Chapter 5
Geomorphological indicators of past sea levels

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5.1 INTRODUCTION

Geomorphic indicators of past sea levels represent a continuum of processes and landforms associated with sea level and sea-level change. In an attempt to honor this continuum, while at the same time discussing geomorphic features separately, this chapter discusses geomorphic indicators of past sea level in the following sequence. First, coral as a geomorphic indicator of past sea level is discussed. The recognition of late Pleistocene uplifted coral platforms as indicators of past sea level was one of the pioneering efforts in sea-level research. Decades later, coral was again the focus of pioneering work in late Holocene sea-level research when coral microatolls were used to reconstruct sea-level history to a much higher resolution of centimeters of sea-level change over the scale of years. Erosional landforms that indicate former sea levels are then discussed, starting with late Pleistocene geomorphic features, marine terraces, and shoreline angles, before describing the more highly resolved sea-level records that can be gleaned from similar features in the late Holocene through the use of the seacliff–shore platform junction. Tidal notches form another set of erosional indicators that are discussed. Finally, the suite of geomorphic landform indicators that are deposits are discussed. Coastal deposits were initially laid down near sea level and such deposits, if subsequently vertically or horizontally separated from the shoreline, may indicate past sea levels. The discussion on deposits focuses on beach ridges.

For each set of geomorphic indicators, the accuracy with which the geomorphic feature indicates a past sea level is discussed. "Indicative meaning" defines the relationship of the geomorphic feature to tidal range and thereby allows the relative sea-level change to be measured by differencing the elevation of the geomorphic feature from a specific reference water level (e.g., mean high tide or some other tidal datum) that formed the feature initially (van de Plassche, 1986). For geomorphic features that indicate a past sea level, knowledge of the tidal datum that the geomorphic feature represents (e.g., mean high tide) and the resolution at which the geomorphic indicator forms at that elevation is required. As an example, consider a survey that shows that a shore platform–seacliff junction (a sea-level indicator) eroded into bedrock forms at about mean high spring to mean high neap tide level. Such an indicative meaning, ranging over meters, is not precise compared to resolving past sea level using biostratigraphic data of sediments, but relative lack of precision is usually the case with geomorphic indicators. The exception is the instance where the geomorphic indicator is an organism (e.g., a coral microatoll) rather than an eroded landform.

Geomorphic indicators of former sea level are overwhelmingly concentrated on coasts that have undergone relative sea-level fall over a time span ranging from the last few thousand years to over hundreds of thousands of years. The reason for the bias is that relative sea-level rise will drown geomorphic features created by former sea levels, and these indicators become submerged, buried by sediment, and/or eroded. In some instances, submerged indicators may remain preserved; an example is a submerged tidal notch. Coastal sediments that aggrade within embayments are employed for sea-level reconstruction for the case of rising sea level in the last few thousand years, and the key role provided by coastal sediments provides good justification for Chapter 4 being devoted entirely to these deposits.

Preservation of geomorphic indicators of former sea level is usually the result of a period of

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slowly changing or static relative sea level, when deposits or erosional landforms are created, followed by a relatively rapid relative sea-level lowering that subsequently subaerially preserves these deposits or erosional landforms. An example would be a shore platform formed in the tidal zone over hundreds–thousands of years that is subsequently coseismically uplifted 1–2 m during an earthquake. Other examples are provided in Sections 5.5–5.7. However, deposits recording static or slowly changing sea level over time can be preserved in cases of coastal progradation along sediment-rich coasts.

Rate of change of relative sea level affects the robustness of geomorphological features indicative of former sea level. In the case of eroding (non-prograding) coasts, a static or very slowly falling sea level will result in the concentration of wave energy and weathering at the same shoreline elevation over geomorphologically significant durations of time (at least thousands of years) such that shoreline features can form and then be preserved if relative sea level falls. In contrast, in the case of ongoing, rapid-falling relative sea level (≥50 mm a^{-1}), preservation of geomorphic indicators of former relative sea level may be unlikely. In high-latitude regions (e.g., Norway, Sweden, Spitzbergen, or the Hudson Bay region) however, beach ridges or shore platforms may form and survive under rapidly falling relative sea level; see Section 5.6.3 on preservation of geomorphic indicators of past sea levels at high latitudes.

5.2 MODEL FOR FORMATION OF LATE PLEISTOCENE EMERGENT SHORE PLATFORMS

Shore platforms form at times when the rate of glacioeustatic sea-level change is equal to the rate of local surface uplift (Mesolella et al., 1969; Bradley and Griggs, 1976). From the definition provided by Milne and Mitrovica (2008), the expression "glacioeustatic sea-level curve" refers to the time-dependent "spatially uniform height shift of the ocean surface to accommodate any mass gained/lost from grounded ice." The relative sea-level change at any specific coastal site is also affected by glacial isostatic adjustment (GIA) that causes the observable sea-level change to be different from the eustatic curve (Milne and Mitrovica, 2008). At such time that the rate of glacioeustatic sea-level change is equal to the rate of local surface uplift (and the rate of sea level change is modified to some extent by GIA), there is a relative sea-level stasis and wide platforms develop.

The details of platform formation versus the tempo of glacioeustatic sea-level rise and subsequent peaking at a highstand are slightly different for erosional marine terrace formation versus constructional reef terrace formation (see fig. 2 in Muhs et al., 2002), but both shore platform types form at, or temporally near, highstands. Upon successive sea-level fall brought on by glacioeustatic sea-level change, the platforms emerge and continue to rise tectonically. By the time of formation of a succeeding glacioeustatic sea-level highstand, the platforms are well above the reach of sea level. In such a manner, topographic "stair steps" of emergent marine platforms develop on both constructional (i.e., coralline; Mesolella et al., 1969; Chappell, 1974) and erosional (i.e., mid latitude; Bradley and Griggs, 1976; Lajoie, 1986) coasts (Fig. 5.1a, b). In the case of the coralline terraced landscape of the Huon Peninsula of New Guinea (Fig. 5.1a), the stair-stepped terraces have been dated and a sea-level curve for the highstands constructed (the "New Guinea sea-level curve") using known ages, highstand elevations, and a constant uplift rate (Chappell and Shackleton, 1986).

For raised platforms the emergent stair step consists of a tread, which is the paleo-shore platform, and a riser landward of the tread, which is the paleo-seacliff; "shoreline angle" or "platform inner edge" are both terms for the junction of the paleo-seacliff and the paleo-platform. In the case of raised late Pleistocene shore platforms, this junction is invariably covered by marine deposits and by colluvium shed off the adjoining seacliff (in Fig. 5.1b note that the platform inner edges in all cases are buried by marine deposits and colluvium). The elevation of the raised junction therefore cannot usually be measured directly but must be inferred by subtracting an assumed, or known, thickness of cover sediment. By contrast, junctions are well exposed in modern coastal settings and for some raised late Holocene platforms. These cases are discussed in Section 5.6.2.

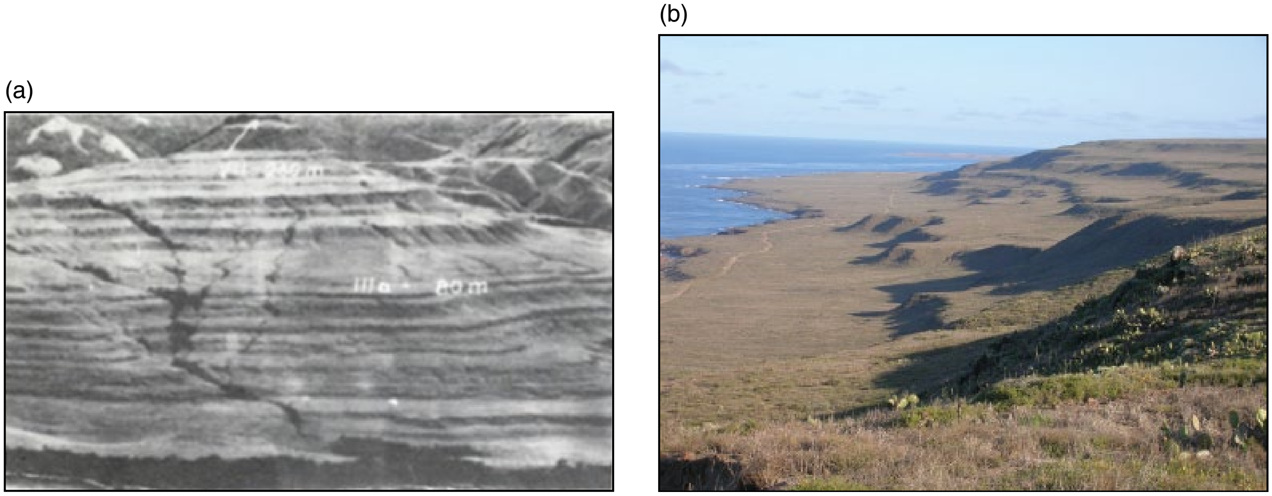

Fig. 5.1. (a) Coralline terraced landscape of Huon Peninsula, New Guinea (6°05' N, 147°34' E). *Source*: Reproduced from Chappell (1974, fig. 2). The white lettering on the image is the original annotation of Chappell (1974) and refers to his terrace identifying number and the elevation of the terrace in meters. *Source*: Reproduced with permission of the Geological Society of America. (b) Erosional marine terraced landscape exemplified by marine terraces on the west coast of San Clemente Island, California (32°53' N, 118°31' W) looking northward (Muhs et al., 2002). The most extensive terrace, in the foreground and middle ground on the left of the photo, is the c. 120 ka marine terrace. The youngest and lowest terrace, limited in extent and minimally exposed along the embayed coastline in the middle ground, is the 80–100 ka marine terrace. The higher and older terraces, in the center and right on the photograph, are mid-Pleistocene marine terraces (200 ka and older). *Source*: Photograph by Daniel Muhs. Reproduced with permission. For color details, please see Plate 8.

