13 Coastal Geomorphology and Sea-Level Change

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Introduction

The intricate shores of the Mediterranean Sea twist and turn for some 46,000 km, with three-quarters of their convoluted length confined to only four countries—Italy, Croatia, Greece, and Turkey. Just over half the coast is rocky, much of it limestone, with the remainder encompassing almost every type of littoral environment (exceptions being coral reefs and mangrove wetlands) (Table 13.1). Such littoral diversity has long made the seaboard of southern Europe, the Levant, and North Africa a fruitful natural laboratory for studying coastal geomorphology and sea-level change. The virtually enclosed sea ensures that wave processes are generally modest and the tidal range is limited (often less than half a metre), a combination that permits observational evidence of many modern shoreline features to be related precisely to mean sea level. Consequently, relative shifts in the position of now relict coastal features can be used to track the rhythms of relative sea-level change and shoreline evolution.

Such rhythms have a bearing on several aspects beyond the physical geography of the Mediterranean basin: they inform archaeological reconstructions of the past settlement and exploitation of a coastal zone that has been an important focus of human activity since Palaeolithic times; they provide testing and fine-tuning for geophysical, geodynamic, and palaeoclimatic models for the region; and they set the backdrop to contemporary societal issues, such as future sea-level rise and coastline adjustments to mass tourism, which threaten the long-term sustainability of the Mediterranean littoral. In this chapter, we review these diverse facets of the Mediterranean coastal realm to provide a synthesis of how these shores have evolved into their present-day appearance.

Morphotectonics of the Mediterranean Seaboard

The Mediterranean occupies the convergence zone between two major tectonic plates, Africa and Europe, with a third, Arabia, pressing from the east. Caught within the collisional vice of these great plates are several minor plates and crustal blocks, most notably Anatolia and Apulia. The result is a complex network of plate tectonic structures that define the general configuration of the seaboard (Figure 13.1). In particular, two major subduction systems partition the Mediterranean basin into a patchwork of minor basins and subsidiary seas (Krijgsman 2002; Chapter 1).

In the eastern Mediterranean, the Hellenic arc subduction system and its former extension towards Cyprus have advanced southwards consuming the ancient Tethyan ocean floor of the Ionian and Levantine Seas, and in doing so has stretched open a major new seaway, the Aegean, in its wake. In the western Mediterranean, the Calabrian arc subduction system has similarly migrated south-eastwards destroying old Tethyan ocean floor ahead of it and rifting open a suite of successive young marine basins behind (the Alboran, Valencia, Balearic-Algerian, and Tyrrhenian) (Figure 13.1). Only the Adriatic is neither ancient ocean nor young rift, but instead is a drowned epicontinental

TABLE 13.1. *Coastal environments around the Mediterranean Sea classified into bedrock coasts and accretion coasts (which include beaches, dunes, marshes, lagoons, estuaries, and deltas)*

Country	Bedrock (km)	%	Accretion (km)	%
Spain	80	3	2,370	92
France	1,090	64	613	36
Italy	3,181	40	4,772	60
Malta	180	100	0	0
Yugoslavia	4,893	80	1,223	20
Albania	125	30	293	70
Greece	10,500	70	4,500	30
Turkey	3,115	60	2,076	40
Cyprus	391	50	391	50
Syria	119	65	64	35
Lebanon	146	65	79	35
Israel	10	5	190	95
Egypt	50	5	900	95
Libya	90	5	1,680	95
Tunisia	260	20	1,040	80
Algeria	600	50	600	50
Morocco	256	50	256	50
Total	**25,086**	**54**	**21,047**	**46**
North coast	**23,164**	**59**	**15,847**	**41**
South and east coasts	**1,922**	**27**	**5,200**	**73**

Note: More than 5% of Spain's coastline may be described as artificial. Yugoslavia includes data for Montenegro, Croatia, and Slovenia. The north coast is defined as Spain to Turkey in the data set above (and includes all the major islands for each country) and the south and east coastline is Syria to Morocco (including Cyprus). Total length of coastline = 46,133 km.

Source: Modified from Grenon and Batisse (1989).

platform. In geodynamic terms, therefore, the coastlines on and immediately inboard of the Hellenic and Calabrian arcs are the tectonically mobile borders of young active basins, while those around the periphery of the Mediterranean are, by and large, tectonically stable vestiges of the Tethyan passive margin (Figure 13.1; Chapter 1).

The geodynamic complexity of the Mediterranean ensures that the three main tectonic types of coastline coexist here: collision, trailing-edge, and marginal-sea coasts (Inman and Nordstrom 1971; Inman 1994; Davis 1996). Collision coasts, which characterize the narrow-shelf, mountainous seaboards of plates colliding with each other are arguably best typified by the Strait of Gibraltar collision zone where the African and European plates directly impinge. Trailing-edge coasts, which develop as wide-shelf plains on the rifted flanks of continents, characterize much of the gently warped and foundered North African margin. However, in general, the strong and pervasive tectonic partitioning in the Mediterranean means that its coastal configuration is best viewed as a nested set of marginal-sea coasts—narrow shelves fronting steep hinterlands along the shores of restricted seas enclosed by major land masses and island chains.

Littoral Cells

The fundamental units of study for coastal evolution, littoral cells, correspond to coastal compartments that delineate complete systems of sediment sources, transport paths, and sinks and within whose boundaries the budget of sediment is balanced (Carter 1988). The sediment dynamics of the Mediterranean littoral are strongly related to the size and character of these marginal-sea coasts. Such coasts, because they front onto smaller water bodies, are typically characterized by more limited fetch and reduced swell dynamics. In these settings, river deltas (Figure 13.2) become especially prominent and serve as important sources of sediment for littoral cells.

In the southern Mediterranean, marginal-sea coasts typically exhibit large sedimentation cells, some up to hundreds of kilometres in length. Indeed, one of the world's largest littoral cells stretches 700 km from Alexandria on the Nile delta to offshore northern Israel (Figure 13.3). Within this Nile littoral cell, sediment is swept eastwards from the delta mouth, 1 million $m^3\,yr^{-1}$ moving by wave-dominated longshore transport and 10 million $m^3\,yr^{-1}$ carried by coastal currents of the east Mediterranean gyre (Inman and Jenkins 1984; Chapter 2). This eastward drift of sediment is locally interrupted by eddy currents at prominent headlands such as the Damietta promontory, temporarily entraining sediment in migrating offshore sandy shoals or ribbons (Murray *et al.* 1981). Further east, the gradual northerly bend into the Levant coastline produces a progressive divergence in sediment flow, with the shallow longshore component becoming increasingly susceptible to wind action and feeding the extensive sand-dunefields along the coasts of the delta, Sinai, Gaza, and Israel (the 'dry' sink), while the shelf currents funnel the bulk into the Akhviz Submarine Canyon north of Haifa and into the Levantine basin (the 'wet' sink). It is no surprise that this extensive regionally simple sediment routeway has evolved along the long-lived and stable margin of the ancient Tethyan basin.

Elsewhere in the Mediterranean, more active and complex geodynamics foster more complicated sedimentation cells. In the west, the Alboran Sea presents a younger and tectonically active basin in which a more restricted and intricate seaway combines with intense surface water gyres (Chapter 2) and strong winds from the Azores high-pressure cell to create smaller and more convoluted sediment transport cells

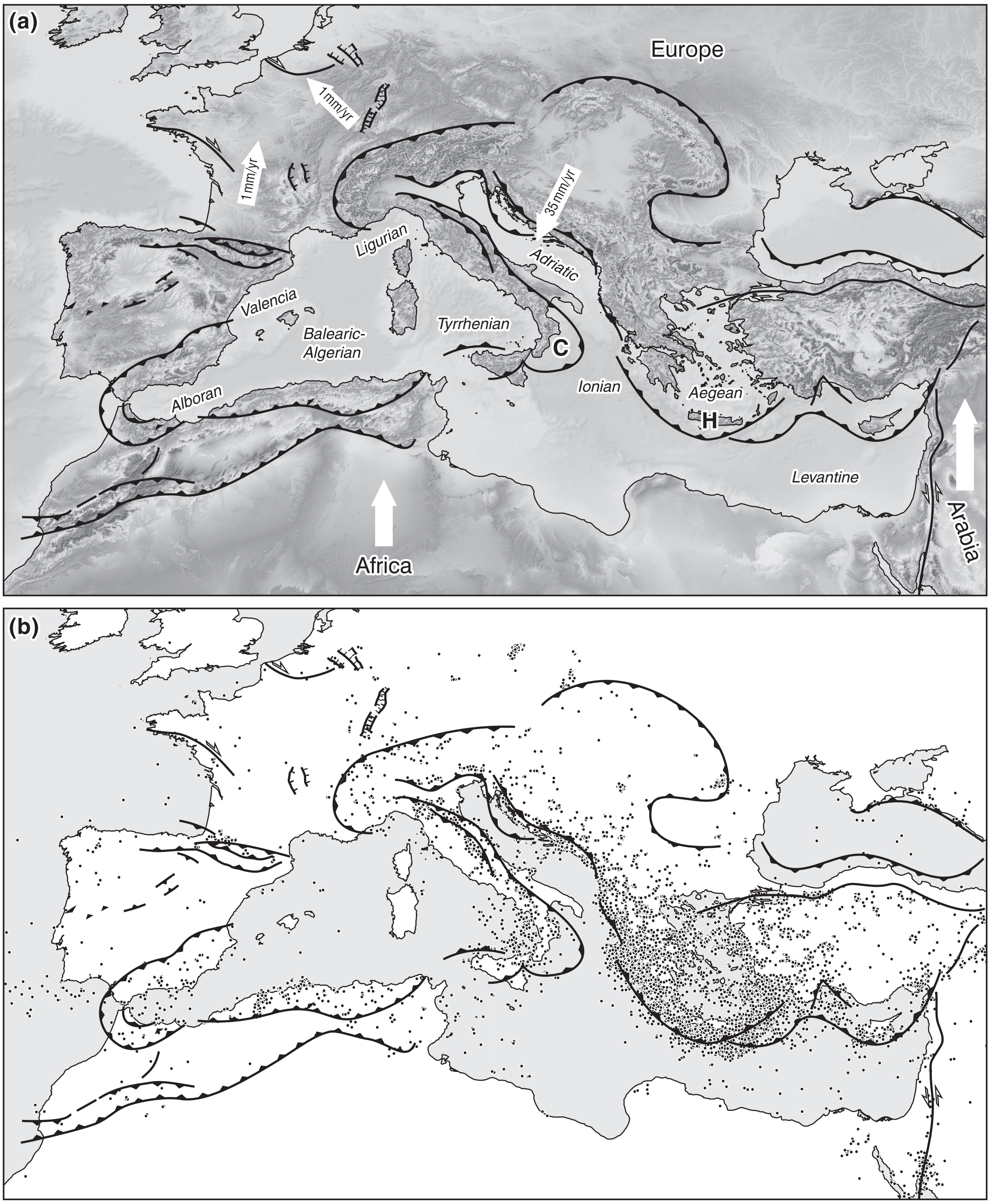
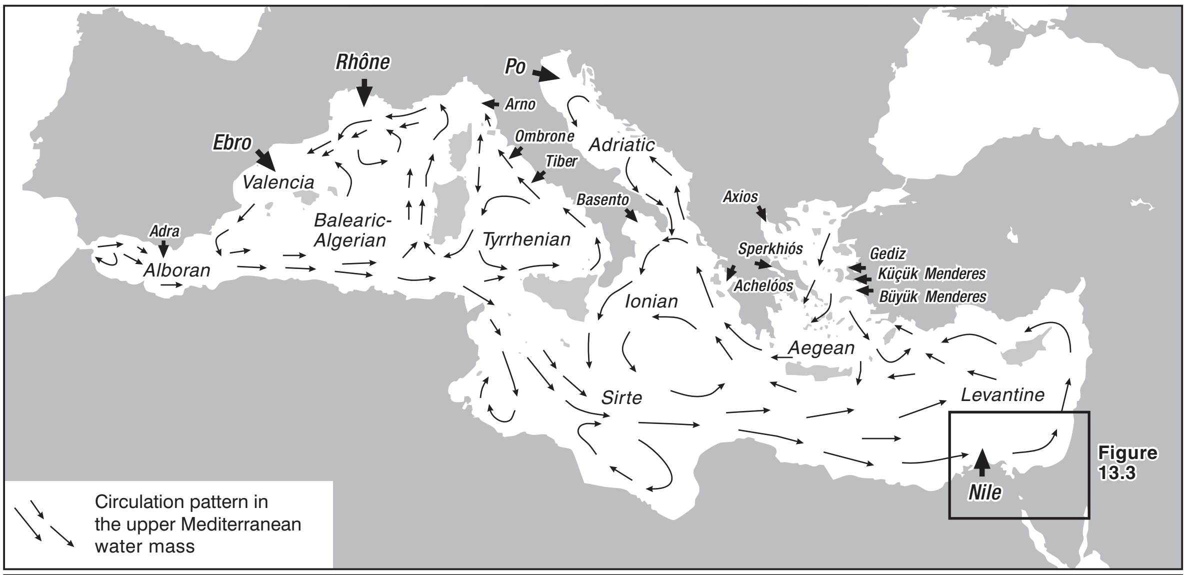


Fig. 13.1. Major tectonic structures (a) and associated seismicity (b) of the Mediterranean region, highlighting how active geodynamic zones define the general coastal configuration of the region. The coastlines on and immediately inboard of the Hellenic (H) and Calabrian (C) arcs are the tectonically mobile edges of young active basins, while the coasts around the rest of the Mediterranean are generally much more tectonically stable remnants of the Tethyan passive margin (Chapter 1).



Delta Characteristics	Ebro	Rhône	Po	Nile
Plio-Quaternary thickness (m)	1,000–1,500	1,000–1,500	5,000	3,500
Holocene thickness	30	35	30	modified
Delta area (km^2)	350	720	770	22,000
Continental shelf area (km^2)	9,000	9,000	20,000	17,000
Shelf width (km)	65	70	—	70
Shelfbreak (m)	100	100	100	250
Onset of delta construction	early Holocene	early Holocene	early Holocene	early Holocene
Annual discharge (tons/year) historical time	—	60×10^6	—	140×10^6
Annual discharge (tons/year) modern time	3.5×10^6	3×10^6	20×10^6	—
Delta-front advance (km/century)	5	4	0.5–7	1.5

Fig. 13.2. Coastal morphodynamics of the Mediterranean basins showing the general near-surface water circulation pattern (see Chapter 2) and the locations (large triangles) and attributes (tabulated data) of the four major delta shelves: Ebro, Rhône, Po, and Nile (after Got *et al.* 1985). Small triangles indicate the location of other deltas discussed in the text. The box shows the area in Figure 13.3.

(Goy *et al.* 2003). Most of the southern coast of Spain trends east–west, roughly parallel to the prevailing winds, and this situation favours longshore currents and littoral drift. However, where prominent bays and headlands break this trend, the sediment transport routes are intercepted and occasionally locally reversed, thereby delimiting second- and third-order littoral cells. Within the even more tectonically fragmented coastal landscape of the circum-Aegean Sea (Figure 13.1), fault-bounded gulfs and uplands produce a complex shelf and nearshore bathymetry that creates variable patterns of sediment routing over distances of a few tens of kilometres to kilometres. These often switch in character across active tectonic structures creating a heterogeneity of sediment sources and sinks and a collage of littoral environments (Leeder *et al.* 1991; Collier *et al.* 1995).

Tectonics, Climate, and Sea Level

As well as partitioning littoral sedimentation systems, geodynamic processes determine whether the Mediterranean shores are emerging or subsiding (Milliman 1992). The fastest geodynamic movements are occurring at rates of 30–40 mm per year horizontally and about 1 or 2 mm per year vertically, which mean that over thousands to millions of years their cumulative effects make dramatic changes to the coastal configuration. The most pervasive effect of this has been the gradual closure of the seaway by Africa–Europe convergence, a process responsible for triggering a regional dessication event, the Messinian Salinity Crisis (Hsü *et al.* 1973), which in turn predicated the large-scale reshaping of the Mediterranean seaboard (Chapter 1).

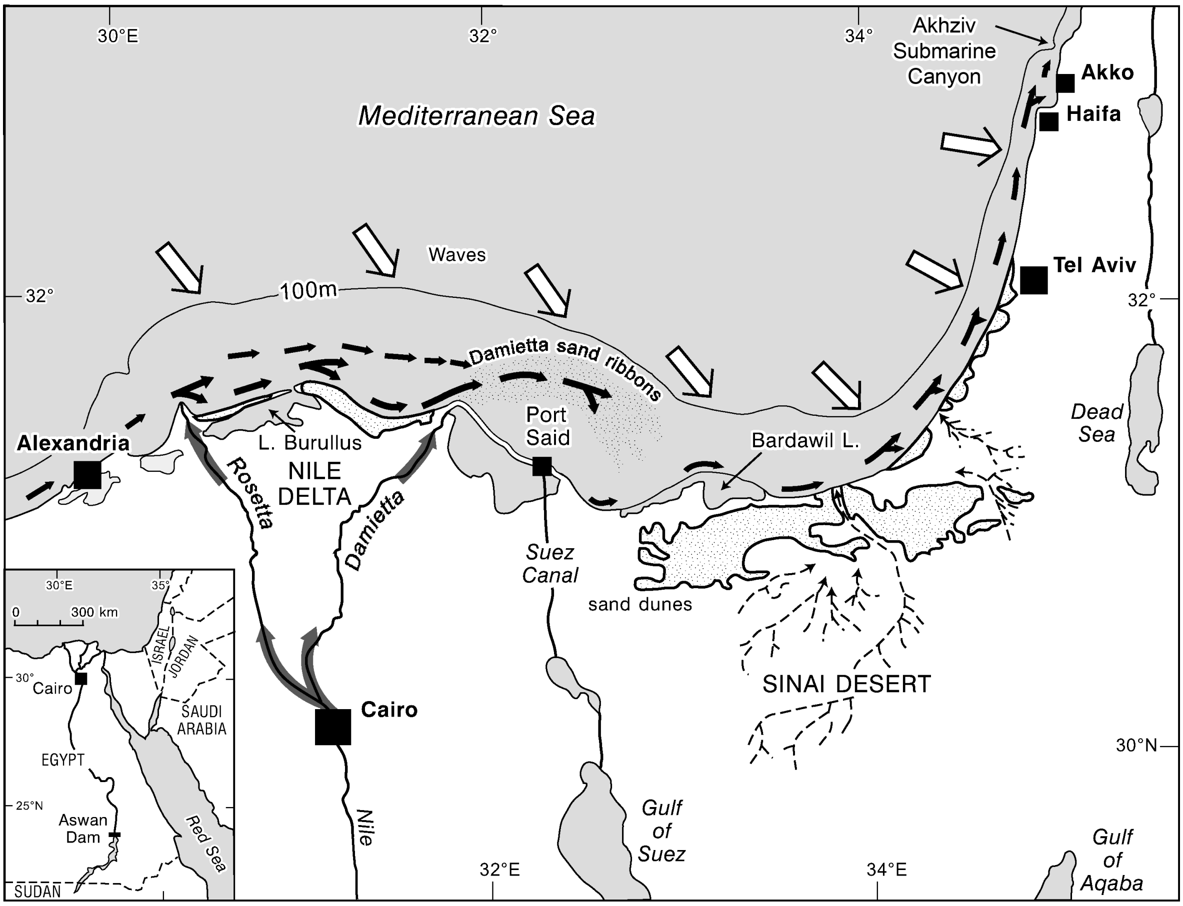


Fig. 13.3. The Nile littoral cell extends along the south-eastern Mediterranean coast from Alexandria, Egypt, to the Akhziv Submarine Canyon, Israel. Sediment transport paths are shown by dark arrows (modified from Inman and Jenkins 1984).

