Geomorphological indicators

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Alessio Rovere (alessio.rovere@unive.it) - Ca' Foscari University of Venice (IT)
Marta Pappalardo – University of Pisa (IT)
Michael J. O'Leary – University of Western Australia (AU)

Abstract. Several physical, chemical, and biological processes shape coastal environments close to sea level. Acting through time, these processes create a variety of coastal landforms. When found outside their environment of formation, these landforms can be used by geoscientists as geomorphological indicators of former relative sea levels. In this chapter, we outline the main processes acting on coastal areas, and link them to distinct types of sea-level indicators, defining general elevation/depth formation ranges for each, which are essential to calculate the paleo relative sea-level position starting from the measured elevation or depth of the landform.

Keywords
1. Abrasion notches
2. Beach ridges
3. Beachrocks
4. Bioerosional indicators
5. Biological rims
6. Coastal caves
7. Fossil coral reefs
8. Marine terraces
9. Relic coastal deposits
10. Sea arches
11. Sea-level changes
12. Shore platforms
13. Speleothems in coastal caves
14. Tafoni and honeycomb
15. Tidal Notches

Key points/objectives box
- Defining the main processes active on coastal environments
- Connecting coastal processes to landforms
- Giving information on how to derive paleo relative sea-level positions from geomorphological indicators

Glossary
Relative sea level – The position of a former sea level as indicated by the elevation or depth of a geomorphological indicator, without any correction for secondary displacement (e.g., due to tectonics).
Modern analog – A landform still actively shaped by modern coastal processes, that can be considered as analog to a relic landform in terms of both processes and elevation or depth where these processes are active.

Indicative range – The elevational or depth range occupied by a modern analog landform.
Reference water level – The midpoint of the indicative range.
Indicative meaning – The combination of indicative range and reference water level, used to calculate the paleo relative sea level and the associated uncertainty from the measured elevation of a geomorphological indicator.

Introduction

Physical, chemical, and biological processes acting close to modern sea level shape a range of different morphologies. When found outside of the elevation or depth range where processes forming them are active (i.e., out of their environment of formation), these features can be used as geomorphological indicators of past changes in relative sea level, provided that their elevation, relationship with former sea level (the so-called indicative meaning) and age are known (Shennan et al., 2015).

To quantify the indicative meaning of any relict coastal feature, it is important to refer to analog landforms to understand the specific elevation (or depth) range at which the feature is shaped on the modern coast. Here the general occurrence of geomorphological sea-level indicators is constrained between the lower end of the shoreface (fair weather wave base, where waves start to interact with the sea bottom) and the upper foreshore (reached by marine spray and storm waves).

In this chapter, we first describe the processes acting at (or close to) the shoreline. Then, we focus on the most common landforms shaped by these processes, that have been used as geomorphological sea-level indicators in the literature.

Processes

Physical

Mechanical erosion. Several types of mechanical processes act close to sea level. Sediments suspended in the water column or loose on the sea bottom are periodically moved by waves. When they encounter a hard substrate in their path, sediment particles exert a mechanical action on it, progressively wearing it by abrasion. Mechanical erosion may be enhanced at the entrance of rock cavities, sea caves, or small bays, where wave flow is increased because of the narrowing of the hydraulic section (e.g., flow passing from open sea to a narrow cave entrance). Mechanical erosion may also happen in the absence of sediments, close to the sea surface. Incoming waves force water though the entrance of cracks, joints or holes created by bioerosion (see below), causing an increase in the air pressure inside them. Such sudden (and repeated) increase in pressure exerts a mechanical stress on the rock, causing its fracture. This process is often called wave pounding, hydraulic erosion or hydraulic fracturing (Huggett, 2007). Another mechanical process acting close to sea level is wetting-drying (hydroclasty). Rocks absorbing water tend to expand, while when they dry out, they tend to deflate. This is a cause of rock deterioration, which makes it more prone to erosion (Trenhaile, 2015). Salt weathering (haloclasty) is instead triggered by sea spray or wave runup over hard substrates depositing saltwater, that seeps into cracks in the rock. When desiccated, the volume of salt crystals increases and exerts a mechanical stress on the rock, breaking it. At high latitudes also frost shattering can be active on so called “periglacial coasts”, where ice rather than salt crystals grow inside cracks exerting mechanical stress.

Sediment transport. Within the coastal zone, sediments are typically transported by wave and tidally driven coastal currents either in suspension (for finer particles) or as a bedload (for coarser particles). Sediments can become part of a depositional environment where grains are transported onshore above wave action (with the aid of wave runup and aeolian processes) or offshore below wave base. The largest volume of sediments is transported between the lower limit of the shoreface and the upper part of the foreshore. The lower shoreface starts where the bathymetry corresponds to half the wavelength of the fair-weather waves, while the upper part of the foreshore is defined by the highest reach of storm waves.

Chemical

Karst. The karst dissolution of soluble limestone rocks occurs where they become subaerially exposed to meteoric waters either though localized uplift or sea level regression. Carbonic acid contained within the meteoric surface and ground waters reacts with and dissolves calcium carbonate, with the rate of dissolution greatest in tropical high precipitation climate zones compared to arid zones. Karst dissolution results in a range of surface (i.e., pinnacle or cockpit), subsurface (cave or passage) or connecting (sinkhole) morphologies. Caves or cavities that have formed during glacial lowstands are subsequently flooded during sea-level transgressive and highstand phases. Caves that become situated in the intertidal zone and exposed to waves can become subject to enhanced mechanical erosion through surging breaking waves. Hyperkarst is a process that may happen on limestone coasts at the marine-freshwater mixing zone, which is usually close to sea level. The mixing zone, which is oversaturated or even supersaturated of dissolved carbonates, is conversely undersaturated with respect to the lithology of the rock (e.g., near submarine freshwater springs). While hyperkarst processes may have a role in dissolving limestone, it is worth noting that the importance of hyperkarst processes in coastal areas spans more than one century (Agassiz, 1903; MacFadyen, 1930; Kelletat, 2005). .