5.3 GEOMORPHIC INDICATORS OF PAST SEA LEVEL AND GLACIAL ISOSTATIC ADJUSTMENT IN THE NEAR AND FAR FIELD

Raised platform inner edges on stair-stepped terraced coastlines form a unique dataset comprising a set of elevations for inner edges, where each inner edge corresponds to a specific occurrence of a global sea-level highstand. Assuming that the tectonic uplift rate responsible for raising the platforms is constant on a given coast and that the New Guinea sea-level curve (Chappell and Shackleton, 1986) is a global sea-level curve that provides relative elevation of these highstands, then for any given stair-stepped marine terrace sequence, a set elevations of platform inner edges uniquely corresponds, one for one, to a set of global eustatic sea-level highstands (e.g., Lajoie, 1986). Following this reasoning, Bull (1985) proposed that by knowing the elevation of shoreline angles of a particular marine terrace flight, one could graphically match a set of terraces one-for-one to a set of specific-aged highstands using an assumed constant uplift rate. The flaw to this reasoning is most elegantly articulated by Lambeck and Nakada (1992) and Muhs et al. (2012, p. 20), who argue against the concept of a global sea-level curve. The flaw entails not including another variable: glacial isostatic adjustment (GIA).

Although a review of GIA is not within the scope of this chapter (and is thoroughly covered elsewhere; e.g., Nakada and Lambeck, 1989; Mitrovica and Peltier, 1991; Milne and Mitrovica, 2008), GIA is a response to load transfer of glacier ice to ocean basins through melting that perturbs the mantle both in the near field, which is proximal to the ice sheets, and in the far field, which includes the equatorial regions that are removed for the immediate effects of ice mass redistribution. In addition, load transfer of glacier ice to ocean basins perturbs the mantle differently beneath coastlines of major continental masses as opposed to beneath coastlines of oceanic islands. The net result is that the factor of global isostatic adjustment varies in space and in time. Although uplifted shoreline angles are geomorphic relics of former sea-level highstands of glacioeustatic sea-level fluctuations, two identically aged shoreline angles at different parts of the globe will have elevations dependent both on local tectonic uplift rate and local glacial isostatic adjustment effects on the immediate coastline. Changes in ocean water temperature and salinity also lead to departures from a global sea-level curve, but these departures are relatively minor compared to GIA (Milne and Mitrovica, 2008).

5.4 GEOMORPHIC INDICATORS OF PAST SEA LEVEL AND THE ISSUE OF INHERITANCE AND REOCCUPATION

Geomorphic indicators are here separated into late Pleistocene and late Holocene as the latter should have more resolved indicative meaning. The late Holocene indicator was formed within a few thousand years of the present time and the well-exposed modern shoreline is an analog for the genesis and formation of the late Holocene indicator landform. However, is the geomorphic indicator late Holocene or late Pleistocene in age? The nature of glacioeustatic sea levels is that, in general, each glacioeustatic sea-level highstand (marine oxygen isotope stages 5, 7, 9, etc.) is about the same height as the previous. For example, on the tectonically stable Florida Keys, reefs dating to marine oxygen isotope stage 5 (~120 ka) and 7 (~200 ka) apparently had paleo-sea levels that were only slightly different from each other, although both were slightly higher than present (Muhs et al., 2011). On coasts with little tectonic uplift, late Holocene sea-level highstands could therefore reoccupy (inherit) the same elevation as the late Pleistocene sea-level highstand, and investigators should be aware of this possible scenario for modern shore platforms (Fig. 5.2).

As a consequence, geomorphic features created in the late Pleistocene can be reoccupied and modified in the late Holocene by sea levels that reach the same elevations (Phillips, 1970; Trenhaile, 1972; Kelsey, 1990; Kelsey and Bockheim, 1994). Notches and shore platform–seacliff junctions can therefore both be polygenetic in that a Holocene relative sea level may occupy a landform (platform or notch) that is actually inherited from the late Pleistocene. Field sea-level researchers need to be aware of this potential, and the coastal site illustrated in Figure 5.2 is a potential case in point.

Fig. 5.2. The seacliff–shore platform junction at Drake's Beach, Point Reyes Peninsula, California at low tide (38°01.5' N, 122°58' W). The platform is prominently displayed, as are the tide drainage channels cutting the platform. Platform and seacliff eroded in siltstone and sandstone of the Miocene–lower Pliocene Purisima Formation. The wide, prominent modern platform, the lack of any older uplifted platforms in the landscape, and the low uplift rates along this section of the coast adjacent to the San Andreas fault all allow for the possibility that the platform may have been originally formed in the late Pleistocene and then inherited and reoccupied in the late Holocene. *Source*: Photograph by S. M. Cashman. Reproduced with permission. For color details, please see Plate 9.

5.5 CORAL AS A GEOMORPHIC INDICATOR OF PAST SEA LEVEL

5.5.1 Constructional (low latitude, coralline) late Pleistocene marine shore platforms

Growth of a coralline fringing reef or a shallow lagoon with a coralline outer barrier ridge produces constructional shoreline platforms. Such platforms are recognized in many low-latitude coastal environments where corals thrive.

In cases where tectonic uplift is ongoing, coralline shore platforms are preserved as stair-stepped coralline terraced landscapes (Fig. 5.1a). Using radiometric dating techniques for coral, such terraced landscapes have been used to date former glacioeustatic highstands of sea level.

Two of the earliest pioneering and well-documented examples of paleo-sea-level histories using terraced coralline landscapes are Barbados and Papua New Guinea. Barbados is a small island (431 km^2) on the eastern margin of the Caribbean Sea with a terraced coral cap that overlies deformed Tertiary strata (Mesolella et al., 1969; Radtke and Schellmann, 2004). Each individual terrace represents an elevated reef tract. The raised terraced reef tracts were assumed to represent major eustatic sea-level highstands and radiometric ages define the times of these highstands (Mesolella et al., 1969, Radtke and Schellmann, 2004). On the Huon Peninsula of New Guinea (Fig. 5.1a), the stair-stepped, terraced landscape consists of an off-lapping sequence of more than 20 coral reefs, and dating of the reefs enabled inference of a eustatic sea-level curve for the late Pleistocene (Bloom et al., 1974; Chappell, 1974; Chappell and Shackleton, 1986).

The New Guinea record provides a good example of the level of accuracy of sea-level reconstruction possible for late Pleistocene sea levels. On the Huon Peninsula, the outer reef crest assemblage of flat, narrow (tens of meters wide) fringing reefs, or the barrier ridge on the outer margin of uplifted shallow lagoons, was considered the most reliable indicator of former sea level. Of course, corals can only grow below sea level and these late Pleistocene sea-level indicators must be lowest tide markers. However, growth up to the proximity of the lowest tide was a sufficiently resolved former sea level in the case where workers were trying to assess late Pleistocene sea-level highstands within the context of 100-meter-plus climatically driven oscillations in sea level. Finally, there is a vast literature on the accumulation of modern carbonate reefs relative to sea level (well summarized in Kim et al., 2012), and workers extensively using coralline platforms for sea-level reconstruction should be familiar with these processes.

5.5.2 Coralline indicators of late Holocene relative sea-level change: microatolls

An exceptionally informative geomorphic indicator of past sea levels in the late Holocene are coral microatolls (Fig. 5.3). Microatolls are so named because, although on the order of 0.5 m to a few meters in diameter, many specimens have raised rims and low centers (Fig. 5.3a) that much resemble the atoll islands first well publicized by Charles Darwin (Darwin, 1842); however, other microatoll specimens have different morphologies (Fig. 5.3b). Although Chapter 8 is entirely devoted to microatolls as a sea-level indicator, it is appropriate to put microatolls in perspective with other geomorphic indicators within this chapter.