Several million years of subduction-related tectonic emergence of the Betic-Rif area had gradually shallowed the narrow seaways that connected the Mediterranean to the Atlantic Ocean (Krijgsman *et al.* 1999; Warny *et al.* 2003; Duggen *et al.* 2003, 2004). Around 5.96 million years ago, this tectonic uplift, aided perhaps by global glacio-eustatic processes (Clauzon *et al.* 1996; Hodell *et al.* 2001), finally closed the Atlantic gateway. Over the course of the next few hundred thousand years, evaporation of the now largely landlocked basin dropped water levels by at least 800 m, and probably as much as 1,300 m, below their modern equivalent (Ben Gai *et al.* 2005). Thick sequences of evaporites were deposited in the hypersaline abyssal plains. Around the shores of the Mediterranean salt lake the plummeting base-level triggered a major phase of river downcutting, creating vast planation surfaces and carving extensive canyon networks. The most formidable river, the Nile, cut a canyon that was three times longer than the Grand Canyon with a similar depth and width (Said 1981), and major rock-cut gorges and cataracts are known to exist beneath the Rhône, the Ebro, and the Po rivers. Pronounced coastal progradation at these canyon mouths would form the earliest submarine cones of the region's great deltas. Today, infilled palaeo-valleys can be found along much of the Mediterranean coast and its continental shelf, vestiges of this Messinian incision (Chapter 1).

This 'great drying' ended about 5.3 million years ago, when the marine gateway to the Atlantic was restored, due to the westward propagation of the Alboran rift basin (Duggen *et al.* 2003, 2004) and/or eastward piracy of the Atlantic waters (Blanc 2002; Loget *et al.* 2005) breaching the Betic-Rifian land bridge at Gibraltar to create the present-day straits.

Since the Messinian Salinity Crisis, the Mediterranean waters have been maintained through a fragile balance between tectonics and climate, much of it dependent on east–west differences in salinity that reflect a water budget deficit in which outputs from evaporation exceed inputs from precipitation and freshwater influx (Chapters 2 and 8). The long-term effects of this in terms of sea-level changes during Plio-Pleistocene times are poorly constrained, because the Mediterranean lacks both the sensitive marine-continental switches of the neighbouring Red Sea (e.g. Siddall *et al.* 2004) and the coral-reef staircases, which in locations such as Barbados, the Huon Peninsula, and Tahiti, have yielded long-term eustatic records (for a review, see Lambeck and Chappell 2001). Nevertheless, some constraints on Pleistocene sea-level oscillations have been provided by stratigraphical sequences in subsiding coastal plains and shallow shelves (e.g. van Andel *et al.* 1990) or in drowned littoral caves (Fornós *et al.* 2002; Tuccimei *et al.* 2003; Antonioli *et al.* 2004). On emerging coastlines, records of sea-level highstands have been derived from flights of marine terraces (e.g. Keraudren and Sorel 1987; Goy and Zazo 1988; Dumas *et al.* 1993; Carobene and Dai Pra 2003; Zazo *et al.* 1999, 2003; Rodríguez-Vidal *et al.* 2004; Uluğ *et al.* 2005).

Coastal Tectonics

As well as revealing long-term (10^4 to 10^6 year) sea-level records, the ancient shorelines of the Mediterranean can also be used as a measure of geodynamic activity. A particularly useful marker is the shoreline that formed during the Last Interglacial period, 120,000–130,000 years ago, the climatic optimum of Marine Isotope Stage (MIS) 5.5. This episode has left an especially indelible trace in the Mediterranean, in part because global sea levels stood 3–6 m higher than present, and in part because the unusually warm climatic conditions favoured the development of distinctive faunal assemblages. The fossil *Strombus bubonius* is an index marker of this highstand (Bordoni and Valensise 1998).

Figure 13.4a shows the elevation of this geomorphic marker across the Mediterranean realm. In the Strait of Gibraltar, direct continental collision has locally raised the MIS 5.5 terrace to 20 m above sea level, but this drops abruptly back to being within a few metres of modern sea level along most of Spain's Atlantic and Mediterranean coasts (Zazo *et al.* 1999, 2003). Other plate-boundary structures coincide with elevated MIS 5.5 terrace elevations, notably where the horizontally slipping North Anatolian Fault intersects the Marmara Sea coastline (Yaltirak *et al.* 2002), and more significantly in north-eastern Sicily and south-western Calabria (Miyauchi *et al.* 1994; Bordoni and Valensise 1998; Ferranti *et al.* 2006), which appear to be on the rise following the Late Pleistocene detachment of the subducted African slab (Westaway 1993). Elsewhere along the Africa–Europe subduction and collisional front, however, MIS 5.5 shorelines are roughly at their original elevation. Perhaps surprisingly, the highest Last Interglacial terraces in the Mediterranean fall not on a plate-bounding but on a prominent interplate rift zone, the Gulf of Corinth fault system (Keraudren and Sorel 1987; Armijo *et al.* 1996). No other rifted gulfs of the circum-Aegean coast attain such high Last Interglacial shoreline elevations (Kelletat *et al.* 1976), and along the opposing Turkish Aegean coast no MIS 5.5 terrace sites have so far been reported (Brückner *et al.* 2004). Along the Mediterranean coast of Turkey, Pirazzoli (1991) attribute higher shorelines to MIS 5.5 but these have not been dated.

The general tendency for Last Interglacial marine terraces to lie within a few metres of present-day sea level indicates general long-term land stability throughout most of the circum-Mediterranean region. The marked variability in terrace elevation highlights the role of local (basin-bounding) rather than regional geodynamics. Both these characteristics are also apparent in the Holocene and modern tide-gauge records from the Mediterranean (Flemming 1978, 1993; Emery *et al.* 1988; Zerbini *et al.* 1996). According to these recent sea-level data, most of the Mediterranean coast appears to be submergent, although mostly at a low rate (−1.2 mm/yr, Emery *et al.* 1988). The zones of little or no sea-level change correspond to the coastal tracts between Gibraltar and Genoa in the west and along southern Turkey in the east. In contrast, the most mobile shores are mostly on or immediately inboard of the Hellenic and Calabrian arc areas of the central northern Mediterranean, where they coincide with zones of high earthquake activity (e.g. Pirazzoli *et al.* 1996*a*; Stewart *et al.* 1997; Chapter 16) or active volcanic centres (Dvorak and Mastrolorenzo 1991; Firth *et al.* 1996; Stiros 2000; Morhange *et al.* 2006; Chapter 15). The overall picture, therefore, is of considerable variation in both the amount and sense of vertical coastal movements along the Mediterranean littoral.

Isostatic Responses to Ice-Ocean Loading

In addition to tectonic deformation, land movements in the Mediterranean littoral are subject to

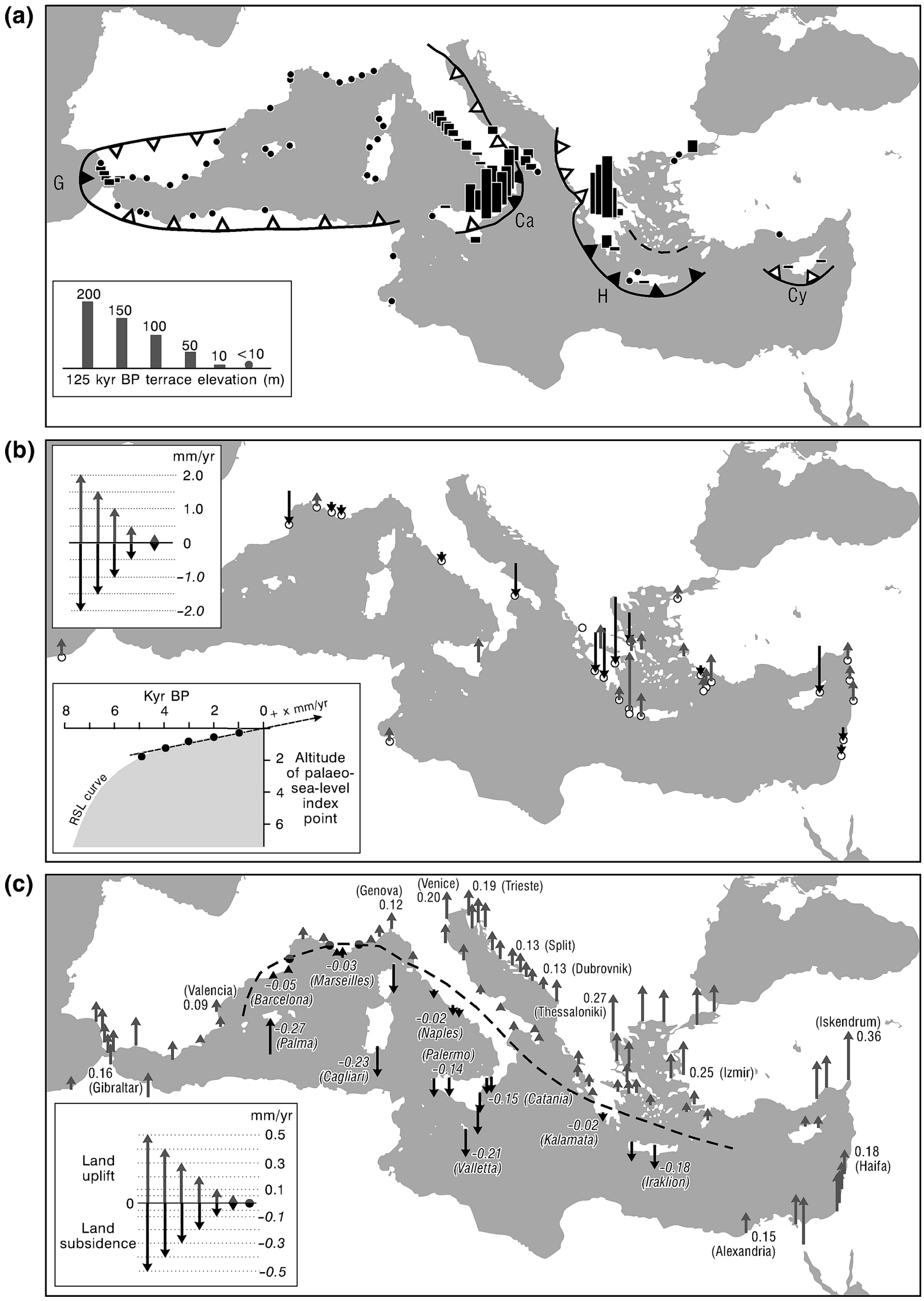


Fig. 13.4. (a) Elevation of the Last Interglacial (Marine Isotope Substage 5.5) shoreline based on a compilation of published data. Terrace data from Antonioli *et al.* (2006), Conchon (1975), Kelletat *et al.* (1976), Bordoni and Valensise (1998), Zazo *et al.* (1999, 2003) and Yaltirak *et al.* (2002). Solid lines trace the main geodynamic arcs (Ca = Calabrian, Cy = Cyprus, G = Gibraltar, H = Hellenic), with solid triangles indicating active subduction fronts and open triangles denoting collisional fronts. (b) Rates of late Holocene crustal movement derived from sea-level curves in Pirazzoli (1991) augmented by more recent sea-level studies. Black downward arrows denote subsiding areas and grey upward arrows indicate emerging areas. (c) Predictions of global isostatic adjustment made for Mediterranean tide-gauge stations, updated from predictions of Peltier (2001) by adopting the new analysis of Peltier (2004) and using values listed at <www.pol.ac.uk/psmsl/peltier/index.html>, accessed 27 October 2008.

glacio-hydro-isostatic crustal effects, whereby the marine basin adjusts to past and ongoing fluctuations in surface loads of ice and water. Various numerical models have emerged which make competing predictions on the shoreline responses to ice-ocean loading (Lambeck 1995; Peltier 1998, 2000; Lambeck and Bard 2000). The intricacies of these ice-history/earth-rheology models are outside the scope of this chapter, but there are essentially two key contributions to land movement. The first, a glacio-isostatic contribution, comes from rebound of the former ice-mass centres of Europe and North America, whereby mantle material squeezed beneath the Mediterranean crust by ice-sheet depression now flows back causing the previously upwarped Mediterranean 'forebulge' to subside. Figure 13.4c shows the predicted effect of this glacial isostatic adjustment in the Mediterranean. The second, a hydro-isostatic contribution, results from meltwater from ice-sheet decay increasing the water load of the global oceans and seas, thereby downwarping the marine basin floors and upwarping their margins (Lambeck *et al.* 2004*a*, 2004*b*).

The application of these models to the Mediterranean littoral is discussed in Pirazzoli (2005), but key points are noted here. In some models, the combined effect of both glacio-isostatic and hydro-isostatic changes is to accentuate general subsidence across much of the central Mediterranean and induce compensatory upwarping of its margins (Lambeck 1995; Lambeck and Purcell 2005). However, according to Peltier (1987) the Mediterranean region is sufficiently remote from the North European centre of glaciation as to be little influenced by the collapse of its forebulge; instead, relative sea-level movements ought to be more significantly influenced by the local water load emplaced on the basin itself. The models also highlight local idiosyncracies. For instance, along parts of the North African and Levantine coast, Lambeck and Purcell (2005) predict that the two isostatic contributions can potentially counteract each other, whilst in the northern Adriatic the Alpine deglaciation probably amplifies the glacial isostatic signal (Lambeck and Purcell 2005). In both areas, accentuated coastal emergence is postulated, making it possible that mid- to late Holocene land uplift may have temporarily outpaced sea-level rise to create locally a highstand shoreline. Figure 13.5 shows sea-level and shoreline predictions for the Mediterranean region for four time slices at 20 ka, 12 ka, 6 ka, and 2 ka

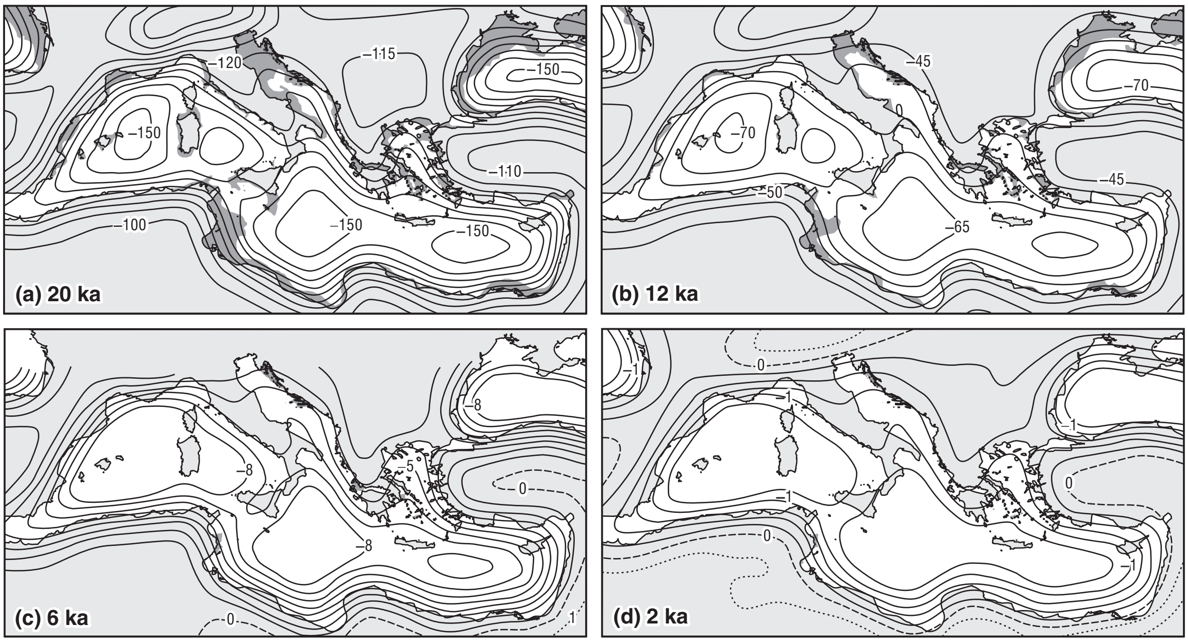


Fig. 13.5. Predicted relative sea levels and shorelines across the Mediterranean region at four epochs: a = 20 ka, b = 12 ka, c = 6 ka, and d = 2 ka. The palaeoshoreline positions are defined by the seaward side of the grey shading. For 20 and 12 ka BP, the contour intervals are 5 m. For 6 and 2 ka BP the solid contours denote negative values, the finely dashed contours denote positive values, and the dashed contours correspond to zero change (modified from Lambeck and Purcell 2005).

respectively. Note the expansion of the coastal plain at the global LGM when sea level was around 120 m lower than today.