Calcium carbonate precipitation. Once dissolved by karst processes, calcium carbonate is precipitated (often as high magnesium calcite or aragonite) from freshwater dripping, seeping, or flowing inside caves. Like karst dissolution, carbonate precipitation can only happen above sea level in the vadose zone, and is most frequent inside caves or overhangs, due to the lower partial pressure of carbon dioxide and higher temperatures in these environments. Calcium carbonate is also precipitated in the so-called "mixing zone", that is usually constrained between the marine phreatic and vadose zones on sedimentary coasts.

Biological

Bioconstruction. Marine organisms such as coralline algae, sponges, bryozoans, vermetids, serpulids, and hermatypic corals can build structures that resist the erosive force of waves and currents. These structures are often referred to as "organic reefs", defined as "*calcareous deposits created by essentially in place sessile organisms*" (Riding, 2002). Organic reefs can be built at any depth by different organisms, but they are particularly common in the euphotic zone both in tropical (e.g., coral reefs) and temperate areas (e.g., coralligenous algae bioconstructions).

Bioerosion. Several marine organisms, from fish to echinoderms to cyanobacteria, can remove hard substratum (either rock or bioconstructions) through physical or chemical means, or a combination of both (Davidson et al., 2018). Physical bioerosion often happens when organisms scrape, etch, bite, bore, or abrade the substrate. Rock boring by marine organisms includes a physical component but is often mediated by the secretion of acidic substances that can partially corrode (or dissolve) substrates containing calcium carbonate. **Bioprotection.** Some marine sessile biotas such as coralline algae, oysters, and possibly coastal biofilms, can build structures that protect rock surfaces exposed to seawater spray and splash from weathering processes, preventing abrasion. Distinct cross shore zonation of bioprotectors span in elevation from the sublittoral to the high supralittoral with factors such as wave energy, tidal range, and water quality influencing species assemblages and their indicative ranges with respect to mean sea level. Each species or species assemblage has a unique indicative range with respect to mean sea level. Their relevance as sea-level indicators is thus dependent on the ecological needs of the specific fossil biota.

Indicators

Tidal notches

Tidal notches (Figure 1 A-E) are meter-scale U-shaped landforms carved on limestone cliffs at or near sea level (Pirazzoli, 1986). Both bioerosion and hyperkarst have been considered as major factors in the origin of tidal notches. Some authors account for a combined action of these processes in tidal notch carving (Antonioli et al., 2015) whereas others point to either one or the other factor acting separately (Higgins, 1980; Schneider and Torunski, 1983). Tidal notches are usually more developed on micro- or meso- tidal areas and with low-to-moderate exposures to direct ocean waves. Under these conditions, geometric measurements of tidal notches (Antonioli et al., 2015) suggest that the maximum inflection point, generally referred to as "apex" or "retreat point", forms very close to mean sea level, with the notch amplitude (i.e., the distance from the floor to the roof) larger than the mean tidal range (MHHW to MLLW) but smaller than the maximum tidal range (HAT to LAT). Under higher surf conditions, the retreat point tends to develop higher than mean sea level (Pirazzoli, 1986), while in calmer waters (e.g., inside caves; Carobene, 2015), the base of the notch may be absent, giving the notch a "reversed L" shape. Tidal notches have been widely used to reconstruct past changes in relative sea level, at various time scales (Pirazzoli et al., 1989; Boulton and Stewart, 2015; Lorscheid et al., 2017).

Abrasion notches

In contrast with tidal notches, that are found exclusively on limestone coasts, abrasion notches (Figure 1 F,G) develop on most lithologies, where the action of the wave and the presence of loose sediments favors mechanical abrasion. Along modern coasts, it is possible to observe abrasion notches from well below sea level (at or above the breaking depth of average waves) to well above it (at or below the maximum reach of average wave runup). For this reason, abrasion notches are considered inaccurate sea-level indicators. If they are developed at the inner margin of a shore platform (see below), though, their indicative range can be reduced, especially when comparisons with modern analogs within the same physiographic unit are possible (e.g., Bini et al., 2014).

