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Holocene Eustatic Changes and Coastal Tectonics in the Northeast Mediterranean: Implications for Models of Crustal Consumption

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HOLOCENE EUSTATIC CHANGES AND COASTAL TECTONICS IN THE NORTHEAST MEDITERRANEAN: IMPLICATIONS FOR MODELS OF CRUSTAL CONSUMPTION

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The African and Eurasian plates are converging at 2.6 cm/a and the Turkey plate is moving west at about 3.0 cm/a. The Aegean is thus a zone of two-dimensional convergence. The boundaries of plates and sub-plates in this area cannot be defined at all simply. Observations of relative changes of sea level were made at 202 ancient harbour sites, some dating back to 2000 B.C., in Greece, Turkey, and Cyprus, and 175 independent estimates of sea level change were made. Measurements of uplifted and submerged recent marine solution notches in Antikythera, Crete, Karpathos, Rhodes, and Cyprus, showed that the islands of the Hellenic Arc are subject to intermittent seismic movements with a periodicity ranging from 157 to 714 a. The archaeological data and the notch data are combined to show that blocks with typical horizontal dimensions of 100 × 50 km at sea level are tilting intermittently as rigid units. The rates of tilt vary from 3.6 to 10.6″/ka.

Stable blocks were identified in south Turkey east of Antalya, north Cyprus, and southwest Turkey from Kusadasi to Marmaris. The active areas are: the Peloponnese, where the peninsulae of Argolis and Messenia are subsiding; the central Peloponnese is probably doming or folding, while a slightly uplifted ridge extends east of south through Kythera and Antikythera; a graben structure separates Antikythera from Crete; west Crete is tilting northeast at a rate of 10.6″/ka; eastern Crete is tilting northwards at 10.3″/ka; Karpathos has subsided about 1.0 m maintaining horizontality; Rhodes is folding about a plunging anticlinal axis trending approximately east–west; in Turkey the Izmir-Cesme peninsula is subsiding, as is the coast from Fethiye to Cape Gelidonya; the south coast of Cyprus is subsiding irregularly.

These data are used to construct a model of the subduction zone beneath the Hellenic Arc. It is concluded that there is no single planar or slightly curved subducted slab, since the zone bends through 90° south of Crete and again near the Anaximander Mountains. The quasi-linear sections of the subduction zone are interrupted by short tear faults and abrupt changes of strike at intervals of 50–100 km along their length. The resulting narrow fingers of slab tend to break, and are not subducted to great depth.

Combination of this model with geophysical data allows construction of a new system of probable plate boundaries. It is suggested that there is no Aegean plate. The north Aegean is an extension of the Turkish plate, while the south Aegean is a region of crustal extension produced by gradual break-up of the southern margin of the Greek–Apulian plate. It is concluded that the Hellenic Arc is not an island arc in the normal sense, but represents a late phase in the destruction of a true island arc and back-arc sea by incipient continent–continent collision.

Introduction

Tectonic models of the origin and evolution of the Central and Eastern Mediterranean have been based on recent seismicity recorded over the last 50 years (Galanopoulos 1967, 1972, 1973; Caputo, Panza & Postpischl 1970, 1972; McKenzie 1970, 1972); reflexion profiling (Ryan 1969; Rabinowitz & Ryan 1970; Sancho *et al.* 1973; Lort & Gray 1974; Wong & Zarudzki 1969); gravity measurements (Harrison 1955; Allan *et al.* 1964; Rabinowitz & Ryan 1970; Woodside & Bowin 1970; Weigel 1973); fault plane solutions (Papazachos & Delibasis 1969; McKenzie 1970, 1972); geological structure of land masses (Aubouin 1965; Aubouin & Dercourt 1965, 1970; Dercourt 1964; Brunn *et al.* 1971); and combinations of specific plate tectonic models of the movements of Africa and Eurasia (Smith 1971; Dewey, Pitman, Ryan & Bonnin 1973).

The northeast Mediterranean, comprising Greece, Turkey, the Aegean Sea, the Hellenic Arc, and Cyprus (figure 1) is the most seismically active area of Europe, although the seismic energy released is low in relation to the plate tectonic boundaries proposed, and to normal island arcs (North 1974). The composition of the crust on the sea floor in the region shows an intermediate nature between oceanic and continental (Woodside & Bowin 1970), while ophiolite series occur uplifted in Cyprus (Gass & Masson-Smith 1963; Vine, Poster & Gass 1973; Miyashiro 1973), southwest Turkey (Brunn *et al.* 1971), and central and northern Greece (Smith 1971). Several large fault systems which are accepted as parts of the boundaries between the African and Eurasian plates approach or bound the margins of the northeast Mediterranean. Specifically the Azores–Gibraltar (Gloria Fault) (Laughton & Whitmarsh 1974) approaches from the west, the Anatolian Fault (Ambraseys 1970, 1975) bounds the northeast margin, while the Dead Sea Rift bounds the eastern margin. The mechanism and structural pattern by which the relative motions along these faults is taken up between 20° and 36° E is not at all clear. Numerous different patterns have been suggested as representing present or defunct plate boundaries in the area (Dewey & Bird 1970; McKenzie 1972; Dewey *et al.* 1973; Galanopoulos 1973, p. 103; Vilminot & Robert 1974, p. 2099), while most authors freely concede that the nature of the area being a buffer zone between major plates is not susceptible to description in terms of a simple model. Particular problems are the westward continuation of the movement of the Anatolian Fault, the description of the Hellenic Arc in terms of the structure of a classic