Microatolls, consisting of several species of the coral genera *Porites* and *Goniastrea*, grow upward to an extent limited by the lowest tides, and the uppermost limit of growth is called the highest level of survival (HLS). Taylor et al. (1987) initially introduced the HLS as an indicator of late Holocene sea-level change. The form and morphology of microatolls can therefore be interpreted in terms of gradual and abrupt changes, both positive and negative, in relative sea level (Fig. 5.3; Taylor et al., 1987, Sieh et al., 1999; Zachariasen et al., 1999, 2000).

Using coastal field sites in Vanuatu and Sumatra, Taylor et al. (1987) and Sieh et al. (1999), respectively, both suggested that microatolls are valuable as geodetic markers of earthquake-induced changes in relative sea level. Corals have color banding that can be used to discern annual growth (Taylor et al., 1987), and developments in U-series dating (Edwards et al., 1988) allow age determination uncertainty for these corals to be on the order of a few years to a few decades. The annual growth record combined with well-resolved ages for selected annular rings means that coral microatolls are not only geomorphically distinct but potentially precise paleogeodetic markers. Some limitations to the use of microatolls as paleogeodetic markers include the tendency for them to live for a limited period of time, so that data often have to be combined from several areas to get a late Holocene relative sea-level record, and the observation that a microstoll's highest level of survival may be influenced by ponding on a reef platform.

Paloegeodesy from microatolls can be as precise as 20 mm of change over a few years to a few decades (Sieh et al., 1999; Zachariasen et al., 2000). Abrupt coral death can be inferred to be a result of an abrupt coseismic elevation shift of the coral to an elevation above the highest level of survival (e.g., Fig. 5.3c; Sieh et al., 2008). Microatoll paloegeodetic records have also been interpreted in terms of long-term paleoseismic behavior of the Sunda megathrust (Sieh et al., 2008) and in terms of a combination of earthquakes and non-tectonic "sea-level anomalies," which are aperiodic oscillations of

Fig. 5.3. Microatolls that were killed, wholly or in part, due to coseismic uplift during the December 2004 Aceh Sumatra earthquake. The three sites, located in the vicinity of 2°50' N, 95°49' E, are described in detail in Meltzner (2010). (a) Microatoll with cup-shaped geometry that implies that rapid submergence was occurring before the abrupt coseismic uplift event that emerged, and killed, the microatoll. Langi site, northern Simeulue Island, Sumatra. *Source*: Photograph by D. H. Natawidjaja. Reproduced with permission of Aron Meltzner. (b) Uneroded concentric ridges and swales characterize this microatoll and attest to annual growth outward once the coral reached the highest level of survival (HLS) dictated by low tide. Outward growth terminated when the coral was killed by coseismic uplift. This specimen is from the Lewak site, northern Simeulue, but similar specimens can be found in the circum-Pacific in tropical regions. *Source*: Photograph by D. H. Natawidjaja. Reproduced with permission of Aron Meltzner. (c) Most of this microatoll was abruptly killed by abrupt relative sea-level fall, although the lowest 5 cm of the colony is still alive in this photo. The fall in relative sea level is measured as the elevation difference between the microatoll's dead flat top (the former HLS) and the highest living corallites today (the new HLS, a few centimeters above water level in this photo). Photo was taken by J. Galetzka at the Lewak site, northern Simeulue, on 5 February 2005. *Source*: Photograph by J. Galetzka. Reproduced with permission of Aron Meltzner. For color details, please see Plate 10.

sea-surface temperatures (Meltzner et al., 2010). By utilizing specimens that can survive vertical crustal displacements yet stay with the zone of survival and grow for long periods of time, a time series of sea-level relative elevation change can be constructed for up to 300 years (Meltzner et al., 2012). Such time series have produced exceptionally detailed (millimeter-scale over a single year) sea-level records (Meltzner et al., 2012; Philibosian et al., 2012) and have produced records for periods back in time as far as 1200–1350 AD (Philibosian et al., 2012).

5.6 EROSIONAL LANDFORMS THAT INDICATE FORMER SEA LEVELS

5.6.1 Late Pleistocene emergent shoreline platforms

In mid-latitude non-tropical coastal areas, shoreline platforms are an erosional response to selective chemical and bioerosional weathering and/or wave-induced abrasion in the tidal zone (for an excellent discussion, see chapter 9 in Trenhaile, 1987). For tectonically uplifting coasts, successions of uplifted shore platforms may be preserved as stair-stepped flights of marine terraces (Fig. 5.1b). Similar to the case for uplifted terrace fights consisting of former carbonate fringing reefs, the erosional stair-stepped marine terraces can be related to a succession of sea-level highstands (Bradley and Griggs, 1976; Pillans, 1983).

The inner edge, or shoreline angle, of an uplifted shore platform is a paleo horizontal contour line (within an error margin of about 1–4 m dictated by lithology-driven and coastal-morphology-driven variation in inner edge position). Variation in elevation of the inner edge of uplifted platforms in a shore-parallel sense is an indication of shore-parallel variation in surface uplift rates. Tracking the elevation variation of uplifted inner edges of marine terraces is an effective means to detect faults, regional tectonic tilts, and long-wavelength (multiple kilometer) folds active in the Quaternary (e.g., Bradley and Griggs, 1976; Pillans, 1983; Kelsey, 1990; Kelsey and Bockheim, 1994).

Elevated late Pleistocene emergent shoreline platforms are therefore useful for calculating long-term uplift rates over tens of thousands of years, and useful for determining rates of regional tilt over the same timescale. Inner edge elevations of uplifted buried platforms are often difficult to measure directly however, except in specific cases of fortuitous road cuttings or borehole data. The inability to directly measure variation or complexity in inner edge elevations along a flight of uplifted shore platforms, as well as the inability to determine timing of inner edge formation, precludes the use of late Pleistocene emergent shoreline platforms for determining vertical elevation changes caused by coseismic or interseismic tectonic movements that occur on the timescale of hours to hundreds of years. It is for this reason that late Holocene erosional platform–cliff junctions are a focus of applied sea-level research.

5.6.2 Late Holocene platform–seacliff junctions

Compared to late Pleistocene shore platforms, the resolution of sea-level changes in the late Holocene is better because the shore platform–seacliff junction is usually exposed (e.g., Fig. 5.4) and amenable to surveying; the junction also may be amenable to dating through cover sediment packages or in-growth-position organisms that bore into the platform. On bedrock coasts, platform junctions tend to be close to the high tidal level, specifically between the elevation of mean high neap to mean high spring tides (Fig. 5.4; Trenhaile, 1972). Nonetheless, the elevation variation of this junction in a shore-parallel direction can be as much as 4 m or more. This variation in height of the cliff–platform junction is partly related to the seasonal variability of the height of spring and neap tides but mostly related to the variation in lithologic resistance to erosion of the platform (junctions tend to be higher in areas of hard than in soft rocks) and to the location of the specific site on a headland versus an embayment (Wright, 1970; Trenhaile, 1972). Causes of this variation are reinforced in some cases and cancel each other out in other cases, that is, junctions are higher than in the adjacent bays on most headlands but the reverse may be true in some cases as higher wave exposure and weathering on headlands preferentially lowers junctions (Trenhaile, 1987).

The most productive application of the platform–cliff junction in sea-level research is to reconstruct sea-level change in the late Holocene. Such studies have focused on the combined problem of sea-level change caused by abrupt vertical tectonic movements and sea-level change, more gradual in nature, as a consequence of glacial isostatic adjustment in the far field away from former ice sheets. For both applications, the inner edge of the late Holocene platform (or the shoreline angle) is the optimum elevation marker of a raised Holocene platform. As discussed above, however, this inner edge forms at a variable elevation relative to the tidal cycle and to the position along the coast.

In reconstructing former late Holocene sea levels, the inherent elevation variability for a platform junction can be avoided by surveying the modern junction relative to the relict, uplifted late Holocene junction at the same coastal site and measuring the difference between the two elevations. By doing so, Hull (1987) was able to constrain the maximum amount of coseismic tectonic uplift (6 m) for an

Fig. 5.4. Modern examples of the shore platform inner edge, which is also called the platform–seacliff junction or the shoreline angle. (a) Platform inner edge eroded into highly erosive, interbedded (3–6 cm thick) middle Miocene sandstone and argillite, northern California, USA. People are astride the inner edge. Note that the well-rounded beach cobbles, which partially cover the platform, are a distinctly different, much more resistant massive sandstone (40°06' N, 124°06.5 W). (b) Close-up view of modern platform inner edge, La Paz Peninsula, Mexico. Platform and seacliff consist of andesitic lahar deposits (24°19' N, 110°12' W). (c) Modern shore platform and platform inner edge at Arisaig, northern coast of Nova Scotia (45°46' N, 62°09' W). The platform is cut in landward-dipping Silurian sandstone and siltstone (Porter et al., 2010). *Source*: Photograph (a) by Harvey Kelsey and photographs (b) and (c) by A. S. Trenhaile. Reproduced with permission. For color details, please see Plate 11.

earthquake about 2300–2750 ^{14}C a BP. The age determination was on shells from intertidal organisms within a 1–6-m-thick marine cover sediment sequence that overlay the late Holocene shore platform. Using similar strategies, emergent late Holocene platforms, often dated by means of associated cover sediment, have been employed to document a stand of sea level higher than present in the latter part of the Holocene in the eastern Pacific and Indian ocean regions (e.g., Woodroffe et al., 1995).

