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Late Pleistocene and Holocene sea-level change in Greece and south-western Turkey: a separation of eustatic, isostatic and tectonic contributions

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SUMMARY

Sea-level change during Late Pleistocene and Holocene times is a combination of eustatic, isostatic and tectonic contributions. In the central Mediterranean Sea region of Greece and western Turkey, the isostatic components are important, due to the changing gravitational potential of the Earth, the readjustment of the crust upon the removal of the large ice sheets and the addition of the meltwater into the oceans, including the Mediterranean. Changes in relative sea-level due to these glacio-hydro-isostatic adjustments have reached amplitudes of 1 mm yr^{-1} during the last few thousand years. A model for the isostatic contribution to sea-level change, including the movement of shorelines due to the combined eustatic–isostatic changes, is developed, based on earth-model and ice-sheet parameters estimated from sea-level observations from other areas. Comparisons of this model with observations of sea-level change permit rates of vertical movement to be estimated for Late Holocene time. Allowance for the isostatic factors makes a significant difference to these rates. The plains of Argolis, Lakonia, Messinia and Navarine, in the southern Peloponnese, for example, appear to be tectonically stable on time-scales of a few thousand years and longer, consistent with the position of the Last Interglacial shoreline close to the present-day sea-level. The observations here, of a rising sea-level, are largely the consequence of the glacio-hydro isostasy and not of a long-term tectonic process. Crete as a whole is subject to tectonic uplift, but at variable rates, and forms part of an arcuate zone of uplift extending from Rhodes and Karpathos in the east to Kithera in the west at rates that locally exceed 4 mm yr^{-1}. The southern shore of the Gulf of Corinthos is also subject to uplift at rates approaching 1.5 mm yr^{-1}. These isostatically corrected estimates for the past few thousand years are in agreement with longer-term estimates based on the position of the Last Interglacial shoreline. Only on the Perachora Peninsula do the tectonic rates over these two time intervals appear to be in disagreement, with the Late Holocene rates being much higher than the long-term rates.

Key words: Greece, Holocene, isostasy, sea-level, tectonic rates.

1 INTRODUCTION

That sea-level in the Greece and Aegean regions has changed significantly in the past 20 000 yr is not a matter for dispute, but the magnitudes and timings of this change, as indicated by published estimates, are a matter of greater controversy. Thus it is generally recognized that during the last glacial maximum, sea-level was at least 100 m lower than it is today, with some estimates ranging up to 165 m (van Andel & Shackleton 1982). It is also accepted that much of the subsequent sea-level rise occurred from about 15 000 yr BP (before present) until mid-Holocene times, but published estimates of this rise have varied greatly. Compare, for example, some of the often-quoted results, obtained primarily from the Atlantic and Gulf of Mexico coasts of the United States of America, and used in discussions of the coastal consequences of sea-level change in the Mediterranean (Shackleton, van Andel & Runnels

1984). There is also a broad agreement that sea-level since mid-Holocene times has been relatively stable, with some authors arguing for small oscillations in level and others arguing for a level that continued to rise during this period (e.g. Bintliff 1977; Flemming 1968). There are a number of reasons for these different results, including the often uncertain nature of the observational evidence itself. Other reasons include the occurrence of vertical tectonic displacements of the crust relative to the sea and inadequate corrections, or no corrections at all, for the glacio-hydro-isostatic adjustments of the sea-floor in response to the changing surface loads of ice sheets and meltwater. For the eastern Mediterranean all these reasons are important.

The best way to establish sea-level change for a particualr location is to infer it from the geological and geomorphological records for the region or, as in the case of the Mediterranean region, from the archaeological records. However, observations that permit a quantitative estimate of the position of the sea surface at a known time seldom exist for the area of interest and, if they do, will probably be a partial temporal record at best. Instead, knowledge of the sea-level record in space and time is usually patchy and what is requied is a physical model for interpolation between the fragments of information, so as to be able to predict sea-level change at any place for any time. This is what the eustatic and glacio-hydro-isostatic model does for tectonically stable areas, providing a relatively smoothly varying and predictable function of sea-level change in time and space. However, when vertical tectonic processes are important, the interpolation model becomes much more difficult, because the rates of the crustal displacements can change abruptly over comparatively short distances; they might not be uniform in time when viewed over periods of the order of millenia either.

Relative sea-level change for the past 20 000 or so years is the sum of two principal contributions: the increase in the ocean volume as the great ice sheets of the late Pleistocene age melted, the eustatic change, and the vertical movements of the crust itself. This latter contribution comprises two components: (i) the vertical movements arising from any active tectonics; and (ii) the vertical isostatic movements associated with any changes in the surface loads of ice and water. The total sea-level change relative to the crust therefore consists of eustatic, isostatic and tectonic signals. The eastern Mediterranean is a region of active tectonics and the region of Greece, and the Aegean Sea, in particular, is one of the most rapidly deforming continental areas on Earth (e.g. McKenzie 1978; Le Pichon & Angelier 1979; Jackson 1994). The region is also unusual in that it is at sea-level and that there is, therefore, the potential for quantifying the rates of vertical movements in recent geological time. However, the region also lies sufficiently near to formerly glaciated northern Europe for the isostatic factor to be significant. Thus the separation of the isostatic and eustatic components from the tectonic component is important if sea-level is to be used as a reference for determining vertical movements.

Several approaches are possible to isolate the various contributions to sea-level change for the eastern Mediterranean region, where no one part of the shore line can be assumed to be tectonically stable on time-scales of tens of thousands of years. A first approach is to use one of the published eustatic sea-level curves as a reference, and to attribute any observed discrepancies from it as being caused by a combination of the isostatic and tectonic processes. A potential problem with this is that many of the published eustatic curves are based on limited data from restricted areas, and as such they might already be contaminated by either tectonic effects or isostatic contributions, and the viability of translating such curves to the Aegean must be questioned (e.g. Flemming 1978; van Andel & Shackleton 1982). Nevertheless, this approach is widely used, largely because the detailed isostatic models have not been available. Another approach that has sometimes been used is the analysis of the sea-level observations to obtain a 'eustatic' function that is only time-dependent and a 'tectonic' function that is both time- and position-dependent (Flemming 1972). Unless the isostatic factor has been corrected for separately, however, this latter function will be the combination of the tectonic and isostatic contributions. Flemming further assumed that the tectonic part at any one location is linear over the past few thousand years, whereas the occurrence of rather well-defined emerged notches observed along some sectors of the Greek shoreline suggests that the tectonic component on and near faults is more likely to be episodic on this time-scale.

A third approach, which is the one adopted here, is to develop models of eustasy and isostasy for tectonically stable areas in other parts of the world and then to use these models to predict the eustatic–isostatic contribution for the area in question. Studies of these contributions have been made for regions where other tectonic factors are believed to be small, so that the models can be fully tested and the relevant model parameters can be evaluated. Departures from this model of observed sea-levels can then be attributed to tectonic factors (and to the limitations of both the model predictions and the observational evidence). The advantage of this approach is that it allows for the spatial variability of the isostatic adjustment and does not assume that the tectonic changes follow any particular spatial or temporal function. The success of this approach does depend on whether successful isostatic models can be developed with parameters that are representative of the region under consideration. If this can be achieved, however, the model can then be used to predict the positions of former shorelines and palaeo-water depths. Alternatively, if the morphology of such shorelines, or features that can be related to palaeo-shorelines, has been observed, then the model can be used to predict the ages of these features.

The main purpose of this paper is to demonstrate the importance of the isostatic corrections for the seas around Greece and western Turkey, and to examine some of the consequences of these corrections on estimates of tectonic rates. The principal data used is the important compilation of archaeological evidence made by Flemming (1978), including the solution-notch evidence discussed by him, as well as some of the data discussed by Pirazzoli *et al.* (1982, 1994), with the emphasis on the regions of Crete and the Peloponnese and, to a lesser degree, south-western Turkey. Section 2 of the paper summarizes the glacio-hydro-isostatic model as applied to the eastern Mediterranean region. The parameters required to quantify this model are discussed in Section 3 and the observational evidence for sea-level change is discussed in Section 4. The comparison of models with the observational data and the estimation of tectonic rates are discussed in Section 5.

2 THE RELATION BETWEEN SEA-LEVEL, ICE SHEETS AND THE SOLID EARTH

In the presence of vertical tectonic motions, the relative sea-level change, $\Delta\zeta(\varphi, t)$, at a site φ and time t can be written schematically as

$$\Delta\zeta(\varphi, t) = \Delta\zeta_e(t) + \Delta\zeta_I(\varphi, t) + \Delta\zeta_T(\varphi, t), \qquad (1)$$

where $\Delta\zeta_e(t)$ is the eustatic sea-level, $\Delta\zeta_I(\varphi, t)$ is the isostatic correction for the response of sea-level to the redistribution of the surface load as a result of the exchange of mass between the ice sheets and the oceans, and $\Delta\zeta_T(\varphi, t)$ is the tectonic contribution. During times of glaciation, water is taken out of the ocean to form the continental ice sheets and, on average, sea-level will fall by an amount

$$\Delta\zeta_e(t) = (\rho_{ice} \times \text{change in ice volume})/(\rho_{water} \times \text{ocean surface area}), \qquad (2)$$

(where ρ is density). This is the eustatic sea-level (esl) change, and it provides a measure of the change in total ice volume through time. Even on a rigid earth, not subject to any isostatic adjustment, the eustatic sea-level provides only a partial description of the actual sea-level, because the gravitational attraction of the newly formed ice sheet tends to pull the water towards the ice mass. Thus, in the vicinity of the ice margin, sea-level actually rises relative to the eustatic change during times of ice-sheet growth, while further away, in order to preserve mass, sea-level falls by more than the eustatic amount. For an ice sheet with the dimensions of the former Fennoscandian ice sheet, the influence of this effect is important well beyond the ice margin.

When the continental ice sheets melt and the ocean volumes increase, the surface load of the planet is modified, as is the stress distribution within the planet. The additional stresses will usually be sufficiently large to induce an instantaneous elastic deformation of the Earth followed by a viscous flow in the mantle. Thus, under the shrinking ice sheet the Earth's surface rebounds and mantle material flows towards the centre of uplift. This behaviour is well illustrated by the crustal adjustment of Fennoscandia and of the Hudson Bay region of Canada, where sea-level is seen to be falling relative to the crust. But smaller scale adjustments occur globally. For example, crustal adjustment remains significant in non-glaciated areas such as the Mediterranean where, mainly in response to the withdrawal of mantle material towards the formerly ice-loaded crust, the crust subsides and relative sea-levels rise.

The concomitant change in the ocean load further modifies the state of stress in the Earth, and contributes to the new flow regime in the mantle and to surface deformation; the incrementally loaded ocean floor tends to subside under the increasing water load, whereas adjacent unglaciated land masses, free from any change in surface load, tend to rise in compensation. The combined contribution of these components of the isostatic correction is a complex spatial pattern of sea-level change during and after deglaciation, which can be written schematically as

$$\Delta\zeta_I(\varphi, t) = \Delta\zeta_r(\varphi, t) + \Delta\zeta_i(\varphi, t) + \Delta\zeta_w(\varphi, t). \qquad (3)$$

The first term is the consequence primarily of the gravitational attraction of the ice sheet on the ocean surface for a rigid earth and will be a function of both position and time. Its influence extends well beyond the formerly glaciated regions, but in the eastern Mediterranean the main influence is from the former Fennoscandian ice sheet. A second contribution to $\Delta\zeta_r$, much smaller for this region, comes from the changing gravitational attraction of the water load. The second term in eq. (3) is the ice-load term $\Delta\zeta_i$, or the glacio-isostatic term, and it describes the incremental change in sea-level resulting from the deformation of the Earth's surface under the changing ice sheets. It includes any contribution from the modification of the gravity field produced by the changing mass distribution of the ice and by the redistributed mass within the mantle. This term is a function of the rheological parameters describing the Earth's response, and of the temporal and spatial distribution of the ice sheets. All the global ice sheets contribute to the changes in sea-level, although, for the eastern Mediterranean region, it is the Scandinavian ice sheet that is responsible for much of the spatial variability in sea-level change.

The third term in eq. (3), the water-load term $\Delta\zeta_w$, or hydro-isostatic term, defines the contribution from the adjustment of the Earth to the new water load and includes the contribution from the associated changes in the gravity field. This term is a function of the Earth's rheology, of the change in sea-level itself by an amount that is spatially variable, and of the shape of the oceans. This last effect includes the time-dependent changes in shoreline positions and changes in the sea-floor geometry, particularly near the ice sheets, that modify the holding capacity of the oceans. Eq. (3) is therefore an integral equation that can be solved iteratively (Johnston 1993; Lambeck 1993b).

In addition to predicting temporal and spatial changes in sea-level, the glacio-hydro-isostatic model also predicts the location of past shorelines and palaeo-water depths, if the present water depth $H_0(\varphi)$ at location φ is known. That is, the water depth at a time t and position φ are related to $\Delta\zeta(\varphi, t)$ by

$$H(\varphi, t) = H_0(\varphi) - \Delta\zeta(\varphi, t). \qquad (4)$$

To solve eqs (3) and (4) for the spatial and temporal distributions of sea-level change during a glacial cycle, several requirements have to be met. One is a model of the ice-sheet growth and decay. Because of the global nature of the problem, all ice sheets must be considered, although the required accuracy of their description will depend on the location of the region under consideration. For localities near former ice sheets, a detailed description is required, whereas for regions away from the ice margins, such as the Mediterranean, approximate spatial descriptions of the changing ice sheets will generally suffice, provided that the overall rate of change of ice volume of each ice sheet is known with reasonable accuracy. The eustatic sea-level function $\Delta\zeta_e(t)$ will be defined by the time dependence of the melting of the combined ice sheets.

A second requirement for the solution of eqs (3) and (4) is a model to describe the planet's response to surface loading. This model needs to define both the elastic response and the time-dependent or viscous response. The former part is described by the density and elastic parameters derived from global seismic models and is well known. Less well known is the viscosity profile, which is usually approximated by a number of discrete layers of different 'effective' linear viscosities, whose values are inferred from the observational data and the rebound model. The true viscosity structure of the Earth is likely to be more complex than is suggested by such a simple model, but the latter does provide a very adequate description of the behaviour of the planet when it is subjected to a changing surface load on time-scales of thousands of years. A third requirement is a description of the coastline geometry, as this enters into the calculation of the water-load term. The modification of this coastline associated with sea-level change must also be considered, particularly if high-accuracy local solutions are sought for a region such as the eastern Mediterranean. Hence, models of shallow water bathymetry and coastal elevations are required for the accurate prediction of palaeo-shorelines using eq. (4). A further requirement is a reliable and accurate numerical scheme for the evaluation of the isostatic terms. Solutions are usually obtained by expanding the surface loads and the ocean geometry into spherical harmonics, and high-degree expansions, out to degree 240 or more, depending on the choice of earth model, are required in order that the estimates of the two load terms, particularly of the water-load term, converge to an extent that is commensurate with observational accuracies of 1 m or better.