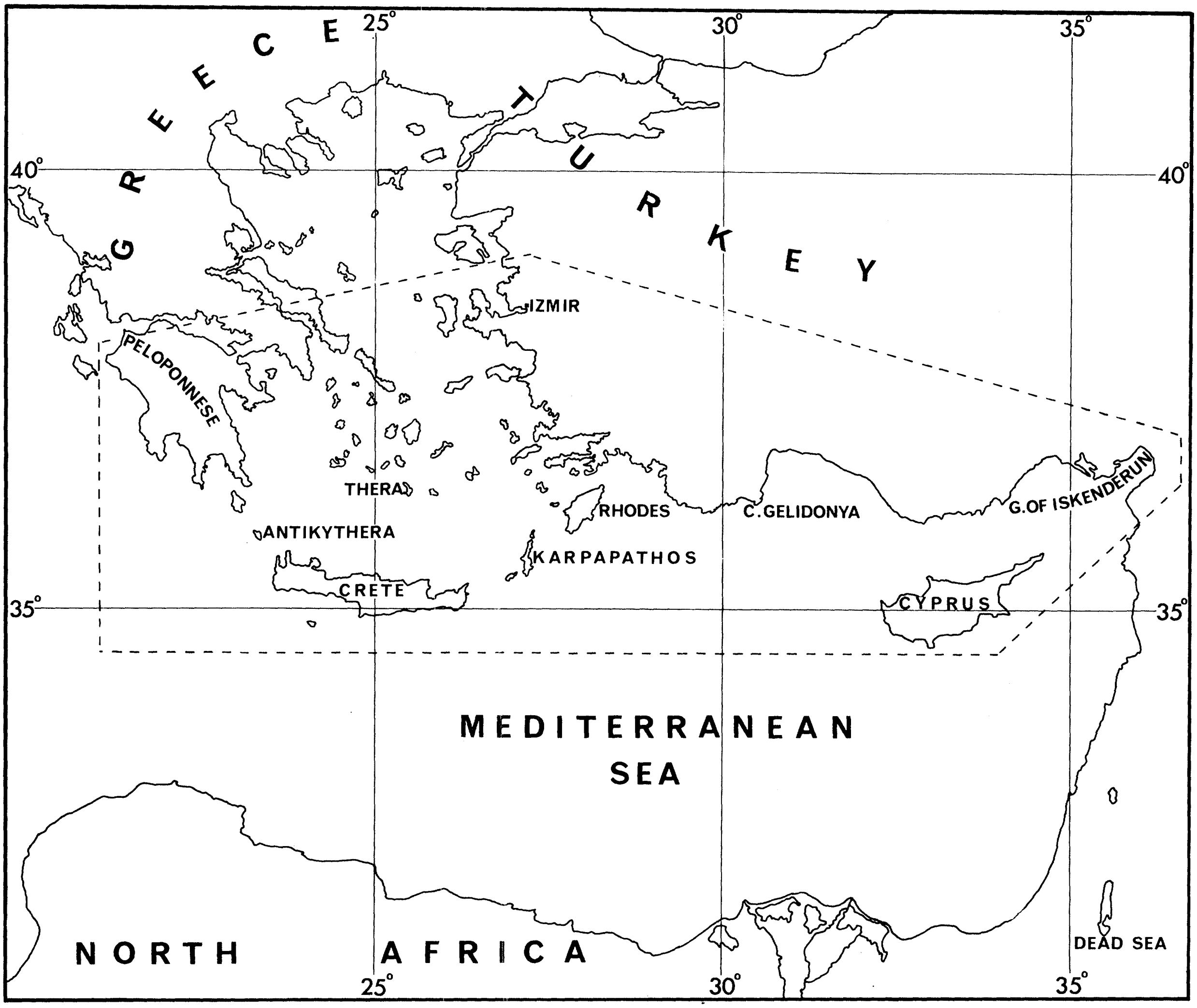

FIGURE 1. The area studied, and principal place names. Data were not obtained from the small islands in the central Aegean.

arc-trench system, the location of a plane or surface of movement south of Turkey between Crete and Cyprus, and the location of the eastern boundary of the so-called 'Aegean plate'.

The present survey arose from a study of eustatic sea levels and earth movements in the western Mediterranean conducted from 1960 to 1965 (Flemming 1969). The study was extended

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into the eastern Mediterranean in 1967, when it became apparent that the methods used could reveal the rates of vertical tectonic movement of coastal areas in great detail.

The field data are several hundred observations of relative change of land and sea level at 202 sites (figures 3 and 4) within the study area (figure 1). Before such data can be considered as evidence of earth movements the displacements relative to present sea level must be corrected for possible eustatic changes of sea level in the last 4000 years. Previous work (Flemming 1969, p. 85) indicated that the net eustatic change in the western Mediterranean in the last 2000 years was less than 0.5 m, and determination of the eustatic factor is further refined in the present study. It is axiomatic that analysis of the eustatic curve and derivation of earth movements must be treated as a single statistical problem. So-called eustatic curves derived in other areas cannot be used because of relatively unknown correction factors which should be applied to allow for hydro-isostasy (Walcott 1972), average differences of tectonic origin, net changes of average atmospheric pressure difference between the areas, etc.

In the general case, field data on relative sea level changes constitute an irregular sample from a noisy spectrum, that is, there is some component of energy or amplitude in any chosen bandwidth (Munk & Cartwright 1966, p. 535). An approximate spectrum of energy or amplitude against period for sea level changes shows peak amplitudes of 30 m at 30 s for wind-gravity waves, 5–10 m at 12–24 h for astronomic tides, and then a long low-energy band in the one day–one year period range. For longer periods of tens to millions of years, corresponding to climatic and glacial control, the amplitude increases progressively with period. The largest amplitude variations are caused by tectono-eustatic control (Menard 1964; Menard & Smith 1966; Rona 1973) and can be of the order of 500–800 m (Flemming & Roberts 1973). The order of magnitude rate of change of level for each period range can be obtained by dividing the amplitude by half the period. On this basis there is a continuous decrease in rate of change with increasing period. Most papers in the last decade or so relate the change of sea level to a land mass which is also assumed to be moving vertically, and carry out an analysis of the probable absolute movement of both elements in so far as they can be statistically separated (see, for example, Emery 1958; Scholl & Stuiver 1967; Mörner 1969; Pirazzoli 1976).

While movement of the land surface is now assumed to be generally non-negligible, it is usually implied that the nature of earth movements, their rate, discontinuous activity, and differential movement between sites, is fundamentally different from sea level change to such an extent that the two phenomena can be separated on more or less common-sense grounds, or by the application of simple statistical tests. In several well known examples, this is true, as in the case of sea level/earth movement studies of the east coast of the United States (Scholl & Stuiver 1967) and the Swedish coast (Mörner 1969) for the last 10000 years. In both these cases there is a longitudinal tilt parallel to the coast over a distance of 1000 km or more. The slope of ancient raised shorelines or the comparative displacement between relative sea level curves at different points along the shore, allow the separation of a most probable eustatic curve from the regional earth movement.

To illustrate the potential confusion of eustatic sea level changes with local or regional earth movements, some data on vertical earth movements are summarized here. Secular vertical rates of movement in southeast Europe and the Ponto–Caspian area measured by geodetic levelling in the last few decades (Mescherikov 1973; Blagovolin, Lilienberg & Pobedonostsev 1975) average 2–4 mm/a for areas generally regarded as stable with values as high as 15–30 mm/a for the Caucasus and Alpine mountain ranges. Major seismic events associated with rapid

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vertical movements are in general preceded and followed by periods of secular movement in the opposite sense which may amount to 30–40 % of the magnitude of the movement during an earthquake (Benioff 1964; Burridge & Knopoff 1967; Okada 1970; Fitch & Scholz 1971; Lensen 1974). These progressive vertical movements may have a lateral extent of several hundred kilometres, while the period between earthquakes is of the order of 100–1000 a (Ambraseys 1971; Lensen 1974). Meade (1973) reports sustained vertical movements at rates of 3.5–7.0 mm/a for sites in California, and 2–4 mm/a for the southeast United States, measured by repeated levelling. The changes were relative to continental bench marks, not to sea level. Near volcanoes such as Vesuvius, Etna, or Solfatara, vertical movements of 1 m can occur in a few days with cumulative magnitudes of 5–10 m over several hundred or thousand years (Lyell 1830, vol. 1, p. 449; Günther 1903*a*, *b*; Zugiani 1972).

In the longer term, vertical earth movements from isostatic or tectonic causes may be about 2–10 m over 1000 years (Flemming 1968*a*, 1972; Mörner 1969; Walcott 1972) while vertical movements of several hundred metres are associated with isostatic adjustment to a complete glacial cycle of several tens of thousands of years. Vertical earth movements of 1000–10000 m or more are associated with major mountain-forming orogenies over periods of several million years.

It is presumed that faulting associated with major earthquakes will, if it occurs during the actual period of observation or levelling, be immediately apparent. Vertical displacements in the range 6–11 m are recorded by Pflaker & Save (1970) and Pflaker (1972, p. 906) for earthquakes in Chile, 1960, and Alaska, 1964, respectively. Lateral scales were of the order of 100 and 800 km respectively. While the actual occurrence of a seismic event is unlikely to be confused with a sea level change, the cumulative effect of small earthquakes over several thousand years may be, and the secular changes which occur between earthquakes can be, confused with a sea level change over a period of a few hundred years.

