigraphy model (eustasy, tectonics, and sedimentation).

Seamount Subduction

Several observations yield insight into what may have been the driving mechanism for the earlier M1 unconformity, ca. 1.95–1.78 Ma, including: M1 has a steeply dipping upper-slope geometry (8°–10°) that cuts down section to Ls4, removing ~900 m thickness of Lower Strata (Fig. 2), and it extends across the middle and lower slopes to the landward extent of the frontal prism, within ~5 km of the trench (Lower Strata included; Fig. 10). It therefore almost extends across the entire slope (excluding the frontal prism), ~30 km in down-dip length, and it extends across the width of CRISP volume, ~11 km. These observations suggest that this erosional event was precipitated by a collapse
of the slope and mass transport to the trench. We interpret that this collapse was driven by underthrusting topography that removed a portion of the lower slope and subsequently oversteepened the middle and upper slopes, analogous to the observed slope bite marks, collapse, and subsequent infilling seen along the Costa Rican margin (Fig. 1). These analogous slope bite marks are linked to seamount subduction (e.g., von Huene et al., 2000).

Several other observations also suggest slope oversteepening and collapse. M1 is cut by a dense, intersecting array of NW-, W-, and NE-striking normal faults, with offsets up to ~100 m; these faults seem to be temporally linked. This suggests that the slope area was under an episode of extension, possibly during uplift (e.g., Dominguez et al., 1998), or alternatively during deep-seated down-slope mass movement accommodated by fault block rotation (McIntosh et al., 1993). The thoroughly channelized M1 upper slope and overlying aggrading Middle Strata channels suggest that once the seamount passed, the upper-slope and/or shelf-edge water depth increased rapidly, and the steep slope gradient served as a pathway for sediments headed down slope, some likely refilling the deflected trench.

If we use the present-day water depths of the upper slope (300–700 m) and the approximate maximum thickness of Lower Strata removed (~900 m), we find a minimum seamount height of 1.2–1.6 km above surrounding bathymetric depths, if any of the slope was subaerially eroded (Fig. 20). This height does not consider downward flexure due to an increased overlying load (i.e., isostasy), extension and thinning of Lower Strata during uplift (although this would be minor), or offscraping of seamount material during frontal prism and/or lower-slope underthrusting. Many analogous conical seamounts of up to ~2 km heights are observed within the seamount zone of CNS generated crust seaward of the present-day trench (von Huene et al., 2000). These conical seamounts will impact the margin at some point, perhaps producing analogous local slope unconformities. However, we must note that if there was no subaerial erosion, we do not have a constraint on its minimum height, other than that it must have been of sufficient size to oversteepen the slope and drive slope collapse. Importantly, we observe that the thickness of middle- and upper-slope material removed is proportional to subsequent sedimentation (M1 truncated ~900 m of Lower Strata and subsequent filling with Middle Strata reach up to ~1000 m), suggesting that this discrete slope erosional event did not contribute substantially to middle- and upper-slope net subsidence (up to ~2.45 km; Fig. 19).

**Plateau Subduction**

We interpret the later slope unconformity, U1, to also be due to underthrust seafloor topography. However, the nature of surface erosion and subsequent subsidence was notably different than the M1 event. The U1 unconformity is relatively flat, parallelising the seafloor. Its 3D surface is generally smooth, with only subtle and spare channels, small normal fault offsets, and gentle uplift and subsidence within the central to eastern portion due to younger landward- and seaward-dipping anticlinal thrusting (including the inner/outer wedge transition thrust). We interpret these observations to mean that the underthrust topography did not remove the lower slope and initiate middle- and upper-slope collapse. The middle- and upper-slope rise and fall was coherent and broad and produced an unconformity that is dipping parallel to the seafloor. The rise and fall also did not result in new accommodation space for sediments (Upper Strata sedimentation rates are generally <0.12 km/m.y.). This parallelism and subsequent lack of sedimentation suggest that the underthrust topography was broad and flat, with shallowly sloped sides. We observe that U1 removed up to ~150 m, likely via subaerial wave-plain erosion, with the greatest amounts of erosion within the upper slope. Inferred wave-plain erosion continues down to depths of ~1.1 km bsl (approximate seaward limit). This places a minimum underthrust topographic height of ~1.1 km, not accounting for the same unknowns as the M1 seamount height.

We interpret U1 as being due to the underthrusting and migration of a broad, flat, and shallowly sloped oceanic plateau, >1 km high, ca. 1.3 Ma (Fig. 20). The underthrusting and migrating plateau would propagate a wave of forearc uplift without removing the lower slope and driving slope collapse (Martinod et al., 2013; Zuemann and Hampel, 2015). An analogous seafloor feature is the NE-trending Quepos Plateau, which is generally ~1.5 km high and ~35 km wide (von Huene et al., 2000). Projecting the current NE trend of the Quepos Plateau through the trench and reconstructing relative Cocos Plateau motion, a proto–Quepos Plateau, or some related feature, possibly underthrust and migrated across the CRISP portion of the margin ca. 1–2 Ma.