Of the various requirements, it is the rheology of the Earth and the history of the ice sheets and their changes in volume during the last glacial cycle that are not well known. Solutions of the sea-level eq. (3) have therefore tended to be iterative ones, in which observations of sea-level from tectonically stable areas are compared with the predictions to improve the relevant model parameters. Because of the spatial variability of the Earth's response to the various ice- and earth-model parameters entering into eq. (3), a separation of the various unknowns is sometimes possible, and internally consistent solutions with useful predictive capabilities can be developed (e.g. Lambeck 1993a,b).

A simple model suffices to illustrate the relative importance of the various contributions to the sea-level change. Consider a model earth that is ocean-covered except for a continental polar cap of radius 1000 km (Fig. 1a). This cap is covered by a longitude-independent ice sheet of dimensions that are comparable to those of the Fennoscandian ice at the time of the last glacial maximum. The earth model has a lithosphere of 65 km effective thickness, an upper mantle viscosity of 4×10^{20} Pa s and a lower (below 670 km depth) mantle viscosity of 10^{22} Pa s. The ice is removed instantaneously at 12 000 yr before present, and the resulting sea-level change is predicted along a meridional section radiating out from the centre of the ice sheet. Greece corresponds to a colatitude of about 25° from the centre of the load.

The rigid term $\Delta\zeta_r$ comprises two parts: the attraction arising from the ice sheet and the attraction arising from any change in the water depths associated with the formation of the ice sheet. The first part is illustrated in Fig. 1(b), which

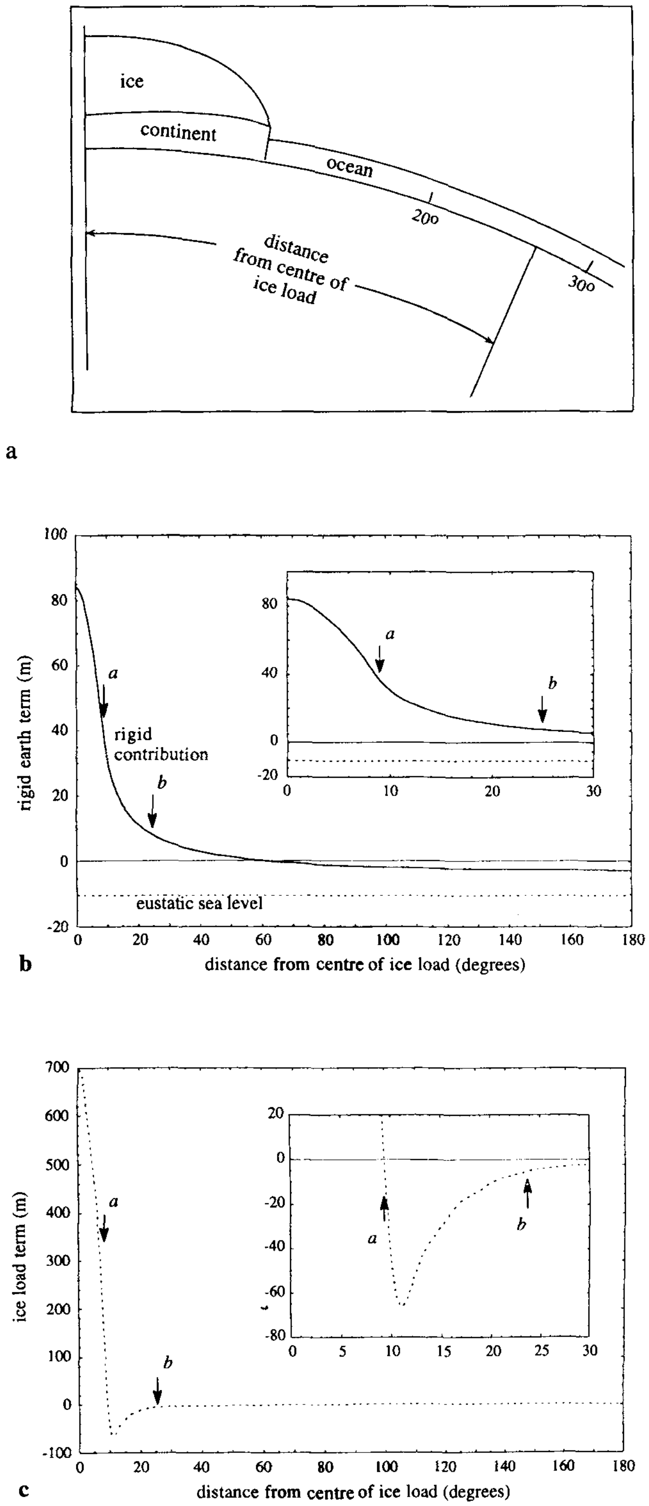


Figure 1. Schematic isostatic corrections for a nominally axisymmetric ice sheet over northern Europe (see text) as a function of distance from the centre of the ice sheet. (a) The configuration of the axisymmetric load which melts instantaneously at 12 000 yr BP. (b) The rigid term $\Delta\zeta_r$ and the corresponding eustatic term at $T > 12\,000$ yr BP. [The arrow *a* marks the location of the ice limit. The Aegean region (arrow *b*) lies at 20–25° from the ice centre.] (c) The present ice-load term, $\Delta\zeta_i$, 12 000 yr after the instantaneous melting of the ice sheet. The insets show the details of the two contributions out to a distance of 30°.

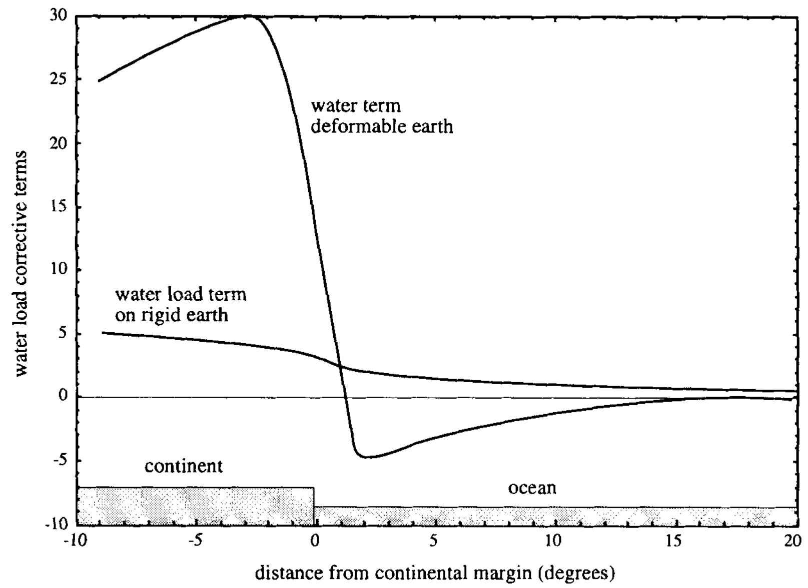


Figure 2. Isostatic and gravitational attraction corrections across a linear continental margin, due to an instantaneous water load of 120 m added into the oceans at 12 000 yr BP. The rigid contribution of the water load corresponds to times before 12 000 yr BP, and the earth deformation term due to the water is for an epoch 12 000 yr after the addition of the water.

gives the sea-level change due to the gravitational attraction of the ice sheet. This term is positive out to a large distance beyond the former ice-sheet margin, including the distances corresponding to Greece and the Aegean Sea, so that sea-levels in this region and at the time of the ice sheet would, in the absence of the other contributions, stand higher than the present level by as much as 7–12 m. Much further away from the ice load, this term becomes negative, so that the average of this contribution over the ocean area is zero; that is, ocean mass is conserved. The ice-load term, including the effect of the crustal deformation and any displacement of mass within the planet, is shown in Fig. 1(c) for a time 12 000 yr after the instantaneous removal of the ice. Beneath the ice, the crust rebounds such that sea-levels formed prior to the deglaciation event will now lie above present sea-level, but, further away, at the distance of the Aegean Sea, shorelines of glacial age will now lie below the present level. Much further away again, this contribution turns positive once more, such that its average value over the oceans is zero. In the Aegean region, the glacio-isostatic term is smaller than the rigid term, and the combined effect is to place the shorelines in glacial times at a shallower depth than would be predicted by the eustatic estimate only.

The water-load term for sites across a continental margin is illustrated in Fig. 2. The assumed total eustatic change is 120 m and occurs instantaneously at 12 000 yr BP. The rigid contribution of the changing water load is not negligible, but the dominant contribution arises from the planet's deformation under the water load, producing a relative fall in sea-level seawards of the coast and a rise in level at the coast and at inland sites, along narrow estuaries or bays penetrating into the interior of the continent. The magnitude of this contribution is quite significant and, in view of the complex coastal geometry in the Aegean and eastern Mediterranean region, considerable spatial variability in sea-level can be expected. For example, coastal observations of sea-level at the time of glaciation could be as much as 10 m less than predicted by the eustatic estimate, in addition to the contribution from the ice load.

These simple models are inadequate for evaluating the isostatic corrections, but they illustrate that the deviations of sea-level from the simple eustatic assumption can be substantial, and spatially variable, not only during the early stages of deglaciation but also during the post-glacial phase. More realistic models require the inclusion of the other ice sheets, an improved description of the geometry of the ice sheets during their retreat, and consideration of the complex coastal geometry. These are discussed in Section 3.3.

3 MODEL PARAMETERS AND SEA-LEVEL PREDICTIONS

The observation equation linking the observed relative sea-level values to the model parameters is (Lambeck 1993b)

$$\Delta\zeta_o(\varphi, t) + \varepsilon_o(\varphi, t) = \Delta\zeta_e(t) + \delta\zeta_e(t) + \beta^j\Delta\zeta_i^j + \Delta\zeta_w, \qquad (5)$$

where $\Delta\zeta_o$ is the observed shoreline elevation (reduced to mean sea-level), above or below present sea-level, at a location φ and a time t for a total of $s = 1 \cdots S$ observations, each of accuracy σ_s; ε_o is the estimation of the observation error; $\Delta\zeta_e$ is the eustatic sea-level function for the totality of the ice sheets; $\delta\zeta_e$ is the correction term to $\Delta\zeta_e$; $\Delta\zeta_i^j$ is the glacio-isostatic correction for the jth ice sheet; β^j is the scale parameter for the height of the jth ice sheet; and $\Delta\zeta_w$ is the hydro-isostatic correction corresponding to the meltwater from the totality of the ice sheets.

The ice-scale parameter β^j is introduced because, although the ice limits are often reasonably well known from geomorphological data, the thickness of the ice sheet is mostly inferred from glaciological models and not from direct observations. These observation equations can be solved by a least-squares process for any of the unknowns, such as the earth-model parameters that are included in the two isostatic terms, the correction to the nominal eustatic function, and the scale factors for the ice sheets. Solutions are usually carried out by region: for Fennoscandinavia or Great Britain, for example, to estimate primarily the ice parameters for these ice sheets together with the Earth's rheological parameters (e.g. Lambeck 1993b), or for the Australian margin, to estimate the eustatic function and the rheological parameters, adopting the β scale parameters determined previously from data near and within the ice limits (e.g. Lambeck & Nakada 1990).

3.1 The eustatic sea-level curve

Published estimates of the eustatic sea-level curve are varied, and some examples are illustrated in Fig. 3(a). There are several reasons for this wide range, including problems associated with the interpretation of the observational evidence. Corals, for example, used to estimate the sea-level curve, provide a lower limit to the curve, while submerged, *in situ* terrestrial vegetation such as fresh-water peats or tree stumps provide an upper limit. Another reason for the discrepancies is the neglect of, or the inadequate correction for, the glacio-hydro-isostatic factors. Thus a number of the frequently quoted estimates used in sea-level studies in the Mediterranean (e.g. van Andel & Shackleton 1982) are based mainly on observations from the Atlantic coast of North America where the relative sea-level is influenced by the glacio-isostatic rebound of the Laurentide region, and the observed levels in lateglacial time will generally lie above the eustatic curve [see, for example, a somewhat analogous case for the coastal site of Kavalla in Fig. 6(e), below]. The adoption of such curves for the Mediterranean Sea will therefore result in an underestimation of the eustatic component. Along continental margins, far from the ice sheets, corrections for the hydro-isostatic contribution can also be significant during lateglacial time, of the order of 20 m across the Australian margin at the time of the last glacial maximum, for example (Lambeck & Nakada 1990). Observations from islands far from both continental margins and ice sheets provide better estimates of the eustatic sea-level, but even these are not immune to the isostatic factors. Here the ocean floor and island can be expected to move vertically together, and the main departure from the eustatic change is due to the redistribution of water in the oceans in response to the isostatic adjustment of the sea-floor, particularly around the glacial areas. Thus the establishment of a global eustatic sea-level curve requires models for the isostatic corrections, and the procedure adopted here is to apply model (5) to several areas of the world where either other tectonic processes producing vertical movements are believed to be insignificant or corrections for these movements can be made, and where either the isostatic corrections are not strongly earth-model-dependent or the earth-model parameters can be adequately separated from the other model parameters.