In general, however, glacio-isostatic model simulations confirm the findings of most sea-level studies, i.e. that postglacial Mediterranean sea levels have never been higher than at the present day. In other words, any signs of Holocene coastal phenomena found elevated above the modern waterline is strong evidence for local tectonic movements. It is a reminder that geological evidence for past sea levels provide the testing ground for these global isostatic models (e.g. Pirazzoli 1998). Thus, in a comparison between several late Holocene sea-level histories and glacio-hydro-isostatic predictions, Pirazzoli (2005) contends that, in non-tectonic areas, hydro-isostatic effects have been, in places, overestimated (e.g. Sardinia) and in other places underestimated (e.g. southern Tunisia). Pirazzoli (2005) disputes predictions of isostatic subsidence along the Hellenic arc and on the Levant coast. Resolving the discrepancies between model and field data will continue to refine our understanding of Mediterranean postglacial shoreline change, and in the following sections, we discuss how such geologically derived sea-level constraints are determined.

Postglacial Sea-Level Changes

Tectonic and glacio-hydro-isostatic effects ensure that there is considerable spatial and temporal variability in land movements within the Mediterranean realm, but the general pattern of postglacial eustatic sea-level rise is adequately known (Lambeck and Chappell 2001). This is known thanks mainly to 'global' sea-level curves based upon coral-reef reference sites (e.g. Barbados, the Huon Peninsula, and Tahiti) which show that world sea levels rose by 120–30 m, largely between 16,000 and 8,000 years BP, thereafter stabilizing at present levels by 6,000 years BP (Pirazzoli 1991).

However, tracking sea-level behaviour following this mid-Holocene stabilization is more difficult, because corals are low-precision reference markers whose vertical range of repartition in modern reefs is of an equivalent magnitude to eustatic variations over the last 6,000 years (Blanchon and Shaw 1995). This mid- to late Holocene period, however, is critical for isolating various non-eustatic dynamic factors such as isostasy, tectonics, and sea-surface topography that may induce significant local sea-level fluctuations (e.g. Mörner 1996). The Mediterranean littoral, however, offers an opportunity to establish such detailed late Holocene sea-level histories because it combines a diversity of potentially high-precision sea-level proxies with a microtidal regime. As is discussed below, the evidence of former shorelines around the Mediterranean region is key to establishing the local relative sea-level trajectories needed to refine tectonic and glacio-isostatic models.

Evidence of Former Sea Levels in the Mediterranean Littoral Zone

A wide range of Mediterranean shoreline indicators can be employed for establishing Holocene sea-level curves ranging from precise 'sea-level index points' that securely tie the elevation of the land–sea interface at a specific period in time, to more equivocal and/or less well-dated indicators of former submergence or emergence (Haslett 2000). General reviews are provided by van de Plaasche (1986) and Pirazzoli (1991), but in brief they include a suite of sedimentological, geomorphological, geoarchaeological, and biological proxies.

Sedimentological Proxies

Sedimentological evidence provides the most widespread but typically least precise record of shifting shorelines. Fossiliferous sands containing gastropods and bivalves equivalent to those of the modern infralittoral zone can be used, though altitudinal errors may be as large as 10 m. Such metre-scale precision can also be expected from cores from coastal sediment sequences using grain size or fossil fauna to identify contrasting littoral environments (such as terrestrial floodplains, freshwater lakes, seasonal pools, brackish lagoons, beaches, marine delta fronts, and prodelta slopes) and assess the extent of altitudinal change. Even where sharp interfaces between intertidal muds and subaerial peats provide reliable sea-level index points, problems of sediment compaction generally maintain altitudinal uncertainty. In Provence, for example, Vella and Provansal (2000) analysed basal peat formations from the eastern limit of the Rhône delta above a non-deformable substrate and determined an altimetric precision of ± 50 cm at best.

Clastic shoreline deposits (beachrock) have been used as sea-level indicators where they contain early diagenetic carbonate cements whose textures can be diagnostic of intertidal cementation; though an error of +1 and −5 m may be expected (Fouache *et al.* 2005*b*). A potentially better positional accuracy can be obtained from the drowning of littoral caves, whereby speleothems precipitated in emergent caves during lowstands subsequently become encrusted by marine biogenic overgrowths when flooded with sea water (Vesica *et al.* 2000;

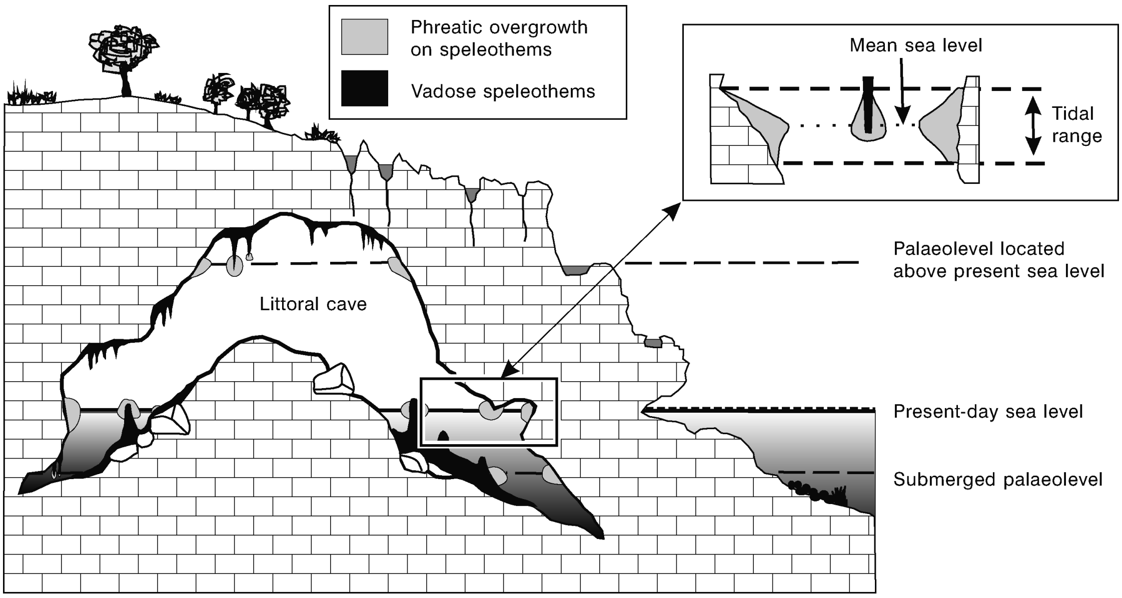


Fig. 13.6. Schematic representation of a littoral karst cave based on observations on Mallorca Island and Capo Caccia (Sardinia). Note the presence of phreatic overgrowths on speleothems related to former and present-day sea levels (modified from Tuccimei *et al.* 2003).

Bard *et al.* 2002; Antonioli *et al.* 2004; Chapter 10) (Figure 13.6). Complexities include the geometry of the cave system and the fact that there is often a hiatus of several millennia between the end of speleothem formation and marine overprinting caused by the influx of fresh groundwater ahead of saline inundation (Surić *et al.* 2005).

Geomorphological Proxies

The geomorphological form of the coast reflects the local balance of bioerosion and bioconstruction. On rocky shores, bioerosion at mean sea-level forms mid-littoral notches and tidal platforms. These are best developed on limestone, where they are carved by boring and grazing organisms and dissolved and abraded by wave action. Being a combination of physical, chemical and biological erosion, the nature of the local context strongly controls their precise form (Pirazzoli 1986*a*; Kershaw and Guo 2001; Naylor and Viles 2002; Trenhaile 2002). The effects of variations of coastal wave energy, for example, produce a continuum of notch forms whose precise relationship to mean sea-level can, in some instances, be difficult to assess (e.g. Kershaw and Guo 2001) (Figure 13.7). Nevertheless, notches and associated platforms are well developed around many carbonate coasts in the Mediterranean, and whether emergent (e.g. Pirazzoli *et al.* 1994*a*, 1996*a*; Stewart *et al.* 1997; Rust and Kershaw 2000) or submergent (Faivre and Fouache 2003; Antonioli *et al.* 2006), they are valuable markers of land movements relative to sea level.

Geoarchaeological Proxies

The long human legacy of the Mediterranean means that cultural features are some of the most reliable shoreline markers for reconstructing sea-level changes over recent millenia. Indeed, arguably the best example of the postglacial sea-level highstand comes from Cosquer Cave, southern France, where Palaeolithic wall paintings depicting horses have partially been eroded by the rising sea water, clearly showing that Holocene sea level has never reached higher than its present-day level (Morhange *et al.* 2001; Figure 13.8). Elsewhere it is archaeological remains that have long provided the reference datum (Flemming 1969, 1978 1993; Blackman 2005): for example, the study of ancient Mediterranean harbours has emerged as an important component of geoarchaeology (Marriner and Morhange 2007).

Ancient rock-cut installations such as quarry floors and piscinae (fish ponds) show that the Israeli coast has been tectonically stable for the last two thousand years

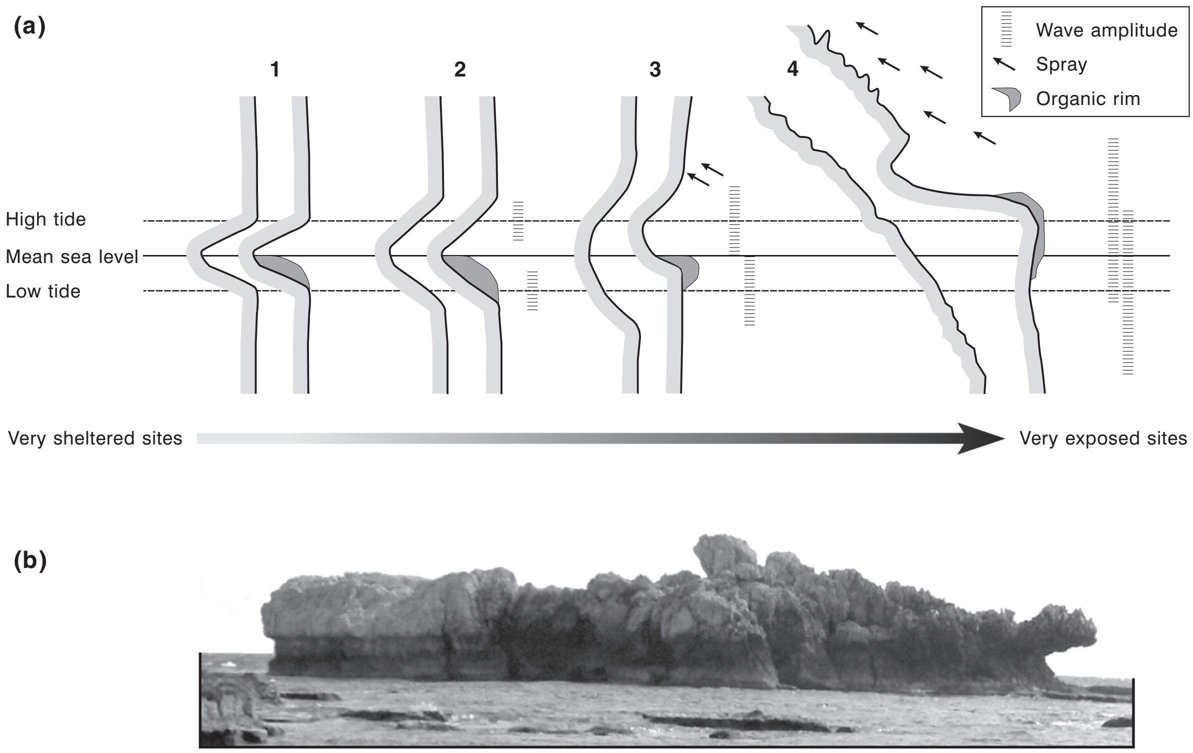


Fig. 13.7. (a) Effects of variations of coastal-wave energy on marine-notch formation (modified from Rust and Kershaw 2000 and Pirazolli 1986*a*). 1 and 2 are wave notches formed in quiet conditions at mean sea level; 3 and 4 are surf notches formed in more turbulent conditions as much as 2 m above sea level. (b) A marine notch at Capo Milazzo in north-eastern Sicily illustrating the contrasting morphology of an open (right) to a sheltered (left) position (*photo*: Iain Stewart).

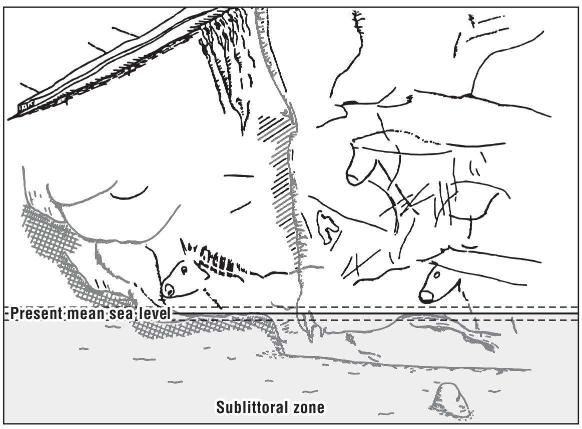


Fig. 13.8. The absence of a Holocene sea level above present datum is supported by the evidence of painted horses on a wall of a half-submerged Palaeolithic cave near Marseilles. Below the current water level of Cosquer Cave the wall paintings are significantly degraded (after Morhange *et al*. 2001).

(Galili and Sharvit 1998; Sivan *et al.* 2001). Similarly, Lambeck *et al.* (2004*b*) have utilized Roman piscinae constructed on or in bedrock along the Tyrrhenian coast of Italy to determine that sea level here two millennia ago was 1–2 m lower than today, a change that they attribute almost entirely to glacio-isostatic subsidence. This result is out of step with previous studies (e.g. Pirazzoli 1976).

Biological Proxies

The above proxy indicators are by no means independent—ancient harbours, for example, are important stratigraphical archives (e.g. Kraft *et al.* 2003; Mariner *et al.* 2006; Marriner *et al.* 2006)—and many Mediterranean sea-level studies typically combine sedimentological, geomorphological, and archaeological indicators. However, few of these indicators are valuable without associated biological indicators. As well as providing often precise reference markers for palaeo-sea-levels, biological proxies also serve as dateable deposits from which to establish sea-level histories. Over the last decade or so, the use of biological sea-level indicators in the study of Mediterranean sea-level changes has gradually evolved from a descriptive to a multidisciplinary approach integrating many of the proxies above (Laborel and Laborel-Deguen 1994, 1996). It is an approach based on the recognition that the vertical distribution of the littoral fauna and flora of rocky shores shows a pattern of superimposed ecological belts, a tendency called biological zonation (Péres 1982).

According to biological zonation, marine benthic animals and plants are finely adapted to very precise ecological conditions such as light intensity, turbidity, water salinity and temperature, and surf exposure. Consequently, changes in local ecological conditions are followed by a concomitant quantitative and qualitative modification of the organisms with replacement by more tolerant forms. Laborel (1986) discusses this biological zonation in detail and shows how it can be used to measure past sea levels. In very general terms, several parallel zones can be recognized (Figure 13.9) and these are outlined as follows:

1. A supralittoral fringe (or supralittoral zone) never or rarely submerged but wetted by surf in which the biomass is very low and mainly represented by boring endolithic cyanobacteria and grazing gastropods.
2. A midlittoral zone submerged by tides and waves on a regular basis, which displays a pattern of parallel algal and faunal belts, with biomass and species diversity increasing downwards. Cyanobacteria, limpets (*Patella* spp.) and Chitons are the main bio-eroders in this zone. Constructional elements such as the rim-building coralline rhodophyte *Lithophyllum byssoides* may develop in the north-west Mediterranean.
3. An infralittoral (or sublittoral) zone whose upper limit is marked by a sudden increase in biodiversity (Boudouresque 1971), thus defining a biological sea level that ranges down to the lower limit of marine phanerogams and photophilous algae, i.e. to a mean depth of about 25–35 m. The upper part of this infralittoral zone (also called 'infralittoral fringe') is densely populated by brown algae (*Cystoseira)*, Coralline Rhodophytes, fixed vermetid gastropod molluscs (such as *Dendropoma* sp.), and cirrhipeds, for example *Balanus* spp. Active erosive agents, such as clionid boring sponges, sea-urchins, and rock-boring mussels (*Lithophaga*, *Hyatella*, *Coralliophaga* spp.), are responsible for rapid underwater erosion of the limestone outcrop.

The limit between the midlittoral and the infralittoral corresponds to the 'biological sea level'(Laborel 1986). The influence of local variations in coastal morphology upon surf exposure explains why this limit may be slightly undulating locally. Aperiodic sea-level oscillations linked to atmospheric pressure or wind variations have little influence upon the marine zonation of living organisms with a lifespan of more than one year. Biological sea level itself is best characterized by

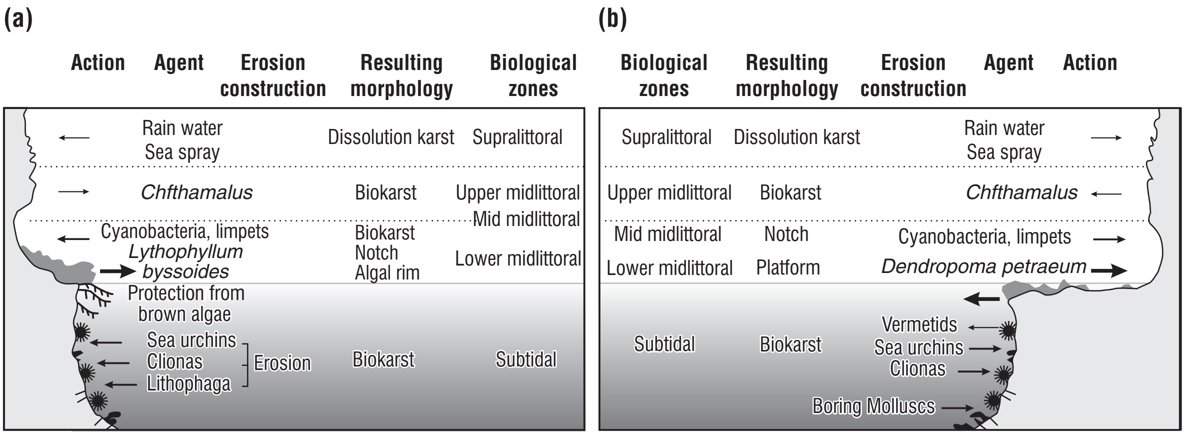


Fig. 13.9. A schematic coastal profile showing the main characteristics of bioconstruction and biodestruction on calcareous coasts in (a) the western Mediterranean, and (b) the eastern Mediterranean (modified from Laborel and Laborel-Deguen 1994).

the development of the few marine species, with a very narrow depth range, located immediately above (e.g. *Lithophyllum byssoides* rim) or below (e.g. *Dendropoma* rim) which are considered the more precise sea-level indicators. When such species are absent, the highest limit of biological perforations by *Cliona* and *Lithophaga* (Laborel and Laborel-Deguen 1994) are also excellent proxies. Consequently, littoral algal and vermetid bioconstructions and the upper limits of bioerosive elements (marine burrows and perforations), and fixed invertebrates (oysters, barnacles, solitary vermetids) are commonly used as biological sea-level indicators.