Another useful strategy for assessing changes in relative sea level is to measure the difference between the inner edge elevation of two uplifted platforms at the same site. Because they are taken at the same site, these differences avoid the problem of inherent variability of the inner edge elevation in a shore-parallel direction along the coast. Using such a differencing strategy, various workers in regions as diverse as New Zealand (Wilson et al., 2006) and the Philippines (Ramos and Tsutsumi, 2010) have determined specific relative sea-level drops, interpreted to be related to coseismic uplifts, to an elevation resolution of less than a meter. In another example, Merritts (1996) inferred the amount of coseismic uplift for each of the multiple earthquakes on the coast in the vicinity of Cape Mendocino, California by measuring, and differencing, along-coast elevation of platform–cliff junctions of late Holocene emergent platforms. However, even for platforms with ages younger than a few thousand years, the uplifted platforms were partly buried by sediment or removed by coastal erosion, and elevation of uplifted platform–cliff junctions had to be extrapolated from limited platform exposures. The ^{14}C ages of the late Holocene coseismically uplifted platforms were determined from *pholads* (mollusks that bore into rock platforms) or from platform cover sediment that either consisted of marine sediment containing shells or colluvial sediment containing wood.

5.6.3 Erosional indicators of past sea levels at high latitudes

Processes that operate on Holocene shore platforms at high latitudes are the same as those that operate at mid–low latitudes, that is, mechanical wave action, chemical weathering, and bioerosional weathering (Trenhaile, 1983). At high latitudes, freeze–thaw cycles are a significant mechanism for triggering recession of sea cliffs. Although the process of freeze-and-thaw may accelerate cliff retreat, the process is not the primary mechanism responsible for planation of shore platforms (Trenhaile, 1983). The indicative meaning of the shoreline angle is therefore unaffected by the importance or lack of importance of frost action.

In areas of isostatic rebound where sea-level fall can be rapid (1 to >10 mm a^{-1}), modern platforms appear to reflect modern sea level (although documentation is mostly qualitative observation rather than surveys; Trenhaile, 1983). Nonetheless, it appears that shore platforms can both form and adjust rapidly to modern sea level along coasts subject to isostatic rebound.

5.6.4 Notches as sea-level indicators

Notches are geomorphically distinct undercuttings, centimeters to a few meters deep, formed in bedrock cliffs on coasts (Fig. 5.5), and are potential indicators of former sea levels. Notches have been extensively investigated for sea-level reconstruction and paleoseismic applications in the Mediterranean region (Pirazzoli, 2005; Evelpidou et al., 2012) where they were first publicized (Higgins, 1980; Pirazzoli, 1980), but notches have also been used for sea-level reconstruction in other parts of the world (Liew et al., 1993).

Similar to the formation of shore platforms, the formation of notches requires periods of stable relative sea level that are longer than the effective period of notch formation. Notches can then be preserved as indicators of former sea level if relative sea level then abruptly rises or falls.

The indicative meaning of notches, that is, the relationship of the formation of notches to the tidal regime, was first critically addressed by Pirazzoli (1986). He proposed that tidal notches can have a well-defined indicative meaning because the notch apex (point of maximum inward growth) is at mean tide level and the upper and lower elevation limits of the notch indicate the range of the tides plus wave splash. Tidal notch formation mechanisms include some combination of bioerosion and dissolution (Pirazzoli, 1986). However, Pirazzoli noted that notches can be formed at or near sea level by other mechanisms including abrasion by clast-laden wave action and enlargement by abrasion erosion along horizontal zones of structural weakness. According to Pirazzoli (1986), such features, although notch-like in morphology, do not have

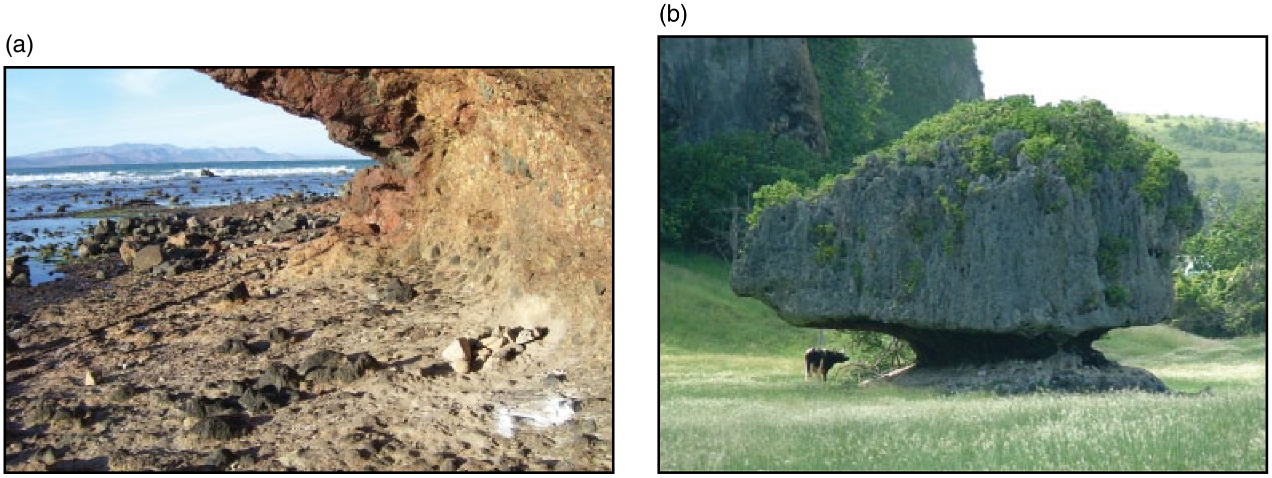

Fig. 5.5. (a) Modern tidal notch in tidal zone, La Paz Peninsula, Mexico; notch is eroded into andesitic lahar deposits (24°19' N, 110°12' W). (b) Uplifted paleo-tidal notch, Barbados Island, western Caribbean; notch is eroded into coralline limestone (13°10' N, 59°33' W). *Source*: Photograph (a) by A. S. Trenhaile and photograph (b) by Anthony Long. Reproduced with permission. For color details, please see Plate 12.

the clearly defined indicative meaning that is a characteristic of tidal notches.

As is the case with shore platforms, relative sea-level reoccupation can occur during notch genesis. Late Holocene tidal (or non-tidal) notches can potentially reoccupy late Pleistocene notches because glacioeustatic sea-level highstands rise to approximately the same levels on stable coasts or coasts with low tectonic uplift rate ($<0.1\,mm\ a^{-1}$) and/or negligible GIA-induced net relative sea-level change from one highstand to the next.

Late Holocene notches are potentially more promising for indicative meaning than late Pleistocene counterparts, because late Holocene notch positions can be directly tied to modern sea levels (compare Fig. 5.5a and b). Recent studies have used this attribute to assess the magnitude of coseismic uplift events (earthquakes) from raised Holocene notches (Liew et al., 1993; Stiros et al., 2000).

Finally, as is the case with all geomorphic indicators of former sea level, the indicator is of little use for sea-level studies if the indicator cannot be assigned an age. In the case of paleo notches, they can be dated either directly or indirectly. The direct approach is to find remnants of dead in-growth-position organisms that bioeroded the notch, for example barnacles or borings that contain the organism that created the bore. The indirect approach is to associate a submerged late Holocene notch with an adjacent coastal salt-marsh that has undergone accretion over the time of notch formation and submergence. In such a case, marsh accretion rates (determined by radiocarbon age determinations) can be related to submerged notches such that the notch provides the inference that the submergence was rapid and the ages provide the chronology of the submergence (e.g., Nixon et al., 2009). In many cases however, relict tidal notches are barren of in-growth-position organisms or are not associated with nearby sediment packages that record the same relative sea-level change.

In summary, the applicability of uplifted, or submerged, notches for sea-level reconstruction relates directly to the accuracy at which these relict notches can be assigned to a paleo-tidal regime operating at a specific time in the past. For investigators who encounter relict notches (paleo-notches) in areas geographically removed from well-studied regions, it is important to assess whether modern notches are forming in the tidal range and determine how the notch geometry relates to the tidal range (i.e., what is the indicative meaning of the notch?). For example, the notch depicted in Figure 5.5a developed in andesitic lahar deposits along a coastal segment of the La Paz Peninsula, Mexico. Elsewhere along this same coast, the tidal zone-upland transition is not a notch but is an inner edge of a platform; that is, there is a continuum of the notch geometry to the platform inner edge geometry. Because platform inner edges on average form at high tide levels, does the notch in this Mexico coast example form at high tide or at mean tide, as suggested in the literature? Only local surveying (in this example, on the La Paz Peninsula coast) can address such uncertainty.