Figure 3(b) illustrates the best estimate of the eustatic curve based on analyses of sea-level change from locations far from the ice sheets, along the continental margin of Australia and from a number of sites in the Pacific region, as well as from sites in Europe. This result is consistent with the upward coral reef growth-rate curve from Barbados (Fairbanks 1989), when the latter is corrected for the isostatic perturbations. Major contributions to this globally integrated rate of melting come from the Laurentian ice sheet, from Fennoscandia and the Barents Sea area, and from Antarctica (Nakada & Lambeck 1988). One characteristic of this sea-level curve is that the bulk of melting ceased at about 6000 yr ago when the last ice from

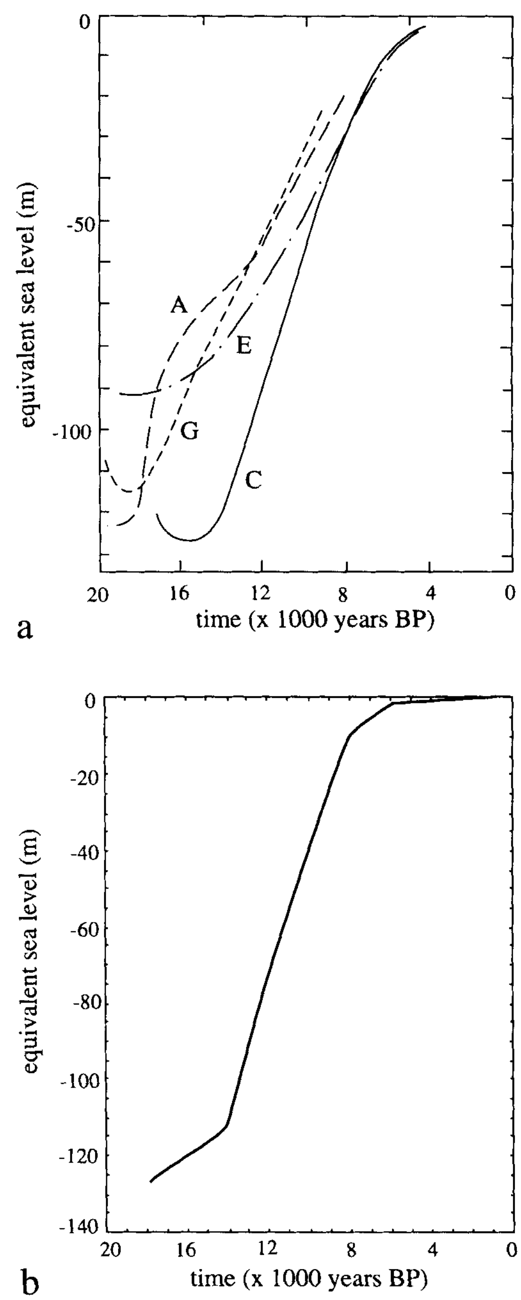


Figure 3. (a) Estimates of eustatic sea-level change, based on observational evidence from the Atlantic and Gulf of Mexico margins of the United States of America (A, C, E) and from the east African coast (G) (from van Andel & Shackleton 1982). (A: Curray 1965; C: Milliman & Emery 1968; E: Dillon & Oldale 1978; G: Delibrias 1974.) (b) Eustatic sea-level curve based on depth–age observations corrected for the glacio-hydro-isostatic contributions.

Laurentia vanished. (The Fennoscandian ice sheet vanished soon after 9000 yr BP.) However, a small amount of additional meltwater continued to be added to the oceans after this time, from a reduction in the Antarctic ice volume, so as to raise the eustatic level by about 2 m during the past 6000 yr (Nakada & Lambeck 1988; Lambeck 1993b). Another characteristic of this eustatic sea-level curve is that the function is devoid of rapid oscillations. Detailed studies of sea-level change during the past 6000 yr from the same or nearby sites indicate that there is little evidence for such oscillations once the major ice sheets have melted (or approached their present volumes) (Chappell 1983). Likewise, estimates of the correction term $\delta\zeta_e$ (eq. 5) for the period between 12 000 and 6000 yr BP suggest that any oscillations in this interval are likely to be less than about 2 m (Lambeck 1993b).

For the precise prediction of the isostatic corrections, the fluctuations in the ice-sheet volumes before the last glacial maximum are required. This information is scanty and reliable models for the individual ice sheets between the times of the last interglacial and the glacial maximum do not exist. Estimates of the eustatic sea-level curve before the last glacial maximum also cannot be established from sea-level information alone because of the paucity of observations preserved in the earlier geological record. Instead, an approximate estimate of this function is obtained from the oxygen isotope records of sea-floor sediments scaled by the more directly observed changes for the past 18 000 yr (Chappell & Shackleton 1986; Shackleton 1987). This results in a globally integrated estimate of the changes in ice volume, but, combined with some local information on the individual ice sheets, it is possible to obtain estimates of the volume changes of the major ice sheets since the last interglacial period that are largely adequate for the isostatic modelling, provided that the predictions of sea-level change are not extended back in time beyond about 18 000 yr BP.

3.2 Earth-model parameters

The elastic moduli of the Earth and the radial density distribution are obtained from the analysis of seismic wave velocities through the planet, and these quantities are well known. The assumed viscosity models are usually simpler, and the parameters are deduced from the sea-level observations themselves. Studies from different localities produce largely consistent results for the earth-model parameters, although there is some evidence that both the upper mantle viscosity and the lithospheric thickness exhibit a degree of lateral variation. Hence the analyses for earth-model parameters using (5) are conducted regionally rather than globally (Nakada & Lambeck 1991). Studies of multilayered models have indicated that a three-layer mantle model, comprising an essentially elastic lithosphere of 60–80 km thickness, an upper mantle (down to the 670 km seismic discontinuity) with an effective viscosity of $(3–5)\times 10^{20}$ Pa s, and a lower mantle with a viscosity of about 10^{22} Pa s, gives an adequate representation of the sea-level response (Lambeck, Johnston & Nakada 1990; Lambeck & Nakada 1990; Lambeck 1993b).

These are effective parameters in the sense that they describe well the response of the planet to surface loading on time-scales of 10^4 yr and are not, therefore, directly comparable with either seismic or long-term tectonic estimates of, for example, lithospheric thickness. In tectonically active areas such as the Aegean Sea, both the lithospheric thickness H_l and the upper mantle viscosity η_{um} can be expected to be lower than the values obtained from the other regional studies. What may constitute appropriate values, however, remains a matter for debate—one can only say for sure that the effective lithospheric thickness will be less than the seismically defined thickness and greater than the thickness estimated from long-term tectonic loading problems. A range of different plausible earth models has been explored, and the results for two of these will be discussed below: (i) the nominal model, within the above range of parameters, with $H_l = 65$ km, $\eta_{um} = 4\times 10^{20}$ Pa s, and $\eta_{lm} = 1.3\times 10^{22}$ Pa s; and (ii) a model with a thin lithosphere and a low upper mantle viscosity, which may be considered to be more appropriate for a tectonically active region, with $H_l = 30$ km, a low-viscosity channel of 5×10^{19} Pa s down to 200 km depth, and a viscosity for the remainder of the mantle equal to that for the first model. Predicted sea-levels for these models differ in observable ways, as discussed below, but the overall trend of the change is the same for both models. Thus, while the rates for vertical tectonic discussed below will be earth-model-dependent, the trends and patterns established should be largely independent of the choice of parameters, at least within the range of the models considered.

3.3 Predicted sea-levels

More realistic estimates of the various contributions to the sea-level change in the Aegean region are illustrated in Figs 4 to 6, based on the nominal ice and earth models discussed above. In these examples the melting of the ice sheets is terminated at 6000 yr BP. Before 18 000 yr BP the northern ice sheet has been assumed to have remained constant for a very long period, such that the stress state at the onset of deglaciation approaches the local isostatic state. In contrast, the Antarctic ice sheet is assumed to have expanded during the period from 29 000 to 18 000 yr BP, before retreating to its present limit by 6000 yr BP, so as to illustrate the consequence of the glacial loading part of the cycle on sea-level.

The temporal and spatial variations in sea-level of isostatic origin are predicted here using the first-iteration formulation. The expansion of the solution is carried out to spherical harmonic degree 240, corresponding to a spatial resolution (half-wavelength) of about 75 km. Tests with higher degree expansions (300) have shown that this is adequate.

Figures 4(a)–(c) illustrate the three contributions, $\Delta\zeta_r$, $\Delta\zeta_i$ and $\Delta\zeta_w$, to the sea-level change associated with the decay of the northern hemisphere ice sheet, and Fig. 4(d) gives the total deviation from the eustatic change. The predictions are for the five sites located in Fig. 4(e). The rigid term is the combined effect of the ice sheet and the changing water load, although, by analogy with the simple model results, this second part is relatively small. The rigid term vanishes at 6000 yr BP, when no further change occurs in the ice sheets. The sites of Skopelos and Izmir lie at about the same distances from the ice centres, and both the rigid and

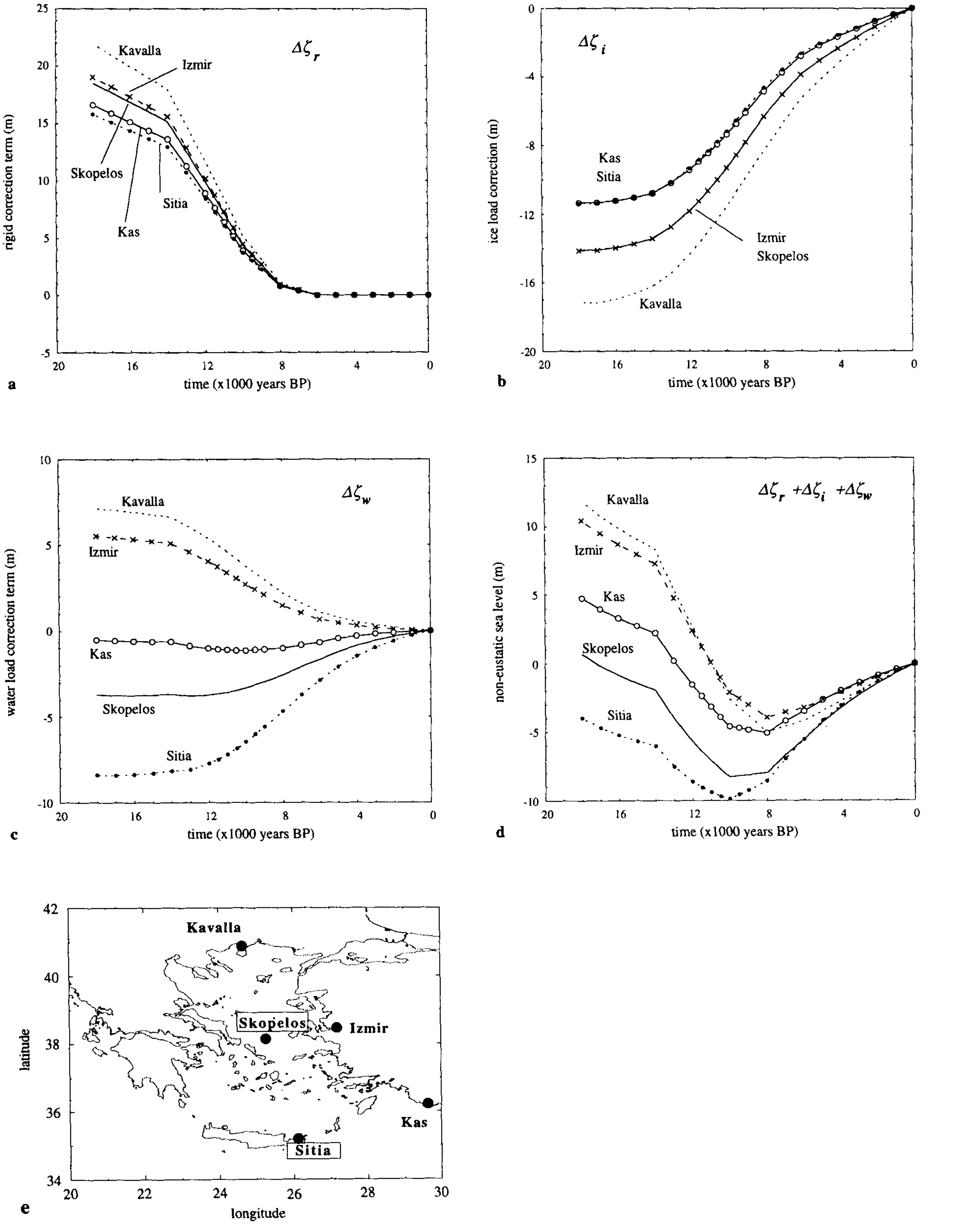


Figure 4. Predicted isostatic corrections for selected sites [see (e) for locations] from the arctic ice sheets. (a) The rigid correction $\Delta\zeta_r$, including the contributions to sea-level from the ice load and the changing water volume (first iteration only) on a rigid earth model; (b) the ice-load terms $\Delta\zeta_i$; (c) the water load terms $\Delta\zeta_w$; (d) the total non-eustatic sea-level correction; (e) location map of the Aegean sites.

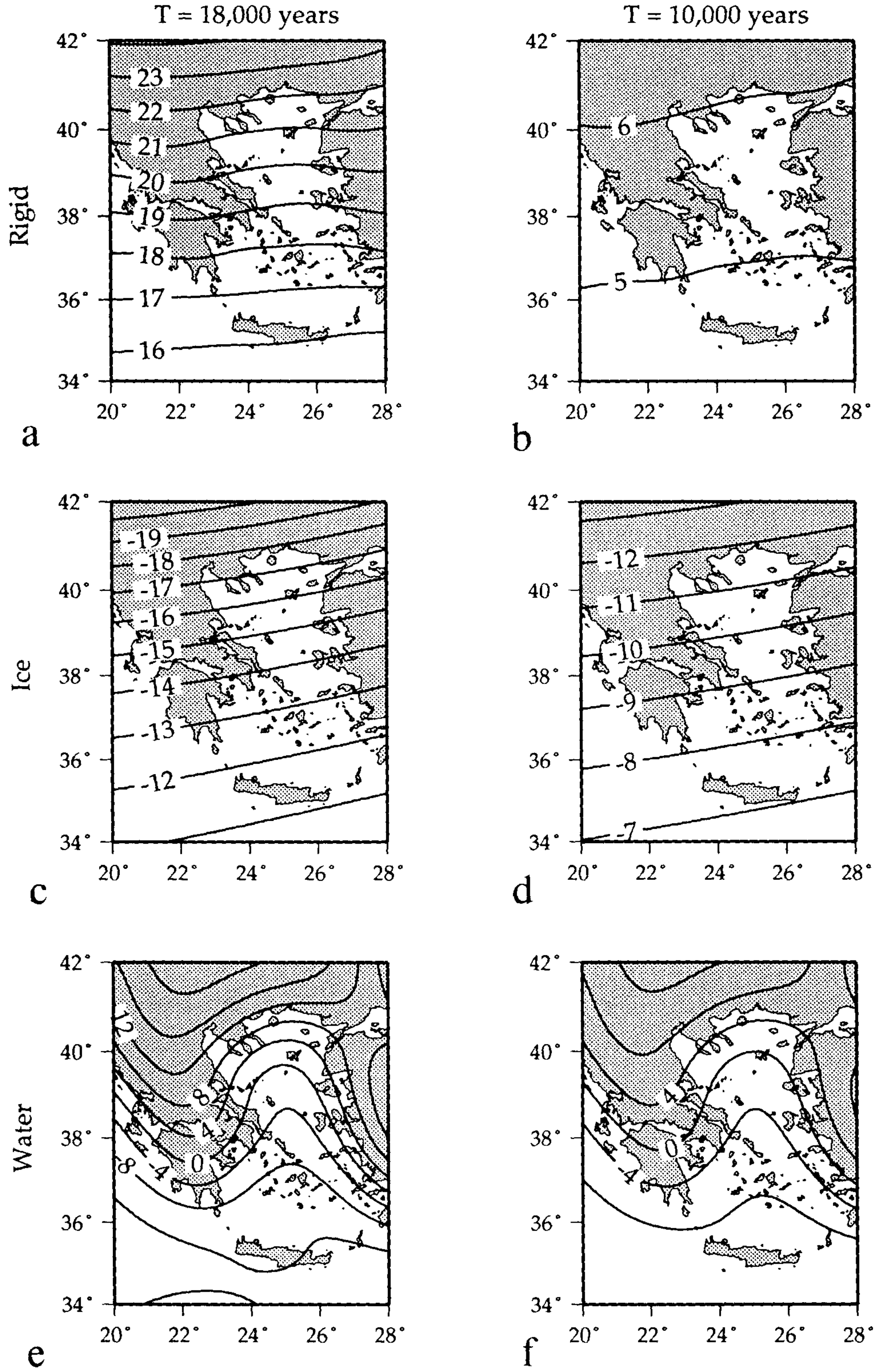


Figure 5. Spatial variation of the rigid (a and b), glacio-isostatic (c and d) and hydro-isostatic (e and f) contributions across Greece and the Aegean region at 18 000 (a, c, e) and 10 000 (b, d and f) yr BP. The first two components are for the combined northern hemisphere ice sheets. The water-load component is for the first-iteration estimate of the combined ice loads from both hemispheres.