Comparison of earth movements and sea level changes shows that for periods from 10 a to 1 Ma the vertical magnitude of movement, and hence rate of movement, is similar to within an order of magnitude for the two processes. For a period greater than one million years the potential magnitude of earth movements continues to increase, while the vertical range of sea level probably does not increase. Throughout the range of periods from one year to one million years the rate of change of both sea level and earth movements is of the order of 0.1–10.0 mm/a. On the lateral scale for periods greater than 10 a, earth movements tend to be correlated over distances of 100 km or more up to thousands of kilometres. These distances are larger than most study areas, and a component of the vertical change is therefore liable to appear in the experiment as uniform and independent of location, and thus easily regarded as eustatic.

It follows that any study intended to determine vertical earth movements and sea level changes is prone to errors arising from the statistical similarity of the two processes. To avoid confusion the size of the area studied, and the density of sampling on the lateral scale, vertically, and on the time scale, should be greater than is common in such studies.

The field method of the present study consists of observing the relation between present sea level and the archaeological remains of harbours and other structures which can be related to the sea level at the time they were built or occupied. The method is described below. Archaeological site maps of the 202 sites are being published elsewhere (Flemming, Czartoryska & Hunter 1973*a*), and the quantitative results from each site only will be listed in the present work. The distribution of the sites in terms of dating and lateral spacing will be analysed in order to assess experimental bias and selectivity. The data will then be analysed in terms of regional trends, local discontinuities, and estimated eustatic change. Finally, the conclusions will be discussed in terms of other estimates of eustatic change, and regional studies by other authors of the geology and tectonic and plate tectonic processes in the area.

Previous papers by the present author and co-authors have described various parts of the region as work progressed from 1967 to 1974 (Flemming 1968*a*, *b*, *c*, 1969, 1972; Flemming *et al.* 1973*a*, *b*). The field methods, the techniques of analysis, and the background knowledge of the structure and tectonics of the region have evolved continuously, and the conclusions and discussions in some of the earlier papers differ in part from the final synthesis and conclusions here. Summaries of the data and data analysis presented in those papers are included in the present study unchanged, but the conclusions and speculations have been modified. Readers who wish to refer to these interim reports should bear this in mind. In particular, the discovery of a large tilted block in western Crete in 1972 has altered the emphasis of analysis from attempting to find smoothed distorted surfaces to interpretation more on the basis of coherent blocks with discontinuities between them.

Field methods

Choice of sites

The tidal range mean high water–mean low water (m.h.w.–m.l.w.) in the eastern Mediterranean averaged for 22 ports varies from 27.3 cm springs to 6 cm neaps. Only two ports, Izmir and Tripoli (Lebanon) exceed 46 cm spring range (*Admiralty Tide Tables* 1972, vol. 1, pp. 392–393). This means that coastal structures such as quays, jetties, roads, slipways, etc. may be built very close to the water line. In sheltered bays and in harbours, working or walking surfaces may be built within 20 cm of mean water level. From the Bronze Age onwards, that is after 3000 B.C., early civilizations were building harbours and other coastal waterline structures in the area under study, and these structures provide datable indicators of the relative level of land and sea at the time of occupation. Data points are frequent, and interpretation is relatively independent of lithology, climate, temperature, and vegetation, although partly dependent upon local topography and exposure to wind fetch and storm waves.

As far as possible every known archaeological coastal site in the area was visited. The initial list of sites was compiled from standard classical atlases (Grundy 1963; Van Der Heyden & Scullard 1959; Lehman-Hartleben 1923) supported by the classical writings of Strabo, Pausanias, Polybius, and archaeological reports of the nineteenth and twentieth centuries, including the surveys of Beaufort (1817, 1820), Pashley (1837), Spratt (1865), Négris (1904), and Pendlebury (1939). The 202 sites listed in table 1 include all those sites for which definite field data were obtained by direct observation of the author and assistants. Twenty-two sites identifiable in the literature were not visited owing to inaccessibility, lack of official permits, or bad weather (table 2).

The site list (table 1) gives the latitude and longitude of sites to one minute as measured from Admiralty charts. The names quoted are, as far as possible, the most commonly used Latin or Greek version of the ancient name, with the modern name given in the version most commonly used by the present government of the country concerned.

TABLE 1

List of all sites for which estimates of relative sea level change could be made. The site numbers are the same as those on the maps (figures 3, 4) and in the text. T is age in thousands of years; Z is best estimate of relative vertical displacement in metres; -ve is submergence; Z/T is rate of relative submergence in metres per 1000 a, -ve is submergence. The weightings refer to the number of points allocated to each estimate on either side of Z (see figure 2); negative values mean more submergence, positive values less submergence.)

site no.	Ancient name	nearest Modern name	lat. N	long. E	age, T ka	best estimate Z	Z/T	weighting in 25 cm intervals −75	−50	−25	Z	+25	+50	+75	total weight
E. Peloponnese															
1	Gytheum	Gythion	36° 45′	22° 34′	1.5	−2.5	−1.4	—	—	—	—	—	—	—	—
2	Trinasus	Trinisi	36° 47′	22° 36′	1.0	−1.0	−1.0	—	—	—	—	—	—	—	—
3	Asopus I	Plitra	36° 41′	22° 50′	3.0	−3.0	−0.7	—	—	—	—	—	—	—	—
3	Asopus II	Plitra	36° 41′	22° 50′	1.5	−2.0	−1.3	—	—	—	—	—	—	—	—
4	—	Arkangelos	36° 37′	22° 53′	2.0	−0.2	−0.1	—	—	—	—	—	—	—	—
5	Onugnathus	Elaphonisos	36° 28′	22° 57′	3.5	−4.0	−1.1	—	—	—	—	—	—	—	—
6	Kythera	Aragora	36° 13′	23° 17′	4.0	0.0	0.0	—	—	—	—	—	—	—	—
7	Antikythera	Potamos	35° 53′	23° 02′	5.0	3.0	0.6	—	—	—	—	—	—	—	—
8	Boeae	Neapolis	36° 31′	23° 02′	1.5	−0.3	−0.2	—	—	—	—	—	—	—	—
9	Minna	Monemvasia	36° 41′	23° 03′	2.0	−1.0	−0.5	—	—	—	—	—	—	—	—
10	Zarax	Limen Ierakas	36° 47′	23° 05′	2.5	−3.0	−1.2	—	—	—	—	—	—	—	—
11	Asine	Tolon	37° 32′	22° 53′	3.0	−2.0	−0.7	—	—	—	—	—	—	—	—
12	Halieis	Porto Cheli	37° 19′	23° 09′	2.4	−5.0	−2.1	—	—	—	—	—	—	—	—
13	—	Lorenzon	37° 17′	23° 12′	1.0	−2.0	−2.0	—	—	—	—	—	—	—	—
14	Epidaurus	Palia Epidauros	37° 38′	23° 09′	2.0	−2.7	−1.3	—	—	—	—	—	—	—	—
15	Cenchreae	Kenchreai	37° 52′	23° 00′	2.5	−2.0	−0.8	—	—	—	—	—	—	—	—
16	Lechaeum	Nr Corinth	37° 56′	22° 52′	2.0	−0.7	−0.3	—	—	—	—	—	—	—	—
W. Peloponnese															
17	Phea	Katakolon	37° 38′	21° 20′	2.0	−1.0	−0.5	—	—	—	—	—	—	—	—
18	Pylus Coryphasium	Navarino Bay	36° 57′	21° 30′	2.0	−1.0	−0.5	—	—	—	—	—	—	—	—
19	Methone	Methoni	36° 49′	21° 42′	1.0	−1.5	−1.5	—	—	—	—	—	—	—	—
20	Asine	Koroni	36° 48′	21° 58′	0.6	−1.2	−2.0	—	—	—	—	—	—	—	—
21	Corone	Petalidi	36° 57′	21° 56′	—	—	—	—	—	—	—	—	—	—	—
22	Cardamyle	Kardamyli	36° 52′	22° 15′	1.0	−1.0	−1.0	—	—	—	—	—	—	—	—
23	Leutra	Stoupa	36° 50′	22° 16′	2.0	−2.5	−1.3	—	—	—	—	—	—	—	—
24	Pephnus	Selenitsa	36° 49′	22° 18′	1.0	−1.0	−1.0	—	—	—	—	—	—	—	—
25	—	Tigani	36° 32′	22° 22′	2.0	−0.5	−2.5	—	—	—	—	—	—	—	—
26	Teuthrone	Kotronas	36° 37′	22° 29′	1.0	−1.0	−1.0	—	—	—	—	—	—	—	—
27	—	Skoutari	36° 40′	22° 30′	1.5	−3.5	−2.2	—	—	—	—	—	—	—	—
28	Acriae	Elea	36° 44′	22° 48′	3.0	−2.0	−0.7	—	—	—	—	—	—	—	—
S.W. Turkey															
29	Elaea	Kazikbaglar	38° 57′	27° 03′	2.3	0.0	0.0	—	2	3	3	—	—	—	8
30	Cyme	Namurt Limani	38° 46′	26° 56′	2.4	−1.0	−0.2	—	1	2	4	2	1	—	10
31	Smyrna I	Bayrakli	38° 28′	27° 11′	4.0	−3.0	−0.8	—	—	—	—	—	—	—	10