Biological Evidence for the Amplitude and Rapidity of Shoreline Change

If a suitable biological indicator is available, and if an accurate study of local ecological conditions has been undertaken, fossil biological sea-level markers can provide reliable constraints on the direction, amplitude, rapidity, complexity, and timing of apparent coastal deformation (ibid.). Of these, indicators of the amplitude and rapidity of relative sea-level displacement are particularly important for discerning tectonic versus eustatic sea-level changes.

The amplitude of shoreline change is derived from the altitude of elevated or submerged shorelines. This parameter is best estimated by direct measurement of the altitudinal difference between the upper limit of the elevated (or submerged) remains and the corresponding upper limit of its modern equivalent. This can easily be done with species such as *Dendropoma* sp. or *Lithophyllum byssoides* whose populations have a very narrow vertical range closely connected to biological sea level. For species with a wide vertical range such as *Lithophaga*, balanids, or solitary vermetids, good results are obtained only when the uppermost limits of both fossil and living populations are well delineated or when the fossil remains are correlated with a morphological sea-level indicator such as a tidal notch. Any direct reference to the present instantaneous water level, whether observed or calculated from tide tables, is thus unnecessary and incorrect. For example, the only Mediterranean reef-coral, *Cladocora caespitose*, can grow from the surface down to more than 40 m, but it does not present any well-marked upper limit nor does it develop distinct sea-level linked build-ups. It should therefore only be used as a proof of submersion but not as a depth proxy and should not be attributed with a precise depth coefficient (Antonioli *et al.* 1999).

When a series of measurements is to be made in a limited area, one must keep in mind as stated above, that a biological sea-level mark is *not* a perfect horizontal line but may be naturally warped, even over very short distances, by local hydrodynamic variations. For that reason, each individual altitudinal measurement must be done on a vertical profile of its own, including both the fossil specimen and its present equivalent (Laborel 1979). The accuracy that can be obtained depends upon both the state of preservation of the upper limit of fossil populations and local ecological conditions. An accuracy of about ±5 cm was obtained in south-west Crete for a series of remarkably well-preserved vermetid rims (Thommeret *et al.* 1981). In Euboea, for elevated populations of *Lithophaga* burrows (Stiros *et al.* 1992), the lowest accuracies observed (around ±50 cm to several metres) are typically found in relation to *Chthamalus* populations.

Biological sea-level indicators are especially useful in determining the rapidity of relative sea-level changes. In the case of slow uplift (less than a few millimetres per year), biological indicators living in the subtidal zone are killed by emersion and their remains are slowly carried up through the midlittoral zone to the supralittoral zone. Small or fragile skeletons quickly disappear by bioerosion and physical abrasion. Massive algal rims and vermetid constructions are not completely destroyed and can be used as proxies. In contrast, good preservation of fragile elevated remains is the best evidence of a rapid uplift.

Determining the rapidity of subsidence is much more difficult than for uplift, because many biological sea-level indicators are either rapidly destroyed by subtidal erosion or covered by new generations of marine organisms. Thus, only bioconstructing species with a very narrow vertical range, such as *Lithophyllum byssoides* or *Dendropoma petraeum*., may be of some use. For such bioconstructed rims, it is often possible to obtain valuable information from the distribution of their drowned remnants, provided they are not very old and have not been removed by biological erosion. Eroded remains below the present rim are often found where subsidence is slow. More rapid change would not provide sufficient time for the development of a bioconstructed rim at intermediate depth, and drowned bioconstructions would thus not appear clearly separated (Laborel and Laborel-Deguen 1994).

Holocene Sea-Level Histories for Contrasting Mediterranean Coasts

As discussed earlier, there is considerable variation in the stability of the Mediterranean coastline, with the

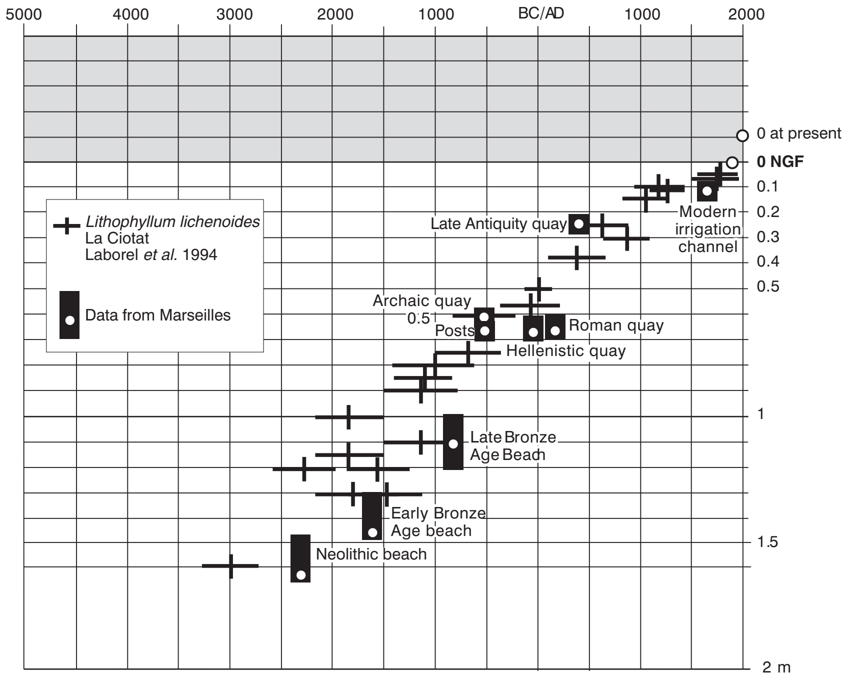


Fig. 13.10. Age-depth diagram from Marseilles's archaeological excavations compared with dated algal rims from nearby rocky cliffs (modified from Morhange *et al.* 2001 and Laborel and Laborel-Deguen 1994). NGF = Nivellement Général de la France.

geodynamically active Calabrian and Aegean seaboards being especially mobile, the rifted margins of the southern, western, and eastern sectors being largely stable, and localized volcanic centres experiencing irregular paroxysms (Figure 13.4b). In the following section, we examine in more detail the relative sea-level histories established for three sites in these contrasting settings.

'Stable' Coast

The site of Marseilles (southern France) is a good example of relative sea-level 'stabilization' in a so-called tectonically 'stable' area (Morhange *et al.* 2001). Allied with recent archaeological excavations of the ancient harbour, biological indicators have yielded high-precision data for the past 5,000 years (Figure 13.10). One of the best biological sea-level indicators used here is the upper limit of barnacles (*Balanus* sp.). They commonly develop upon quay walls in clear or polluted waters, stopping abruptly at biological sea level. When their upper limit is continuous, a precision of plus or minus a few centimetres can be obtained. Wherever barnacle-bearing hard surfaces were not available, the upper-limit of subtidal beach onlap layers was used as a sea-level indicator with a precision of ±0.2 m.

Data obtained in Marseilles fit well with indicators from rocky coasts such as *Lithophyllum* rims (Laborel and Laborel-Deguen 1994). The age-depth diagram shows a regular sea-level rise up to about AD 500 followed by a period of stability. Total rise has been less than 1.5 m since 4,500 years cal. BP. Observations do not show any Holocene level higher than present (Figure 13.10). The rate of mean sea-level rise was 0.4 mm/yr between 4,500 years cal. BP and AD 500, and 0.2 mm/yr thereafter. During the twentieth century, the rate of sea-level rise increased to *c.*1.5 mm/yr, most likely in connection with global warming. In other words, in settings where geodynamic activity is low or insignificant, background eustatic trends can be isolated.

Volcanic Coasts

More complicated coastal elevation changes can occur in areas affected by volcano-tectonic deformation (e.g. Firth *et al.* 1996). The most detailed record of the complex mobility of volcanically active shores can be shown by the archaeological ruins of Pozzuoli on the Phlegrean Fields in the Bay of Naples. The bioeroded columns of the port's Roman market were made famous as the frontispiece of Charles Lyell's *Principles of Geology*, and

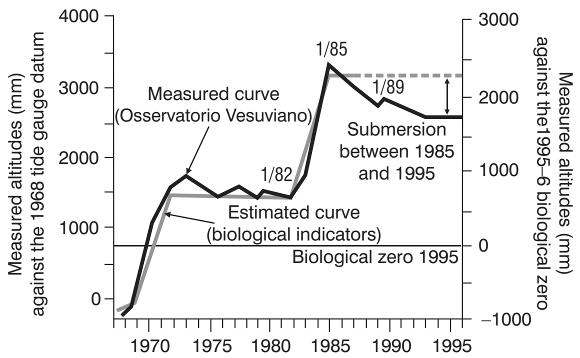


Fig. 13.11. Measured relative sea-level changes in the old harbour of Pozzuoli compared to estimated relative sea-level changes using biological indicators (modified from Morhange *et al.* 2006).

since that time have served as a palaeo-tide gauge to track the deformation history for the area (Dvorak and Mastrolorenzo 1991; Orsi *et al.* 1999; Morhange *et al.* 2006).

Relative sea-level movements at Pozzuoli are far more intense than the average 50-cm sea-level rise recorded in the north-west Mediterranean sea since Roman times (Pirazzoli 1976). Moreover, the post-Roman sea-level history reveals three cycles of submersion and emergence (Figure 13.11). The first submersion accompanied a period of marine transgression which ended around AD 400–550 without any volcanic eruption. A second submersion affected the site in the early Middle Ages, around AD 700–900, though again no eruptive activity followed this inundation. The third submersion occurred in the late Middle Ages, when it was followed by a well-documented pulse of land uplift that culminated in a major eruption in 1538. More recently, the two major events of 1969–72 and 1982–4 resulted in a total uplift of *c.*3.5 m, although no eruption occurred during these periods. Indeed, during the last 2,000 years, non-eruptive coastal uplift episodes have been the rule rather than the exception. Clearly, here, the dominant sea-level change signal is linked to phases of volcanic activity and quiescence (see Chapter 15).

Tectonically Active Coast

In contrast to volcanically active coastlines where land movements are discernible over months to years, equivalent shoreline movements on earthquake-prone coasts tend to be instantaneous. Abrupt littoral emergence or submergence in these environments has been widely used to identify past large earthquakes in the Holocene coastal record (e.g. Stiros *et al.* 1992; Pirazzoli *et al.* 1994*a*, 1994*b*, 1996*a*; Stewart 1996; Stewart *et al.* 1997). Such 'coastal palaeoseismology' studies reveal that seismically active coasts in the Mediterranean realm typically experience decimetre- to metre-scale seismic jerks every few centuries to millennia.

Arguably the most remarkable example of this seismotectonic action has been reported from the shores of western Crete and nearby Antikythira island, where a series of elevated stepped shorelines are carved into the limestone cliffs supporting well-preserved Vermetid formations. The highest of these contains very fragile shell skeletons that would have been rapidly destroyed in the surf zone (Thommeret *et al.* 1981). The sea-level history derived from these shorelines is especially complex (Figure 13.12). It reveals that, between 4,000 and 1,700 years BP, ten successive increments of small but abrupt subsidence affected a huge block of lithosphere about 150 km long without noticeable tilting. This phase of subsidence was followed about 1,530 years BP by an abrupt uplift, by about 9 m, of the south-west corner of Crete and a north-eastward tilting of the western

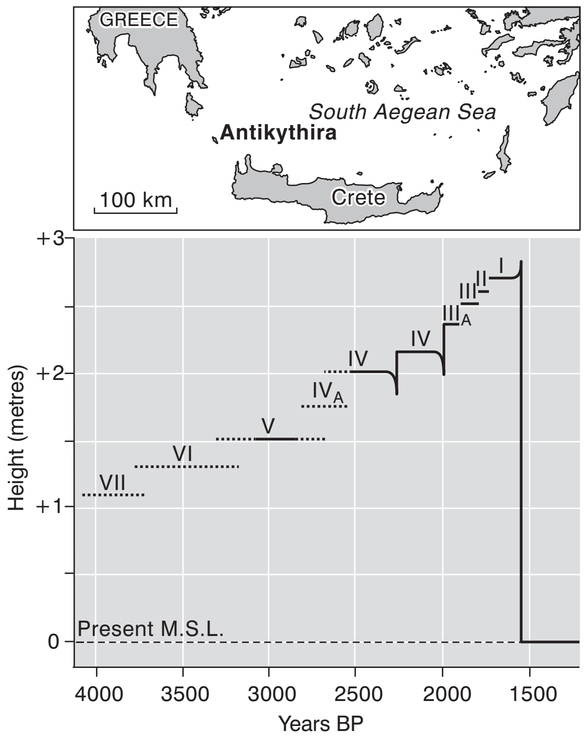


Fig. 13.12. Recent relative sea-level variations in Antikythira island (Greece), from Pirazzoli (1986*b*) and Pirazzoli *et al.* (1996*b*).

part of the island. In north-west Crete, the Roman harbour of Phalasarna was raised by over 6 m (Pirazzoli *et al.* 1992; Chapter 16).

Relative sea-level change histories at other coastal sites from western Greece to the Levant coasts also appear to exhibit marked vertical tectonic displacements—some of them several metres in amplitude—between 1,750 and 2,000 years BP (Pirazzoli 1986*b*; Pirazzoli *et al.* 1996*b*). These studies attribute the main movement to a great earthquake in AD 365 (Stiros and Papageorgiou 2001), though recalibration of the original radiocarbon dates on elevated littoral fauna led Price *et al.* (2002) to shift the main emergence event to the 6th century AD, another period of seismic unrest on Crete (Di Vita 1996). The tectonic mechanism for such dramatic coastal displacements also remains uncertain, but they have been ascribed to an exceptional burst of tectonism that occurred on a regional scale in post-Roman times—the 'Early Byzantine Tectonic Paroxsym' (Pirazzoli 1986*b*; Chapter 16). Nevertheless, the literally Byzantine history of Holocene shoreline development in western Crete highlights the potential complexity of land movements at plate boundary zones.

Tsunamis as a Coastal Process

The high seismicity of the Mediterranean region means that much of its coastline is subject to recurrent tsunamis (Soloviev 1990; Chapter 17). Unsurprisingly, the main areas of observed tsunami incidence coincide with the principal earthquake belts (Figure 13.13). However, in addition, volcanic-induced tsunamis affect the Tyrrhenian and Aegean seas, and the steep margins of the north-western Mediterranean basin mean that tsunamigenic submarine slumps are potentially commonplace. The result is that while some coastal zones are more susceptible than others, virtually any

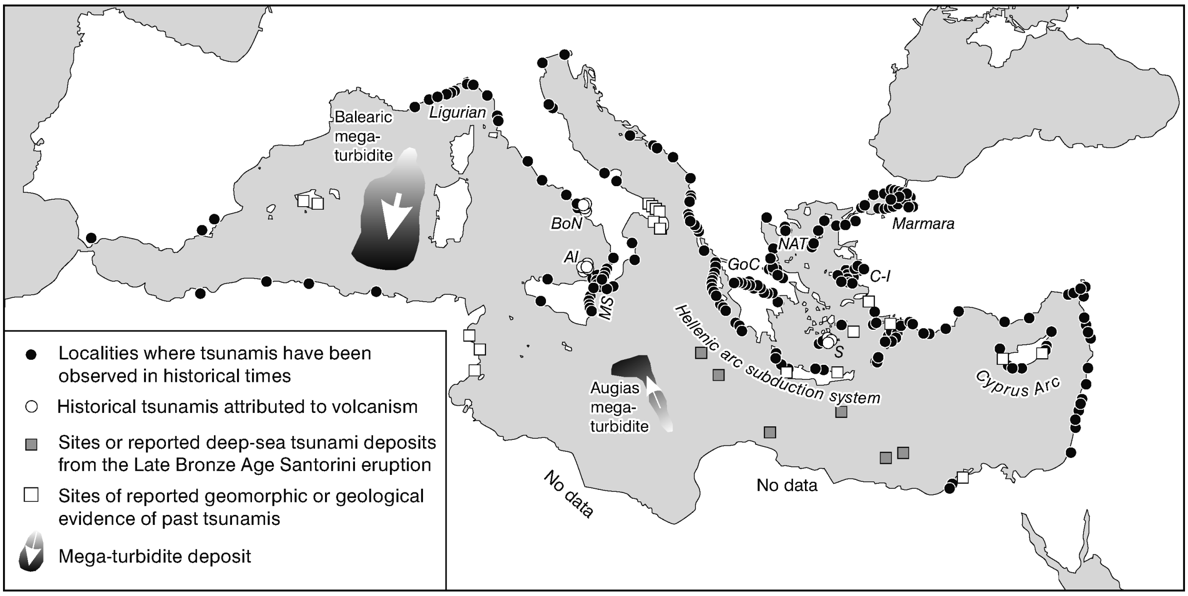


Fig. 13.13. Tsunami activity in the Mediterrenean Sea. Historical data compiled from the following sources: Ambraseys (1962) for the Levant; Soloviev (1990) for the western Mediterranean; Tinti and Maramai (1996) for the French Côte d'Azur and Italy; Papadopoulos (1998) for Greece and the Aegean; Altnok and Ersoy (2000) and Yalçıner *et al.* (2002) for Turkey. Palaeo-tsunami sites are compiled from Wood (1996) for Mallorca, Tunisia, and Egypt; Mastronuzzi and Sansò (2000, 2004) for Apulia, Italy; Dominey-Howes *et al.* (2000) for the central Aegean; Minoura *et al.* (2000) for Crete and western Turkey, and Kelletat and Schellman (2002) for Cyprus. Italicized text indicates the principal tsunamigenic zones, including the Bay of Naples (BoN), Aeolian Islands (AI), Messina Straits (MS), Gulf of Corinth (GoC), Santorini (S), North Aegean trough (NAT), and Chios-Izmir zone (C-I). Mega-turbidites are from Rothwell *et al.* (2000) in the Balearic basin, and Hieke (2000) in the Ionian Sea. Deep-sea homogenite deposits in the eastern Mediterranean are from Cita *et al.* (1996) and Cita and Aloisi (2000). The lack of recorded tsunamis in some parts of the region, most notably the North African coast, undoubtably reflects a lack of records rather than a real absence of incidence. Data for the Atlantic and the Black Sea are not shown. See Chapter 17.