ice-load terms are very similar, as is also the case for the Kas and Sitia sites. These predictions do illustrate well the N–S trend that can be expected in the sea-level change from the northern site of Kavalla (Thrace) to Skopelos and to the southern site of Sitia (Crete). The principal contribution to both terms comes from the Scandinavian ice sheet, but the Laurentian ice also makes a significant contribution, even though its centre lies about 65° from the Aegean (*cf.* Fig. 2).

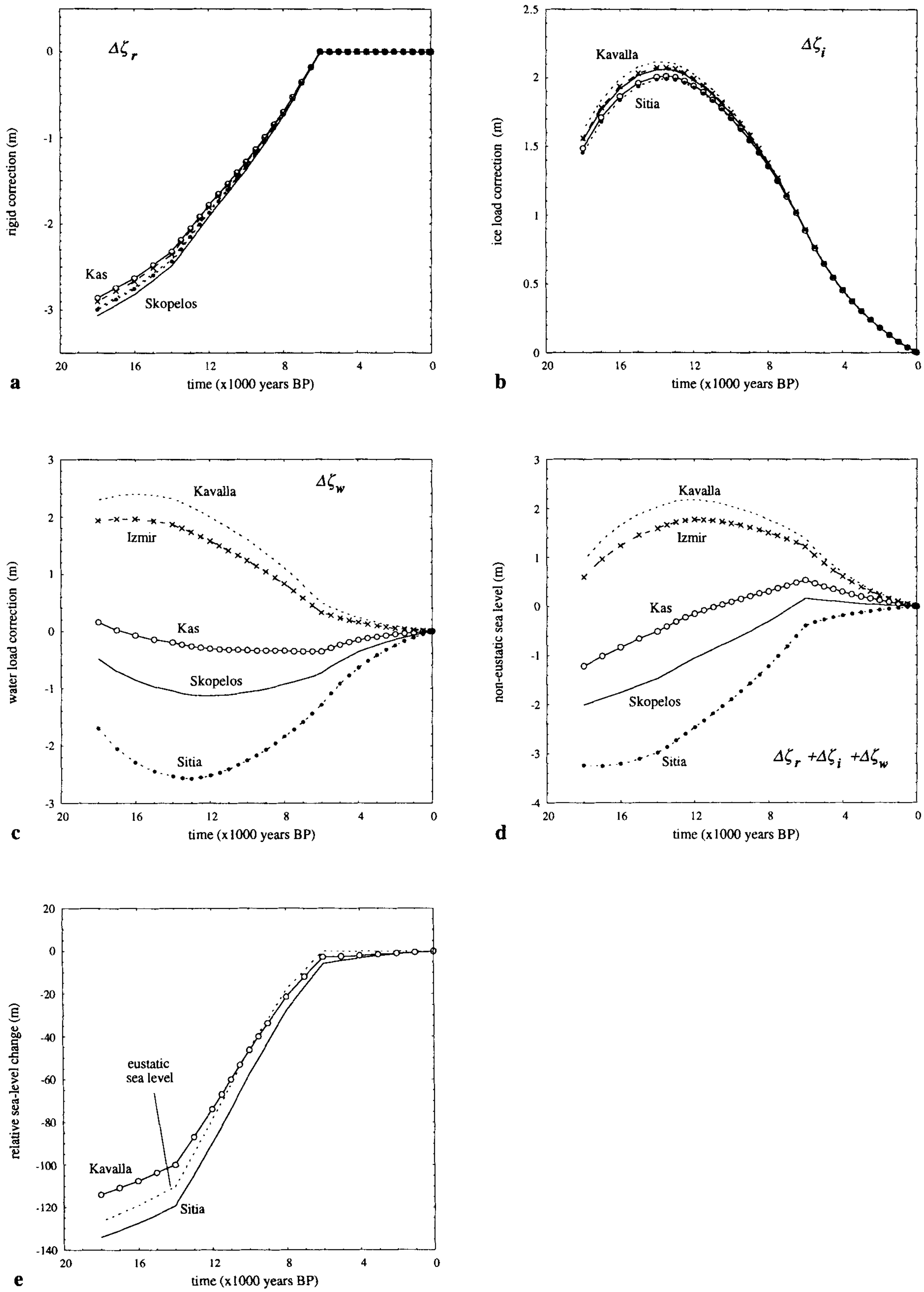


Figure 6. Predicted isostatic correction for selected Aegean sites from changes in the Antarctic ice sheet. (a) The rigid term; (b) the ice-load term; (c) the water-load term; (d) the total non-eustatic sea-level correction; (e) the total relative sea-level change, from both hemispheres, at two of the sites compared with the nominal eustatic sea-level function.

The glacio-isostatic contributions at the five sites are illustrated in Fig. 4(b). They also exhibit a N–S trend, and the predictions for Kas and Sitia, for example, are nearly identical, being at approximately the same distance from the centre of the dominant ice load over northern Europe. The N–S trend is, however, of the opposite sign to that produced by the rigid contribution and, for the early glacial stage, the two partially cancel each other out for these particular distances from the ice. Once the ice sheet is small or vanishes the glacial term dominates.

The water-load term (Fig. 4c) for these realistic models behaves much like the simple model described earlier. Kavalla and Izmir are essentially margin sites and the water-load correction is positive, in contrast to the negative values for the island sites of Skopelos and Sitia. Of these two, the effect for the former site is smaller, because the confined nature of the Aegean Sea prevents the development of the full mid-ocean load effect. At Kas, about which the water load is distributed through an azimuth of about 240°, the response to the water load is between the inland and continental margin responses. The total deviatoric part arising from the northern ice sheets is given in Fig. 4(d). The early pattern, for $T \lesssim 10\,000$ yr BP, is dominated to a large degree by the rigid contribution, whereas the pattern for $\lesssim 8000$ yr BP is more the consequence of the ice-load term, with its viscous response to the changing ice load. The N–S (e.g. Kavalla, Skopelos, Sitia) spatial variability is well illustrated here, but the E–W variability (Skopelos–Izmir and Sitia–Kas), resulting from the different water-load terms, is equally important, and the combined effect is to produce a complex spatial and temporal pattern of sea-level change for these sites. This is well illustrated in Fig. 5 in which the variation of the three contributions is illustrated across Greece and the Aegean Sea for the epoch of maximum glaciation. The rigid term shows the pronounced N–S trend, upon which is superimposed the smaller effect of the gravitational attraction of the water load in the form of the relatively shallower contours over the larger land bodies. Likewise, the glacial isostatic term exhibits a pronounced N–S trend, although it is of opposite sign to that of the rigid part and thus partially cancels out the spatial variability over the region, but not the average value of the corrective terms. After about 10 000 yr BP, when the rigid term is either small or zero, the overall regional N–S trend becomes more pronounced because of the now-dominant (but reduced in amplitude) glacial load term. The water-load contribution represents the greatest spatial variability and will dominate the pattern of sea-level change for this and subsequent epochs. This term also exhibits a pronounced N–S trend in response to the loading of the Mediterranean sea-floor to the south, but the consequence of the loading of the smaller Aegean and Ionian Seas is also seen. It is interesting to note that the southern Peloponnese response is very like that of an island.

Figure 6 illustrates comparable results for the distant Antarctic ice sheet, which lies at a distance of about 120° from the eastern Mediterranean. Here, the contributions from the rigid and ice-load terms are nearly constant over the region and are relatively small in amplitude (*cf.* Fig. 1 for a distance of about 120°.) The signs of these terms are, however, opposite to the corresponding terms for the northern ice sheet, which is consistent with the predictions arising from the simple model (Figs 1 and 2) and is primarily the consequence of the need to conserve mass between the ice and the water. It is interesting to note that the maximum ice-load term occurs after the time of maximum glaciation and that it is out of phase with the rigid term. This arises from the assumed increase in ice volume up to 18 000 yr BP, followed by the immediate melting phase such that the Earth's viscous response to the maximum load is delayed and occurs after the peak in the ice load has occurred. If this is also the case with the northern hemisphere ice sheets, then the time of maximum fall in sea-level will occur a few thousand years after the time of maximum glaciation. The water-load term exhibits the same behaviour for the Antarctic ice model as for the northern ice sheets, except that the amplitudes are decreased in proportion to the eustatic contributions from the two hemispheres. Also, these corrections exhibit the same phase lag as the Antarctic ice-load terms.

Figure 6(e) illustrates the total predicted sea-levels for the two sites that show the greatest spatial variability, Kavalla and Sitia. The results are based on the preliminary eustatic and rebound models and include the ice sheets from both hemispheres. At mainland coastal sites, such as Kavalla, the sea-level curve in lateglacial time (>10 000 yr BP) generally lies above the eustatic curve, whereas at the island sites this relation is reversed. At all sites in the Aegean area the post-glacial sea-levels ($T \lesssim 8000$ yr BP) are characterized by a slow rise, the result of $\Delta\zeta_r$ now being very small ($8000 \lesssim T \lesssim 6000$ yr BP) or zero ($T \lesssim 6000$ yr BP), such that the dominant contributions now are the ice-load term and, to a lesser extent, the water-load term. The predicted sea-levels for the last few thousand years are therefore characterized by a slow rise, and, in the absence of vertical tectonics and a reduction in ocean volume, no levels above the present sea-level are expected to have developed anywhere in the region since the onset of the last glaciation.

The above results are for the earth model with the 'globally representative' parameters. The corrective terms for the alternative, low-viscosity, thin-lithosphere model are compared with the nominal earth-model predictions in Fig. 7. Because of the trade-off between the lithospheric thickness and the upper mantle viscosity frequently encountered in these loading problems, the general pattern of the sea-level change remains essentially unchanged, particularly for the past 6000 yr for which most of the observational evidence occurs (see Figs 7c and f). For Kavalla, the difference in the predictions is a consequence of both the ice- and water-load terms, and is of a magnitude where it should be possible to discriminate between the two models, with good-quality observations going back into lateglacial times. For the island site of Sitia, the differences between the two models for the ice- and water-load terms tend to cancel each other and the total corrections are comparable. However, of greater significance is that the trends of the corrective terms for the two models are very similar and that the past sea-levels at these sites remain below their present level throughout Holocene and latest Pleistocene times.

Figure 8 illustrates the predicted sea-levels relative to the present sea-level across the region, in the form of contours of equal sea-level position at 6000 yr BP. These predictions are based on the above-discussed earth- and ice-model