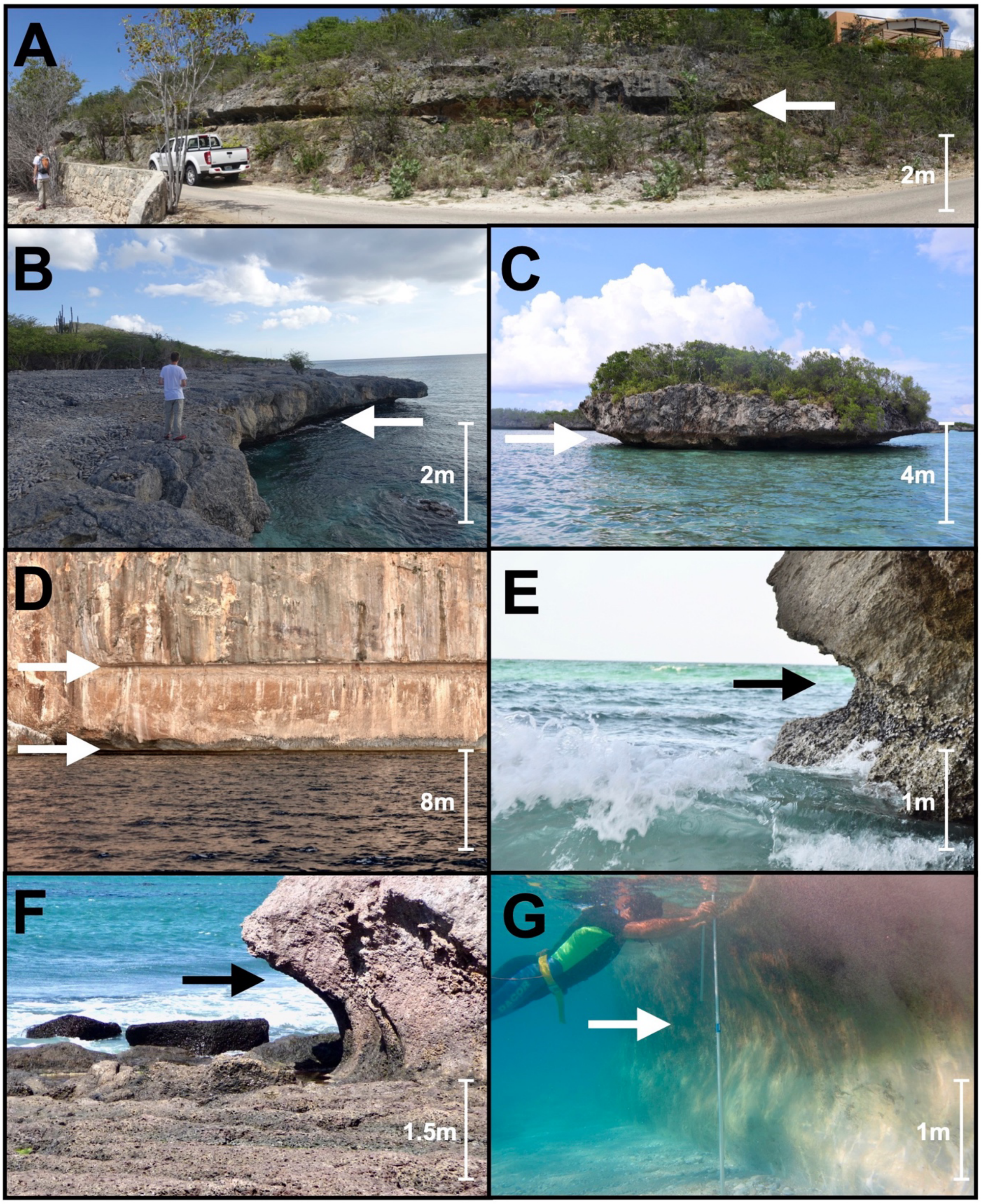

Figure 1 A) Fossil tidal notch (arrow), shaped during the Last Interglacial (125 ka) along the leeward side of the island of Bonaire, Lower Caribbean. Photo by A. Rovere via Wikimedia Commons (CC BY 4.0). B) Modern tidal notch (arrow), located few tens of meters from the location where the photo in Panel A was taken. Photo by A. Rovere via Wikimedia Commons (CC BY 4.0). C) "Mushroom rock", bordered by modern tidal notches (arrow) carved on Pleistocene limestones on the island of Aldabra, Seychelles. Photo by A. Rovere via Wikimedia Commons (CC BY 4.0). D) Relic (upper arrow) and modern (lower arrow) tidal notches along the coasts of the Golfo di Orosei, Sardinia, Italy. E) Tidal notch (arrow) carved at Salalah, Sultanate of Oman. Photo by M. Pappalardo. F) Modern abrasion notch carved above sea level at Puerto Deseado, Santa Cruz Province, Argentina. Photo by M. Pappalardo. G) Modern abrasion notch (arrow) carved underwater at Capo Noli, Liguria, Italy. Reprinted from Rovere et al., 2016 with permission from Elsevier.

Large-scale erosional landforms

Typically, tidal and abrasion notches represent the first stage in the process of rock coast retreat, as they are weakness points on the rock surface, in which mechanical erosion is enhanced. Similarly, mechanical erosion may act on fractures, joints or bedding planes in the rock and starts widening them. The final consequence of these processes is the formation of caves, sea stacks and arches (Figure 2 A-D), all forming within the same range as abrasion notches (from breaking depth of average waves to maximum runup height). These morphologies may be found as relic landscapes but are rarely used as sea-level indicators. In limestone bedrocks caves karstic in origin may have been open towards the sea and have been partly rehandled by waves action They may preserve deposits, bioconstructions, or other morphological elements that can be used for higher-accuracy paleo relative sea-level reconstructions.

Tafoni and honeycomb

Tafoni and honeycomb morphologies (Figure 2 E,F) are centimeter-to-meter scale landforms initiated by haloclasty on rocky coasts and evolving through a combination of processes triggered by thermal stress (Tingstad, 2008). They may develop well above sea level and have therefore little use as sea-level indicators, and they may be only used as “terrestrial limiting” constraints. This means that, if they are found as relic morphologies away from the potential effects of salt weathering, they indicate that paleo relative sea level at the time of their formation was below the location where they are found.