TABLE 13.2. *Amplitude, duration, permanence, and length of coast affected by various types of rapid relative sea-level change in the Aegean*

Process or type of change	Amplitude	Duration of change	Permanence of change	Length of coastline affected
Local tsunami	Up to 30 m[a]	Minutes	<1 hour	Tens to hundreds of km
Storm surges	1 m[b]	Hours	Hours to days	Tens to hundreds of km
Coseismic subsidence or uplift (moderate magnitude earthquakes)	0.3–0.8 m (this study) 1–3 m (Gulf of Corinth)[c]	Seconds	Centuries	Hundreds of metres to tens of kilometres
Semi-diurnal tides	0.6 m (Gulf of Evvia) Typically 0.1 m or less[d]	*c.*6 hours	Hours	Entire Aegean region
Marine flooding from barrier breaching (natural or artificial coastal barriers)	Decimetres[e]	Hours	Days to centuries	Hundreds of metres

[a] Maximum reported value for 1956 south Aegean tsunami (Ambraseys 1960).
[b] Value for Z_{50} extreme sea level based on tide gauge records at N. Halkis, Gulf of Evvia (Tsimplis and Blackman 1997).
[c] Based on estimates in Hubert *et al.* (1996) and Soter (1998).
[d] Zoi-Morou (1981).
[e] Depends on age of barrier, regional sea-level history, coastal morphology, etc.

Source: Cundy *et al.* (2000).

stretch of Mediterranean littoral is prone to tsunami attack.

Given their likely historical propensity (Soloviev *et al.* 2000), tsunamis are expected to be an important agent of abrupt coastal change. Yet detecting and assessing the past impacts of tsunamis on coast evolution is difficult, not least because a number of nearshore mechanisms can cause rapid transient flooding and associated erosion and deposition (Table 13.2) (Cundy 2005). Biostratigraphical studies of coastal sediment sequences have been used to identify likely tsunami flooding events (e.g. Dawson 1996; Dominey-Howes *et al.* 1998), but in many respects their deposits remain indistinguishable from those produced by barrier breaching or storm-surge flooding (Cundy *et al.* 2000; Dominey-Howes *et al.* 2000). Furthermore, the long-term (geological) preservation potential of tsunami deposits is poor (Dawson and Stewart 2007). One potentially long-lived coastal expression of large tsunamis, however, may be littoral boulder fields—deposits that require very high-energy emplacement conditions. Mega-clast deposits attributed to tsunami action have been described from both Holocene coastal environments in Apulia (Mastronuzzi and Sansò 2000, 2004), Cyprus (Kelletat and Schellman 2002; Whelan and Kelletat 2002), and Sicily (Monaco *et al.* 2006), while possible Late Pleistocene examples are described from Egypt, Tunisia, and Mallorca (Wood 2000).

The tendency for Mediterranean shores to have experienced sudden and extensive inundation episodes was not limited to past tsunamis. As we discuss in the following section, postglacial transgression resulted in the abrupt drowning of many low-lying parts of the Mediterranean coastal zone, with major implications for its Palaeolithic and Mesolithic coastal dwellers.

Postglacial Shoreline Development and Human Activity

At the height of the global Last Glacial Maximum, a 300-km wide coastal plain was emergent from Tunis to Tripoli, and North Africa was disconnected from the northern Mediterranean only by a few narrow marine straits between Sicily and Tunisia (van Andel 1989). A series of exposed seamounts served as ‘stepping-stones’ within the Strait of Gibraltar (Collina-Girard 2001, 2002), and islands such as Elba, Malta, and most of the Greek Ionian islands were linked to the continent by land bridges. The Gulf of Lions was a continental plain linked with Valencia in present-day Spain, and roughly half the Aegean and Adriatic seas were dry. Between 16,000 and 6,000 years BP, rising postglacial waters inundated continental shelves, submerging low-lying islands, cutting land bridges, and creating many of the islands of the modern Mediterranean. For example, during this period the Greek Cyclades metamorphosed from a wide, low-lying coastal plain into a pair of land masses (one centred on Paros and Andiparos, the other on Mykonos–Tinos–Andros), and then into its present scattered rocky archipelago (Lambeck 1996). The effects were even more striking in the eastern Adriatic, where the flooding of karst topography created one of the world's most indented coasts and a submarine shelf of submerged caves (Chapter 10).

The Mediterranean's coastal caves and plains were important loci for Palaeolithic peoples during the last cold stage (e.g. Gamble 1986), so postglacial coastline changes had dramatic effects on human activity along its shores (Petit-Maire 2003). The broad coastal shelves that constituted the prime Palaeolithic hunting grounds were incrementally submerged, with a probable succession of standstills and spasmodic events (Collina-Girard 1997, 1998; Laborel *et al.* 1999), so that much of the early coastal human occupation sites of the Mediterranean now lie offshore (Flemming 1998).

Pulses of meltwater releases from the great continental ice sheets led to irregular postglacial sea-level rise (Blanchon and Shaw 1995), and consequently episodes of rapid marine encroachment onto coastal plains. The rate at which sea levels rose meant that at times the shores of some coastal lowlands were retreating at several hundred metres per year, forcing seashore communities continually to relocate further inland. The progressive reduction of the relatively hospitable coastal plain environment caused a concomitant loss of resources, squeezed coastal inhabitants into narrower and less conducive littoral zones, and rerouted human traffic patterns (van Andel 1989; Lambeck 1996). Human adaptation to these shifting shores is illustrated by archaeological records from the Late Palaeolithic remains at Franchthi Cave on the Argolid Peninsula of southern Greece (Figure 13.14). At the peak of the last glaciation the coast lay at least 6 km west of Franchthi, but by 12,000–10,000 years BP marine molluscs and small fish bones began appearing in the remains (Jacobsen 1976) providing evidence of an encroaching sea.

At times, the rising Mediterranean waters may have had especially dramatic effects. Ryan *et al.* (1997) argued that the final meltwater rise event around 7,200 years BP was responsible for Mediterranean waters overtopping the narrow land bridge of the Bosphorus Straits and cascading into the freshwater Black Sea basin, inundating burgeoning Neolithic settlements along its shores. The hypothesis of the so-called 'Black Sea flood' has been developed further (e.g. Ryan and Pitman 2000; Ryan *et al.* 2003; Siddall *et al.* 2004), but the timing, route, and abrupt nature of this marine influx is much disputed (Görür *et al.* 2001; Aksu *et al.* 2002; Major *et al.* 2002; Kerey *et al.* 2004). Some workers see in this inundation, and others that flooded the various marginal-sea basins, gulfs, and lakes at different times during the postglacial transgression, the root of deluge myths that are deep-seated in eastern Mediterranean culture (Ryan and Pitman 2000; Petit Maire 2004; Wyatt 2004). However, a more gradual but almost equally dramatic reshaping of the Mediterranean's coastal geography would take place once the bulk of postglacial sea-level rise had abated.

The Formation and Growth of Mediterranean Deltas

Many Mediterranean coasts today comprise long, monotonous sandy beaches interrupted by occasional rocky headlands. Inlets and harbours are rare, except on islands. But this has not always been the case. In antiquity, the shores of the Mediterranean were studded with bays and lagoons, and even large rivers such as the Ebro had harbours at their mouths instead of deltas. The gradual disappearance of this complicated indented coastline may have begun with its submergence by rising postglacial waters, but it was largely accomplished once the last meltwater pulse event had occurred about 8,000 years BP. As sea levels slowly stabilized, sediment supply was able to keep pace with or overwhelm eustatic change, allowing rivers with significant bed loads to establish permanent gravel barrier-beaches and spits around river mouths. Behind these barriers, enormous volumes of clay and silt draped and then buried the drowned glacial coastline. The result was that around the Mediterranean Sea, as on coasts worldwide, deltas began rapidly to advance seawards (Stanley and Warne 1994, 1997).

As they grew in extent, the Mediterranean deltas began to subside under their own weight (typically at a few millimetres a year), creating additional space to accommodate ever more sediment. Around the largest deltas, such as the Nile and Rhone, isostatic subsidence, combined with compaction and failure by faulting and slumping, caused harbours to become entombed. The great weight of the Nile delta, for example, has caused its shores to be lowered by 4–5 m in the last 2,500 years (Stanley 1989). Together with localized soft-sediment slumping, this has left Greek port cities such as Herakleion and eastern Canopus entirely submerged in water depths of up to 8 m (Stanley *et al.* 2004). More often, however, it was simply the advance and lateral migration of the delta fringes that overwhelmed successive bays and inlets with alluvial sediment and transformed them into marshes, wetlands, and floodplains (Chapter 9). In the process, many of the ports and harbours of antiquity gradually lost their connection to the open sea, silted up, and became abandoned (Kraft *et al.* 2003;

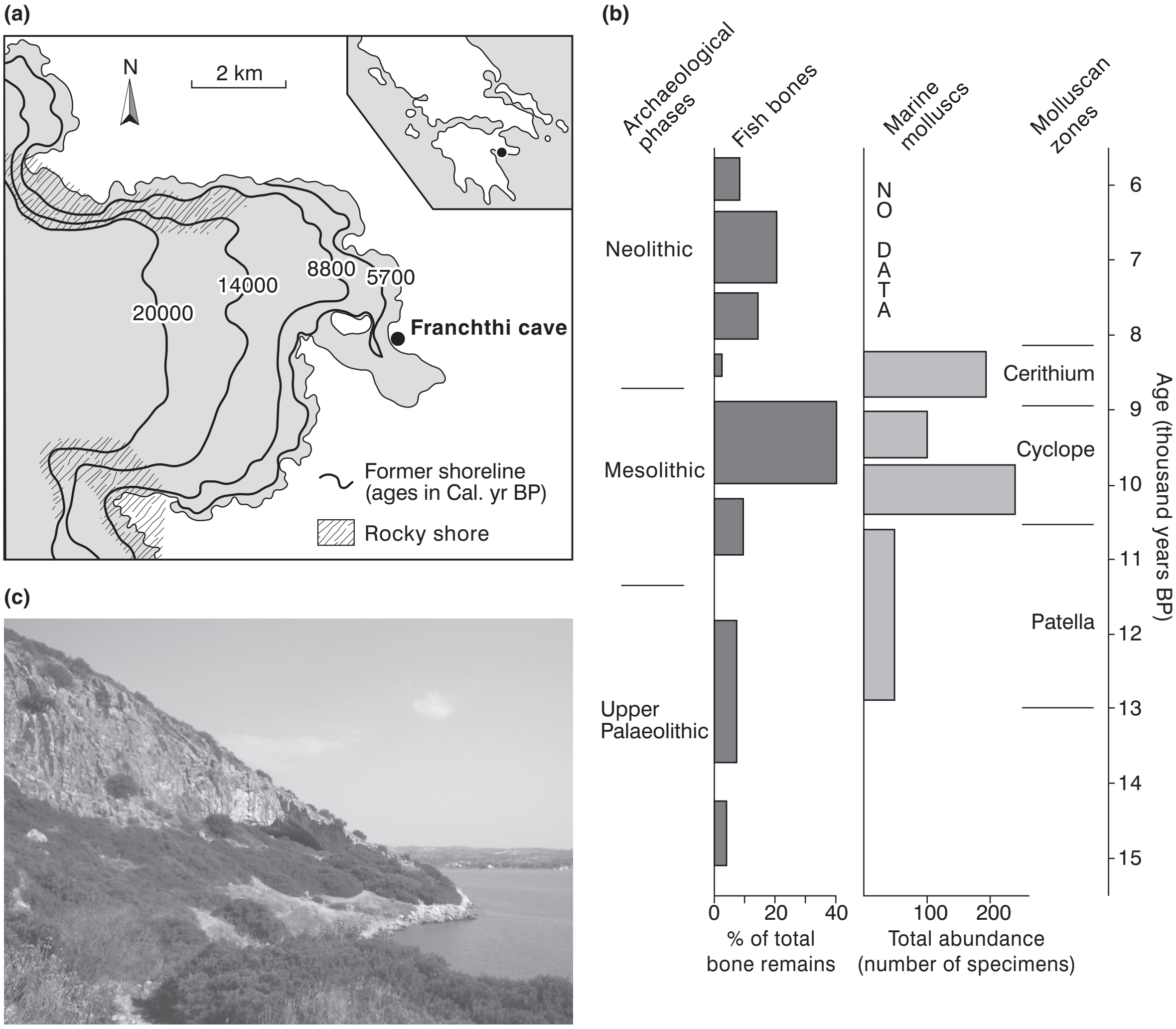


Fig. 13.14. Franchthi Cave in the south-east Peloponnese, Greece showing (a) sea-level transgression since the global Last Glacial Maximum (after van Andel *et al.* 1980) and (b) the sequence of marine resource exploitation recovered from the archaeological excavations (after Payne 1975 and Shackleton and van Andel 1980). Layout of (a) and (b) after Roberts (1988). (c) Photograph of the site taken in 2006 by Mike Morley. The floor of the cave now lies about 10 m above modern sea level.

Brückner *et al.* 2004, 2005; Fouache *et al.* 2005*a*; Marriner and Morhange 2007).

The apparent close association in Classical antiquity between the construction of harbour cities and their demise by delta growth has led many historians and geologists to implicate human-induced denudation of catchment slopes as the root cause of accelerated coastal accretion in late Holocene times. Palaeoenvironmental studies of some ancient harbours clearly detect the anthropogenic influence on sedimentation rates following harbour construction (Morhange *et al.* 2003; Marriner *et al.*, 2006). Others, such as Grove and Rackman (2001), however, argue against strong anthropogenic forcing, pointing out that in those coastal lowlands where past human activities have been most intense (e.g. the Anatolian rivers) there is no tendency for the deltas to be disproportionately large, and noting that not all Aegean deltas have slowed their advance in post-Classical times (e.g. Kraft *et al.* 1977).

For some workers, the late Holocene progradation of deltas is a climatic artefact, reflecting a change

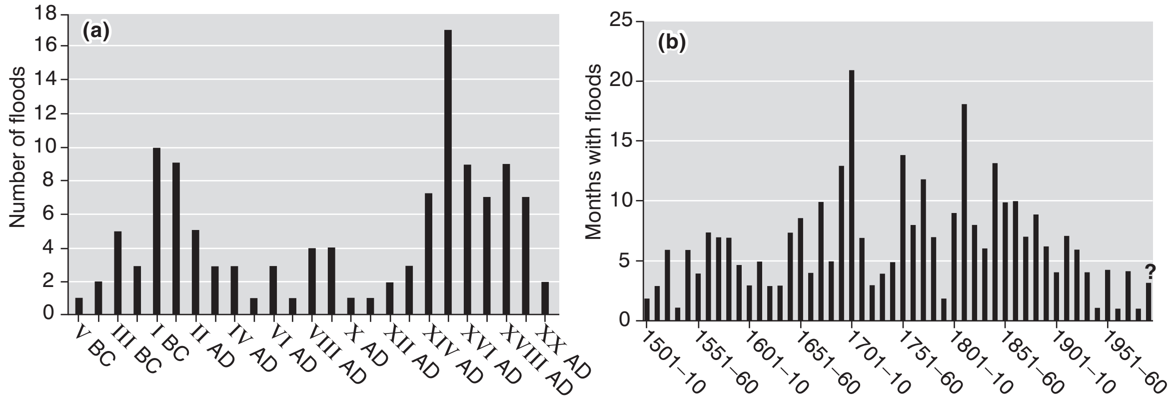


Fig. 13.15. Historical records of coastal flooding for (a) the River Tiber and (b) the River Rhône (see Chapter 18).

towards the present Mediterranean climatic conditions and an increasingly open landscape subject to erosion by episodic storms (Dubar and Anthony 1995; Chapter 11). Many deltas, particularly those in the western and central Mediterranean, experienced their greatest advances during the centuries of climatic deterioration of the Little Ice Age (late sixteenth to early eighteenth centuries AD). Thus Grove and Rackman (2001) report that Rome's Tiber River delta, which by the Roman period had engulfed the harbour port of Ostia, stalled its further advance until the sixteenth century when frequent bouts of severe flooding (Figure 13.15) drove it forward again. In the seventeenth century the closely charted Po delta advanced at an unparallel rate of 4–9 km along a 30–35-km front (Fabbri 1985). The seventeenth and eighteenth centuries witnessed the Rhône delta also undergo a major advance (50 m annually between 1587 and 1711) and metamorphosis, changing from a meandering to a braided channel and avulsing to its present-day position (Arnaud-Fassetta and Provansal 1999; Vella *et al.* 2005). Historical records of coastal flooding in the lower reaches of the Rhone and Tiber river systems are shown in Figure 13.15. The Arno river near Pisa and the Spanish deltas of the southern Costas and the Ebro basin also appear to have grown most actively during Little Ice Age times (Chapter 11; Grove and Rackman 2001 and references therein). Few deltas in the east changed significantly during this period, while some in the west, such as the Costa Brava deltas draining the eastern Pyrenees, were similarly unresponsive.