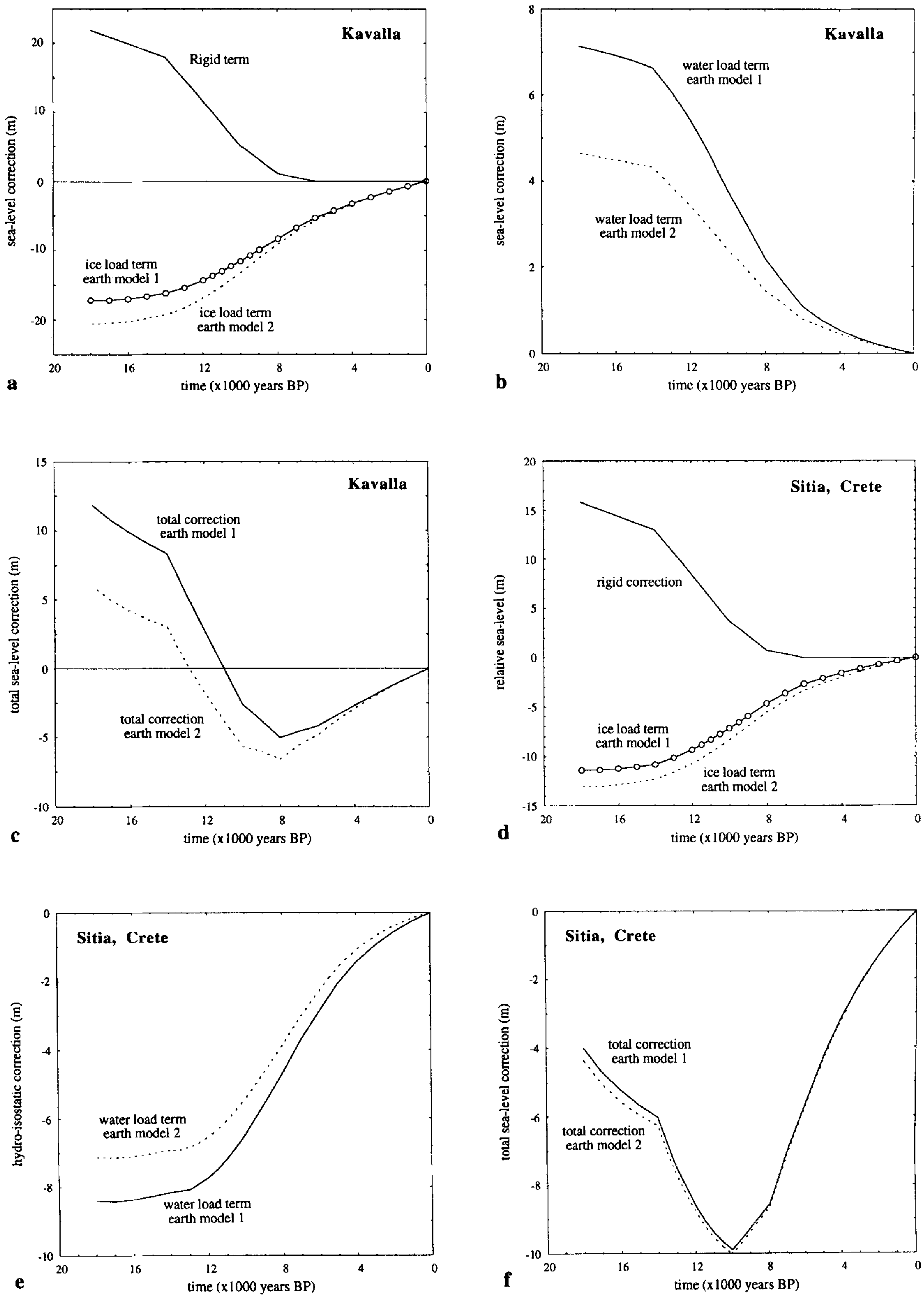
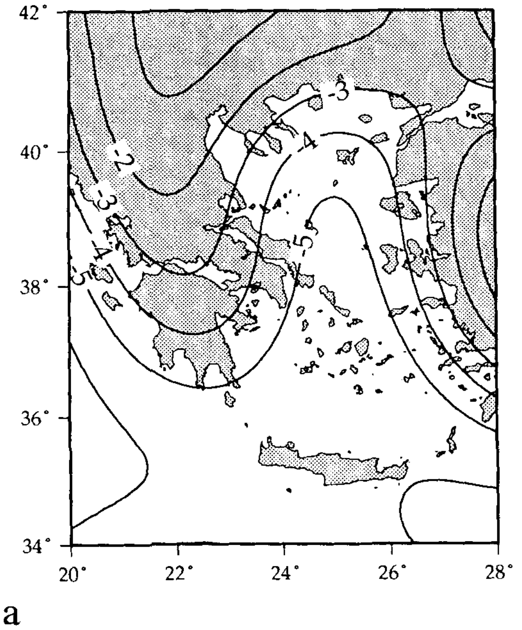


Figure 7. The corrective terms to the equivalent sea-level function for the nominal (earth model 1) and tectonic earth-model parameters (thin lithosphere, low upper mantle viscosity; earth model 2) at Kavalla (Thrace) and Sitia (Crete), for the northern hemisphere ice sheet. (a) The rigid term and the ice-load terms for the two models; (b) the water-load terms; and (c) the total corrective term, at Kavalla. (d) The rigid and ice-load terms; (e) the water-load term; and (f) the total corrective term, for Sitia.



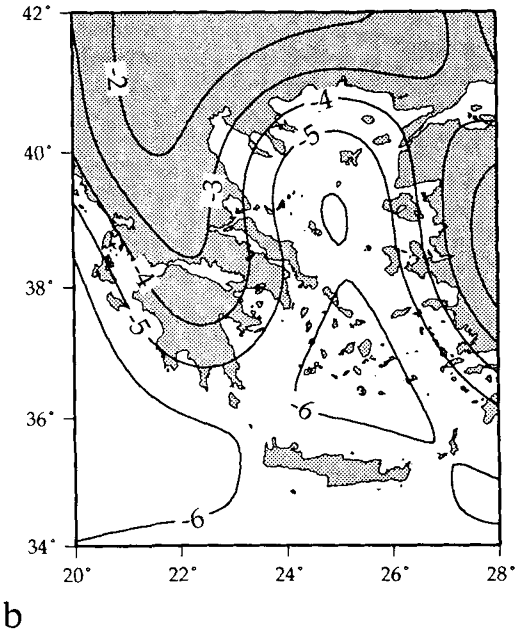


Figure 8. Relative sea-levels predicted in the Aegean region at 6000 yr BP for the first-iteration solution. The contours represent the sea-level position at this time relative to the present sea-level, (a) for the nominal earth-model, and (b) for the tectonic earth-model parameters.

parameters. Fig. 8(a) gives the results for the nominal earth model and Fig. 8(b) gives comparable results for the 'tectonic' earth model. The two results are broadly consistent and at no epoch, particularly during the past 6000 yr, are sea-levels predicted to have risen above their present levels; the region as a whole, in the absence of vertical tectonics, appears to have experienced gradual land subsidence at rates ranging from about -0.6 to $-1.2\,\mathrm{mm\,yr^{-1}}$ for the past 6000 yr for the nominal earth model, and about -0.5 to $-1.4\,\mathrm{mm\,yr^{-1}}$ for the 'tectonic' earth model.

4 OBSERVATIONAL EVIDENCE FOR RELATIVE SEA-LEVEL CHANGE

Evidence for past sea-levels usually comes from a variety of geological indicators, but in the eastern Mediterranean basin, archaeological data sources are particularly important for reconstructing the sea-level record over the past few thousand years. In contrast, the geological evidence for the positions of shorelines since the last glacial maximum is limited in the region and largely confined to Holocene times.

4.1 Archaeological evidence

The archaeological evidence for sea-level change in the eastern Mediterranean has been examined by numerous authors and most comprehensively by Flemming (1972, 1978), particularly for the coast of the Peloponnese and south-western Turkey. The evidence consists of the positions of archaeological structures that bear a specific relation to the position of the sea at the time of their construction. Because the tidal range is typically 20 cm or less, constructions, particularly in sheltered bays and harbours, have been made very close to the water line and their present positions can provide a precise indication of any change in sea-level. Flemming (1972) believes that the principal source of error in such inferences lies in the interpretation of the relationship between the structure and sea-level at the time of construction, and he estimates that this varies from site to site by 0.25 to 1.0 m, depending on the exposure of the site to the waves at the time of construction: the more exposed a site, for example, the higher will be the original working surface of a quay or building foundations. Other sources of error include those associated with the present height measuring process, such as errors arising from waves (~5 cm), tides (~20 cm) and atmospheric pressure variation (~10 cm). More difficult to evaluate is the possible effect of substratum compaction and local subsidence subsequent to construction. The other major source of error comes from the age determination of the archaeological construction. Dating of the period of occupation of a submerged archaeological site usually depends on pottery or inscriptions found at the locality and associated with the structure. The duration of occupation of the site is, however, usually several hundred years and often more than a thousand years, and the age determinations can be quite uncertain. Flemming (1978) gives many of the ages to the nearest 500 yr (i.e. an error of ±250 yr) but, because the average rates of sea-level change for the past few thousand years can be expected to be small throughout the region, the associated uncertainty in height introduced by the age error is generally only of the order of 20 cm. Flemming's estimate of the overall rms error of the observation is about 40 cm, and an upper limit, for the less well-defined structures, would be about 105 cm. Fig. 9 illustrates the location of the sites discussed by Flemming (1978) for south-western Turkey, Crete, the Peloponnese and Rhodes. No information is included north of latitude 39° and the data set does not include any information from the central Aegean Islands, particularly from the relatively stable islands within the Cyclades group.

Archaeological estimates of the past sea-levels have also

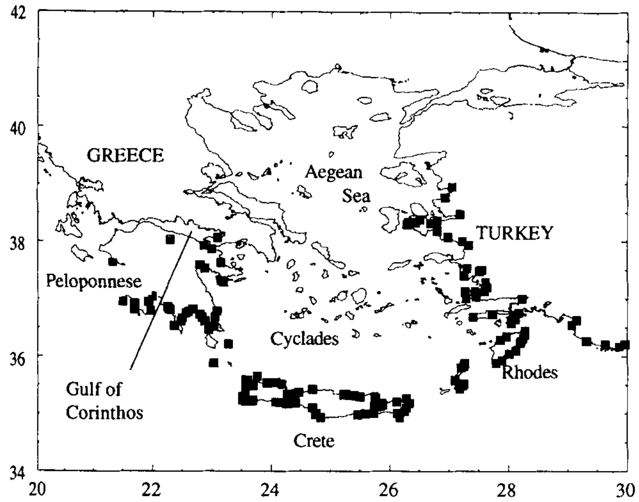


Figure 9. Location map of the sea-level sites in Greece and western Turkey discussed by Flemming (1978), with additional sites around the Gulf of Corinthos.

been published for a number of other sites, including the North Peloponnesian coast (Papageorgiou *et al.* 1993) (included in Fig. 9) and the Cycladean island of Antiparos (Morrison 1968). This latter site is particularly significant because the islands in this group are generally believed to be tectonically stable. Most of the evidence is from Bronze Age or younger sites, but evidence from a few older sites has also been reported, mainly in the form of upper limits to past sea-levels. Offshore from Franchthi Cave, near Kilada in the southern Argolis Peninsula, a Neolithic site points to the sea-level having been at least 11 m lower than present at some time between 7610 ± 150 and 6220 ± 130 yr BP (Jacobsen & Farrand 1987; van Andel 1987) (see Fig. 11a). At Kyra Panagia (or Pélagos) in the northern Sporadhes, relative sea-level was at least 10 m below the present level during 6000–7000 yr BP (Flemming 1983).

Figure 10(a) illustrates the observed height–age relationship of Flemming's (1978) data set for Greece and south-western Turkey. With the exception of a number of data points clustering around 2000 and 5000 yr BP, the majority of the evidence is suggestive of a relative rise in sea-level, that is, of either a rising sea-level relative to the crust or land subsidence. The majority of the sites are from the mainland coat and near-mainland interiors, and Fig. 10(b) illustrates the age–height relationship for these 'mainland' localities. Here the evidence is largely one of a relative sea-level rise. The principal exception occurs for the data points from Rhodes at 5000 yr BP, which correspond to the solution-notch data included in Flemming's (1978) compilation (see Section 4.2). Fig. 10(c) shows comparable results for the offshore island localities of Crete, Antikithira and Karpathos. Here the cluster of uplifted points at 2200 yr BP corresponds to solution-notch data from western Crete.

4.2 Geological evidence

The geological evidence for past shorelines in the Aegean and the surrounding region takes a number of forms, including the depths of submerged terrestrial or lagoonal vegetation and sediments (e.g. Kraft & Rapp 1975),

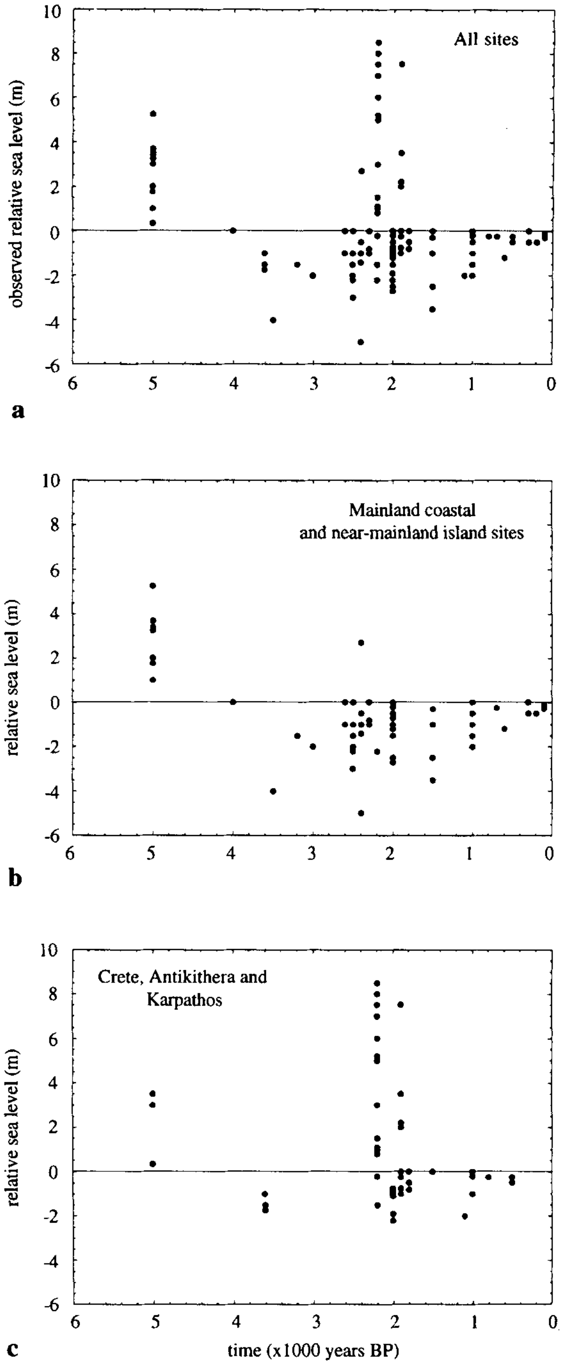


Figure 10. Observed height–age relationships of shoreline features discussed by Flemming (1978). (a) all data; (b) mainland and near-mainland island sites; (c) offshore island sites, including Crete.

inferences drawn from seismic reflectors in shallow offshore sediments (e.g. van Andel & Lianos 1984), and the age–height relation of marine-solution notches (e.g. Flemming 1978; Pirazzoli *et al.* 1994). The problems associated with the interpretation of this evidence are similar to those of the archaeological evidence, and include (i) the need to establish the relation of the particular indicator to sea-level at the time of its formation, (ii) the relation of the submerged or elevated shoreline indicator to present mean sea-level, and (iii) the dating of the sea-level indicators. The last requirement is usually met using radiocarbon methods, the time-scale that is also used to define the advance and retreat of the ice sheets and hence of the eustatic sea-level function. Thus archaeological dates should be reduced to the same reference time frame as the radiocarbon time-scale, or vice versa.