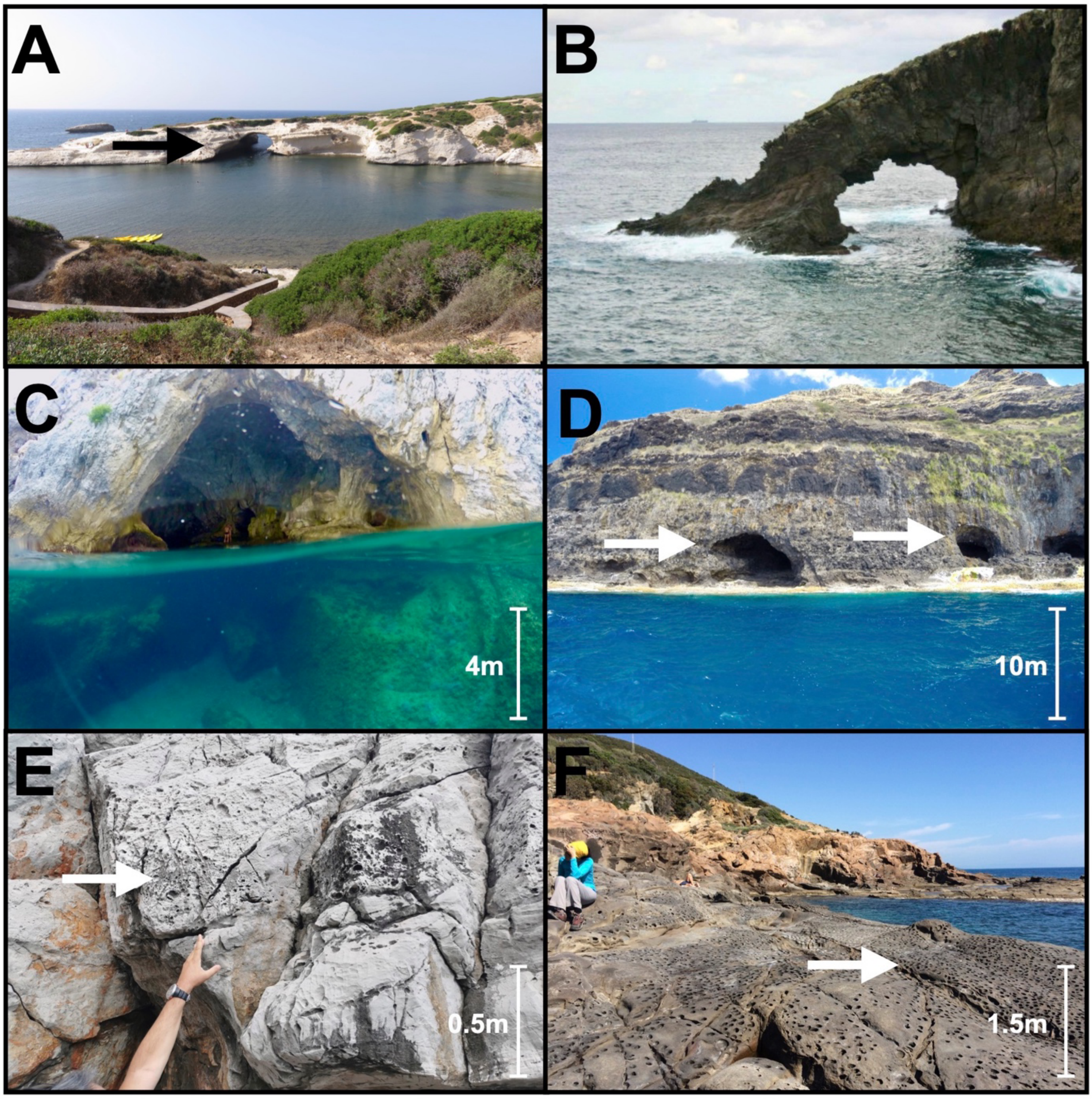

Figure 2 A) Sea arch (arrow) carved by modern waves on marly limestones at S'Archittu, Sardinia, Italy. Photo by E. Casella, used here with permission from the original author. B) Sea arch at Pantelleria, Sicily, Italy. Photo by M. Pappalardo). C) Coastal cave created by karst processes on limestones in Bergeggi, Liguria, Italy, and further shaped by coastal processes. Photo by A. Rovere via Wikimedia Commons (CC BY 4.0). D) Coastal caves (arrow), likely shaped by Pleistocene sea-level highstands carved into volcanic rocks in Santa Maria Island, Azores, Portugal. Photo by A. Rovere via Wikimedia Commons (CC BY 4.0). D) Tafoni and honeycomb holes (arrow) located few meters above sea level at Cap Ferrat, Cote d'Azur, France. Photo by Karla Rubio Sandoval, used here with permission from the original author. F) Tafoni and honeycomb holes (arrow) carved slightly above sea level at Calafuria, Tuscany, Italy. Photo by M. Pappalardo.

Shore platforms

Shore platforms (Figure 3 A,B) are horizontal or quasi-horizontal surfaces shaped by mechanical erosion, subaerial weathering and bioerosion (Stephenson, 2000) at or near sea level on rock coasts. The width of shore platforms depends primarily on lithological, structural, and environmental conditions (i.e., wave energy, exposure, and tide amplitude), and may range from tens to hundreds of meters. The defining elements of a shore platform are its inner margin and its seaward edge. The inner margin is the knickpoint connecting the sub-horizontal bedrock and the cliff. At locations where wave action is particularly intense and there are loose sediments on the shore platform, a wave ramp may develop as a beveled surface connecting the shore platform to the cliff. The outer edge of a shore platform is defined as "*the point where active erosion of the bedrock ceases at or landward of wave base*" (Kennedy, 2015). In general, shore platforms develop around mean sea level, with the inner margin most often shaped between MSL and MHHW. The outer edge is often located between MLLW and the breaking depth of incoming waves, where mechanical erosion on the rock is reduced. Typical smaller-scale morphologies that may characterize shore platforms are bioerosion marks or morphologies related to mechanical erosion, such as potholes.

Marine terraces

Marine terraces (Figure 3 C,D) are flat surfaces interrupting the continuity of the coastal landscape and are created by the interplay of marine erosional and depositional processes. They are usually larger in size than shore platforms (they may be up to few kilometers wide) and are often mantled by deposits of marine origin. In general, a marine terrace may include layered successions of deposits formed between the highest reach of storm waves (Storm Wave Swash Height, SWSH) and the breaking depth of average waves or even in the offshore. Close to the inner margin of marine terraces it is not unusual to find fossil beach deposits or other smaller-scale morphologies that may help refine this broad range. Marine terraces are among the most widely used sea-level indicators for which concerns past interglacials, as they can be preserved for a long time in coastal landscapes. In particular, stairs of marine terraces characterize areas with relevant uplift rates, such as the Pacific Coast of South America (Freisleben et al., 2021) or New Zealand (Ryan et al., 2021).