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Table 1 (*cont.*)

site no.	Ancient name	nearest Modern name	lat. N	long. E	age, T ka	best estimate Z	Z/T	weighting in 25 cm intervals −75	−50	−25	Z	+25	+50	+75	total weight
S.W. Turkey (*cont.*)															
31	Smyrna II	Bayrakli	38° 28′	27° 11′	2.5	−1.0	−0.4	—	1	3	6	—	—	—	10
32	Clazomenae	Urla Iskelesi	38° 22′	26° 48′	2.3	−1.0	−0.4	—	—	—	3	3	2	1/1	10
33	—	Urla Beach	38° 19′	26° 41′	0.3	−0.5	−1.7	—	2	2	4	1	1	—	10
34	Erithrae I	Ildir	38° 23′	26° 30′	0.1	−0.3	−3.0	—	—	—	5	5	—	—	10
34	Erithrae II	Ildir	38° 23′	26° 30′	0.2	−0.5	−2.5	—	—	—	5	5	—	—	10
34	Erithrae III	Ildir	38° 23′	26° 30′	2.4	−1.4	−0.6	—	—	1	6	2	1	—	10
34a	—	Yali	38° 20′	26° 26′	1.0	−0.5	−0.5	—	1	3	6	—	—	—	10
35	—	Ilica	38° 20′	26° 22′	2.4	−0.5	−0.3	—	2	2	2	2	2	—	10
36	—	Cesme	38° 20′	26° 19′	0.1	−0.2	−0.2	—	—	—	5	—	—	—	5
37	—	Ciftlik	38° 18′	26° 17′	0.1	−0.1	−0.1	—	—	5	5	—	—	—	10
38	Teos I	Sigacik	38° 11′	26° 48′	2.3	−0.8	−0.4	—	—	1	7	2	—	—	10
38	Teos II	Sigacik	38° 12′	26° 48′	2.3	−0.8	−0.4	—	—	—	7	2	1	—	10
38	Teos III	Sigacik	38° 12′	26° 48′	0.1	−0.1	−1.0	—	—	5	5	—	—	—	10
39	Lebedus	Kisik	38° 05′	26° 59′	—	—	—	—	—	—	—	—	—	—	—
40	Notium	Nr Ahmetbeyli	38° 00′	27° 13′	2.6	−1.0	−0.4	—	1	2	4	2	1	—	10
41	Ephesus	Nr Selcuk	37° 56′	27° 20′	2.0	—	—	—	—	—	—	—	—	—	—
42	Miletus I	Balat	37° 32′	27° 18′	2.0	−1.5	−0.8	—	—	—	5	3	2	—	10
42	Miletus II	Balat	37° 32′	27° 18′	3.2	−1.5	−0.5	—	1	2	5	2	—	—	10
43	Heraclea Latmus	Kapikiri	37° 30′	27° 33′	2.5	−1.5	−0.6	—	1	1	1	1	1	—	5
44	—	Ghioucker Island	37° 29′	27° 31′	—	—	—	—	—	—	—	—	—	—	—
45	Panormus	Kauala Limani	37° 24′	27° 15′	2.6	0.0	0.0	—	1	2	4	2	1	—	10
46	Iasus I	Asin Kale	37° 17′	27° 37′	1.0	0.0	0.0	—	—	3	7	—	—	—	10
46	Iasus II	Asin Kale	37° 17′	27° 37′	2.0	−0.5	−0.3	—	—	—	6	3	1	—	10
47	Bargylia	Varvil, Asar	37° 12′	27° 38′	2.0	−0.2	−0.1	—	—	2	7	1	—	—	10
48	Caryanda I	Guvercinlik	37° 08′	27° 36′	0.3	0.0	0.0	—	—	2	8	—	—	—	10
48	Caryanda II	Guvercinlik	37° 08′	27° 36′	1.5	−0.3	−0.2	—	—	3	7	—	—	—	10
49	—	Gol	37° 08′	27° 26′	—	—	—	—	—	—	—	—	—	—	—
50	Myndus	Gumusluk	37° 08′	27° 16′	2.0	−1.2	−0.6	—	—	1	8	1	—	—	10
51	—	Karatoprak	37° 00′	27° 17′	1.0	−0.5	−0.5	—	1	2	5	2	—	—	10
52	Halincarnassus	Bodrum	37° 02′	27° 27′	2.4	−1.0	−0.4	—	1	2	6	1	—	—	10
53	Cedreae	Tasbuku	37° 00′	28° 14′	1.5	−0.3	−0.2	—	—	1	6	3	—	—	10
54	(New) Cnidus	Takir	36° 41′	27° 24′	2.3	0.0	0.0	—	—	1	7	1	1	—	10
55	(Old) Cnidus	Datca	36° 44′	27° 44′	2.5	0.0	0.0	—	—	1	8	1	—	—	10
56	—	Orhaniye	36° 46′	28° 10′	1.0	0.0	0.0	—	—	5	5	—	—	—	10
57	—	Bozburun	36° 42′	28° 05′	1.0	−1.0	−1.0	—	—	—	4	3	3	—	10
58	Loryma	Oplotheke	36° 35′	28° 03′	1.5	−1.0	−0.7	—	—	—	6	3	1	—	10
59	—	Saranda	36° 39′	28° 07′	1.0	−0.5	−0.5	—	—	—	5	5	—	—	10
60	—	Fethiye	36° 38′	29° 09′	—	—	—	—	—	—	—	—	—	—	—
61	Patara	Gelemic	36° 16′	29° 19′	—	—	—	—	—	—	—	—	—	—	—
62	Antiphellus I	Kas	36° 13′	29° 38′	2.2	−2.2	−1.0	—	1	1	8	—	—	—	10
62	Antiphellus II	Kas	36° 13′	29° 38′	1.0	−1.0	−1.0	—	1	1	7	1	—	—	10