Figure 11(a) illustrates results obtained by Kraft & Rapp (1975) for Messini, a location at the head of the Gulf of Messenia in the Peloponnese (see Fig. 16 for the location). The indicators used here are freshwater swamp grasses whose modern counterparts grow in back swamps up to a few metres above the present sea-level. Hence these observations represent upper limits and the sea-level curve must lie below these data points. The data from this site suggest that, if the swamp grasses are valid indicators of sea-level, a rapid rise of almost 10 m occurred at about 5000 yr BP, followed by a gradual rise to the present value. Somewhat different results have been obtained for other Peloponnese localities. Near Tiryns, on the Argolis plain, Kraft, Aschenbrenner & Rapp (1977) have provided three data points, of which one (A on Fig. 11a), consisting of shallow marine sediments containing molluscs, is a measure of the lower limit of sea-level, and the other two, from swamp deposits, are indicative of upper limits. Together, these data points are indicative of a sea-level curve that lies well above the Messini result between about 5000 and 7000 yr BP. For the plain of Elos, at the head of the Gulf of Laconia, Kraft *et al.* (1977) identified two sea-level indicators, representing shallow marine (point B) and backswamp sediments, with the data point at about 5000 yr BP being comparable to the Tiryns results. A more complete record has been obtained for the bay of Navarine near Pilos, where the record includes a number of shallow marine sediments which are indicative of a lower limit to the sea-level, and submerged fluvial floodplain sediments which provide upper limits (Kraft, Rapp & Aschenbrenner 1980). However, even here the constraints imposed on the sea-level curve are not stringent, and for the southern Peloponnese region as a whole the available data points are few, so that, even in the absence of vertical tectonics, the data would be insufficient for constructing a reliable regional sea-level curve.

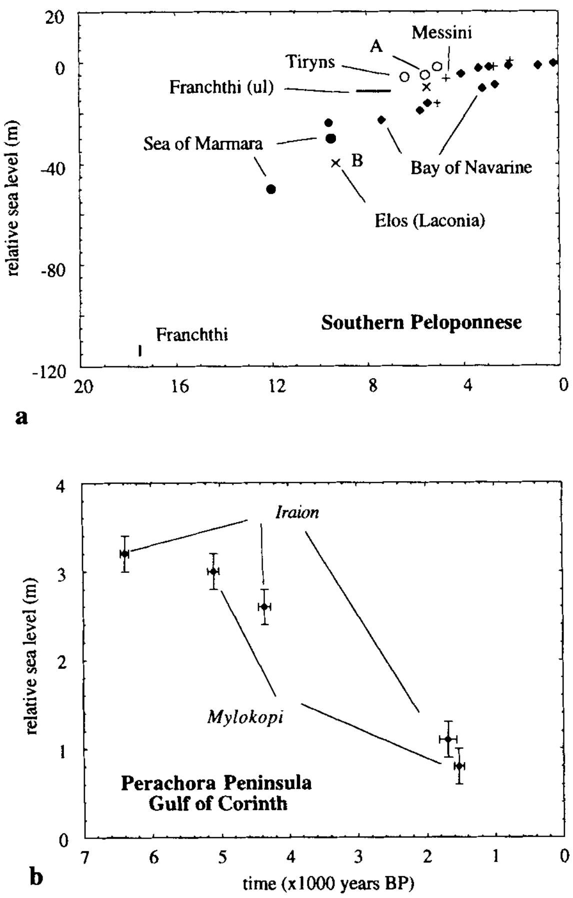


Figure 11. (a) Depth–age relationships of sea-level indicators for Messini (Kraft & Rapp 1975), Tiryns and Elos (Kraft *et al.* 1977), Navarine Bay (Kraft *et al.* 1980), and data from the Franchthi area in Argolis (van Andel & Lianos 1984); 'ul' refers to upper limit estimates; see Fig. 16 for locations). (b) Height–age relationships at Iraion and Mylokopi on the Perachora Peninsula (see Fig. 16 for locations).

An estimate of the sea-level position during the last glacial maximum has been made by van Andel & Lianos (1984) for a location in the southern Argolis Peninsula near Franchthi Cave (Fig. 11a). From sub-bottom profiler records they identified a well-defined basal reflector that changes to a fan-shaped wedge of discontinuous reflectors at a depth of 115–118 m, which is indicative of marine conditions. No direct age determination of this feature has been possible, but they attribute it to the time of the last glacial maximum, partly because the level is comparable to other estimates of water depths at the time of the last glacial maximum, and partly because it has been possible to trace the basal reflector to a Late Pleistocene surface on the coastal plain.

Other important sources of information are the marine-solution notches found in limestone cliffs at numerous localities, in, for example, the Gulf of Corinth (Pirazzoli *et al.* 1994), Rhodes and Karpathos (Flemming 1978), Crete (Thommeret *et al.* 1981), and Evvoia (Stiros *et al.* 1992). These notches are most evident in uplifted sections of coast, but at some localities they occur below the present sea-level, such as on Karpathos and Evvoia. The marine borers (*Lithophaga*) grow up to mean sea-level and their cumulative action over a period of time produces a notch within the tidal zone. In a slowly uplifting environment, the notch progressively widens as the borers move down with falling sea-level. In contrast, the preservation of well-defined raised fossil notches is indicative of rapid and episodic uplift by amplitudes that exceed the tidal range (Pirazzoli *et al.* 1982). The present elevation of such a notch above the present formation height is, however, the cumulative effect of the initial uplift event and any subsequent episodic and slow vertical movements. In areas where the notches can be followed over considerable horizontal distances along the coast, Flemming (1978) estimated their ages by correlating them with archaeological remains at the same height that were initially at sea-level. These notches, particularly in Crete and Rhodes, have provided continuity of evidence between otherwise widely spaced archaeological sites. Such data points have been included in Flemming's (1978) results for the East Mediterranean sea-level data and are included in the site location map (Fig. 9).

Marine notch data have also been presented by Pirazzoli *et al.* (1994) for the Perachora Peninsula in the Gulf of

Corinth. Here the age–height relationships of several notches at different heights have been determined by radiocarbon-dating of fossil marine borers and barnacles (Fig. 11b). The preservation of the distinct notches requires that uplift occurs in discrete steps, but the ensemble of the data indicates a more-or-less uniform uplift when averaged over about 7000 yr. Other solution-notch observations have been published for the northern coast of the Peloponnese (Papageorgiou *et al.* 1993) and for several sites on the island of Evvoia (Stiros *et al.* 1992).

5 A COMPARISON OF OBSERVATIONS WITH PREDICTIONS FOR SEA-LEVEL CHANGE

In comparing the predicted and observed sea-levels, a more comprehensive model has been used, which includes (i) the earlier loading cycles of both the northern hemisphere and Antarctic ice sheets, (ii) the second-iteration water-load corrections, including the effect of the time dependence of the shorelines, and (iii) a small correction to the eustatic sea-level in which the ocean volume increased after 6000 yr BP by a small amount, primarily between about 6000 and 4000 yr BP, consistent with the results from studies in other parts of the globe (e.g. Lambeck 1993b). The results illustrated in Figs 12 to 18 are, unless otherwise indicated, for the nominal earth model. They are, however, also representative of the results obtained for the tectonic earth model, although the estimated tectonic rates, but not the trends and general conclusions, may differ for the two models, the tectonic models leading to tectonic uplift rates that are about 5 per cent greater.

The predicted sea-levels for all the observation locations and times discussed above are illustrated in Fig. 12 for the interval 0–7000 yr BP. The predictions all lie well below the eustatic sea-level function used in this model, illustrating that, in the absence of tectonics, the region as a whole is one of relative submergence because of the glacio-isostatic processes. This result is essentially unchanged for the tectonic earth model (Fig. 12b), as well as for models based on the nominal eustatic sea-level function in which no ice melting occurs after 6000 yr, and no sea-levels are predicted to have occurred above present sea-level at any of the sites (Fig. 12c). These predictions emphasize the earlier conclusions that (i) a single sea-level curve should not be constructed from the observations for the region, even in the absence of tectonics, and (ii) if such a curve is constructed, it does not provide a reliable estimate of the eustatic sea-level, even when the oscillations or irregularities produced by the spatial variability are smoothed out.

The relation between the observed and predicted sea-levels is illustrated in Fig. 13. If the models and data were both perfect, these points would fall on a straight line of unit gradient (the line AA′ in Fig. 13). Significant clustering about this line occurs, as do significant departures from it, departures that exceed the estimates of observational accuracy as well as any limitations of the isostatic model. Nominal uncertainties (including observational and model errors) from this linear relation are illustrated by the dashed lines. Points outside of and above these limits indicate sites where tectonic uplift has been predominant, and two clusters of points refer mainly to western Crete and

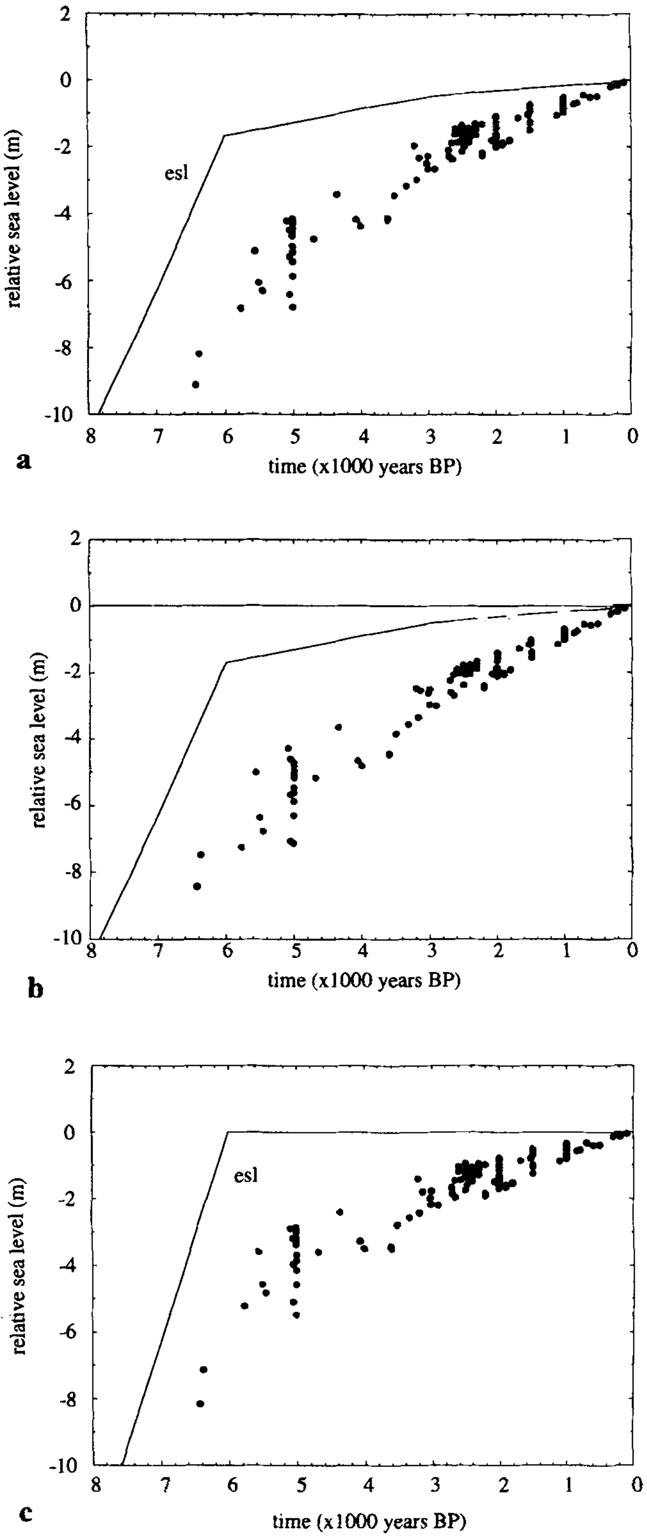


Figure 12. Predicted sea-levels for zero tectonic motion at all locations and epochs from 0 to 7000 yr BP of the archaeological and geological observations discussed in the text. Predictions in (a) and (b) are based on the complete glacio-hydro-isostatic theory and on the eustatic sea-level curve illustrated in Fig. 3(b). Predictions in (c) are based on the nominal eustatic model in which all melting ceased at 6000 yr BP. Predictions in (a) and (c) are for the nominal earth model and those in (b) are for the tectonic earth model.

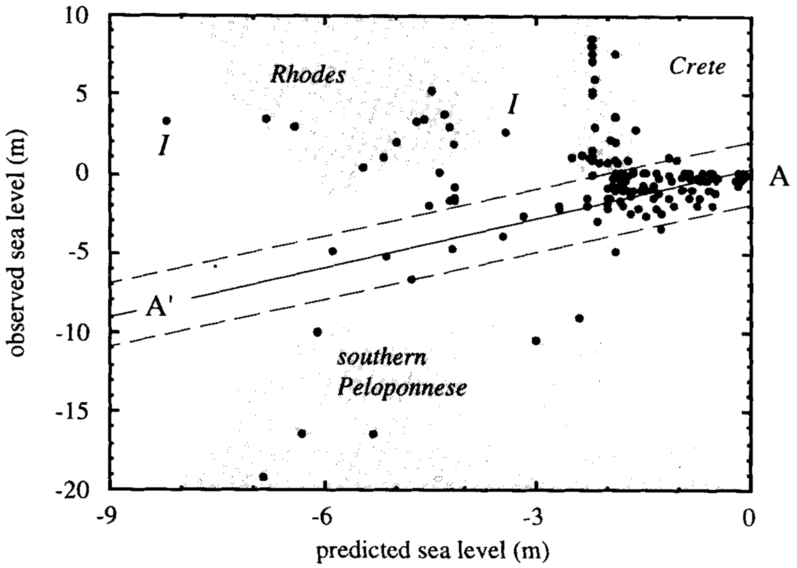


Figure 13. Observed versus predicted sea-levels for the times and locations of the observed values for the past 7000 yr. In the absence of tectonic movements and observational and modelling errors, all points would fall on the line AA'. Points below this line are indicative of tectonic subsidence and points above it are indicative of uplift. Points of significant vertical tectonic movement lie outside the zones between the two dashed lines parallel to AA'. The shaded areas show the principal localities of sites at which the rates of vertical tectonic movement are significant. The points denoted by *I* refer to the Iraion site on the Perachora Peninsula.

Rhodes. Points below the limits, fewer in number and occurring mostly at southern Peloponnese sites, are indicative of subsidence.