5.7 EMERGENT HOLOCENE SHORELINE DEPOSITS: BEACH RIDGES

Emergent coastal deposits are geomorphic indicators of past sea levels; because such deposits are ubiquitous globally, they are potentially one of the richest sources of information on former relative sea levels. In a broad sense, all coastal deposits are associated with shorelines and thus associated with present or past sea levels, but deposits most often ascribed to have indicative meaning (that is, indicate a past sea level) are variously referred to as beach ridges, storm berms, and strandlines.

But confounding any discussion of beach ridges, storm berms, and strandlines as indicators of past sea level is terminology. There is much well-intentioned, but cumulatively confusing, literature that attempts to orchestrate a common language for beach ridges, storm berms, and strandlines (e.g., Roy et al., 1994; Otvos, 2000; Hesp et al., 2005). Because of the ongoing debate among experts as to what terms refer to what coastal features, beach ridges and related terms are defined here before discussing the extent to which these geomorphic features indicate former sea levels.

Beach ridges are shore-parallel ridges that consist of marine sediment and are formed by wave swash. Berms are features that can be generated by tides and are ephemeral. Storm berms, in contrast, are not ephemeral and are generated by wave swash during storms and high tide with higher elevation of deposition, possibly enhanced by storm surge. Storm berms are therefore one end-member of beach ridges formed by wave swash, and I refer to storm berms as beach ridges. Rate of beach ridge addition is a function of sediment supply, sea-level trends, and storm climate (Roy et al., 1994). In contrast, foredunes are dunes constructed by aeolian processes; if excavated, foredunes are distinct sedimentologically and do not have foreset beds characteristic of wave swash. Beach ridges that are no longer active can be capped by aeolian deposits.

Strandplains are broad accumulations of sediment formed in parallel or semi-parallel ridges oriented approximately parallel to the coastline (Roy et al., 1994; Hein et al., 2012). Strands are linear geomorphic features parallel to the modern shoreline that define former shorelines. Strandplains may represent a composite of wave-swash-created beach ridges, foredunes or even emergent offshore bar features. The progradational nature of strandplains allows them to serve as recorders of paleoenvironmental change (Wells, 1996; Goy et al., 2003; Hein et al., 2012). For instance, regressive strandplain sequences can chronicle sea-level fall in the late Holocene. At Pinheira, southern Brazil, falling sea level led to forced coastal progradation. The result was a 5-km-wide strandplain that prograded as offshore sediment was reworked and supplied. The plain consists of a succession of closely spaced arcuate, low, 1–4-m-high beach and dune ridges (Hein et al., 2012). In order to investigate the deposit geometry that accompanies such a progradation, Hein et al. (2012) demonstrate the utility of the active-source imaging technique of ground-penetrating radar (GPR) for investigating subsurface stratigraphy and sea-level history on coastal plains.

Given the above definitions, then beach ridges (including "storm berms") are wave-swash-created features that, if separated from the active beach by relative sea-level fall or by coastal progradation, can be sea-level indicators. Beach ridges cannot form above spring high tides or above storm wave conditions; beach ridges can therefore document relative sea-level fall if the features are meters above spring high tide level or storm swash levels. Wells (1996), following Badukov (1986), suggests determination of mean sea level for beach ridges can be derived from the location of the swash zone on the beach face of a beach ridge. Tide range determination involved excavation of the lower slope of the beach ridge at the base of the former beach face to identify the foreset beds of the swash zone (Wells, 1996). If the reference water level for beach ridges is within the lower slope of the beach ridge, then the indicative range is large because such a reference water level could be the upper swash zone of spring high tides or the relatively higher upper swash zone of less frequent storms. A beach ridge may therefore indicate a sea level datum only over a broad vertical range, and that range may be greater than the height difference of adjacent beach ridges.

The elevations of two adjacent paleo-beach ridges can be differenced to assess relative sea-level fall between two beach ridges. Elevation differencing of the base of beach ridges and not the crests is however desirable, because paleo-beach ridge elevation can be affected by

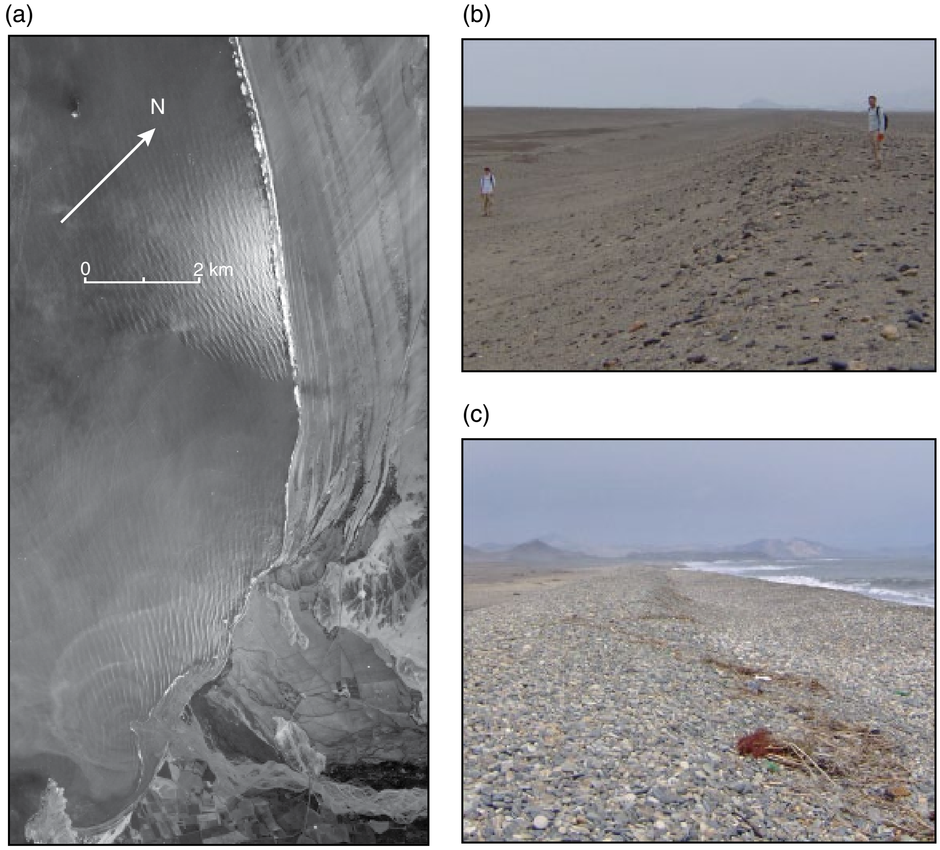

Fig. 5.6. (a) Mosaic of three vertical aerial photographs depicting beach ridges on the progradational beach-ridge plain of the Santa beach ridge complex, latitude 8° S, Peru. Santa River empties to Pacific Ocean to the southeast of the beach-ridge plain. Note the plume of sediment at the river mouth that is being transported north. Source of photograph: Servicio Aerofotográfico Nacional (Peru), Project 147-66, 20 March 1967. *Source*: Photographs mosaicked and contributed by D. Sandweiss. Reproduced with permission. (b) Abandoned inactive beach ridge, Santa beach ridge complex, latitude 8° S, Peru. View to north-northwest. (c) Modern beach ridge, seaward edge of Santa River progradational plain, latitude 8° S, Peru. View to south-southeast. *Source*: Photographs (b) and (c) by David Reid. Reproduced with permission. For color details, please see Plate 13.

sediment supply, storminess, and aeolian caps (Orford et al., 2003).

On progradational plains, beach ridges often occur as multiple parallel ridges with swales between them. There is not any one cause for the isolation of a beach ridge and the deposition of another in a more seaward position. Various case studies and review articles (Sandweiss, 1986; Tanner, 1995; Wells, 1996; Goy et al., 2003; Orford et al., 2003) argue that beach ridges can become isolated by a storm as it may make the ridge less accessible to wave swash, by sea-level change of the order 1–1.5 m, or by a climate-induced change in sediment supply. Each of these factors can reinforce the others, leading to the isolation of a beach ridge.