TABLE 1 (*cont.*)

site no.	Ancient name	nearest Modern name	lat. N	long. E	age, T ka	best estimate Z	Z/T	weighting in 25 cm intervals −75	−50	−25	Z	+25	+50	+75	total weight
S.W. Turkey (*cont.*)															
63	Andriace-Myra	Kale	36° 13′	29° 58′	1.0	−1.0	−1.0	—	1	2	4	2	1	—	10
64	—	Gemili or St Nicholas Island	36° 33′	29° 04′	1.0	−2.0	−2.0	—	—	—	5	5	—	—	10
65	Aperlae	Kekova	36° 10′	29° 52′	2.5	−2.2	−0.9	1	2	2	5	—	—	—	10
S. Turkey															
66	Phaselis	Tekirova	36° 31′	30° 34′	2.5	−0.2	−0.1	—	—	2	6	2	—	—	10
67	Lara	Magydus	36° 50′	30° 48′	1.7	0.0	0.0	—	2	3	4	1	—	—	10
68	Side	Selimiye	36° 46′	31° 24′	2.2	0.0	0.0	—	—	—	10	—	—	—	10
69	—	Kara Burun	36° 39′	31° 41′	2.0	0.0	0.0	—	—	2	8	—	—	—	10
70	Ptolemais	Figla Burun	36° 36′	31° 47′	—	—	—	—	—	—	—	—	—	—	—
71	—	Alanya	36° 32′	32° 01′	0.7	0.0	0.0	—	—	1	8	1	—	—	10
72	Laertes	—	36° 28′	32° 08′	—	—	—	—	—	—	—	—	—	—	—
73	Syedra	Demirtas	36° 26′	32° 11′	1.5	0.0	0.0	—	1	1	6	1	1	—	10
74	Sellinus	Gazipasa	36° 16′	32° 18′	1.8	0.0	0.0	—	—	1	8	1	—	—	10
75	Hamaxia	Aydap	36° 19′	32° 16′	1.5	0.0	0.0	—	—	—	6	1	1	1/1	10
76	Antiocha ad Cragum	Gunen Koy	36° 08′	32° 27′	—	—	—	—	—	—	—	—	—	—	—
77	Mellisa	Melec	36° 03′	32° 42′	—	—	—	—	—	—	—	—	—	—	—
78	Anemurium	Anamur	36° 02′	32° 49′	1.5	0.0	0.0	1	1	2	2	2	1	1	10
79	Mamuriye	Mamaure	36° 05′	32° 55′	0.6	0.0	0.0	—	1	2	6	1	—	—	10
80	Arsinoe	Bozyazi	36° 06′	32° 59′	1.5	0.0	0.0	—	—	2	8	—	—	—	10
81	—	Softa Kalesi	36° 05′	33° 02′	—	—	—	—	—	—	—	—	—	—	—
82	Mellaxia	Soguksu Limani	36° 08′	33° 19′	2.0	0.0	0.0	—	—	3	4	3	—	—	10
83	Celendris	Gelindere	36° 08′	33° 21′	0.5	0.0	0.0	—	—	3	6	1	—	—	10
84	Holmus	Ovacik	36° 10′	33° 42′	0.6	0.0	0.0	—	—	3	7	—	—	—	10
85	—	Bagsakada	36° 16′	33° 51′	0.6	0.0	0.0	—	—	2	6	2	—	—	10
86	—	Agalimeni Fortress	36° 16′	33° 52′	0.6	0.0	0.0	—	1	1	7	1	—	—	10
87	—	Monastir	36° 19′	33° 52′	—	—	—	—	—	—	—	—	—	—	—
88	Persente	Susanoglu	36° 25′	34° 03′	1.5	0.0	0.0	—	—	2	6	2	—	—	10
89	Corycus	Kizkalesi	36° 27′	34° 11′	1.5	0.0	0.0	—	—	2	8	—	—	—	10
90	Elaeusa	Ayas	36° 28′	34° 12′	—	—	—	—	—	—	—	—	—	—	—
91	Soli	Viransehir	36° 37′	34° 33′	2.0	0.0	0.0	—	1	2	6	1	—	—	10
92	Megarus	Karatas	36° 34′	35° 23′	2.0	0.0	0.0	—	—	3	7	—	—	—	10
93	Aegeae	Yumurtalik	36° 34′	35° 47′	2.0	0.0	0.0	—	1	3	6	—	—	—	10
94	Baiae	Payas	36° 46′	36° 12′	0.5	0.0	0.0	—	1	3	6	—	—	—	10
95	—	Bonnel	36° 25′	35° 53′	2.0	0.0	0.0	—	1	2	4	2	1	—	10
96	Seleucia Pieria	Magaracik	36° 04′	35° 58′	1.8	0.0	0.0	—	1	2	6	1	—	—	10
Crete															
97	Amnisos	Matium	35° 20′	25° 13′	3.6	−1.5	−0.4	—	1	2	5	2	1	—	10
98	Nirou Khani	Ayia Theodoroi	35° 20′	25° 16′	3.6	−1.75	−0.5	1	2	2	5	—	—	—	10

TABLE 1 (cont.)