Average tectonic rates $\dot{\Delta}\zeta_T(t)$ for a locality are defined as

$$\dot{\Delta}\zeta_T(t) = \{\Delta\zeta_o(t) - [\dot{\Delta}\zeta_e(t) + \Delta\zeta_I(t)]\}/t, \tag{6}$$

and the adopted corresponding accuracy estimates are

$$\sigma_{\dot{\Delta}\zeta}(t) = \sigma_{\Delta\zeta}(t)/t \approx 0.40/t \text{ m}, \tag{7}$$

based on the above estimates of the archaeological evidence. Data points for which

$$\dot{\Delta}\zeta_T(t)/\sigma_{\dot{\Delta}\zeta}(t) \geq 3 \tag{8}$$

are considered to provide statistically significant estimates of tectonic uplift or subsidence, particularly when a number of data points from nearby locations give similar results. This latter criterion can, however, be misleading, as some of the observations, particularly the solution-notch information from the same general region, may not provide independent estimates of the tectonic rates.

Figure 14 illustrates the estimates of the uplift rates for Crete, where the observational data base extends back to about 2500 yr BP with a few isolated older points. (For sites where observations from different epochs are available, the estimates of the tectonic rates are based on a weighted least-squares adjustment of the individual isostatically corrected relative sea-level estimates, with the requirement that the linear trend with time pass through the origin to within ±0.2 m. The estimate of the accuracy of this rate is the weighted least-squares variance scaled by the variance of unit weight.) Most of the island is subject to tectonic uplift and the estimated rates are significant at the 3 sd (standard

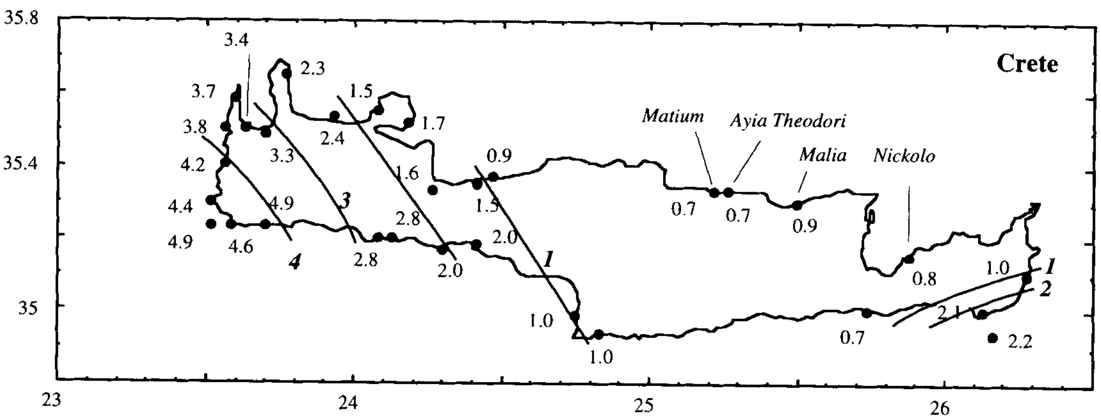


Figure 14. Estimates of the tectonic rates of vertical movements for Crete, for the Flemming (1978) sites at which the uplift rates are statistically significant. Also given are contours of constant vertical tectonic rates. Units are mm yr^{-1}. The figure is based on the nominal earth model and the eustatic sea-level function of Fig. 3(b).

deviation) or higher level for most of the western part, where the rates approach 5 mm yr^{-1} in the south-west corner. For the south-east corner of Crete, significant uplift occurs at rates of up to 2 mm yr^{-1}. Very similar results are obtained for the tectonic earth model parameters, with rates that are about 5 per cent greater than for the nominal earth model. The results do not change significantly if the nominal eustatic sea-level function is adopted; uplift is predicted throughout the island, although amplitudes are reduced by about 10–15 per cent. Much of the north coast is also subject to tectonic uplift, at rates that are significant according to the above criterion, even though the archaeological indicators at some of the sites now lie below sea-level. At such sites, such as Matium, Ayia Theodori, Malia and Nickolo, the tectonic rates are smaller than the isostatic rates of opposite sign, with the net result of a rising sea-level relative to the land. Multiple observations at different epochs at the same site are available only near Nickolo, where two estimates of uplift point to a more-or-less constant uplift rate over the past 3500 yr (Fig. 15a).

Similar results for the Peloponnese are illustrated in Fig. 16. Tectonic uplift is significant at the 3 sd level at Mavra Litharia, on the southern shore of the Gulf of Corinthos, as well as at Iraion and Mylokopi on the nearby Perachora Peninsula at rates of about 1.5 ± 0.2 mm yr^{-1}. At Corinthos itself, the estimated rate of uplift is small, 0.3 ± 0.3 mm yr^{-1}, and statistically insignificant. Further south, the evidence from the Argolis Peninsula points primarily to subsidence, with the estimates from Porto Cheli, nearby Lorenzon and Palia Epidaurus being significant at the 2.5 sd or higher level. Fig. 15(b) illustrates the comparison of the model predictions with the observations of Kraft *et al.* (1977) for Tiryns, at the head of the Argolis Gulf. Here, the two are in general agreement, with the alluvial flood plain and swamp deposits lying above the predicted sea-level curve; no significant vertical tectonic displacement appears to be required here.

The only other areas where the estimated uplift rises significantly above the noise level is for the islands of Kythera and Antikythera, and for the southern limits of the peninsulas flanking the Lakonian Gulf (at the site of Tigani, Arkangelos and Neapolis, with the uplift estimates being significant at the 2 sd or greater level). In contrast, the northern shorelines of the Lakonian Gulf appear to be subject to subsidence. Thus the estimates for Skoutari, Gythion and Plitra, are significant at the 2 sd or greater level, but elsewhere the subsidence rates are generally small and statistically insignificant. For example, as illustrated in Fig. 15(c), the predictions for the Lakonian Plain are broadly consistent with the two geological observations of Kraft *et al.* (1977). Likewise, for the northern part of the Messina Gulf area, the estimates for the vertical movements point to crustal subsidence, but nowhere are the rates for the individual sites statistically significant. Fig. 15(d) illustrates the results for the Messinia Plain from Kraft & Rapp (1975). The observational evidence here is from backswamps that formed a few metres above sea-level, so that the observations provide an upper limit to the sea-level curve. Of the four data points, only the oldest is significantly different from the model predictions and, if correct, it points to a rapid relative subsidence between about 5000 and 4500 yr BP. With the exception of this data point, the observational evidence is consistent with an absence of tectonic movements for the region.

To the west, at Navarine Bay, the shallow marine deposits of H.E. Wright (quoted in Kraft *et al.* 1980) are consistent with the model predictions (Fig. 15e). The shallow marine sediments reported by Kraft *et al.* for the same locality lie below the predicted sea-level curve with the exception of the data point at about 9500 yr BP, which lies at a much shallower depth than predicted. Kraft *et al.* discuss the possibility that their core samples may have been contaminated by older materials, but they do not consider this to be a major factor in this case, and, if this observation is indeed indicative of the relative sea-level position, it would require a very substantial ($\gtrsim 25$ m) tectonic uplift in early Holocene times, followed by tectonic stability for about the past 4000 yr.

Taken together, the southern Peloponnese observations are consistent with the glacio-hydro-isostatic model and, as for the Crete results, the results are largely independent of whether the eustatic sea-level corrective term is included (Fig. 15). The predicted time series for these locations are all quite similar (Fig. 15f), with a rapid rise from about −40 m at 9000 yr BP to about −6 m below the present level at 6000 yr BP. The major discrepancy for the 9000 yr BP data point from the Bay of Navarine is not seen for the similarly aged data from the Lakonian plain, and either this observation is erroneous or major tectonic uplift of about 15 m occurred between about 9500 and 7500 yr BP. The other major discrepancy between observations and predictions, the approximately 5000 yr BP data point from the Messinia plain, is not seen at the other sites for comparable times. Both discrepancies, if real, appear to have been isolated events, without the occurrence of significant subsequent tectonic movements.

An important observation that confirms some of the above conclusions is the position of the Tyrrhenian or Last Interglacial shoreline at some of these localities. At the heads of the Messinia, Lakonia and Argolid Gulfs these shorelines are found to occur at only a few metres above the present sea-level (Kelletat *et al.* 1976; van Andel 1987), where they are expected, if ocean volumes during the Last Interglacial were about the same as today (Lambeck & Nakada 1992). Subsidence rates of less than 0.1 mm yr^{-1} would have placed these shorelines well below the present sea-level, so that their occurrences at 1–5 m elevation points to a long-term stability of these sites. Along the southern shore of the Gulf of Corinthos, well-elevated Tyrrhenian (and older) shorelines and marine sediments have been identified, with maximum heights occurring in the central area near Mavra Litharia (Stiros 1988). According to Keraudren & Sorel (1987), such sediments occur at between 150 and 170 m to the east of Mavra Litharia and at about 30 m near Corinth, suggesting average long-term uplift rates of between 1.25 and 1.42 mm yr^{-1} at the former site and 0.25 mm yr^{-1} near Corinth, values that are wholly consistent with the isostatically adjusted Late Holocene rates for the same localities. On the Perachora Peninsula, however, the Tyrrhenian shoreline has been reported to occur at only about 30 m (Pirazzoli *et al.* 1994), suggesting that average rates on time scales of 10^5 yr are about 0.25 mm yr^{-1} compared with rates of about 1.5 mm yr^{-1} inferred for the Late Holocene. Perachora Peninsula is significantly faulted

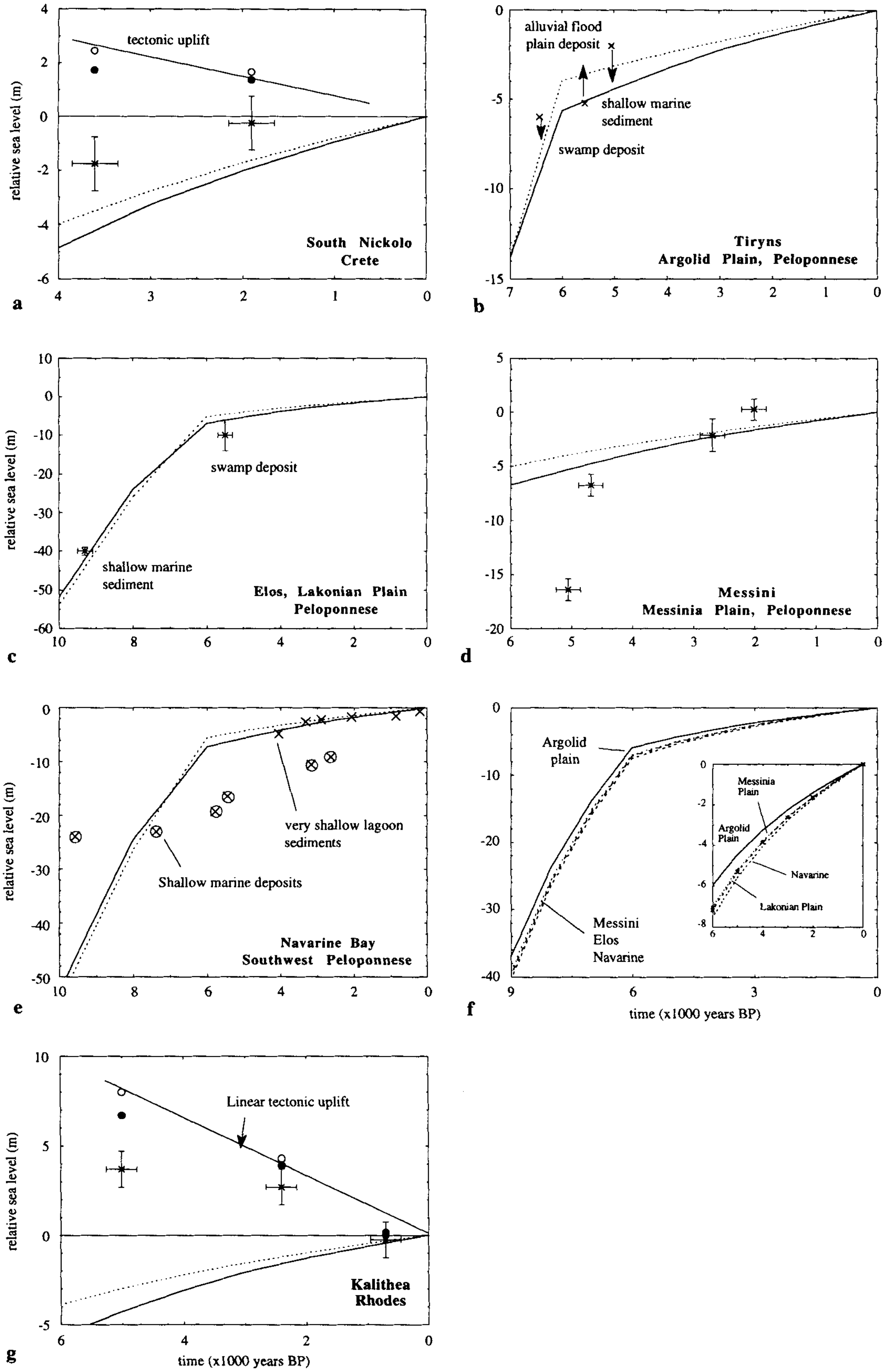


Figure 15. Comparison of observed (individual points, with or without error bars) and predicted sea-levels at different localities in Crete (a), the Peloponnese (b)–(e), and Rhodes (g). The predictions are for the nominal earth model, with (solid line) and without (dashed line) the eustatic corrective term. For the Crete and Rhodes sites, the predicted tectonic contributions to relative sea-level are also shown (open circles). Open and solid circles give the estimates for the model with and without the eustatic correction, respectively. In (f) the predicted total sea-level changes for the four Peloponnese sites are compared.