**Pleistocene Slope Shortening**

Rapid Early Pleistocene subsidence and the two major slope erosional events are conflated with progressive slope shortening. Observed middle- and upper-slope shortening progressed during younger Middle Strata deposition (Figs. 11–12, 14, and 18) and Upper Strata deposition (Figs. 15–18) and increases drastically to the SE, resulting in a relative counterclockwise rotation of fold axes and local SE Upper Strata basins (folds A–H in Fig. 10; Fig. 15). We find that the shortening is decoupled from net subsidence and thus invoke another prominent incoming topographic feature as driving middle- and upper-slope shortening.

**Cocos Ridge Subduction**

The subduction of the Cocos Ridge has been cited as the driver of widespread Costa Rican tectonic erosion and outer forearc subsidence (e.g., Vannucchi et al., 2003). However, recent studies have shown the initiation of Cocos Ridge subduction (2–3 Ma) to be temporally inconsistent with outer forearc subsidence offshore Nicoya, where rapid subsidence began ~5–6.5 Ma (e.g., Vannucchi et al., 2001; Morell, 2015; Morell, 2016). We observe that middle- to upper- slope shortening was low or absent (changes slightly along strike) during rapid Early Pleistocene outer forearc subsidence, with folding and thrusting developing after both L1 (ca. 2.5–2.3 Ma) and M1 (ca. 1.95–1.78 Ma)
erosional events (middle- and upper-slope shortening generally postdates M5 and M8 unconformities, ca. 1.5 Ma and ca. 1.3 Ma; Figs. 9, 17, and 18). We also observe that middle- and upper-slope shortening drastically increases to the SE (Fig. 16), and the youngest, most active thrusts are essentially localized to the SE portion of CRISP, nearest to the impinging Cocos Ridge, resulting in local, SE Upper Strata basins up to ~500 m thick (between folds A and B; Figs. 15 and 17) and rotation of middle- and upper-slope fold axes (WSW trending) relative to the inner/outer wedge transition, frontal prism thrust faults, and trench axis (NW striking and NW trending; see folds A–F in Fig. 5 and A–H in Fig. 10). These observations imply two important Cocos Ridge findings: (1) the Cocos Ridge is driving progressive slope (outer wedge) shortening at CRISP, starting ca. 1.5 Ma, as it slowly migrates from the SE to the NW along the trench with continued subduction (its geometry is slightly clockwise from relative plate motion; Fig. 1), and (2) broad, net outer forearc subsidence is not linked to the Cocos Ridge impact with the trench nor its continued subduction (e.g., Upper Strata, which account for the past ~1.3 m.y. of slope sedimentation, generally have low sedimentation rates of <0.12 km/m.y. with average thicknesses of <150 m; Fig. 15).

## CONCLUSIONS

Three-dimensional seismic sequence stratigraphy mapped within the CRISP seismic-reflection volume offshore Costa Rica spatiotemporally constrains several Pleistocene outer forearc processes, including: sedimentation rates and patterns, vertical motions, and shortening. Three significant shelf and/or slope erosional events at ca. 2.5–2.3 Ma, 1.95–1.78 Ma, and 1.78–1.19 Ma, each with notable differences in spatial extent, volume removed, and subsequent margin response, caused abrupt shifts in sedimentation patterns that fall outside the depositional sequence model. These shifts, coupled with vertical motions and observed shortening, reveal three primary mechanisms for Pleistocene to present shelf and slope sedimentation patterns: (1) regional subaerial erosion and rapid subsidence linked to the southeastward Panama Fracture Zone triple-junction migration and associated abrupt bathymetric variations and plate kinematic changes; (2) transient, kilometer-scale uplift and subsidence due to inferred subducting plate topography passing through the subduction system; and (3) progressive outer wedge shortening accommodated by landward- and seaward-dipping thrust faults and fold development due to the impinging aseismic Cocos Ridge. Furthermore, we find that the present-day wedge geometry (to within ~3 km along strike) has been maintained through the Pleistocene, in contrast to modeled landward margin retreat. We also observe that deformation, i.e., extension and shortening, is decoupled from net margin subsidence. Our findings do not require basal erosion, and they suggest that the vertical motions of the Costa Rican outer forearc are not the result of a particular continuous process, but rather are a summation of plate to plate changes (e.g., passage of a fracture zone triple junction) and episodic events (e.g., subducting plate topography).

We have demonstrated that vertical motions and deformation offshore Costa Rica are driven in pulses of activity. Importantly, we observe that most of the Pleistocene net outer forearc subsidence occurred within an ~0.5 m.y. window during the Early Pleistocene, coincident with a migrating fracture zone and plate kinematics changes, analogous to recent findings offshore Japan. We also link anomalous slope unconformities with subducting topography and variable slope responses (e.g., M1 and Middle Strata versus U1 and Upper Strata) to topography size and shape (e.g., seamount versus plateau).

We document middle- and upper-slope shortening that postdates rapid Early Pleistocene outer forearc subsidence and that shortening increases to the SE and drives counterclockwise rotation of middle- and upper-slope fold axes, providing a clear link to the impinging and slowly NW migrating Cocos Ridge. These findings demonstrate the advantages of 3D seismic-reflection imaging, including having the ability to constrain the relative timing and rates of sedimentation, vertical motions, and shortening, and coupled with drilling, can provide very well resolved basin and margin histories.

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