site no.	Ancient name	nearest Modern name	lat. N	long. E	age, T ka	best estimate Z	Z/T	weighting in 25 cm intervals −75	−50	−25	Z	+25	+50	+75	total weight
Crete (cont.)															
99	Chersonissos	Khersonisos	35° 19′	25° 24′	1.9	−1.0	−0.5	—	—	1	5	4	—	—	10
100	Malia	Malia	35° 18′	25° 30′	3.6	−1.0	−0.3	—	1	2	4	2	1	—	10
101	—	Spinalonga	35° 18′	25° 45′	2.0	−1.0	−0.5	—	—	1	8	1	—	—	10
102	—	Port Kolokithia	35° 16′	25° 46′	1.0	−1.0	−1.0	—	—	2	5	2	1	—	10
103	Olous	Elounda	35° 16′	25° 45′	2.0	−1.9	−1.0	—	—	1	6	2	1	—	10
104	Minoa	Pyrgos	35° 07′	25° 44′	2.2	−1.5	−0.7	—	2	2	2	2	2	—	10
105	—	Gournia	35° 07′	25° 48′	—	—	—	—	—	—	—	—	—	—	—
106	Psira	Psyra	35° 11′	25° 52′	1.9	−0.75	−0.4	1	1	1	1	1	1	1	7
107	Tholos	Kavousi	35° 08′	25° 51′	—	—	—	—	—	—	—	—	—	—	—
108	Mokhlos I	S. Nickolo	35° 09′	25° 53′	1.9	−0.25	−0.1	—	—	1	6	3	—	—	10
108	Mokhlos II	S. Nickolo	35° 09′	25° 53′	3.6	−1.75	−0.5	—	—	2	2	1	—	—	5
109	Eteia	Sitia	35° 12′	26° 07′	1.8	−0.8	−0.4	—	—	2	7	1	—	—	10
110	Itanos	Erimopoulis	35° 16′	26° 17′	2.0	−2.2	−1.1	—	—	1	5	2	2	—	10
111	—	Capa Plaka	35° 11′	26° 20′	1.0	−0.2	−0.2	—	—	—	5	3	2	—	10
112	Zakros	Kato Zakros	35° 06′	26° 17′	1.5	0.0	0.0	—	—	—	5	2	1	1/1	10
113	Ampelos	—	35° 02′	26° 15′	—	—	—	—	—	—	—	—	—	—	—
114	Leuce	Kuphonisi	34° 56′	26° 10′	1.9	2.2	1.2	—	—	2	5	1	—	—	8
115	—	Akri Goudara	35° 00′	26° 08′	5.0	3.5	0.7	—	—	1	8	1	—	—	10
116	Hierpatra	Hierapetra	35° 00′	25° 44′	1.8	−0.5	−0.3	—	—	—	5	3	2	—	10
117	—	Muertos	35° 00′	25° 35′	—	—	—	—	—	—	—	—	—	—	—
118	—	Moni Arvi	34° 59′	25° 28′	0.5	−0.25	−0.5	—	—	1	3	1	—	—	5
119	Lasea	Kalos Limeni	34° 56′	24° 50′	1.8	0.0	0.0	—	—	1	7	2	—	—	10
120	Metallon I	Matala	34° 59′	24° 45′	1.9	0.0	0.0	—	—	2	6	2	—	—	10
120	Metallon II	Matala	34° 59′	24° 45′	1.1	−2.0	−1.8	—	—	—	—	—	—	—	—
121	Komo	Messara Bay	35° 00′	24° 46′	—	—	—	—	—	—	—	—	—	—	—
122	Sulia	Ayia Galini	35° 06′	24° 42′	—	—	—	—	—	—	—	—	—	—	—
123	—	Plakios	35° 11′	24° 25′	1.9	2.0	0.9	—	—	2	6	2	—	—	10
124	—	Ayia Marina	35° 10′	24° 18′	1.9	2.0	0.9	—	—	2	6	2	—	—	10
125	—	Frangocastelli	35° 11′	24° 15′	—	—	—	—	—	—	—	—	—	—	—
126	—	Sphakia	35° 12′	24° 08′	1.9	3.5	1.6	—	—	2	6	2	—	—	10
127	Loutron	Loutro	35° 12′	24° 05′	1.9	3.5	1.6	—	—	2	6	2	—	—	10
128	Kalamydes	—	35° 14′	23° 42′	1.9	7.5	3.4	—	—	2	6	2	—	—	10
129	—	Paleochora	35° 13′	23° 40′	—	—	—	—	—	—	—	—	—	—	—
130	—	Pelikaniotikos	35° 13′	23° 38′	—	—	—	—	—	—	—	—	—	—	—
131	—	Plakaki Krio	35° 14′	23° 35′	2.2	8.0	3.6	—	—	2	6	2	—	—	10
132	Biennus	Krio	35° 13′	23° 34′	—	—	—	—	—	—	—	—	—	—	—
133	Musagores	Elaphonisi	35° 14′	23° 31′	2.2	8.5	3.9	—	—	2	6	2	—	—	10
134	—	Khrysoskalitissa	35° 18′	23° 31′	2.2	7.5	3.4	—	—	2	6	2	—	—	10
135	Sphinarion	Sphinari	35° 24′	23° 34′	2.2	7.0	3.2	—	—	2	6	2	—	—	10
136	Phalasarna	Kutri	35° 30′	23° 34′	2.2	6.6	3.0	—	—	2	6	2	—	—	10
137	—	Gramvousa	35° 35′	23° 34′	—	—	—	—	—	—	—	—	—	—	—

TABLE 1 (*cont.*)

site no.	Ancient name	nearest Modern name	lat. N	long. E	age, T ka	best estimate Z	Z/T	weighting in 25 cm intervals −75	−50	−25	Z	+25	+50	+75	total weight
Crete (*cont.*)															
138	Agneion	Ayios Sastis	35° 35′	23° 36′	2.2	6.0	2.7	—	—	2	6	2	—	—	10
139	Kisamos	Kissamou	35° 30′	23° 38′	2.2	5.2	2.4	—	—	2	6	2	—	—	10
140	Napia	Methymna	35° 29′	23° 42′	2.2	5.0	2.3	—	—	2	6	2	—	—	10
141	Ohiktinaia	Kantzlieres	35° 39′	23° 46′	2.2	3.0	1.4	—	—	2	6	2	—	—	10
142	Burdroae	Theodoro	35° 32′	23° 56′	2.2	3.0	1.4	—	—	2	6	2	—	—	10
143	—	Minoan Site?	—	—	—	—	—	—	—	—	—	—	—	—	—
144	—	Chor Fakia	35° 33′	24° 05′	2.2	1.1	0.5	—	—	2	6	2	—	—	10
145	Marathi	Ormos Loutraki	35° 31′	24° 11′	2.2	1.5	0.7	—	—	2	6	2	—	—	10
146	Ormos Marathi	Souda	35° 31′	24° 11′	—	—	—	—	—	—	—	—	—	—	—
147	Amphimalla	Georgiopolis	35° 20′	24° 16′	2.2	1.5	0.7	—	—	2	6	2	—	—	10
148	—	—	35° 20′	24° 21′	2.2	1.0	0.45	—	—	2	6	2	—	—	10
149	—	—	35° 21′	24° 25′	2.2	0.8	0.36	—	—	2	6	2	—	—	10
150	Rethymne	Rethymnon	35° 22′	24° 28′	2.2	−0.2	−0.1	—	—	—	5	5	—	—	10
151	Panormus	Khandra Kayo	35° 25′	24° 42′	0.5	−0.5	−1.0	—	—	2	5	3	—	—	10
Cyprus															
152	—	C. Kormakiti	35° 24′	32° 56′	—	0.0	0.0	—	—	—	—	—	—	—	—
153	Lapethus	Lambousa	35° 37′	33° 13′	2.0	0.0	0.0	—	—	—	8	2	—	—	10
154	—	Snake Island	35° 21′	33° 18′	—	—	—	—	—	—	—	—	—	—	—
155	Cerynia	Kyrenia	35° 21′	33° 19′	2.3	0.0	0.0	—	—	—	—	—	—	—	—
156	—	Vrisi	35° 20′	33° 25′	3.0	0.0	0.0	—	—	—	—	—	—	—	—
157	—	Platymeis	35° 20′	33° 30′	—	—	—	—	—	—	—	—	—	—	—
158	—	Ayia Marina	35° 22′	33° 40′	—	—	—	—	—	—	—	—	—	—	—
159	—	Ayios Thyrsos	35° 34′	34° 16′	2.0	−0.25	−0.13	—	—	—	—	—	—	—	—
160	Carpasia	Ayios Philon Aphendrika	35° 38′	34° 23′	2.0	0.0	0.0	—	—	—	—	—	—	—	—
161	Dinaretum	Apostolos Andrea	35° 42′	34° 35′	—	—	—	—	—	—	—	—	—	—	—
162	—	Koroni	35° 37′	34° 29′	—	—	—	—	—	—	—	—	—	—	—
163	Salamis I	Salamis	35° 11′	33° 55′	1.7	−2.0	−1.18	—	—	—	—	—	—	—	—
163	Salamis II	Salamis	35° 11′	33° 55′	1.00	−0.3	−0.3	—	—	—	—	—	—	—	—
164	Ammochostos	Famagusta	35° 8′	33° 57′	2.0	−0.5	−0.25	—	—	—	—	—	—	—	—
165	Pedalium Promontorium	Cape Greco	34° 58′	34° 05′	—	—	—	—	—	—	—	—	—	—	—
166	Dades	Ayios Yeorgios	34° 58′	33° 55′	2.0	−0.2	−0.1	—	—	—	—	—	—	—	—
167	Kitium	Larnaca	34° 55′	33° 39′	—	—	—	—	—	—	—	—	—	—	—
168	—	Bronze Age Harbour	34° 54′	33° 39′	—	—	—	—	—	—	—	—	—	—	—
169	Palaea	Petounda Pt	34° 47′	33° 30′	—	—	—	—	—	—	—	—	—	—	—
170	Amathos I	Amathus	34° 43′	33° 08′	2.5	−0.75	−0.3	—	—	—	—	—	—	—	—
170	Amathos II	Amathus	34° 43′	33° 08′	0.5	−0.2	−0.13	—	—	—	—	—	—	—	—
171	Curias	Venetian Canal	34° 37′	33° 01′	0.5	−0.1	−0.3	—	—	—	—	—	—	—	—
172	—	Dreamers Bay	34° 34′	32° 54′	1.8	−2.0	−1.1	—	—	—	—	—	—	—	—