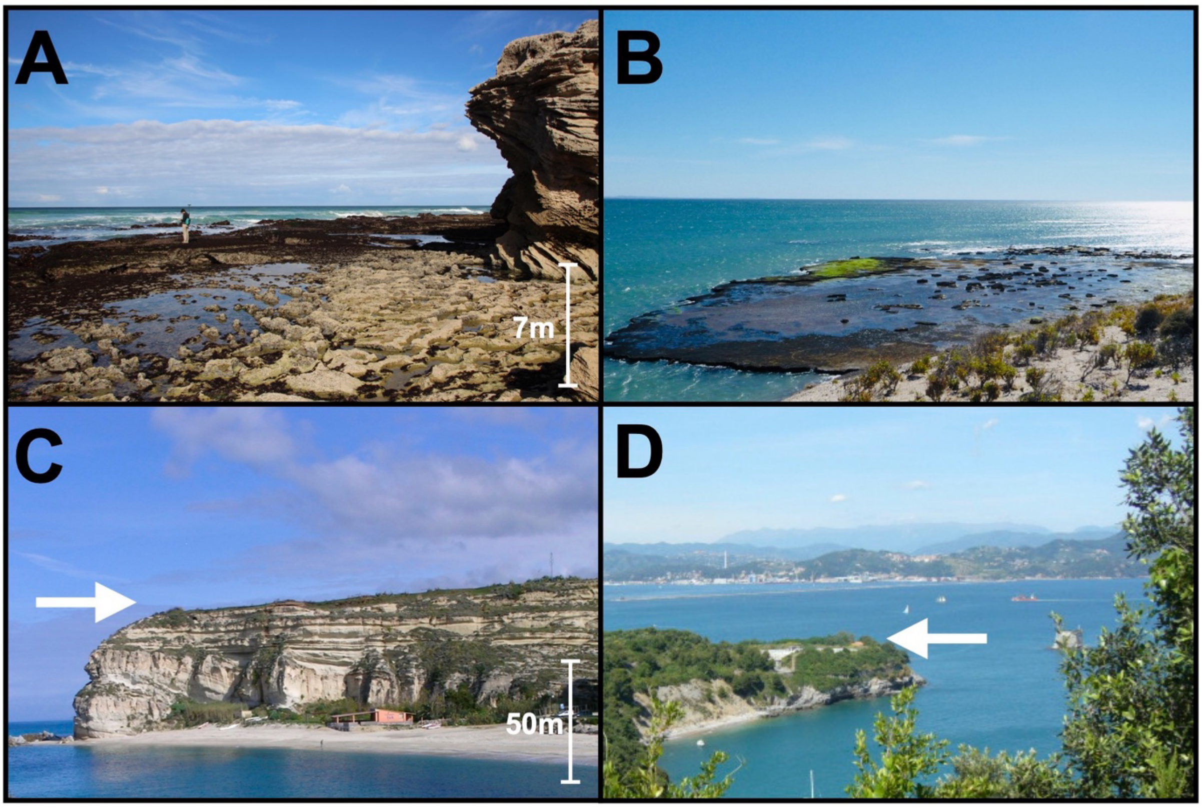

Figure 3 A) Shore platform carved into Pleistocene aeolianites in the De Hoop Nature Reserve, Republic of South Africa. Reprinted from Rovere et al., 2016 with permission from Elsevier. B) Modern shore platform carved at sea level at Caleta Olivia, Santa Cruz province, Argentina. Photo by M. Pappalardo. C) Marine terrace (arrow) at Riaci, Calabria, Italy. Photo by

A. Rovere via Wikimedia Commons (CC BY 4.0). D) Marine terrace (arrow)at Palmaria Island, Liguria, Italy. Photo by M. Pappalardo).

Speleothems

Speleothems are formed inside caves by the precipitation of calcium carbonate, which can only happen in subaerial environments. In general, submerged vadose speleothems (SVS) can be used as terrestrial limiting points, as the only information they bear is that the speleothem has formed in subaerial conditions (Surić, 2018). However, the presence, within the SVS, of marine bioconstruction layers may indicate episodic submergence (Dutton et al., 2009), and such layers may be then used as marine limiting points. Under particular environmental conditions, such as short distance (<300 m) from the coast and just below the water table (Ginés et al., 2012; Dumitru et al., 2021), Phreatic Overgrowths on Speleothems (POS) may develop. These are elongated speleothem morphologies that may form within the tidal range and can be dated with U-series if they contain limited calcite. While they have been found only in few areas globally (Majorca, Sardinia, Japan, Christmas Island, Mexico, Cuba), they are regarded as very precise sea-level indicators, also capable of giving insights into Pliocene relative sea-level changes (Dumitru et al., 2019).

Coastal deposits

Along soft coasts, particularly those with a net positive sediment budget, hydrodynamic processes will transport mobile sediment onshore creating nearshore bars, beaches, and beach ridges. Along transgressive coastlines, sediments may be isolated from wave action and cemented through the precipitation of calcium carbonate, creating an indurated coastal deposit. Fossil coastal deposits have characteristic bedforms that form in the subtidal (herringbone cross beds), intertidal (planar bedding) and supratidal (cross bedding) zones. Among them, beachrocks represent accurate indicators as they usually form at the interface between the marine phreatic and marine vadose zone (corresponding to the intertidal), however they may also form slightly above or below this zone (Mauz et al., 2015). Relic coastal deposits may be also found (either indurated or unconsolidated) in the facies of dune fields, imbricated boulder beaches, beach ridges, cheniers, lagoonal deposits, and intertidal planar-laminated sediments (Figure 4 A-D). The connection of each of these facies with a former sea level can be interpreted by comparison with modern counterparts. When no distinctive elements are present to constrain it, the indicative range of a beach deposit may be set to the general limits of fair-weather coastal transport, i.e., from the breaking depth of ordinary waves offshore to the ordinary berm onshore (Lorscheid and Rovere, 2019). Among the most widely used coastal deposits for paleo sea-level studies, there are saltmarshes (Shennan et al., 1983), where it is possible to exploit either the vertical zonation of foraminifera, or the presence of peculiar sedimentary layers within cores (e.g., peat layers) to reconstruct the former position of sea level (e.g., Leorri et al., 2010; Engelhart et al., 2015).