The evidence that not all Mediterranean deltas show a marked response to the Little Ice Age simply underscores the complex and variable pattern of Holocene delta progradation (Grove and Rackman 2001). Detailed studies, such as that by Arnaud-Fassetta and Provansal (1999) highlight how an intricate interplay of catchment changes drive delta evolution and growth. The largest river catchments feed the biggest and most rapidly expanding deltas (Figure 13.2). Where those catchments drain tectonically active hinterlands or highly erodible lithologies, the resulting deltas can be anomalously large. Thus the Akheloos delta, the largest on the Ionian Sea coast, drains soft flysch terrain in a seismically active region (Piper and Panagos 1981). Tectonically subsiding gulfs and shelves also favour delta development, with subsidence creating accommodation space and preservation conditions for deltaic sediments to accumulate. In contrast, excessively deep basins, frequent storms, or strong currents limit the extent of deltas, creating undersized deltas such as the Var and Ebro. The largest deltas in the Mediterranean are major depocentres containing some of the thickest Pliocene-Quaternary records in the region (Figure 13.2). The Po delta, for example, covers an area of approximately 770 km^2, has a Holocene record of 30 m in thickness and a total Plio-Quaternary sediment record of *c.*5,000 m.

Regardless of the driving forces of late Holocene coastal accretion, and despite the concomitant loss of cultural heritage, the great delta plains provide the modern Mediterranean with extensive tracts of fertile farmland, prime real estate for urban sprawl, tourist infrastructure and airports, and some of the best wildlife habitats in Europe (Chapters 9 and 23). But today these important coastal environments are under threat, since most of the Mediterranean deltas are no longer advancing, but are in retreat.

Humans as Agents of Coastal Change

In modern times the amount of sediment reaching the Mediterranean via the Nile River had been drastically cut, most specifically with the construction of the High Dam at Aswan, and the problem has been compounded by uncontrolled coastal development (Frihy 2001). Historical studies of the Nile shoreline changes show how a delta front that had advanced steadily throughout the eighteenth century had, by the end of that century, gone into reverse. Thus on the Rosetta promontory (Figure 13.3) this retreat has accelerated from rates of 18 m yr^{-1} at the beginning of the nineteenth century to 230 m yr^{-1} in the 1990s (Fanos 1995; Ahmed 2002). The net shortfall in Nile sediment input has a knock-on effect on the Israeli Mediterranean littoral where recent decades have witnessed the aeolianite coastal cliffs eroding back at rates of 20 cm yr^{-1} (Zviely and Klein 2004). Throughout the twentieth century these soft cliffs have contributed approximately 24×10^6 m^3 of sediment to the budget of Israeli beaches, offsetting the loss of Nile sand. A further source of sand supply to this coast is similarly declining as reductions in agricultural and pastoral activity allow natural vegetation to recolonize and stabilize the dunefields of this coast (Tsoar and Blumberg 2002; Chapter 14).

To the west, the Tunisian coastline is similarly retreating everywhere, except for some localized areas at the mouths of active wadis or in sheltered bays and spits (Oueslati 1995). Again the main culprit is marine erosion from human interventions on the coast, but sea-level rise is also a factor. Many ancient (Punic and Roman) port and harbour installations are submerged by a few tens of centimetres, and the widespread intrusion of seawater is favouring the extension of sebkhas (flat supratidal environments of sedimentation devoid of vegetation) and chotts (marginal parts of sebkhas colonized by salt-tolerant vegetation) and the transformation of formerly inhabited and cultivated fields into unproductive land. These environmental changes are most acute in the northern part of the Gulf of Gabès, where inundation is accentuated by active tectonic subsidence (Paskoff and Oueslati 1991).

It is a similar story along many parts of the northern Mediterranean. Shipwrecks along the Gulf of Lions littoral track the progressive landward displacement of the beach barrier (Barusseau *et al.* 1996). The reduced supply of sediment to the Ebro delta coast in recent years had led to a negative sediment budget, erosional retreat of the shoreline (Jiménez and Sánchez-Arcilla 1993), and coarsening of the beachface sediments through preferential winnowing of the finer fractions (Guillén and Palanques 1996). In this region, coastal erosion is threatening the existence of the great dunefields that form the seaward fringes of the Guadalquivir, Rhône, Ebro, Acheleous, and Goksu deltas, which all rely on unhampered sand supply for their survival.

Where mobile dunefields have encroached on agricultural and urban areas they have commonly been stabilized by vegetation and woodland. In some places, however, overgrazing, trampling, and other human interference has destabilized them, leading to wind erosion problems. According to van der Meulen and Salman (1996) almost 75 per cent of the Mediterranean's coastal dunes had, in the preceding thirty years, been damaged or destroyed, largely as a result of tourism. Along the Spanish and French Mediterranean shores, 75–80 per cent of the coastal dunes have succumbed to these pressures (Corre 1991), and they are becoming similarly stressed in Turkey, Greece, Israel, and Tunisia (Chapter 14). The gradual disappearance of the dunes means that sandy beaches are also vanishing at an increasing rate.

Eroding delta fronts and stabilizing dune systems are not the only processes that are starving Mediterranean beaches of sediment. The prime reason is that many of the region's high-discharge rivers are dammed and clogged by flood-control schemes (Chapters 8 and 11), so sands and gravels are no longer making it to the coast. Many of these dams are needed to underpin the dramatic rise in water demand from the increasing mass tourism around the shores of the Mediterranean, a region which sees its 130 million coastal population swelled seasonally (Chapter 21). To cater for such influxes, tourist developments such as marinas and promenade waterfronts have taken more and more control of the littoral zone. Along much of the Mediterranean's shores, the longshore distribution of what little sediment reaches the coast is impeded and diverted into a new set of temporary sediment sinks. In all corners of the Mediterranean realm, tourist development is placing the littoral zone under unprecedented pressures (e.g. Frihy *et al.* 1996; Garcia and Servera 2003; Burak *et al.* 2004; Chapter 23).

New coastal management practices have emerged to tackle the changing character of the Mediterranean's soft shores. In the 1990s, Spain's Costa del Sol authorities switched from stabilizing an eroding coastline with groyne fields to replenishing them artificially (Malvárez

Garcia *et al.* 2002). Such beach nourishment schemes have had mixed success because their aim was to maintain sufficient beach width for recreational purposes rather than to recreate the exact position of the former shoreline or to reduce the impact of flooding (ibid.). Often the new man-made beaches are not robust enough to stand up to the winter storms that frequently buffet them. Furthermore, the offshore sediment is often far more coarse, so many beaches have become highly consolidated or excessively shelly, giving too abrasive a surface for recreational activities. Nevertheless, the beach nourishment approach promoted in Spain is now the preferred coastal adaptation in countries such as Turkey, while others, such as Greece, prefer 'hard-engineered' protection schemes (Hanson *et al.* 2002). In Italy, coastal retreat is being tackled by a mixed set of practices, mostly minor beach nourishment schemes but some major coastal engineering interventions. The largest of these are the construction of a barrier beach in the Venetian Lagoon and a barrier beach front to the city itself, and similar large barrier beach projects at Ravenna in the Po River delta (Hanson *et al.* 2002). Along this Adriatic coast, the problem is not just disappearing beaches but the threat of imminent flooding.

Although the Mediterranean is generally characterized by minor tidal and surge variability, more extreme flooding can characterize semi-enclosed gulfs and restricted marine settings, such as the north Aegean sea (Tsimplis and Blackman 1997). The problem is most acute, however, in the low-lying and tectonically subsiding northern shores of Adriatic Italy (Nicholls and Hoozemans 1996). Here, the city of Venice has long been flooded by storm surges, but the frequency of surges in the Venetian Lagoon has dramatically increased since the 1960s, reaching about two flooding events per year, the highest value since AD 792 (Camuffo and Sturaro 2004). The change reflects a relative sea-level rise of 0.3 m resulting from land subsidence due to a combination of onshore groundwater withdrawal and offshore methane extraction (e.g. Brunetti *et al.* 1998). Much of this subsidence appears to have ameliorated in recent years (Tosi *et al.* 2002), but the city of Venice remains at risk (for a review, see Fletcher and Spencer 2005). The reason for this continuing threat is accelerated global sea-level rise (Pirazzoli 2006).

Although twentieth-century sea-level-rise trends are arguably overestimated (Cabannes *et al.* 2001) or underestimated (Douglas and Peltier 2002), global studies continue to indicate that world sea-level rise is accelerating (Church and White 2006). Satellite and tide-gauge data indicate that Mediterranean waters are currently on the up, with satellite altimetry data showing that between 1992 and 1996 the mean rate of sea-level rise over the Mediterranean Sea was 7 $\pm 1.5\ \mathrm{mm\,yr^{-1}}$ (Cazenave *et al.* 2002). The sea-level changes were spatially variable (e.g. 20–30 $\mathrm{mm\ yr^{-1}}$ in the Levantine basin and falling sea levels in the Ionian Sea), indicating that the sea-level rise was the result of thermal expansion—the differential heating of surface water layers. During roughly the same period (1994–7), a rapid rise of western Mediterranean sea level by more than 10 $\mathrm{mm\ year^{-1}}$ has been at the expense of a 40 per cent fall in the sea-level drop between the Atlantic and the Mediterranean from 1994 to 1997. This highlights the fluctuating exchange flow through the Strait of Gibraltar which in turn is driven indirectly by changing conditions in the Mediterranean (Tetjana *et al.* 2000). Given uncertainties concerning future warming scenarios and ocean response, a global rise in sea level ranging from about 0.2–0.9 m by the year 2100 appears possible, with the estimates being current towards the upper end of this range.

Under such higher water-level conditions, low-lying portions of the Mediterranean coastline face heightened flooding threats in the coming century. Quantifying the extent of its vulnerability is difficult, not least because of the restless character of its land movements (Nicholls and Hoozemans 1996). Nevertheless, microtidal coasts such as the Mediterranean are especially vulnerable to rising water levels because the elevational growth range of coastal vegetation is tied to tidal range (Day *et al.* 1995). These authors also argue that sediment deficits already observed along soft deltaic shores are likely to become more acute with an acceleration in sea-level rise, because structural adaptations to protect against more frequent river floods and storm surges will further decouple beaches from their sediment sources. At 1990 figures, ten million people lived within the region's 1-in-1,000-year storm surge level, and about one million people experienced coastal flooding in a typical year. But, if sea-level rise continues at the modern pace of 1.5–2 $\mathrm{mm\ yr^{-1}}$ witnessed over the last century (Miller and Douglas 2004), ever more people fall within the flood-risk zone. With the inundation of tectonically subsiding or stable coastal lowlands comes the concomitant loss of land, biodiversity, and fisheries (Nicholls and Hoozemans 1996). Nicholls *et al.* (1999), for example, warn that Mediterranean coastal wetlands could disappear almost completely by the 2080s due to sea-level rise alone (Chapter 9). At risk are the great delta plains that once harboured the coastal dwellers in Palaeolithic and Mesolithic times, and which today form much of the Mediterranean's industrial, agricultural, and tourist heartland.

Conclusions

The impacts of present and future climate change on the Mediterranean coastal zone are difficult to gauge because, in addition to the inherent uncertainties in eustatic change estimations, this fragmented and heterogeneous mosaic of marginal seas ensures a complex response to tectonic, climatic, and eustatic forcing mechanisms. To address this important societal issue requires more studies that refine our understanding of the tectonic and glacio-hydro-isostatic dynamics of its seaboard and the sedimentological and geomorphological dynamics of its shoreline. Such research efforts increasingly demand interdisciplinary collaborations between geologists, geographers, oceanographers, climate scientists, archaeologists, biologists, and a suite of other practitioners, a need perfectly exemplified by this volume.

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References

Ahmed, M. H. (2002), Multi-temporal conflict of the Nile Delta coastal changes, Egypt, in *Littoral 2002, The Changing Coast*. EuroCoast, Portugal.

Aksu, A. E., Mudie, P. J., Rochon, A., Kaminski, M., Abrajano, T., and Yaşar, D. (2002), Persistent Holocene outflow from the Black Sea to the eastern Mediterranean contradicts the Noah's Flood hypothesis. *GSA Today*, 1–9 May.

Altınok, Y. and Ersoy, Ş. (2000), Tsunamis observed on and near the Turkish Coast. *Natural Hazards* 21: 185–205.

Ambraseys, N. N. (1962), Data for the investigation of seismic sea waves in the eastern Mediterranean. *Bulletin of the Seismological Society of America* 52: 895–913.

Antonioli, F., Chemello, R., Improta, S., and Riggio, S. (1999), *Dendropoma* lower intertidal reef formations and their palaeoclimatological significance, NW Sicily. *Marine Geology* 161: 155–70.

—— Bard, E., Potter, E.-K., Silenzi, S., and Improta, S. (2004), 215-ka History of sea-level oscillations from marine and continental layers in Argentarola Cave speleothems. *Global and Planetary Change* 43: 57–78.

—— Kershaw, S., Renda, P., Rust, D., Belluomini, G., Cerasoli, M., Radtke, U., and Silenzi, S. (2006), Elevation of the last interglacial highstand in Sicily (Italy): A benchmark of coastal tectonics. *Quaternary International* 145–6: 3–18.

Armijo, R., Meyer, B., King, G. C. P., Rigo, A., and Papanastassiou, D. (1996), Quaternary evolution of the Gulf of Corinth rift and its implications for the Late Cenozoic evolution of the Aegean. *Geophysical Journal International* 126: 11–53.

Arnaud-Fassetta, G. and Provansal, M. (1999), High frequency variations of water flux and sediment discharge during the Little Ice Age (1586–1725 AD) in the Rhone Delta (Mediterranean France). Relationship to catchment. *Hydrobiologia* 410: 241–50.

Bard, E., Hamelin, B., Arnold, M., Montaggioni, L., Cabioch, G., Faure G., and Rougerie, F. (1996), Deglacial sea-level record from Tahiti corals and the timing of global meltwater discharge. *Nature* 382: 241–4.

—— Antonioli, F., and Silenzi, S. (2002), Sea level during the penultimate interglacial period based on submerged stalagmite from Argentarola Cave (Italy). *Earth and Planetary Science Letters* 196: 135–46.

Barusseau, J. P., Akouango, É., Bâ, M., Deschamps, C., and Golf, A. (1996), Evidence for short-term retreat of the barrier shorelines. *Quaternary Science Reviews* 15: 763–71.

Ben Gai, Y., Ben-Avraham, Z., Buchbinder, B., and Kendall, C. G. St. C. (2005), Post-Messinian evolution of the Southeastern Levant Basin based on two-dimensional stratigraphic simulation. *Marine Geology* 221: 359–71.

Blackman, D. J. (2005), Archaeological evidence for sea level changes. *Zeitschrift für Geomorphologie* 137: 61–70.

Blanc, P. L. (2002), The opening of the Plio-Quaternary Gibraltar Strait: assessing the size of the cataclysm. *Geodinamica Acta* 15: 303–17.

Blanchon, P. and Shaw, J. (1995), Reef drowning during the last deglaciation: evidence for catastrophic sea-level rise and ice-sheet collapse. *Geology* 23: 4–8.

Bordoni, P. and Valensise, G. (1998), Deformation of the 125 ka marine terrace in Italy: tectonic implications, in I. S. Stewart and C. Vita-Finzi (eds.), *Coastal Tectonics*. Geological Society, London, Special Publication 146: 71–110.

Boudouresque, C. F. (1971), Contribution à l'étude phytosociologique des peuplements algaux des côtes varoises. *Vegetatio* 22: 83–184.

Brückner, H. (2002), Holocene landscape evolution of Buyuk Menderes. *Zeitschrift für Geomorphologie* 127: 47–65.

—— Müllenhoff, M., van der Borg, K., and Vött, A. (2004), Holocene coastal evolution of western Anatolia—the interplay between natural factors and human impact, in *Human Records of Recent Geological Evolution in the Mediterranean Basin—Historical and Archaeological Evidence*. CIESM Workshop Monograph 24: 51–6.

—— Vött, A., Schriever, M., and Handl, M. (2005), Holocene delta progradation in the eastern Mediteraanean—case studies in their historical context. *Méditerranée* 1/2: 95–106.

Brunetti, A., Denèfle, M., Fontugne, M., Hatté, C., and Pirazzoli, P. A. (1998), Sea-level and subsidence data from a Late Holocene back-barrier lagoon (Valle Standiana, Ravenna, Italy). *Marine Geology* 150: 29–37.

Burak, S., Doğan, E., and Gazioğlu, C. (2004), Impact of urbanisation and tourism on coastal environment. *Ocean and Coastal Management* 47: 515–27.

Cabannes, C., Cazenave, A., and Le Provost, C. (2001), Sea level rise during past 40 years determined from satellite and in situ observations. *Science* 294: 840–2.

Camuffo, D. and Sturaro, G. (2004), Use of proxy-documentary and instrumental data to assess the risk factors leading to sea flooding in Venice. *Global and Planetary Change* 40: 93–103.

Carobene, L. and Dai Pra, G. (2003), Middle and Upper Pleistocene sea-level highstands along the Tyrrehenian coast of Basilicata (southern Italy). *Il Quaternario* 4: 173–202.

Carter, R. W. G. (1988), *Coastal Environments*. Academic Press, New York.

Cazenave, A., Bonnefond, P., Mercier, F., Dominh, K., and Toumazou, V. (2002), Sea level variations in the Mediterranean Sea and Black Sea from satellite altimetry and tide guages. *Global and Planetary Change* 34: 59–86.

Church, J. A. and White, N. J. (2006), A 20th century acceleration in global sea-level rise. *Geophysical Research Letters* 33.

Clauzon, G., Suc, J.-P., Gautier, F., Berger, A., and Loutre, M.-F. (1996), Alternative interpretation of the Messinian salinity crisis: controversy resolved? *Geology* 24: 363–6.

Collier, R. E. L., Leeder, M. R., and Jackson, J. A. (1995), Quaternary drainage development, sediment fluxes and extensional tectonics in Greece, in J. Lewin, M. G. Macklin, and J. C. Woodward (eds.), *Mediterranean Quaternary River Environments*. Balkema, Rotterdam, 31–44.