TABLE 1 (*cont.*)

site no.	Ancient name	nearest Modern name	lat. N	long. E	age, T / ka	best estimate Z	Z/T	weighing in 25 cm intervals: −75	−50	−25	Z	+25	+50	+75	total weight
Cyprus (*cont.*)															
173	Curium	Episkopi	34° 40′	32° 54′	—	—	—	—	—	—	—	—	—	—	—
174	—	Mandria	34° 42′	32° 32′	—	—	—	—	—	—	—	—	—	—	—
		Petra tou Romiou	34° 40′	32° 38′	—	—	—	—	—	—	—	—	—	—	—
175	Paphos	Paphos	34° 46′	32° 25′	2.0	−0.3	−0.15	—	—	—	—	—	—	—	—
176	Drepanum	Drepanum	34° 54′	32° 19′	2.0	−0.3	−0.15	—	—	—	—	—	—	—	—
177	Lara	Lara	34° 57′	32° 19′	—	—	—	—	—	—	—	—	—	—	—
178	—	Yeranisou	35° 01′	32° 18′	—	—	—	—	—	—	—	—	—	—	—
179	Acamus	C. Arnauti	35° 5′	32° 17′	—	—	—	—	—	—	—	—	—	—	—
180	—	Ayios Nikolos	35° 6′	32° 18′	2.0	−0.75	−0.38	—	—	—	—	—	—	—	—
181	Marium Polis	Lachi	35° 4′	32° 24′	2.0	−0.75	−0.38	—	—	—	—	—	—	—	—
182	—	C. Kokkino	35° 11′	32° 41′	—	—	—	—	—	—	—	—	—	—	—
183	Soli	Karavastasi	35° 9′	32° 50′	—	—	—	—	—	—	—	—	—	—	—
Rhodes															
184	Bema	Langona or Skala Kamerensis	36° 17′	27° 52′	2.0	0.0	0.0	—	—	—	—	—	—	—	—
185	Camiras	Ayios Minas	36° 21′	27° 58′	1.0	0.0	0.0	—	—	—	—	—	—	—	—
186	Rhodus	Rhodes/Ileo Koreo	36° 27′	28° 17′	5.0	1.75	0.35	—	—	—	—	—	—	—	—
187	—	Zimbule	36° 26′	28° 17′	0.7	−0.25	−0.36	—	—	—	—	—	—	—	—
188	Calithea I	Kalithea	36° 23′	28° 17′	5.0	3.7	0.74	—	—	—	—	—	—	—	—
188	Calithea II	Kalithea	36° 23′	28° 17′	2.4	2.7	1.12	—	—	—	—	—	—	—	—
188	Calithea III	Kalithea	36° 23′	28° 17′	0.7	−0.25	−0.36	—	—	—	—	—	—	—	—
189	Afandea	Aphandou	36° 19′	28° 15′	5.0	5.24	1.05	—	—	—	—	—	—	—	—
190	—	Cape Teodoco	36° 16′	28° 13′	5.0	3.4	0.68	—	—	—	—	—	—	—	—
191	—	Tsambika	36° 14′	28° 12′	5.0	3.25	0.65	—	—	—	—	—	—	—	—
192	Lindus	Lindos	36° 06′	28° 08′	5.0	2.0	0.40	—	—	—	—	—	—	—	—
193	—	Merminga	36° 03′	28° 01′	5.0	1.0	0.20	—	—	—	—	—	—	—	—
194	—	C. Istros	35° 56′	27° 54′	5.0	0.33	0.07	—	—	—	—	—	—	—	—
195	Mnasyrion Prom.	C. Praso Nisi	35° 53′	27° 48′	—	—	—	—	—	—	—	—	—	—	—
Karpathos															
196	—	Lefcos	35° 35′	27° 06′	2.0	−1.0	−0.50	—	—	—	—	—	—	—	—
197	—	Vurgunda	35° 48′	27° 12′	2.0	−0.75	−0.38	—	—	—	—	—	—	—	—
198	—	Tristomo	35° 50′	27° 15′	0.8	−0.25	−0.31	—	—	—	—	—	—	—	—
199	Palatea	Palatea	35° 53′	27° 16′	1.0	0.0	0.0	—	—	—	—	—	—	—	—
200	Potideion	Pighadia	35° 31′	27° 15′	—	—	—	—	—	—	—	—	—	—	—
201	—	Vathi Potamos	35° 29′	27° 14′	2.0	−1.1	−0.55	—	—	—	—	—	—	—	—
202	—	Makriyalo I	35° 26′	27° 11′	1.0	0.0	0.0	—	—	—	—	—	—	—	—
202	—	Makriyalo II	35° 26′	27° 11′	2.0	−0.9	−0.45	—	—	—	—	—	—	—	—

Table 2. Sites identified from the literature which contain potentially useful data, but which were not visited during the present study

Peloponnese	Kassos
Calauria	Kassos
Cyphanta	Rhodes
Prasieae	Vroulia
Side, plus sites on the Gulf of Corinth	Kerama
Crete	S.W. Turkey
Cap Sidhero	Losta
Dium	Ceramus
Priansos	Caunus
Lebena	Olympus
Suja	Cyprus
Lissos	Cape Arnauti island
Poikilassos	Tou Petra island
Karpathos	Elea
Arkasa	

Derivation of relative change of sea level at a site

In its simplest form, the archaeological method of detecting relative sea level change was used by Hamilton (1776–9), Smyth (1854), Günther (1903 *a*, *b*), Hafemann (1960) and many others (see Flemming 1969, pp. 3–10 for a bibliography). Blackman (1973) has given the most detailed analysis from the standpoint of a classicist and archaeologist of the original functions and relations to sea level of mooring stones, quay surfaces, slipways, docks, rock-cut tanks and channels. These items provide exceptionally accurate means to improve the identification of levels at certain sites. In general, structures are classed as follows:

(1) those structures built on land, including houses, tombs, roads, city walls, wells, drains, gutters, storage tanks, water tanks, olive presses, religious and public buildings, steps, quarries, passages and tunnels, mooring stones and rings, floors and mosaics;

(2) constructions with their foundations necessarily built in the water include slipways, breakwaters, moles, jetties, quays, docks, fish tanks (piscinae), ship channels, and salt pans or salinas;

(3) structures which could have their foundations on land or in the sea include solid towers and parts of city walls, roads on causeways, lighthouse towers, villas and baths built on solid foundations close to or in the sea.