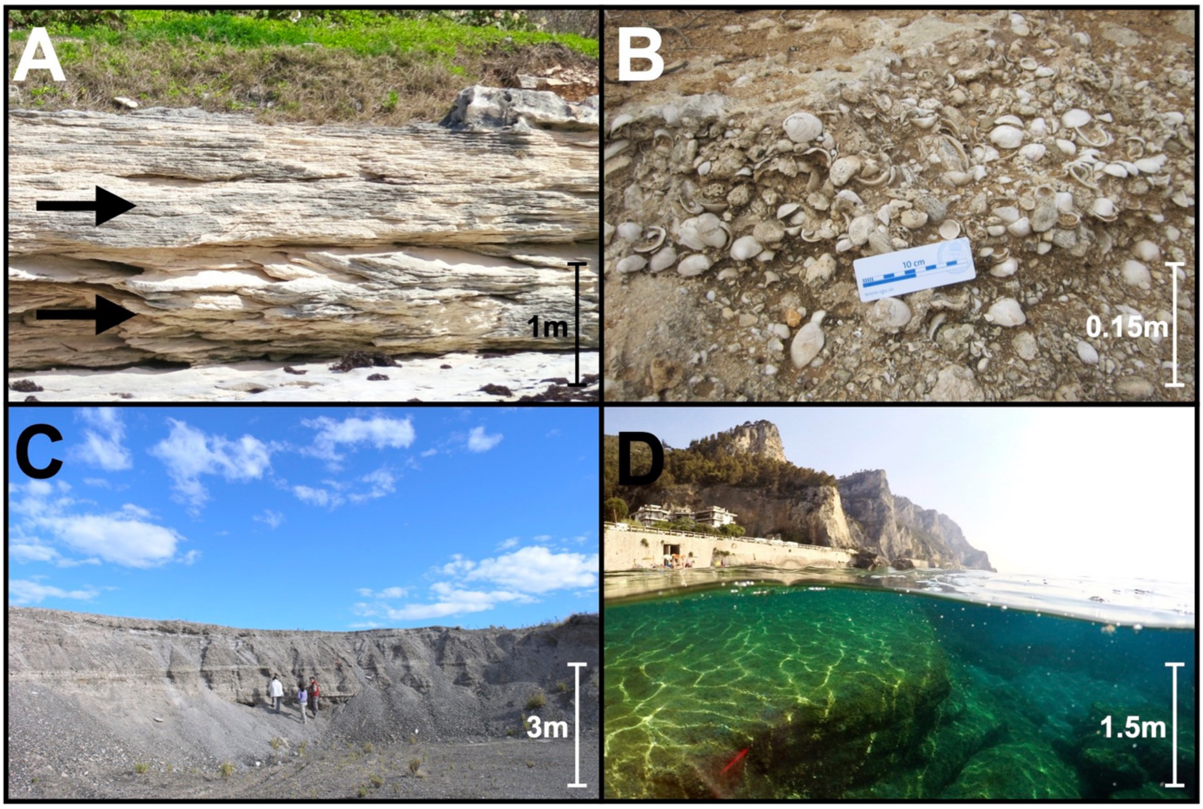

Figure 4 A) Shallow water (lower arrow) to intertidal (upper arrow) Pleistocene beach deposits in Bermuda. Photo by A. Rovere via Wikimedia Commons (CC BY 4.0). B) Cemented Pleistocene beach deposit on Pianosa Island, Tuscany, Italy. Photo by A. Rovere via Wikimedia Commons (CC BY 4.0). C) Back side of a beach ridge cut by roadworks at Camarones, Chubut Province, Argentina. Reprinted from Rovere et al., 2016 with permission from Elsevier. D) Submerged Holocene beachrock in Varigotti, Liguria, Italy. Photo by A. Rovere via Wikimedia Commons (CC BY 4.0).

Coral reefs

Tropical reefs are biological structures whose framework is constructed from Scleractinian corals, and further cemented and in filled by crustose coralline algae and reef derived sediments respectively (Figure 5 A). Reef growth is initially controlled by the underlying antecedent substrate, the rate of sea level rise and the ambient environmental conditions. The combination of these factors and can lead to reefs drowning (i.e., becoming stranded below the photic zone), catching or keeping up, or back stepping with sea level rise. Following sea level stabilization reefs will continue to aggrade vertically utilizing available vertical accommodation space or prograde laterally where reefs have reached base level forming reef flats. Here base level is defined as mean low water spring level (MLWS) which represents the vertical limit in which a coral can endure subaerial exposure. Massive corals growing at or around MLWS level tend to exhibit a circular microatoll morphology, whereby the colonial polyps on the corals upper surface become stressed through subaerial exposure and die with lateral growth occurring on the coral's margins. This microatoll growth form provides a very narrow indicative range (Woodroffe and McLean, 1990). Due to a corals reliance on Zooxanthellae (a photosynthetic symbiont) for nutrition, its life habit is constrained to the water columns photic zone which can vary from a few meters to many 10's of meters depending on the water quality (turbidity), and if taken in their totality can result in very large indicative range (Hibbert et al., 2016). Instead, by directly comparing reef geomorphic elements that typically have a tight indicative range (e.g., a reef flat, notch-visor, beach) rather than individual corals, it becomes possible to place a much higher confidence in the position of paleo sea level compared to an individual coral. At landscape scale, coral reefs may shape tropical coasts with staircase sequences created by the interplay of coral reef accretion and marine erosion, called reef terraces (Figure 5 B).