The Santa beach ridge complex in coastal Peru, at latitude 8°S, consisting of as much as 4 km of coastal progradational plain and as many as 15 individual beach ridges (Fig. 5.6), is a good example of the complexity of identifying the cause of abandonment of one beach ridge and the establishment of the next seaward ridge. Although Wells (1996) justifiably argues from sedimentologic evidence that the ridges are marine, establishing a sea-level record from a set of coast-parallel beach ridges is problematic because: (1) the ridges are difficult to date (Sandweiss et al., 1998); and (2) in Peru the combination of earthquakes and El-Niño-triggered storms can mobilize sediment from coastal drainages that result in episodes of high sediment delivery to the coast (Sandweiss, 1986). But each sediment-influx-related El Niño event does not engender a new ridge (Rogers et al., 2004). And sea level in the last 4000 years has been static or has dropped by 1–2 m or less (Sandweiss, 1986; Wells, 1996). Therefore, using beach ridge elevation to chronicle sea-level change in coastal Peru

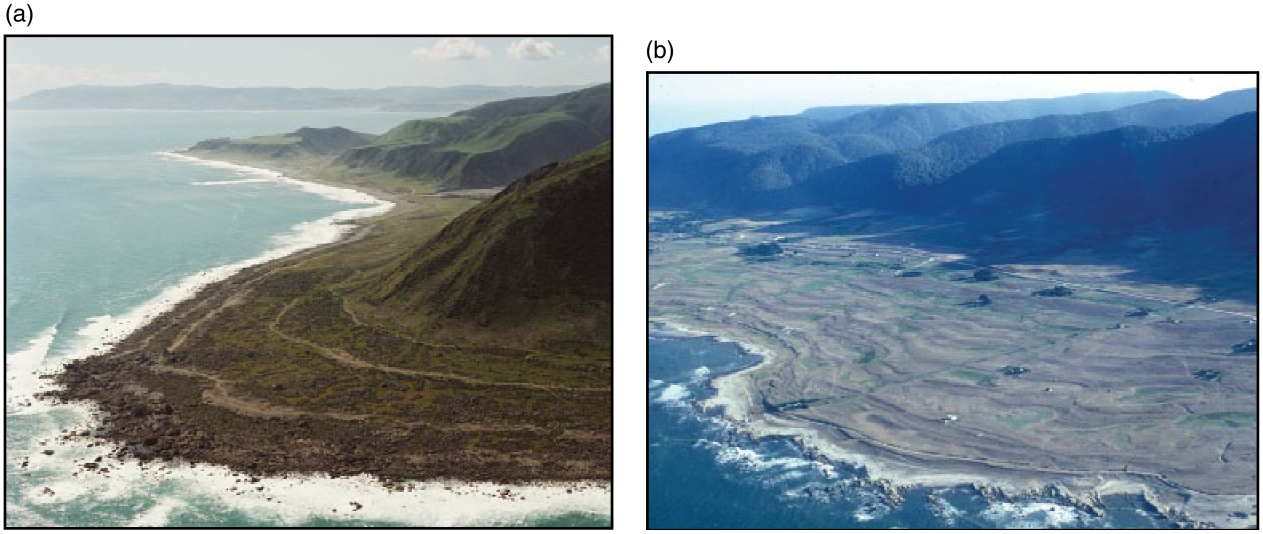

Fig. 5.7. (a) Uplifted beach ridges at Cape Turakirae, South Island, New Zealand (41°26' S, 174°55' E). Uplift of all ridges is inferred to be associated with coseismic displacement on the nearby Wairarapa fault, and the most recent, historic beach ridge uplift occurred during coseismic displacement on this fault in 1855. Individual uplifts range in magnitude from 3.5 to 7.1 m (McSaveney et al., 2006; Little et al., 2009). The two stair-stepped treads of marine terraces in the middle ground are oxygen isotope stage 5 (80–125 ka) and stage 7 (~200 ka) uplifted shore platforms (McSaveney et al., 2006). *Source*: Photograph by Lloyd Homer. Reproduced with permission of GNS Science, New Zealand. (b) Late Holocene uplifted beach ridges along the central-east coast of Isla Mocha, which is an island at latitude 38° S off the coast of Chile (see Nelson and Manley, 1992). Isla Mocha sits above the Chilean subduction zone. *Source*: Photograph by Alan R. Nelson. Reproduced with permission. For color details, please see Plate 14.

is imprecise because the uncertainty in the indicative meaning of the base elevation of a beach ridge is at least as large or larger than the multi-thousand-year period of relative sea-level change. Relative sea-level change alone is therefore not a convincing driver for ridge isolation. Isolation of individual ridges of the Santa beach ridge complex and establishment of the next landward ridge may well be triggered by climate-induced change in sediment delivery to the coastal zone; isolation probably occurs in conjunction with other factors, which could include vertical tectonic displacements, increased storminess and/or minor (of the order decimeters over hundreds of years) GIA-related relative sea-level fall.

On tectonically active coasts where there is minimal or no progradational beach plain, beach ridges can become isolated by tectonic, usually coseismic, uplift (Wellman, 1969; Nelson and Manley, 1992; McSaveney et al., 2006). Workers can identify the mechanism of beach ridge isolation as tectonic in these cases because storms and high tides, while creating the beach ridges in the first place, are not a mechanism to isolate them. These coasts are not progradational plains, and therefore a change in sediment supply is not a potential factor leading to beach ridge isolation. For instance, Turakirae Head on the south end of the North Island of New Zealand prominently features four beach ridges that are vertically separated from each other by 3.4–7.1 m (Fig. 5.7a; Aston, 1912; Wellman, 1969; McSaveney et al., 2006). The crest of the modern beach ridge is located 2–7 m above mean sea level and presumably marks the run-up limit of present-day storm waves. The elevation difference of adjacent raised ridges is a measure of coseismic uplift that caused the stranding of each higher elevation ridge (Fig. 5.7a; Little et al., 2009).

In a similar case study, Nelson and Manley (1992) document beach ridges on the coast of Isla Mocha offshore of mainland Chile and above the Chilean subduction zone megathrust. A set of beach ridges was coseismically uplifted in the 1960 Chilean subduction zone earthquake, and there are older, higher beach ridges that may have been isolated from sea level by the same mechanism (Fig. 5.7b).

Although shore-parallel lines of raised beach ridges on a tectonically active coast can be a signature of abrupt relative sea-level fall brought on by coseismic uplift of the land, aseismic uplift, if fast enough, can elevate beach ridges formed by infrequent storm events and can therefore leave a record of relative sea-level fall. Nelson and Manley (1992) also identify a beach ridge that has been isolated above sea level along the same coast since the 1960 subduction zone earthquake. They infer that this beach ridge was deposited by storms and isolated by unusually rapid aseismic tectonic uplift related to strain on upper plate faults within the subduction zone margin.

In addition, beach ridges can also be stranded above highest tide levels by GIA-induced relative sea-level fall. Such raised beaches are common in equatorial regions (e.g., Fletcher and Jones, 1992, Woodroffe et al., 1995; Grossman and Fletcher, 1998) where ocean syphoning (Mitrovica and Peltier, 1991; Mitrovica and Milne, 2002) has occurred.

Beach ridge formation in high latitudes, with examples from the Arctic, deserves separate mention because raised beach ridges in high latitudes are ubiquitous and are the most common coastal landform (Sanjaume and Tolgensbakk, 2009). A prominent reason for the ubiquity of beach ridges as evidence of former shorelines in high latitudes is that retreat of ice resulted in isostatic rebound and relative sea-level fall. In Spitzbergen (the main island of Svalbard, latitude 79°N), more than 100 beach ridges formed in this manner; however, in addition to fall in relative sea level, requirements for beach ridge formation include adequate exposure of the shore to wind fetch and seasonally ice-free conditions so that waves can form ridges (Brückner et al., 2002). Although beach ridges are useful as an indicator of sea-level change in the Arctic (Long et al., 2012), the number of preserved beach ridges on neighboring sections of coast may vary due to intrinsic factors. For instance, the Varanger Peninsula of northern Norway (70°40’ N) is an example of a region where, even with isostatic rebound being the same among neighboring bays, the number of raised beach ridges varies from one neighboring bay to the next because local ridge formation is controlled more by local sediment supply, offshore gradient, and site-specific coastal evolution than by more regional factors such as storminess and rates of regional sediment supply (Sanjaume and Tolgensbakk, 2009).

Because beach ridge formation in high latitudes has persisted throughout the Holocene, that is, persisted into time periods younger than the period of major ice melting and consequent isotatic adjustments, other factors in addition to isostatic uplift play a role in the formation and preservation of high-latitude beach ridges. Beach ridges can form in the latter part of the Holocene on progradational plains, and the main factors determining whether ridge progradation takes place or not are wave climate and/or high sediment input to the beach (Mason and Jordon, 1993; St-Hilaire-Gravel et al., 2010). And varying summer ice extent may be a main determinant of beach ridge morphology and preservation. Reduced summer sea ice results in more open water that in turn leads to higher wave energy for prolonged periods; reduced ice, more fetch, and higher wave energy therefore leads to better-developed beach ridges (Lowther Island, Canadian Arctic, 74°30’N; St-Hilaire-Gravel et al., 2010). If lack of ice cover, increased fetch and consequent higher wave energy can produce beach ridges on progradational plains, then the main factors influencing Holocene beach ridge isolation and character may be factors other than relative sea-level fall.