Collina-Girard, J. (1997), Analyse de profils sous-marins levés en plongée sur les côtes provençales, implications eustatiques et néo-tectoniques. *CR Acad. Sc. Paris* 325: 955–9.

—— (1998), Paliers bathymétriques observés en plongée au littoral occidental de la Corse. Conséquences néotectoniques. *CR Acad. Sci. Paris* 327: 121–6.

—— (2001), L'Atlantide devant le détroit de Gibraltar? Mythe et géologie. *CR Acad. Sci. Paris, Earth and Planetary Sciences* 333: 233–40.

—— (2002), La Crise Finiglaciaire à Gibraltar et l'Atlantide: Tradition orale et géologie. *Préhistoire Anthroplogie Méditerranéennes* 10/11: 53–60.

Corre, J.-J. (1991), The sand dunes and their vegetation along the Mediterranean coast of France. Their likely response to climatic change. *Landscape Ecology* 6: 65–75.

Cundy, A. (2005), Recent rapid sea-level change in the eastern Mediterranean and the coastal sedimentary record. *Zeitschrift für Geomorphologie*, Suppl.137: 29–35.

—— Kortekaas, S., Dewez, T., Stewart, I. S., Collins, P. E. F., Croudace, I. W., Maroukian, H., Papanastassiou, D., Gaki-Papanastassiou, P., Pavlopoulos, K., and Dawson, A. (2000), Coastal wetlands as recorders of earthquake subsidence in the Aegean: a case study of the 1894 Gulf of Atalanti earthquakes, central Greece. *Marine Geology* 170: 3–26.

Davis Jr., R. A. (1996), *Coasts*. Prentice Hall, Upper Saddle River, NJ.

Dawson, A. G. (1996), The geological significance of tsunamis. *Zeitschrift für Geomorphologie* 102: 199–210.

—— and Stewart, I. S. (2007), Tsunami deposits in the geological record. *Sedimentary Geology* 200: 166–83.

Day, J. W., Pont, D., Hensel, P. F., and Ibanez, C. (1995), Impacts of sea-level rise on deltas in the Gulf of Mexico and the Mediterranean: the importance of pulsing events to sustainability. *Estuaries* 18: 636–47.

De Guidi, C., Catalano, S., Monaco, C., and Tortoni, L., (2003), Morphological evidence of Holocene coseismic deformation in the Taormina region (NE Sicily). *Journal of Geodynamics* 36: 193–211.

Di Vita, A. (1996), Earthquakes and civil strife at Gortyn (Crete) in the period between Justinian and Constant II (6–7th century AD), in S. Stiros and R. E. Jones (eds.), *Archaeoseismology*, British School at Athens Fitch Laboratory Occasional Paper 7: 45–50.

Dominey-Howes, D. T. M., Cundy, A. B., and Croudace, I. W. (2000), High energy marine flood deposits on Astypalaea Island, Greece: Possible evidence for the AD 1956 southern Aegean tsunami. *Marine Geology* 163: 303–15.

—— Dawson, A., and Smith, D. (1998), Late Holocene coastal tectonics at Falasarna, W. Crete: a sedimentary study, in I. S. Stewart and C. Vita-Finzi (eds.), *Coastal Tectonics*, Geological Society, London, Special Publication 146: 343–52.

Douglas B. C. and Peltier, W. R. (2002), The Puzzle of Global Sea-Level Rise. *Physics Today* 55: 35–40.

Dubar, M. and Anthony, E. J. (1995), Holocene environmental change and river-mouth sedimentation in the Baie des Anges, French Riviera. *Quaternary Research* 43: 329–43.

Duggen, S., Hoernle, K., van den Bogaard, P., Rüpke, L., and Philips Morgan, J. C. (2003), Deep roots for the Messinian salinity crisis. *Nature* 422: 602–6.

—— —— —— and Harris, C. (2004), Magmatic evolution of the Alboran region: The role of subduction in forming the western Mediterranean and causing the Messinian Salinity Crisis. *Earth and Planetary Science Letters* 218: 91–108.

Dumas, B., Gueremy, P., Lhenaff, R., and Raffy, J. (1993), Rapid uplift, stepped marine terraces and raised shorelines on the Calabrian coast of the Messina Straits, Italy. *Earth Surface Processes and Landforms* 18: 241–56.

Dvorak, J. J. and Mastrolorenzo, G. (1991), The mechanisms of recent vertical crustal movements in Campi Flegrei caldera, southern Italy. *Special Paper of the Geological Society of America* 263: 1–47.

Emery, K. O., Aubrey, D. G., and Goldsmith, V. (1988), Coastal neo-tectonics of the Mediterranean from tide-gauge records. *Marine Geology* 81: 41–52.

Fabbri, P. (1985), Coastline variations in the Po delta since 2500 BP. *Zeitschrift für Geomorphologie* 57: 155–67.

Fairbanks R. G. (1989), A 17,000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature* 342: 637–2.

Faivre, S. and Fouache, E. (2003), Some tectonic influences on the Croatian shoreline evolution in the last 2000 years. *Zeitschrift für Geomorphologie* 47: 521–37.

Fanos, A. M. (1995), The impact of human activities on the erosion and accretion of the Nile Delta coast. *Journal of Coastal Research* 11: 821–33.

Ferranti, L., Antonioli, F., Mauz, B., Amorosi, A., Dai Pra, G., Mastronuzzi, G., Monaco, C., Orrù, P., Pappalardo, M., and Radtke, U. (2006), Markers of the last interglacial sea-level high stand along the coast of Italy: Tectonic implications. *Quaternary International* 145/6: 30–54.

Firth, C., Stewart, I., McGuire, W. J., Kershaw, S., and Vita-Finzi, C. (1996), Coastal elevation changes in eastern Sicily; implications for volcano instability, in W. J. McGuire, A. P. Jones, and J. Neuberg (eds.), *Volcano Instability on the Earth and Other Planets.* Geological Society, London, Special Publication 110: 153–67.

Flemming, N. C. (1969), Archaeological evidence for eustatic change in sea level and earth movements in the Western Mediterranean in the last 2000 years. *Geological Society of America, Special Paper* 109.

—— (1993), Predictions of relative coastal sea-level change in the Mediterranean based on archaeological, historical and tide-gauge data in L. Jeftic, J. D. Milliman, and G. Sestini (eds.), *Climate Change and the Mediterranean*. Edward Arnold, London, 247–81.

——(1978), Holocene eustatic changes and coastal tectonics in the north-east Mediterranean: implications for models of crustal consumption. *Philosophical Transactions of the Royal Society of London* A289: 405–58.

——(1998), Archaeological evidence for vertical movement on the continental shelf during the Palaeolithic, Neolithic and Bronze Age periods, in I. S. Stewart and C. Vita-Finzi (eds.), *Coastal Tectonics*. Geological Society, London, Special Publication 146: 129–46.

Fletcher, C. A. and Spencer, T. (2005), *Flooding and Environmental Challenges for Venice and its Lagoon*. Cambridge University Press, Cambridge.

Fornós, J. J., Gelabert, B., Ginés, À., Ginés, J., Tuccimei, P., and Vesica, P. (2002), Phreatic overgrowths on speleothems: a useful tool in structural geology in littoral karstic landscapes. The example of eastern Mallorca (Balearic Islands). *Geodinamica Acta* 15: 113–25.

Fouache, É., Dalongeville, R., Kunesch, S., Suc, J.-P., Subally, D., Prieur, A., and Lozouet, P. (2005), The environmental setting of the harbor of the Classical sites of Oeniades on the Acheloos delta, Greece. *Geoarchaeology* 20: 285–302.

——Desruelles, S., Pavlopoulos, K., Dalgonville, R., Coquinot, Y., Peulvast, J.-P., and Potdevin, J.-L. (2005), Using beachrocks as sea level indicators in the insular group of Mykonos, Delos and Rhenia (Cyclades, Greece). *Zeitschrift für Geomorphologie* 137: 37–43.

Frihy, O. E. (2001), The necessity of environmental impact assessment (EIA) in implementing coastal projects: lessons learned from the Egyptian Mediterranean coast. *Ocean and Coastal Management* 44: 489–516.

——Dewidar, K. M., and El Raey, M. M. (1996), Evaluation of coastal problems at Alexandria. *Ocean and Coastal Management* 30: 281–95.

Galili, E. and Sharvit, J. (1998), Ancient coastal installations and the tectonic stability of the Israeli coast in historical times, in I. S. Stewart and C. Vita-Finzi (eds.), *Coastal Tectonics*. Geological Society, London, Special Publication 146: 147–63.

Gamble, C. (1986), *The Palaeolithic Settlement of Europe*. Cambridge University Press, Cambridge.

Garcia, C. and Servera, J. (2003), Impacts of tourist development on water demand and beach degradation on the island of Mallorca (Spain). *Geografiska Annaler* 85A: 287–300.

Görür, N., Cağatay, M. N., Emre, Ö., Alpar, B., Sakinç, M., ÃÏslamoğlu, Y., Algan, O., Erkal, T., Keçer, M., Akkök, and Karlık, G. (2001), Is the abrupt drowning of the Black Sea shelf at 7150 yr BP a myth? *Marine Geology* 176: 65–73.

Got, H., Aloisi, J.-C., Monaco., A. (1985), Sedimentary processes in Mediterranean deltas and shelves, in S. J. Stanley and F.-C. Wetzel (eds.), *Geological Evolution of the Mediterranean Basin*. Springer, New York, 355–76.

Goy, J. L. and Zazo, C. (1988), Sequences of Quaternary marine levels in Elche Basin (eastern Betic Cordillera, Spain). *Palaegeography, Palaeoclimatology, Palaeoecology* 68: 301–10.

——Zazo, C., and Dabrio, C. J. (2003), A beach-ridge progradation complex reflecting periodical sea-level and climate variability during the Holocene (Gulf of Almería, Western Mediterranean). *Geomorphology* 50: 251–68.

Grenon, M. and Batisse, M. (eds.) (1989), *Futures for the Mediterranean Basin: The Blue Plan*. Oxford University Press, Oxford.

Grove, A. T. and Rackman, O. (2001), *The Nature of Mediterranean Europe: An Ecological History*. London, Yale University Press.

Guillén, J. and Palanques, A. (1996), Short- and medium-term grain size changes in deltaic beaches (Ebro Delta, NW Mediterranean). *Sedimentary Geology* 101: 55–67.

Hanson, H., Brampton, A., Capobianco, M., Dette, H. H., Hamm, L., Laustrup, C., Lechuga, A., and Spanhoff, R. (2002), Beach nourishment projects, practices and objectives—a European overview. *Coastal Engineering* 47: 81–111.

Haslett, S. K. (2000), *Coastal Systems*. Routledge, London.

Hodell, D. A., Curtis, J. H., Siero, F. J., and Raymo, M. E. (2001), Correlation of late Miocene to early Pliocene sequences between the Mediterranean and North Atlantic. *Palaeooceanography* 16: 164–78.

Hsü, K. J., Ryan, W. B. F., and Cita, M. B. (1973). Late Miocene dessication of the Mediterranean. *Nature* 242: 240–4.

Inman, D. L. (1994), Types of coastal zones, in *Environmental Science in the Coastal Zone*. Washington, DC, National Research Council, 67–84.

——and Jenkins, S. A. (1984), The Nile littoral cell and man's impact on the coastal zone of the southeastern Mediterranean, in *Proceedings 19th Coastal Engineering Conference, American Society of Civil Engineers*, 2: 1600–17.

——and Nordstrom, C. E. (1971), On the tectonic and morphologic classification of coasts. *Journal of Geology* 79: 1–21.

Jacobsen, T. W. (1976), 17,000 years of Greek prehistory. *Scientific American* 234: 76–87.

Jiménez, J. A. and Sánchez-Arcilla, A. (1993), Medium-term coastal response at the Ebro delta, Spain. *Marine Geology* 114: 105–18.

Kelletat, D. and Schellman, G. (2002), Tsunami in Cyprus: field evidences and ^{14}C dating results. *Zeitschrift für Geomorphologie*, Suppl. 46: 19–34.

——Kowalczyk, G., Schröder, B., and Winter, K. P (1976), A synoptic view on the neotectonic development of the Peloponnesian coastal regions. *Zeitschrift der Deutschen Gesellschaft für Geowissenschaften* 127: 447–65.

Keraudren, B. and Sorel, D. (1987), The terraces of Corinth (Greece)—a detailed record of eustatic sea-level variations during the last 500,000 years. *Marine Geology* 77: 99–107.

Kerey, I. E., Meriç, E., Tunoğlu, C., Kelling, G., Brenner, R. L., and Doğan, A. U. (2004), Black Sea–Marmara Sea Quaternary connections: new data from the Bosphorus, Istanbul, Turkey. *Palaeogeography, Palaeoclimatology, Palaeoecology* 204: 277–95.

Kershaw, S. (2000), Quaternary reefs of northeastern Sicily: Structure and growth controls in an unstable tectonic setting. *Journal of Coastal Research* 16/4: 1037–62.

——and Guo, L. (2001), Marine notches in coastal cliffs; indicators of relative sea-level change, Perachora peninsula, central Greece. *Marine Geology* 179: 213–28.

Kraft, J. C., Aschenbrenner, S. E., and Rapp Jr., G., (1977), Application of Holocene stratigraphy to paleogeographic reconstructions of coastal Aegean archaeological sites. *Science* 195: 941–7.

——Rapp, G., Kayan, I., and Luce, J. V. (2003), Harbor areas at ancient Troy: sedimentology and geomorphology complement Homer's *Iliad*. *Geology* 31: 163–6.

Krijgsman, W. (2002), The Mediterranean: Mare Nostrum of Earth Sciences. *Earth and Planetary Science Letters* 205: 1–12.

——Hilgen, F. J., Raffi, I., Siero, F. J., and Wilson, D. S. (1999), Chronology, causes and progression of the Messinian salinity crisis. *Nature* 400: 652–5.

Laborel, J. (1979), Fixed marine organisms as biological indicators for the study of recent sea level and climatic variations along the Brazilian tropical coast. *Proc. 1978 International Symposium on Coastal Evolution in the Quaternary*. São Paulo, 193–211.

—— (1986) Vermetid gastropods as sea level indicators, in O. van de Plaasche (ed.), *Sea Level Research, a Manual for Collection and Evaluation of Data*. Geo Books, Norwich, 28–310.

—— and Laborel-Deguen, F. (1994), Biological indicators of relative sea-level variation and of co-seismic displacements in the Mediterranean area. *Journal of Coastal Research* 10: 395–415.

—— —— (1996), Biological indicators of Holocene sea-level and climatic variations on rocky coasts of tropical and subtropical regions, *Quaternary International* 31: 53–60.

—— Morhange, C., Lafont, R., Le Campion, J., Laborel-Deguen, F., and Sartoretto, S. (1994), Biological evidence of sea-level rise during the last 4500 years on the rocky coasts of continental southwestern France and Corsica. *Marine Geology* 120: 203–23.

—— Morhange, C., Collina-Girard, J., and Laborel-Deguen, F. (1999), Littoral bio-erosion, a tool for the study of sea level variations during the Holocene. *Bulletin of the Geological Society of Denmark* 45: 144–8.

Lambeck, K. (1995), Late Pleistocene and Holocene sea-level change in Greece and south-western Turkey: a separation of eustatic, isostatic and tectonic contributions. *Geophysical Journal International* 122: 1022–44.

—— (1996), Sea-level change and shore-line evolution in Aegean Greece since Upper Palaeolithic time. *Antiquity* 70: 588–611.

—— and Bard, E. (2000), Sea-level change along the French Mediterranean coast since the time of the Last Glacial Maximum. *Earth and Planetary Science Letters* 175: 203–22.

—— and Chappell, J. (2001), Sea level change through the last glacial cycle. *Science* 292: 679–86

—— and Purcell, A. (2005), Sea-level change in the Mediterranean since the LGM: Model predictions for tectonically stable areas. *Quaternary Science Reviews* 24: 1969–88.

—— Antonioli, F., Purcell, A., and Silenzi, S. (2004*a*), Sea-level change along the Italian coast for the past 10,000 yrs. *Quaternary Science Reviews* 23: 1567–98.

—— Anzidei, M., Antonioli, F., Benini, A., and Esposito, A. (2004*b*), Sea level in Roman times in the Central Mediterranean and implications for recent change. *Earth and Planetary Science Letters* 224: 563–75.

Leeder, M. R., Seger, M., and Stark, C. P. (1991), Sedimentology and tectonic geomorphology adjacent to active and inactive normal faults in the Megara Bain and Alkonides Gulf, Central Greece. *Journal of the Geological Society, London* 148: 331–43.

Loget, N., van den Driessche, J., and Davy, P. (2005), How did the Messinian Salinity Crisis end? *Terra Nova* 17: 414–18.

Major, C., Ryan, W., Lericolais, G., and Hajdas, I. (2002), Constraints on Black Sea outflow to the Sea of Marmara during the last glacial–interglacial transition. *Marine Geology* 190: 19–34.

Malvárez Garcia, G., Pollard, J., and Domínguez Rodríguez, R. (2000), Origins, management and measurement of stress on the coast of southern Spain. *Coastal Management* 28: 215–34.

—— —— Hughes, R. (2002), Coastal zone management on the Costa del Sol: a small business perspective. *Journal of Coastal Research* 36: 470–82.

Marriner, N. and Morhange, C. (2006), The 'Ancient Harbour Parasequence': Anthropogenic forcing of the stratigraphic highstand record. *Sedimentary Geology* 186: 13–17.

—— —— (2007), Geoscience of ancient Mediterranean harbours. *Earth-Science Reviews* 80: 137–94.

—— Morhange, C., Doumet-Serhal, C., and Carbonel, P. (2006), Geoscience discovers Phoenicia's buried harbors. *Geology* 34: 1–4.

Mastronuzzi, G. and Sansò, P. (2000), Boulder transport by catastrophic waves along the Ionian coast of Apulia (southern Italy). *Marine Geology* 170: 93–103.

—— —— (2004), Large boulder accumulations by extreme waves along the Adriatic coast of southern Apulia (Italy). *Quaternary International* 120: 173–84.

Miller, L. and Douglas, B. C. (2004), Mass and volume contributions to twentieth- century global sea level rise. *Nature* 428: 406–9.