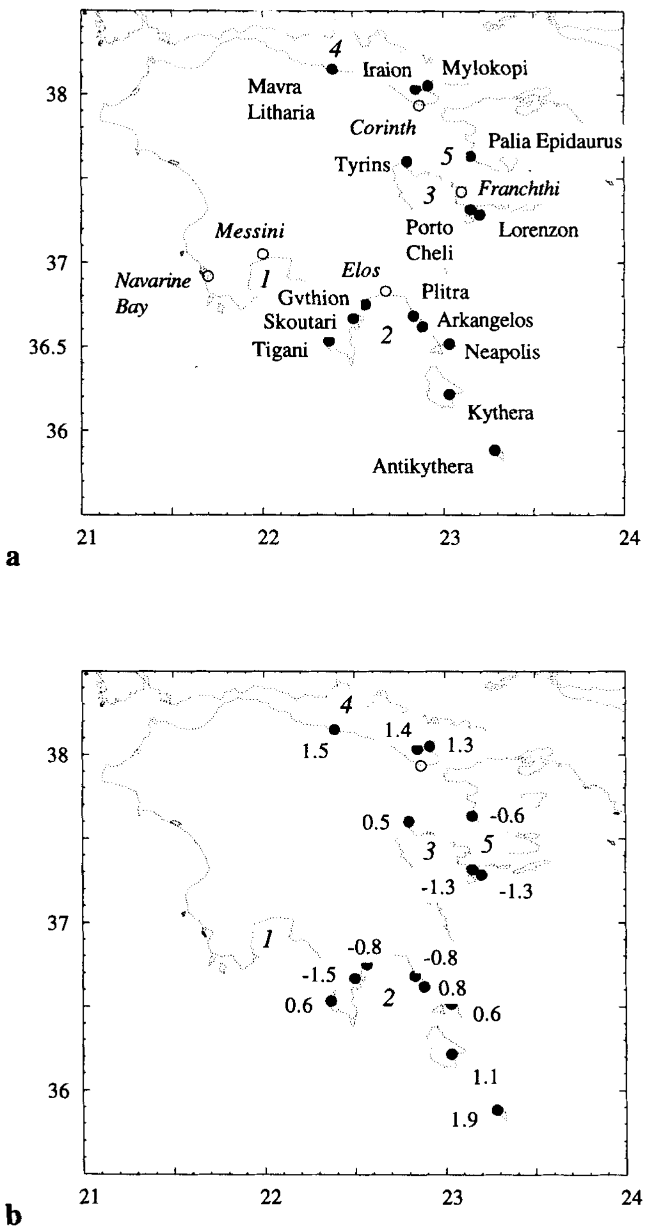


Figure 16. (a) Location of sea-level sites in the Peloponnese, and (b) estimates of the tectonic rates of vertical motion at sites where these rates are statistically significant (in mm yr^{-1}). 1: Gulf of Messinia; 2: Lakonian Gulf; 3: Argolis Gulf; 4: Gulf of Corinthos; 5: Argolis Peninsula.

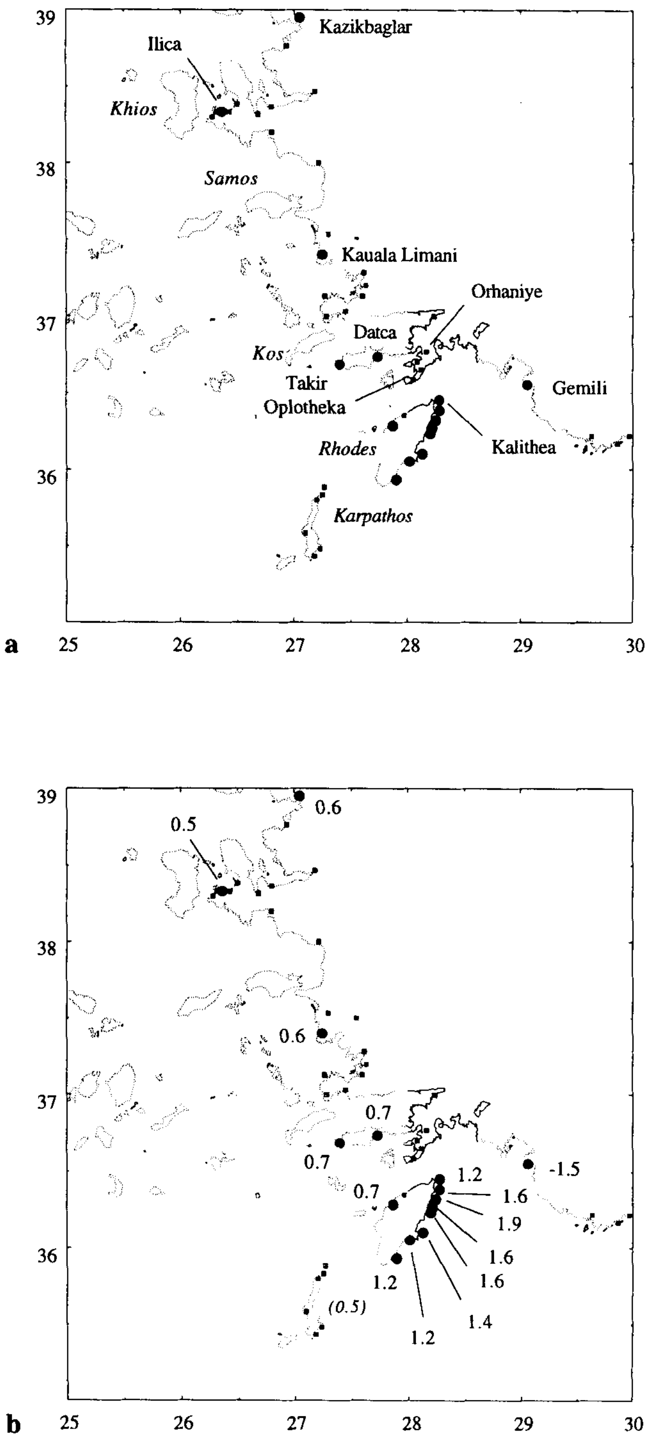


Figure 17. As Fig. 16, but for the Turkish coast. (a) gives the locations of all sites discussed by Flemming (1978). (b) gives the estimates of tectonic uplift or subsidence at those sites where the values are believed to be statistically significant. The value for Karpathos, in parentheses in (b), refers to the mean uplift rate of the island sites.

and different uplift rates for Late Holocene times have been noted by Pirazzoli *et al.* (1994) for sites separated by only short distances, and part of this discrepancy may be a consequence of such spatial variability in the tectonic uplift rates. Further work is warranted.

Of the Cycladian Islands, the sea-level at Antiparos must have occurred below -5 m at about 5000 yr BP (Morrison 1968), consistent both with the isostatic predictions and with an absence of tectonic movement. For Rhodes the situation is different. Here the tectonic trend is one of uplift, at an average rate of about 1.5 mm yr^{-1} along the entire east coast (Fig. 17). This estimate is dominated by the raised solution notch dated at 5000 yr BP, and the individual estimates are strongly correlated. However, the estimate is supported by an archaeological estimate for a different epoch at Kalithea, and uplift appears to have been uniform on a time-scale of a few thousand years (Fig. 15g). To the south of Rhodes, the island of Karpathos is also subject to uplift. The individual estimates themselves are not statistically significant according to the above criterion (eq. 8), but the values are consistent for all sites and the average rate is about 0.5 mm yr^{-1}. Again, the estimates are based on the solution-notch data and depend strongly on the adopted age of formation of this feature. Immediately to the north of Rhodes, the data from four sites from Oplotheka to Orphaniye do not indicate a continuation of the uplift pattern established for Rhodes. Instead, the individual estimates are indicative of both uplift and subsidence occurring at locations separated by distances of less than a few tens of kilometres, with neither the uplift rates nor their weighted mean being statistically significant. To the west, data from the sites of Datca and Takir are indicative of uplift. Elsewhere along the Aegean Coast of Turkey there are few individual estimates of uplift rates that are statistically significant, although the general pattern is one of uplift. Between Kos and Samos, for example, there is a persistent uplift trend with an average value of about 0.3 mm yr^{-1}, but even this average value is barely significant. Likewise, the peninsula between Izmir and Khios appears to have been uplifted at a comparable rate, although three imprecise estimates for recent rates (sites dated at 100 and 300 yrs BP) point to subsidence.

6 CONCLUSION

Glacio-hydro-isostatic factors make an important contribution to the spatial and temporal variations in Holocene sea-level in the seas around Greece and western Turkey. This includes the response of the sea to the gravitational attraction of the ice sheet at glacial maximum and during lateglacial times, and the adjustment of the crust to both the removal of the last great ice sheet over northern Europe and the increased water load in the Mediterranean basin resulting from the melting of the global ice sheets. Less important, but not insignificant, are the crustal adjustments to the melting of the more distant ice sheets. The combined effect is a complex pattern of isostatic sea-level change, with magnitudes that are important when compared to changes in relative sea–land positions produced by vertical tectonic processes.

The theory for glacio-hydro-isostasy is sufficiently well advanced to enable records of sea-level change in tectonically active areas to be corrected for these effects. The overall pattern of the predicted isostatic contribution for the region is one that leads to a rising sea-level throughout Holocene times, even if melting of the great ice sheets had ceased by 6000 yr BP. This is in contrast to the characteristic isostatic signal of a gradually falling sea-level over the last 5000 or 6000 yr for locations much further away from the former ice sheets. The actual rates of the glacio-hydro-isostatic changes are earth-model-dependent. There is some evidence that the effective parameters describing the lithospheric thickness and mantle viscosity may be geographically variable (Nakada & Lambeck 1991), and the values of these parameters may be substantially less than the corresponding values for more stable regions. However, largely because of the fortuitous nature of the trade-off between earth-model parameters, in particular lithospheric thickness and upper mantle viscosity, thin lithosphere and low-viscosity models yield comparable results (e.g. Fig. 8). The overall pattern remains unchanged, and for Greece and the Aegean coast of Turkey, sea-levels, in the absence of tectonics, would exhibit a slow rise or a coastal submergence at rates of the order 1 mm yr^{-1} for at least the past 6000 yr. This rate is comparable to the tectonic rates, and a separation of the two factors is important if reliable estimates for the latter are sought. If the average tectonic rates can be independently assessed from the comparison of Holocene and Last Interglacial shorelines in the region, then the more recent data can be used to estimate the effective parameters for the region, but the currently available data are largely inadequate for this.

Although the emphasis of this paper has been on demonstrating the importance of the isostatic factors, rather than on a critical evaluation of the observational evidence or of tectonic models, the results do point to a need to modify some of the published interpretations. The observed submergence of large sections of the Greek and western Turkey coastlines is not, for example, necessarily evidence for coastal subsidence by tectonic processes, but may instead be largely the result of the isostatic process. Thus the vertical subsidence inferred by Flemming (1978) for sites at the heads of the Lakonian and Messinian Gulfs and for the Navarine Bay of the southern Peloponnese is interpreted here as a consequence of the glacio-hydro-isostatic adjustment during Late Pleistocene and Holocene times, and the observed trends are not representative of trends on much longer time-scales. This is generally consistent with the occurrence in these localities of the Last Interglacial Tyrrhenian Shoreline at a few metres above present sea-level (Kelletat *et al.* 1976; van Andel 1987), for if the observed relative sea-level change in Late Holocene time is of tectonic origin, then shorelines formed at the time of the Last Interglacial would now be at depths of 100 m and more below present sea-levels unless the tectonic subsidence rates were temporally variable. When the observed rates of vertical movements have been corrected for the isostatic factors, significant tectonic subsidence appears to be restricted to only a few localities in the Peloponnese, the southern end of the Argolid Peninsula, and the eastern and western sides of the northern part of the Lakonian Gulf (Fig. 16). Elsewhere, observed estimates for tectonic subsidence are small and statistically insignificant, although the preservation and observation of evidence for sea-level change may well bias the observational record to tectonic uplift.

In regions where the tectonic and isostatic rates are of comparable magnitude, the interpretations of the observational evidence can be quite different if the latter component is neglected. For example, in Crete it has been suggested that the central area of the north coast has been subsiding, in contrast to the substantial uplift elsewhere on the island (Flemming 1978). However, once corrected for the isostatic factors, the data actually point to tectonic uplift here, albeit at a reduced rate compared to western and south-eastern Crete (Fig. 14), and the evidence points to tectonic uplift occurring along an arc from Rhodes and Karpathos in the east, along the length of Crete, to Kithera and Antikythera, and to the southern Peloponnese (Fig. 18).

At Mavra Litharia in the northern Peloponnese, the evidence points to vertical uplift rates of about 1.5 mm yr^{-1},

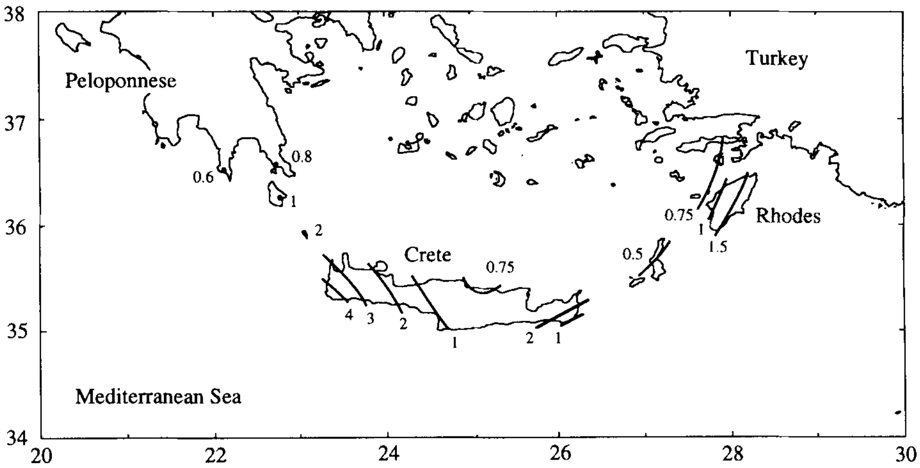


Figure 18. Uplift rates (mm yr^{-1}) for the southern Peloponnese, Crete, Karpathos and Rhodes for the nominal earth model.

but considerable spatial variability appears to occur. For Corinth, for example, subsidence has been reported (Flemming 1978), but when corrected for the isostatic factors this becomes slow uplift at a rate of about 0.3 mm yr^{-1}. The Late Holocene record here is not sufficiently accurate to produce a statistically significant estimate, but the rate is consistent with longer term rates inferred from the position of the nearby Tyrrhenian shoreline at about 30 m elevation. Likewise the rate inferred for Mavra Litharia is consistent with the longer term rates inferred from the Tyrrhenian and older shorelines (Keraudren & Sorel 1987; Stiros 1988). The one area where this agreement between Late Holocene and Last Interglacial evidence appears to be inconsistent is the Perachora Peninsula near the eastern end of the Gulf of Corinth. Here, the rates for the last few thousand years, estimated from solution-notch data, are significantly higher than those inferred from the position of the Tyrrhenian shoreline in the region (e.g. Pirazzoli *et al.* 1994), and, if the evidence has been correctly interpreted, then either the vertical tectonic movements are not uniform in time or the discrepancy is a consequence of considerable spatial variability in tectonic uplift.

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