Finally, for those cases in the Arctic where beach ridges can be used as an indicator of regional sea-level change, beach ridge chronology is essential to track history of relative sea-level change. For northern Spitzbergen in the high Arctic (79°N), age determinations from juvenile articulated shells incorporated into beach ridges during storms provide reliable sample age determinations that compare well with other chronologies derived from driftwood or whalebone (Long et al., 2012).

As a final note on emergent Holocene shoreline deposits, transgressive barrier island systems have received much attention; on the east coast of the US, these depositional systems form in the context of rising sea level throughout the Holocene. For transgressive barrier island systems however, rates of sea-level rise are most effectively documented through saltmarsh stratigraphy, which is discussed in Chapter 5.

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REFERENCES

Aston, B.C. (1912) The raised beaches of Cape Turakirae. Transactions and Proceedings of the New Zealand Institute, 44, 208–213.

Badukov, D. D. (1986) Ancient shorelines as indicators of sea level. Journal of Coastal Research, 2, 147–157.

Bloom, A.L., Broecker, W.S., Chappell, J.M.A., Matthews, R.K., and Mesolella, K.J. (1974) Quaternary sea level fluctuations on a tectonic coast: new 230Th/234U dates from the Huon Peninsula, New Guinea. Quaternary Research, 4, 185–205.

Bradley, W.C., and Griggs, G.B. (1976) Form, genesis and deformation of central California wave-cut platforms. Geological Society of America Bulletin, 87, 433–449.

Brückner, H., Schellmann, G., and van der Borg, K. (2002) Uplifted beach ridges in northern Spitsbergen as indicators for glacio-isostasy and palaeo-oceanography. Zeitschrift für Geomorphologie, 46, 309–336.

Bull, W.B. (1985) Correlation of flights of global marine terraces. In: *Tectonic Geomorphology* (eds Morisawa, M., and Hack, J.T.), Allen & Unwin, Boston, pp. 129–152.

Chappell, J.M.A. (1974) Geology of coral terraces, Huon Peninsula, New Guinea: A study of Quaternary tectonic movements and sea level changes. Geological Society of America Bulletin, 85, 553–570.

Chappell, J., and Shackleton, N.J. (1986) Oxygen isotopes and sea level. Nature, 324, 137–140.

Darwin, C. (1842) *The Structure and Distribution of Coral Reefs.* Smith, Elder and Co., London.

Edwards, R.L., Taylor, F., and Wasserburg, G. (1988) Dating earthquakes with high-precision thorium-230 ages of very young corals. Earth and Planetary Science Letters, 90, 371–381.

Evelpidou, N., Kampolis, I, Pirazzoli, P.A., and Vassilopouos, A. (2012) Global sea-level rise and the disappearance of tidal notches. Global and Planetary Change, 92–93, 248–256.

Fletcher, C.H., and Jones, A.T. (1996) Sea-level highstand recorded in Holocene shoreline deposits on Oahu, Hawaii. Journal of Sedimentary Research, 66, 632–641.

Goy, J.L., Zazo, C., and Dabrio, C.J. (2003) A beach-ridge progradational complex reflecting periodical sea-level and climate variability during the Holocene, Gulf of Almeria, western Mediterranean. Geomorphology, 50, 251–268.

Grossman, E.E., and Fletcher, C.H. (1998) Sea level higher than present 3500 years ago on the northern main Hawaiian Islands. Geology, 26, 363–366.

Hein, C.J., Fitzgerald, D.M., Cleary, W.J., Albernaz, M.B., De Menezes, J., and Klein, A. (2012) Evidence for a transgressive barrier within a regressive strandplain system: Implications for complex coastal response to environmental change. Sedimentology, 60(2), 469–502.

Hesp, P.A., Dillenburg, S.R., Barboza, E.G., Tomazelli, L.J., Ayup-Zouain, R.N., Esteves, L.S., Gruber, N.L.S., Toldo, E.E., Tabajara, L., and Clerot, L. (2005) Beach ridges, foredunes or transgressive dunefields? Definitions and an examination of the Torres to Tramandaí barrier system, Southern Brazil. Annals of the Brazilian Academy of Sciences, 77, 493–508.

Higgins, C.H. (1980) Nips, notches, and solution of coastal limestone: an overview of the problem with examples from Greece. Estuarine and Coastal Marine Science, 10, 15–30.

Hull, A.G. (1987) A late Holocene marine terrace on the Kidnappers coast, North Island, New Zealand: Some implications for shore platform development processes and uplift mechanism. Quaternary Research, 28, 183–195.

Kelsey, H.M. (1990) Late Quaternary deformation of marine terraces on the Cascadia subduction zone near Cape Blanco, Oregon. Tectonics, 9, 983–1014.

Kelsey, H.M., and Bockheim, J.G. (1994) Coastal landscape evolution as a function of eustasy and surface uplift rate, Cascadia margin, southern Oregon. Geological Society of America Bulletin, 106, 840–854.

Kim, K., Fouke, B., Petter, A., Quinn, T., Kerans, C., and Taylor, F. (2012) Sea-level rise, depth-dependent carbonate sedimentation and the paradox of drowned platforms. Sedimentology, 59, 1677–1694.

Lajoie, K.R. (1986) Coastal Tectonics. In: *Active Tectonics* (ed. Wallace, R.E.), National Academic Press, Washington, DC, pp. 95–124.

Lambeck, K., and Nakada, M. (1992) Constraints on the age and duration of the last interglacial period and on sea-level variations. Nature, 357, 125–128.

Liew, P.M., Pirazzoli, P.A., Hsieh, M.L., Arnold, M., Barusseau, J.P., Fontugne, M., and Giresse, P. (1993) Holocene tectonic uplift deduced from elevated shorelines, eastern Coastal Range of Taiwan. Tectonophysics, 222, 55–68.

Little, T.A., Van Dissen, R., Schermer, E., and Carne, R. (2009) Late Holocene surface ruptures on the southern Wairarapa fault, New Zealand: Link between earthquakes and the uplifting of beach ridges on a rocky coast. Lithosphere, 1, 4–28.

Long, A.J., Strzelecki, M.C., Lloyd, J.M., and Bryant C. (2012) Dating High Arctic Holocene relative sea level changes using juvenile articulated marine shells in raised beaches. Quaternary Science Reviews, 48, 61–66.

Mason, O., and Jordan, J. (1993) Heightened North Pacific storminess during synchronous late Holocene erosion of northwest Alaska beach ridges. Quaternary Research, 40, 55–69.

McSaveney, M.J., Graham, I.J., Begg, J.G., Beau, A.G., Hull, A.G., Kim, K., and Zondervan, A. (2006) Late Holocene uplift of beach ridges at Turakirae Head, south Wellington coast, New Zealand. New Zealand Journal of Geology and Geophysics, 49, 337–358.

Meltzner, A.J. (2010) Earthquake recurrence, clustering, and persistent segmentation near the southern end of the 2004 Sunda megathrust rupture. PhD thesis, Californian Institute of Technology, Pasadena.

Meltzner, A.J., Sieh, K., Chiang, H.-W., Shen, C.-C., Suwargadi, B.W., Natawidjaja, D.H., Philibosian, B.E., Briggs, R.W., and Galetzka, J. (2010) Coral evidence for earthquake recurrence and an A.D. 1390-1455 cluster at the south end of the 2004 Aceh-Andaman rupture. Journal of Geophysical Research, 115, B10402, doi: 10.1029/2010JB007499.

Meltzner, A.J., Sieh, K., Chiang, H.-W., Shen, C.-C., Suwargadi, B.W., Natawidjaja, D.H., Philibosian, B. and Briggs, R.W. (2012) Persistent termini of 2004- and 2005-like ruptures of the Sunda megathrust. Journal of Geophysical Research, 117, B04405, doi: 10.1029/2011JB008888.

Merritts, D.J. (1996) The Mendocino triple junction: Active faults, episodic coastal emergence, and rapid uplift. Journal of Geophysical Research, 101, 6051–6070.

Mesolella, K.J., Matthews, R.K., Broecker, W.S., and Thurber, D.L. (1969) The astronomical theory of climatic change, Barbados data. Journal of Geology, 77, 250–274.

Milne, G.A., and Mitrovica, J.X. (2008) Searching for eustasy in deglacial sea-level histories. Quaternary Science Reviews, 27, 2292–2302.

Mitrovica, J.X., and Peltier, W.R. (1991) On postglacial geoid subsidence over the equatorial oceans. Journal of Geophysical Research, 96, 20053–20071.

Mitrovica, J.X., and Milne, G.A. (2002) On the origin of late Holocene sea-level highstands within equatorial ocean basins. Quaternary Science Reviews, 21, 2179–2190.