Milliman, J. D. (1992), Sea-level response to climate change and tectonics in the Mediterranean Sea, in L. Jeftic (ed.), *Climatic Change and the Mediterranean*. Edward Arnold, London, 45–57.

Miyauchi, J., Dai Pra, G., and Sylos Labini, S. (1994), Geochronology of Pleistocene marine terraces and regional tectonics in the Tyrrhenian coast of South Calabria, Italy. *Il Quaternario* 7: 17–34.

Monaco, C., Scicchitano, G., and Tortorici, L. (2006), Large boulder accumulations by tsunami waves along the south-eastern coast of Sicily. *Geophysical Research Abstracts* 8: 02643.

Morhange, C., Laborel, J., and Hesnard, A. (2001), Changes of relative sea level during the past 5000 years in the ancient harbour of Marseilles, Southern France. *Palaeogeography, Palaeoclimatology, Palaeooecology* 166: 319–29.

—— Blanc, F., Bourcier, M., Carbonel, P., Prone, A., Schmitt-Mercury, S., Vivent, D., and Hesnard, A. (2003), Biosedimentology of the late Holocene deposits of the ancient harbor of Marseilles (Southern France, Mediterranean sea). *The Holocene* 13/4: 593–604.

—— Marriner, N., Laborel, L., Todesco, M., and Oberlin, C. (2006), Rapid sea-level movements and noneruptive crustal deformations in the Phlegrean Fields caldera, Italy. *Geology* 34: 93–6.

Mörner, N.-A. (1996), Sea level variability. *Zeitschrift für Geomorphologie*, NS 102: 223–32.

Murray, S., Coleman, J., Roberts, H. H., and Salama, M. (1981), Accelerated currents and sediment transport off the Damietta Nile promontory. *Nature* 292: 51–4.

Naylor, L. and Viles, H. A. (2002), A new technique for evaluating short-term rates of bioerosion and bioprotection. *Geomorphology* 47: 31–44.

Nicholls, R. J. and Hoozemans, F. M. J. (1996), The Mediterranean: vulnerability to coastal implications of climate change. *Ocean and Coastal Management* 31: 105–32.

—— —— and Marchand, M. (1999), Increasing flood risk and wetland losses due to global sea-level rise: regional and global analyses. *Global Environmental Change* 9: S69–S87.

Orsi, G., Civetta, L., Del Gaudio, C., de Vita, S., Di Vito, M. A., Isaia, R., Petrazzuoli, S. M., Ricciardi, G. P., and Ricco, C. (1999), Short-term ground deformations and seismicity in the resurgent Camp Flegrei caldera (Italy): an example of active block-resurgence in a densely populated area. *Journal of Volcanology and Geothermal Research* 91: 415–51.

Oueslati, A. (1995), The evolution of low Tunisian coasts in historical times: from progradation to erosion and salinisation. *Quaternary International* 29/30: 41–7.

Papadopoulos, G. A. (1998), *A Tsunami Catalogue of Greece and Adjacent Seas*. EC Project GITEC-TWO, Institute of Geodynamics, National Observatory of Athens, 8.

Paskoff, R. and Oueslati, A. (1991), Modifications of coastal conditions in the Gulf of Gabès (Southern Tunisia) since Classical Antiquity. *Zeitschrift für Geomorphologie*, Suppl. 81: 149–62.

Peltier, W. R. (1987), Mechanisms of relative sea-level change and the geophysical responses to ice-water loading, in R. J. N. Devoy (ed.), *Sea Surface Studies*. Croom Helm, London, 57–94.

—— (1998), Global glacial adjustment and coastal tectonics, in I. S. Stewart and C. Vita-Finzi (eds.), *Coastal Tectonics*. Geological Society, London, Special Publication 146: 1–30.

—— (2000), On eustatic sea level history: Last Glacial Maximum to Holocene. *Quaternary Science Reviews* 21: 377–96.

—— (2001), Global slacial isostatic adjustment and modern instrumental records of relative sea-level history, in B. Douglas, M. S. Kearney, and S. P. Leatherman (eds.), *Sea Level Rise: History and Consequences*. Academic Press, San Diego, 65–96.

—— (2004), Global glacial isostatic adjustment and the surface of the ice-age Earth: The ICE-5G(VM2) model and GRACE. *Annual Reviews of Earth and Planetary Sciences* 32: 111–49.

Péres, J. M. (1982), Zooations and organismic assemblages, *in* O. Kinne (ed.), *Marine Ecology: A Comprehensive, Integrated Treatise on Life in Oceans and Coastal Waters*. Wiley, New York, i.

Petit Maire, N. (2004), Climatic crises and man in the Mediterranean basin: the last 20,000 years, in *Human Records of Recent Geological Evolution in the Mediterranean Basin—Historical and Archaeological Evidence*. CIESM Workshop monograph 24: 17–24.

Piper, D. J. W. and Panagos, A. G. (1981), Growth patterns of the Acheloos and Evinos deltas, western Greece. *Sedimentary Geology* 28: 111–32.

Pirazzoli, P. 1976. Sea level variations in the Northwest Mediterranean during Roman times, *Science*, 194, 519–521.

—— (1986*a*), Marine notches, in O. van de Plaasche (ed.), *Sea-level Research: A Manual for the Collection and Evaluation of Data*. Geo Books, Norwich, 361–400.

—— (1986*b*), The Early Byzantine Tectonic Paroxysm. *Zeitschrift für Geomorphologie*, NS 62: 31–49.

—— (1991), *World Atlas of Holocene Sea-Level Changes*, Oceanography Series. Elsevier, Amsterdam.

—— (1998), A comparison between postglacial isostatic predictions and late Holocene sea-level field data from Mediterranean and Iranian coastal areas. *GeoResearch Forum* 3/4: 401–20.

—— (2005), A review of possible eustatic, isostatic and tectonic contributions in eight late-Holocene relative sea-level histories from the Mediterranean area. *Quaternary Science Reviews* 24: 1989–2001.

—— (2006), The projected 'MOSE' barriers against flooding in Venice (Italy) and the expected global sea-level rise. *Journal of Marine Environmental Engineering* 8: 247–61.

—— Ausseil-Badie, J., Giresse, P., Hadjidaki, E., and Arnold, M. (1992), Historical environmental changes at Phalasarna harbour, West Crete. *Geoarchaeology* 7: 373–92.

—— Stiros, S. C., Arnold, M., Laborel, J., Laborel-Deguen, F., and Papageorgiou, S. (1994*a*), Episodic uplift deduced from Holocene shorelines in the Perachora Peninsula, Corinth area, Greece. *Tectonophysics* 229: 201–9.

—— —— Laborel, J., Laborel-Deguen, F., Arnold, M., Papageorgiou, S., and Morhange, C. (1994*b*), Late Holocene shoreline changes related to palaeoseismic events in the Ionian Islands, Greece. *The Holocene* 4: 397–405.

—— Laborel, J., and Stiros, S. C. (1996*a*), Coastal indicators of rapid uplift and subsidence: examples from Crete and other Eastern Mediterranean sites, *Zeitschrift für Geomorphologie*, Suppl. NS 102: 21–35.

—— —— —— (1996*b*), Earthquake clustering in the Eastern Mediterranean during historical times, *Journal of Geophysical Research* 101 B3: 6083–97.

—— —— Saliege, J. F., Erol, O., Kayan, I., and Person, A. (1991), Holocene raised shorelines on the Hatay coast (Turkey) Paleoecological and tectonic implications, *Marine Geology* 96: 295–311.

Price, S., Higham, T., Nixon, L., and Moody, J. (2002), Relative sea-level changes in Crete: Reassessment of radiocarbon dates from Sphakia and west Crete. *Annual of the British School at Athens* 97: 171–200.

Rodríguez-Vidal, J., Cáceres, L. M., Finlayson, J. C., Garcia, F. J., and Martínez-Aguirre, A. (2004), Neotectonics and shoreline history of the Rock of Gibraltar, southern Iberia. *Quaternary Science Reviews* 23: 2017–29.

Rust, D. and Kershaw, S. (2000), Holocene tectonic uplift patterns in northeastern Sicily: evidence from marine notches in coastal outcrops. *Marine Geology* 167: 105–26.

Ryan, W. B. F. and Pitman, W. C. (2000), *Noah's Flood: The New Scientific Discoveries About the Event that Changed History*. Touchstone, New York.

—— —— Major, C. O., Shimkus, K., Moskalenko, V., Jones, G. A., Dimitrov, P., Görür, N., Sakinç, M., and Yüce, H. (1997), An abrupt drowning of the Black Sea Shelf. *Marine Geology* 138: 119–26.

—— Major, C. O., Lericolais, G., and Goldstein, S. L. (2003), Catastrophic flooding of the Black Sea. *Annual Reviews in Earth and Planetary Sciences* 31: 525–54.

Said, R. (1981), The *Geological Evolution of the River Nile*. Springer, New York.

Siddall, M., Pratt, L. J., Helfrich, K. R., and Gosan, L. (2004), Testing the physical oceanographic implications of the suggested sudden Black Sea infill 8400 years ago. *Palooceanography* 19: doi:10.1029/2003PA000903.

Sivan, D., Wdowinski, S., Lambeck, K., Galili, E., and Raban, A. (2004), Holocene sea-level changes along the Mediterranean coast of Israel, based on archaeological observations and numerical model. *Palaeogeography, Palaeoclimatology, Palaeooecology* 167: 101–17.

Soloviev, S. L. (1990), Tsunamigenic zones in the Mediterranean Sea. *Natural Hazards* 3: 183–202.

—— Solovieva, O. N., Go, C. N., Kim, K. S., and Shchetnikov, N. A. (2000), *Tsunamis in the Mediterranean Sea, 2000 BC–2000 AD*. Kluwer Academic, Dordrecht.

Soter, S. (1998), Holocene uplift and subsidence of the Helike Delta, Gulf of Corinth, Greece, in I. S. Stewart and C. Vita-Finzi (eds.), *Coastal Tectonics*. Geological Society, London, Special Publication 146: 41–56.

Stanley, D. J. (1989), Subsidence of the northeastern Nile Delta: Rapid rates, possible causes, and consequences. *Science* 240: 497–500.

—— and Warne, A. G. (1994), Worldwide initiation of Holocene marine deltas: deceleration of sea-level rise as a principle factor. *Science* 265: 228–31.

—— —— (1997), Holocene sea-level change and early human utilisation of deltas. *GSA Today* 7/12: 1–7.

—— Goddio, F., Jorstad, T. F., and Schnepp, G. (2004), Submergence of ancient Greek cities off Egypt's Nile Delta. *GSA Today* 14: 4–10.

Stewart, I. (1996), Holocene uplift and palaeoseismicity on the Eliki Fault, western Gulf of Corinth, Greece. *Annali di Geofisica* 39: 575–88.

—— Cundy, A., Kershaw, S., and Firth, C. (1997), Holocene coastal uplift in the Taormina area, northeastern Sicily: implications for the southern prolongation of the Calabrian seismogenic belt. *Journal of Geodynamics* 24: 37–50.

Stiros, S. C. (2000), Fault pattern of Nisyros Island volcano (Aegean Sea): structural and archaeological evidence, in W. J. McGuire, D. R. Griffiths, P. L. Hancock, I. S. Stewart, and C. Vita-Finzi (eds.), *The Archaeology of Geological Catastrophes*. Geological Society, London, Special Publication 171: 385–97.

—— and Papageorgiou, S. (2001), Seismicity of Western Crete and destruction of the town of Kisamos at AD 365: Archaeological evidence. *Journal of Seismology* 5: 381–97.

—— Arnold, M., Pirazzoli, P. A., Laborel, J., Laborel, F., and Papageorgiou, S. (1992), Historical coseismic uplift on Euboea Island, Greece. *Earth and Planetary Science Letters* 108: 109–17.

Surić, M., Juračić, M., Horvatiničić, N. and Bronić, I. K. (2005), Late Pleistocene-Holocene sea-level rise and the pattern of coastal karst inundation: records from submerged speleothems along the Eastern Adriatic Coast (Croatia). *Marine Geology* 214: 163–75.

Tetjana, R., Garrett, C., and Le Traon, P. Y. (2000), Western Mediterranean sea-level rise: Changing exchange flow through the Strait of Gibraltar. *Geophysical Research Letters* 27: 2949–52.

Thommeret, Y., Thommeret, J., Laborel, J., Montaggioni, L. F., and Pirazzoli, P. A. (1981), Late Holocene shoreline changes and seismo-tectonic displacements in western Crete (Greece). *Zeitschrift für Geomorphologie*, NS 40: 127–49.

Tinti, S. and Maramai, A. (1996), Catalogue of tsunamis generated in Italy and in Côte d'Azur, France: a step towards a unified catalogue of tsunamis in Europe. *Annali di Geofisica* 39: 1253–99.

Tosi, L., Carbognin, L., Teatini, P., Strozzi, T., and Wegmüller, U. (2002), Evidence of the present relative land stability of Venice, Italy, from land, sea, and space observations. *Geophysical Research Letters* 29: doi:1029/2001GL013211.

Trenhaile, A. S. (2002), Rock coasts, with particular emphasis on shore platforms. *Geomorphology* 48: 7–22.

Tsimplis, M. N. and Blackman, D. (1997), Extreme sea-level distribution and return periods in the Aegean and Ionian Seas. *Estuarine, Coastal and Shelf Science* 44: 79–89.

Tsoar, H. and Blumberg, D. G. (2002), Formation of parabolic dunes from barchan and transverse dunes along Israel's Mediterranean coast. *Earth Surface Processes and Landforms* 27: 1147–61.

Tuccimei, P., Fornós, J. J., Ginés, À., Ginés, J., Gràcia, F., and Mucedda, M. (2003), Sea level change at Capo Caccia (Sardinia) and Malorca (Balearic Islands) during oxygen isotope substage 5e, based on Th/U datings of phreatic overgrowths on speleothems, in *Puglia 2003—Final Conference Project IGCP* 437.

Uluğ, A., Duman, M., Ersoy, Ş., Özel, E., and Avcı, M. (2005), Late Quaternary sea-level change, sedimentation and neotectonics of the Gulf of Gökova, southeastern Aegean Sea. *Marine Geology* 221: 381–95.

van Andel, T. H. (1989), Late Quaternary sea-level changes and archaeology. *Antiquity* 63: 733–46.

—— Jacobsen, T. W., Jolly, J. B., and Lianos, N. (1980), Late Quaternary history of the coastal zone near Franchthi Cave, southern Argolid, Greece. *Journal of Field Archaeology* 7: 389–402.

—— Zangger, E., and Perissoratis, C. (1990), Quaternary transgressive/regressive cycles in the Gulf of Argos, Greece. *Quaternary Research* 34: 317–29.

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

van der Meulen, F. and Salman, A. H. P. M. (1996), Management of Mediterranean coastal dunes. *Ocean and Coastal Management* 30: 177–95.

Vella, C. and Provansal, M. (2000), Relative sea-level rise and neotectonic events during the last 6500 yr on the southern eastern Rhône delta, France. *Marine Geology* 170: 27–39.

—— Fleury, T.-J., Raccasi, G., Provansal, M., Sabatier, F., and Bourcier, M. (2005), Evolution of the Rhône delta plain in the Holocene. *Marine Geology* 222/3: 235–65.

Vesica, P. L., Tuccimei, P., Turi, B., Fornós, J. J., Ginés, À., and Ginés, J. (2000), Late Pleistocene palaeoclimates and sea-level change in the Mediterranean as inferred from stable isotope and U-series studies of overgrowths on speleothems, Mallorca, Spain. *Quaternary Science Reviews* 19: 865–79.

Warny, S. A., Bart, P. J., and Suc, J.-P. (2003), Timing and progression of climatic, tectonic and glacioeustatic influences on the Messinian Salinity Crisis. *Palaegeography, Palaeoclimatology, Palaeoecology* 202: 59–66.

Westaway, R. (1993), Quaternary uplift of southern Italy. *Journal of Geophysical Research* 87: 21,741–72.

Whelan, F. and Kelletat, D. (2002), Geomorphic evidence and relative and absolute results for tsunami events in Cyprus. *Science of Tsunami Hazards* 20: 3–18.

Wood, P. B. (1996), Dating and Origin of Late Quaternary Catastrophic Shoreline Activity around the Mediterranean Sea. Unpublished Ph.D. thesis, Royal Holloway College, University of London.

Wyatt, T. (2004), Geochronology and myth—are gods catastrophes? in *Human Records of Recent Geological Evolution in the Mediterranean Basin—Historical and Archaeological Evidence*. CIESM Workshop monograph 24: 119–30.

Yaltirak, C., Sakinç, M., Aksu, A. E., Hiscott, R. N., Galleb, B., and Ulgen, U. B. (2002), Late Pleistocene uplift history along the southwestern Marmara Sea determined from raised coastal deposits and global sea-level variations. *Marine Geology* 190: 282–305.

Yalçıner, A. C., Alpar, B., Altınok, Y., Özbay, İ., and Inamura, F. (2002), Tsunamis in the Sea of Marmara: Historical documents for the past, models for the future. *Marine Geology* 190: 445–63.

Zazo, C., Silva, P. G., Goy, J. L., Hillaire-Marcel, C., Ghaleb, B., Lario, J., Bardají, T., and González, A. (1999), Coastal uplift in continental collision plate boundaries: data from the Last

Interglacial marine terrace of the Gibraltar Straits area (south Spain). *Tectonophysics* 301: 95–109.

—— Goy, J. L., Dabrio, C. J., Bardají, T., Hillaire-Marcel, C., Ghaleb, B., González-Delgado, J. Á., and Soler, V. (2003), Pleistocene raised marine terraces of the Spanish and Atlantic coasts: records of coastal uplift, sea-level highstands and climate changes. *Marine Geology* 194: 103–33.

Zerbini, S. (and 16 others) (1996), Sea level in the Mediterranean: a first step towards separating crustal movements and absolute sea-level variations. *Global and Planetary Change* 14: 1–48.

Zviely, D. and Klein, M. (2004), Coastal cliff retreat rates at Beit-Yannay, Israel, in the 20th century. *Earth Surface Processes and Landforms* 29: 175–84.

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