Biological rims

Outside the tropics, other kinds of biological buildups may be present close to modern sea level. In general, any fossil marine biota (e.g., barnacles or oysters) found in living position fixed to a natural or manufactured hard substrate can be used as a sea-level indicator. The accuracy in indicating past sea level depends on the biological and ecological properties of the single species, that determines its depth or elevational living range (Laborel and Laborel-Deguen, 1996). In the Mediterranean Sea, vermetid reefs (Laborel, 1986) are currently thriving in the warmest part of the basin (Levant, Northern Africa, continental Spain, and the Southern Mediterranean islands). Vermetid reefs (Figure 5 C) are bioconstructions built up by the gastropods *Vermetus triquetrus* and *Dendropoma petrae*um, in association with some coralline algae (e.g., *Lythophyllum byssoides*). They form seaward-protruding narrow platforms along rocky coasts. The living part of the reef is traditionally considered constrained to the intertidal. Dating of submerged relic rims was used to constrain Late Holocene sea-level rise (Sisma-Ventura et al., 2020). Also, coralline algae, such as *Lythophyllum* spp., are reported as reef-builders, and were used as Holocene sea-level indicators in a limited number of cases (Vacchi et al., 2016). Vermetid reefs dating to the Last Interglacial are comparatively rare (Sivan et al., 2016). Local physiographic factors, such as exposure to wave energy may enable the growth of biological rims also slightly above MHHW. Therefore, it is always advisable to analyse local modern analogs to establish the indicative range of biological rims.

Bioerosion marks

Although bioerosion effects are mostly visible at the microscopic scale, some boring species may produce macroscopic marks. Boring sponges (mostly in the Clionaidae family) produce mm-scale holes in the rock that are informative as marine limiting points. Boring mussels belong to a variety of genuses, producing boreholes with different shapes. In the Mediterranean, fossil boreholes are most often represented by *Lithophaga lithophaga* (Figure 5 D) and, more seldom, by *Petricola* spp. and *Coralliophaga* spp. These species may live down to several meters depth, but their density is high only in the first few meters below MLLW. For this reason, macro bioerosion trace fossils are considered reliable indicators of intertidal or shallow subtidal marine paleo-environments when they occur as a definite and dense band of holes with a sharp upper limit (Laborel and Laborel-Deguen, 1994). One of the key issues of bioerosion marks is to place chronological constraints on the timing of formation, in particular when they were formed during past interglacials. In this case, in fact, even if a shell is preserved inside the hole in living position, a suitable geochronological method for dating is unavailable. Minimum ages for macrobioerosion traces can be obtained dating continental carbonate concretions sealing the holes. Care should be devoted to convergence with other coastal landforms, such as tafoni or microkarst dissolution holes. Traces of grazing sessile biota, such as sea urchins and limpets, are very rarely preserved and difficult to detect without ambiguity. In case they are, though, their position is roughly correspondent to MHHW.

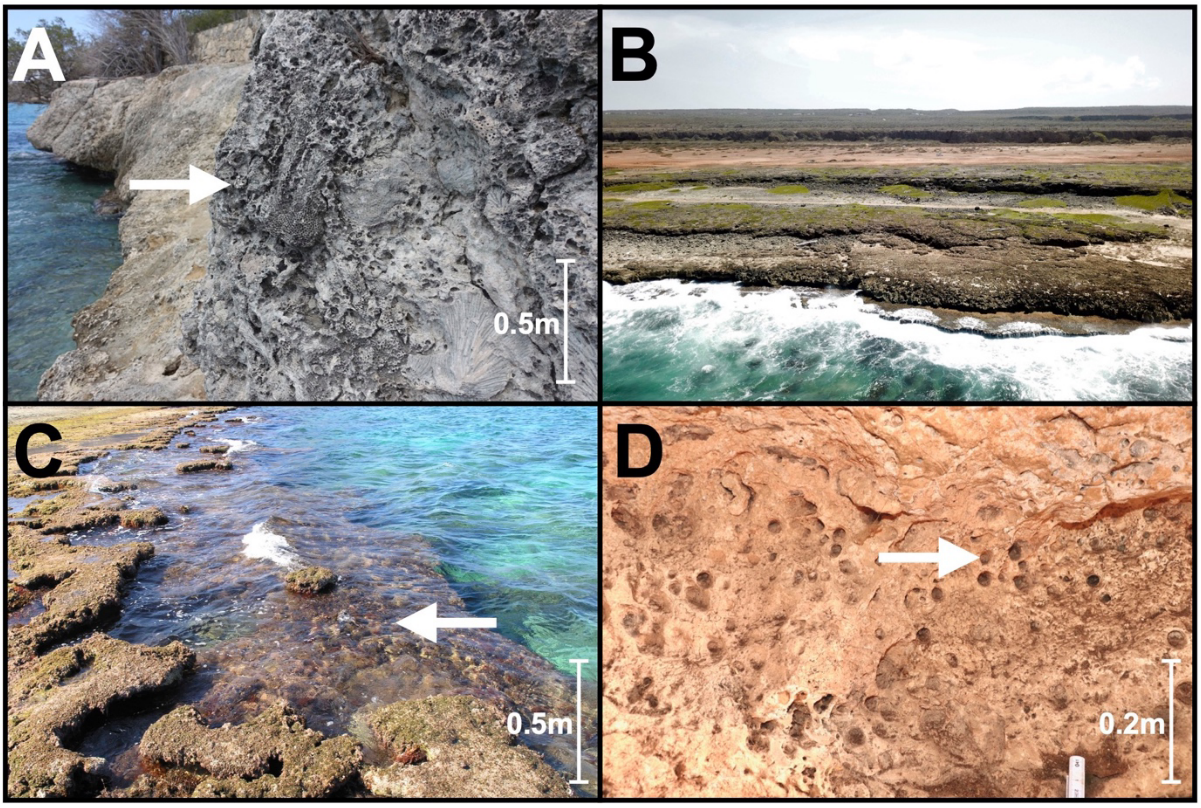

Figure 5 A) Fossil Pleistocene reef (arrow) few meters above modern sea level on the leeward side of the island of Bonaire, Lower Antilles. Photo by A. Rovere via Wikimedia Commons (CC BY 4.0). B) Sequence of Pleistocene reef terraces on the windward side of the island of Curacao, Lower Antilles. Photo by A. Rovere via Wikimedia Commons (CC BY 4.0). D) Living vermetid reefs at San Vito Lo Capo, Sicily, Italy. Photo by A. Rovere via Wikimedia Commons (CC BY 4.0). D) L. lithophaga *boreholes (arrow) in the Caviglione cave, near Ventimiglia, Liguria, Italy. Photo by A. Rovere.*