Muhs, D.R., Simmons, K.R., Kennedy, G.L., and Rockwell, T.K. (2002) The last interglacial period on the Pacific Coast of North America: Timing and paleoclimate. Geological Society of America Bulletin, 114, 569–592.

Muhs, D.R., Simmons, K.R., Schumann, R. and Halley, R.B. (2011) Sea-level history of the past two interglacial periods: new evidence from U-series dating of reef corals from south Florida. Quaternary Science Reviews, 30, 570–590.

Muhs, D.R., Simmons, K.R., Schumann, R.R., Groves, L.T., Mitrovica, J.X., and Laurel, D. (2012) Sea-level history during the Last Interglacial complex on San Nicolas Island, California: implications for glacial isostatic adjustment processes, paleozoogeography and tectonics. Quaternary Science Reviews, 37, 1–25.

Nakada, M., and Lambeck, K. (1989) Late Pleistocene and Holocene sea-level change in the Australian region and mantle rheology. Geophysical Journal International, 96, 497–517.

Nelson, A.R., and Manley, W.F. (1992) Holocene coseismic and aseismic uplift of Isla Mocha, south-central Chile. Quaternary International, 15/16, 61–76.

Nixon, F.C., Reinhardt, E.G., and Rothaus, R. (2009) Foraminifera and tidal notches; dating neotectonic events at Korphos, Greece. Marine Geology, 257, 41–53.

Orford, J.D., Murdy, J.M., and Wintle, A.J. (2003) Prograded Holocene beach ridges with superimposed dunes in north-east Ireland: mechanisms and timescales of fine and coarse beach sediment decoupling and deposition. Marine Geology, 194, 47–64.

Otvos, E.G. (2000) Beach ridges – definitions and significance. Geomorphology, 32, 83–108.

Philibosian, B., Sieh, K., Natawidjaja, D.H., Chiang, H.-W., Shen, C.-C., Suwargadi, B.W., Hill, E.M., and Edwards R.L. (2012) An ancient shallow slip event on the Mentawai segment of the Sunda megathrust, Sumatra. Journal Geophysical Research, 117, B05401, doi: 10.1029/2011JB009075.

Phillips, B.A.M. (1970) The significance of inheritance in the interpretation of marine and lacustrine coastal histories. Lakehead University Review, 3, 36–45.

Pillans, B. (1983) Upper Quaternary marine terrace chronology and deformation, South Taranaki, New Zealand. Geology, 11, 292–297.

Pirazzoli, P.A. (1980) Formes de corrosion marine et vestiges archéologiques submerges: interpretation néotectonique de quelques exemples en Grèce et en Yougoslavie. Annales de l'Institut Océanographique, 56, 101–111.

Pirazzoli, P.A. (1986) Marine notches in sea-level research. In: *A Manual for the Collection and Evaluation of Data* (ed. van de Plassche, O.), Geo Books, Norwich, pp. 361–400.

Pirazzoli, P.A. (2005) Marine erosion features and bioconstructions as indicators of tectonic movements, with special attention to the eastern Mediterranean area. Zeitschrift für Geomorphologie, Supplement 137, 71–77.

Porter, N.J., Trenhaile, A.S., Prestanski, K., and Kanyaya, J.I. (2010) Patterns of surface downwearing on shore platforms in eastern Canada. Earth Surface Processes and Landforms, 35, 1793–1810.

Radtke, U., and Schellmann, G. (2004) *The Marine Quaternary of Barbados*, Kolner Geographische Arbeiten, Heft 81, 137 pp., Geographisches Institut der Universitat zu Koln, Germany.

Ramos, N.T., and Tsutsumi, H. (2012) Evidence of large prehistoric offshore earthquakes deduced from uplift Holocene marine terraces in Pangasinan Province, Luzon Island, Philippines. Tectonophysics, 495, 145–158.

Rogers, S.S., Sandweiss, D.H., Maascht, K.A., Belknapt, D.F., and Agouris, P. (2004) Coastal change and beach ridges along the northwest coast of Peru: Image and GIS analysis of the Chira, Piura, and Colan beach-ridge plains. Journal of Coastal Research, 20, 1102–1125.

Roy, P.S., Cowell, P.J., Ferland, M.A., and Thom, B.G. (1994) Wave-dominated coasts. In: *Coastal Evolution, Late Quaternary Shoreline Morphodynamics* (eds Carter, R.W.G., and Woodroffe, C. D.), Cambridge University Press, Cambridge, pp. 121–186.

Sandweiss, D.H. (1986) The beach ridges at Santa, Peru: El Nino, uplift, and prehistory. Geoarchaeology: An International Journal, 1, 17–28.

Sandweiss, D.H., Maasch, K.A., Belknap, D.F., Richardson, J.B., and Rollins, H.B. (1998) Discussion of: Lisa E. Wells, 1996. The Santa Beach Ridge Complex, Journal of Coastal Research, 12, 1–17. Journal of Coastal Research, 14, 367–373.

Sanjaume, E. and Tolgensbakk, J. (2009) Beach ridges from the Varanger Peninsula (Arctic Norwegian coast): Characteristics and significance. Geomorphology, 104, 82–92.

Sieh, K., Ward, S., Natawidjaja, D., and Suwargadi, B. (1999) Crustal deformation at the Sumatran subduction zone revealed by coral rings. Geophysical Research Letters, 26, 3141–3144.

Sieh, K., Natawidjaja, D.H., Meltzner, A.J., Shen, C.-C., Cheng, H., Li, K.-S., Suwargadi, B.W., Galetzka, J., Philibosian, B., and Edwards, R.L. (2008) Earthquake supercycles inferred from sea-level changes recorded in the corals of West Sumatra. Science, 322, 1674–1678, doi: 10.1126/science.1163589.

Stiros, S.C., Laborel, J., Laborel-Deguen, F., Papageorgiou, S., Evin, J., and Pirazzoli, P.A. (2000) Seismic coastal uplift in a region of subsidence: Holocene raised shorelines of Samos Island, Aegean Sea, Greece. Marine Geology, 170, 41–58.

St-Hilaire-Gravel, D., Bell, T.J., and Forbes D.L. (2010) Raised gravel beaches as proxy indicators of past sea-ice and wave conditions, Lowther Island, Canadian Arctic Archipelago. Arctic 63, 213–226.

Tanner, W.F. (1995) Origin of beach ridges and swales. Marine Geology, 129, 149–161.

Taylor, F.W., Frohlich, C., Lecolle, J., and Strecker, M. (1987) Analysis of partially emerged corals and reef terraces in the central Vanuatu arc: Comparison of contemporary coseismic and nonseismic with Quaternary vertical movements. Journal of Geophysical Research, 92, 4905–4933.

Trenhaile, A.S. (1972) The shore platforms of the Vale of Glamorgan, Wales. Transactions of the Institute of British Geographers, 56, 127–144.

Trenhaile, A.S. (1983) The development of shore platforms in high latitudes. In: *Shorelines and Isostasy* (eds Smith, D.E., and Dawson, A.G.), Academic Press, London, Institute of British Geographers Special Publication 16, pp. 77–93.

Trenhaile, A.S. (1987) *The Geomorphology of Rock Coasts.* Oxford University Press, Clarendon, Oxford, UK.

van de Plassche, O. (ed.) (1986) *Sea-Level Research: A Manual for the Collection and Evaluation of Data*, Geo Books, Norwich.

Wellman, H.W. (1969) Tilted marine beach ridges at Cape Turakirae, New Zealand. Tuatara, 17, 82–93.

Wells, L.E. (1996) The Santa Beach ridge complex: sea-level and progradational history of an open gravel coast in central Peru. Journal of Coastal Research, 12, 1–17.

Wilson, K., Berryman, K.B., Litchfield, N., and Little, T. (2006) A revision of mid–late Holocene marine terrace distribution and chronology at the Pakarae River mouth, North Island, New Zealand. New Zealand Journal of Geology & Geophysics, 49, 477–489.

Woodroffe, C.D., Murray-Wallace C.V., Bryant, E.A., Brooke, B., Heijnis, H., and Price, D.M. (1995) Late Quaternary sea-level highstands in the Tasman Sea: evidence from Lord Howe Island. Marine Geology, 125, 61–72.

Wright, L.W. (1970) Variation in the level of the cliff/shore platform junction along the south coast of Great Britain. Marine Geology, 9, 347–353.

Zachariasen, J., Sieh, K., Taylor, F., Edwards, R., and Hantoro, W. (1999) Submergence and uplift associated with the giant 1833 Sumatran subduction earthquake: Evidence from coral microatolls. Journal of Geophysical Research, 104, 895–919.

Zachariasen, J., Sieh, K., Taylor, F., and Hantoro, W. (2000) Modern vertical deformation at the Sumatran subduction zone: Paleogeodetic insights from coral microatolls. Bulletin Seismological Society of America, 90, 897–913.