From geomorphological indicators to past sea levels

Geomorphological sea-level indicators are studied with the goal of reconstructing past changes in relative sea level, which may be in turn used to either gauge barystatic (i.e., eustatic) sea level changes or vertical land motions of different nature. Quantifying relative sea-level changes from a geomorphological sea-level indicator should always follow the steps summarized hereafter.

Step 1: elevation measurement. Once a geomorphological sea-level indicator is identified, it is necessary to measure its elevation or depth with the highest possible accuracy, preferably with sub-metric vertical positioning errors. Methods to achieve this level of accuracy may vary depending on the type of indicator. For indicators covering broad areas (hundreds of meters to kilometers), such as wide shore platforms, marine terraces or coral reef terraces, precise elevation may be gathered via LIDAR (Laser Imaging Detection and Ranging), drone or plane photogrammetry, multibeam or echosounder (the latter two for submerged terraces). For outcrop-scale (tens of meters) indicators, precise elevations may be gathered via differential GNSS surveys, total station, or levelling. Each elevation measurement should be associated with at least three main properties: i) error, expressed either as 1- or 2- sigma; ii) technique used; iii) sea-level datum to which the elevation is referred. Regarding the latter, in order of preference it is advisable to use orthometric heights referred to mean tidal level from long-term mareograph records, official tidal benchmarks, regional or national geoids, or global geoids. In case none of these solutions is available, it is possible to refer elevation measurements to an ad-hoc pressure sensor collecting data for a limited time (e.g., 4 weeks) benchmarked with differential GNSS. However, obtaining a sea-level datum in this way may carry an additional elevation error, which should be considered. For some geomorphological indicators closely related to tidal datums (e.g., microatolls), elevation is sometimes referred to the height of living counterparts (i.e., a biological sea-level datum). However, also for these indicators, it is advisable to report orthometric elevations alongside those referred to the biological sea-level datum.

Step 2: indicative meaning. After the elevation of an indicator has been measured, it is necessary to understand its relationship to the position of the sea at the time if formed, quantifying its indicative meaning (Shennan et al., 2015). In practical terms, this is achieved by either identifying and surveying modern analogs to the relic geomorphic indicator, or identifying which processes are responsible for the formation of a given geomorphic indicator or establishing the modern living range of fossil bioconstructions. This process is aimed at finding the upper and lower limits (U_L and L_L) of the environment of formation of the geomorphological sea-level indicator. These are then used to calculate its indicative range (IR) and the reference water level (RWL) as follows.

$$RWL = \frac{(U_L + L_L)}{2} \quad \text{(eq.1)}$$

$$IR = U_L - L_L \quad \text{(eq.2)}$$

It is worth noting that the same indicator will likely be characterized by different RWL and IR at distinct locations, depending on the intensity of the processes responsible for the formation of its modern analog (e.g., tidal ranges or wave height). Under certain circumstances, it is also worth investigating whether environmental conditions (e.g., waves, or tides) might have been different in the past, and take these changes into account in the definition of U_L and L_L.

Step 3: paleo relative sea level. Once the elevation (E) of a geomorphic indicator has been measured and the IR and RWL have been calculated as outlined above, the paleo relative sea level (RSL) and its associated uncertainty (σ_{RSL}) are derived using the following equations.

$$RSL = E - RWL \quad \text{(eq.3)}$$

$$\sigma_{RSL} = \sqrt{{\sigma_E}^2 + (IR/2)^2} \quad \text{(eq.4)}$$

Step 4: age determination. The next important step is to attribute an age to the geomorphological sea-level indicator, via absolute or relative dating, or via bio- or chrono-stratigraphy. Absolute and relative dating methods are most often applied to material of biological, sedimentary or geological nature within the indicator. Absolute ages are determined using different geochemical techniques to calculate an age with associated uncertainty. Among the most used absolute dating techniques are radiocarbon, U-series and luminescence. Relative dating techniques (such as amino acid racemization) do not provide absolute ages, but rather are used to discern the relative chronology among different sites and can be correlated with samples dated with absolute dating. Amino acid racemization can also be calibrated to provide numerical ages. Geomorphological indicators formed in past interglacials may also be assigned a general timing of formation via chronostratigraphy or biostratigraphy. This encompasses either correlating a site with another whose age has been constrained using absolute dating techniques or identifying within the site a biological indicator (e.g., a particular species living only at a restricted time in the area of interest), or geological elements (e.g., tephras) that allow establishing a timing of formation.

Step 5: paleo relative sea level. The paleo relative sea level as calculated from eq.4 and eq.5 corresponds to the sum of different factors. At geological time scales, these are barystatic sea level (often referred to as “eustatic”, due to change in mass of the oceans, excluding dynamic changes, Gregory et al., 2019), and vertical land motions (VLM), caused by geological processes (e.g., tectonics, sediment compaction, glacial isostatic adjustment). Most studies attempt to either infer barystatic changes by modeling or estimating the different contributions to VLM or disentangle some type of VLM (e.g., tectonics) using independent knowledge on barystatic changes and other VLM types (e.g., glacial isostatic adjustment and sediment compaction).

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