Identification of several structures in each class at one site allows the most accurate determination of relative sea level at the time of occupation of the site, while the presence of structures in one class only provides an approximate estimate, or sets an upper or lower limit to a possible range of values. The varying reliability of estimates of this kind can be quantified by allocating different probability weights to different estimates as described in the sections on sources of vertical error, and probability of estimates.

Field survey methods

Field mapping techniques were extremely simple. Sketch plans were drawn onto outlines enlarged from large scale charts and maps. The general location of buildings and topographic features were determined by measuring with plastic 50 m tape, pacing, photography, or estimates by eye. The actual proportions of buildings were measured with a 50 m tape or 2 m ranging pole whenever possible. Much of the work was done by snorkel swimmers and occasionally by divers using compressed air breathing apparatus. However, lateral scale does not matter for the present purposes, provided that the dimensions can be measured sufficiently reliably to confirm the probable function of a building, or the reasonable layout of a harbour. The sketch maps published elsewhere (Flemming *et al.* 1973*a*, *b*) are intended to give unambiguous evidence as to the location and nature of the data, and to provide a basis for further work by archaeologists.

Sources of vertical error

Vertical measurements were made with the greatest accuracy possible in the time, using tapes or poles marked in centimetres.

The three principal sources of error in the observation of present sea level arise from wind waves, tides, and variations in atmospheric pressure. In summer the wave amplitudes are often less than 20 cm and in sheltered bays the water may be glass calm. When waves are present, a mean may be taken of the upper and lower limit on a pole or tape. The error in observing mean level is 5 cm or less. The rate at which work was conducted, with seldom more than half a day actually in the water on each site, meant that tidal measurements were only possible at a few sites. Thus there is apparently unavoidable error of about 20 cm. However, this is not quite as clear-cut as it seems. If a structure such as the mosaic floor of a room is flooded nowadays at any state of the tide, one can be absolutely sure that the relative change of level must have been at least sufficient to raise that floor above the observed water level, and probably considerably more to allow for wave action. Thus there is a definite limit at one side of the possible range of estimates.

Variations in atmospheric pressure, whether diurnal or seasonal, can cause sea level changes of 0.3–1.0 m in the Mediterranean, especially when associated with persistent strong winds such as the Mistral in the Golfe de Lions. The *Admiralty pilots* (1955, pp. 28–38; 1961, vol. 49, pp. 46–57) show that the barometric pressure at ten ports in the area of the Hellenic Arc, South Turkey and Cyprus, is on average 7.8 mbar† lower in mid-summer than in mid-winter. Munk & Macdonald (1960, p. 101) and Rossiter (1967) show that the change of sea level arising from a 1 mbar change in atmospheric pressure is about 1 cm. Warners (1957, figs 1–3 and pp. 14–19) shows that from May to August the atmospheric pressure drops sharply in the eastern Mediterranean compared with the central and western Mediterranean. The relative drop is about 4 mbar. Depending on response time, the sea level in summer could thus be about 4 cm higher than in winter. Striem (1974, p. 62) shows that the mean sea level (m.s.l.) at Ashdod on the coast of Israel is 6 cm higher on average for the 6 months May–October than for the 6 months November–April. In the case of the Israel coast however, this includes a rise of m.s.l. of 17 cm between May and August. The root sum square error from waves, tide, and atmospheric uncertainties is then 24 cm.

Accuracy of dating

Accurate dating of the period of occupation of a site or structure depends ultimately on pottery or inscriptions. However, the duration of occupation of a city was usually several hundred years and often more than a thousand. As the mean of the modulus of the rate of change of relative

† 1 bar = 10^5 Pa.

vertical level detected in this area is 0.68 m/ka (table 3), the typical change of relative level during such a period of occupation of 200 years would be of the order of 10–40 cm, with an average of about 20 cm.

Variations in fetch and exposure

The local plan and relief of the coast determine the size of waves which are likely to impinge on any structure. Depending on its function, a structure will therefore be built higher above still water level if it is more exposed. For example, the working surface of a quay or the foundations of a house built on unconsolidated ground, are likely to be at least 1.5 m and 2.0 m above water level respectively if waves of 1 m amplitude can occur at that point on the shore. All estimates of the original sea level deduced from a single structure must include an estimate of clearance from still water level to allow for exposure to waves. Blackman (1973) discusses the precise relation of quay surfaces and mooring stones to the water level. The error in applying such calculations to single structures may be as much as 0.5 m, but where several structures place upper and lower limits on possible sea levels, the overall error from this factor is probably less than 20 cm.

Summary of vertical errors

The root sum square error arising from waves (5 cm), tides (20 cm), atmospheric pressure variation (10 cm), dating error (20 cm), and variation in exposure (20 cm), is therefore 37 cm.

Marine solution notches

The possibility of correlating marine solution notches on calcareous rocks with nearby ancient ruins was realized by Spratt (1865, vol. 2, p. 123) and independently by Günther (1903*b*, p. 5). Solution notches at sea level in partly metamorphosed limestone cliffs in the Mediterranean area are typically 30–50 cm high, and 20–50 cm deep into the rock, and survive for several tens of thousands of years unless the cliff itself is subsequently eroded by wave action at its base (Flemming 1965, 1968*d*). Notches at present sea level are sometimes associated with an organic calcareous growth of 'trottoir' which has been described by Blanc & Molinier (1955). Hafemann (1965, p. 637 *et seq.*) and Pirazolli & Thommeret (1977) report algal trottoir and worm tubules associated with raised notches and caves giving dates ranging from 1600 to 2400 B.P. No trottoir samples were taken in the present survey.

Solution notches on vertical cliffs which continue several metres above and below sea level are distinct from wave erosion notches which form at a cliff-terrace junction. Both types of notches have been used by many authors (see, for example, Stearns 1974) for measuring relative changes of sea level over appreciable periods of the Pleistocene. A refinement of the technique is introduced by the occurrence of micro-solution notches, with a height of 10–20 cm, and a depth of 5 cm or less. Closely spaced notches of this form were first observed in Antikythera in 1967 above sea level (Flemming *et al.* 1973*a*, p. 10), and were subsequently found above sea level in western Crete, Rhodes, and Cyprus, and below sea level in Karpathos. The displacement of the mid-point of the notch relative to the present sea level notch was measured in all cases with a plastic tape or ranging pole marked in centimetres. The systems of small repeated notches will be referred to as 'ripple notches' in the present text.

Beachrock

Beachrock is a rapidly forming rock composed of beach sand and gravel cemented by a calcareous cement of aragonite or magnesium-calcite precipitated from seawater. Beachrock sometimes covers several kilometres of beach in continuous sheets within 1 m above and below the waterline, as at Hierapetra (116)† and east of Siteia (109) in Crete. Occurrence of beachrock was reported by Beaufort (1817, 1820) in south Turkey and plotted on the Admiralty chart of Crete compiled by Spratt, using the descriptive term 'petrified beach'. A modern review of beachrock formation processes in the Mediterranean is provided by Alexandersson (1972) with an extensive bibliography. Alexandersson (1972, p. 205) quotes Schmalz (1969) to show that lithification of beachrock may be completed in less than 10–15 a.

Blanc (1958) described remains of Roman harbour walls at Leptis Magna on the coast of Libya which have been included in a calcareous cemented rock which he called trottoir. The cemented mass includes a large quantity of small molluscan shells and worm tubes, but it is not clear how much of the cementation process is itself organic. The material seems to show characteristics intermediate between purely organic trottoir and inorganic beachrock. In any case, material has formed around Roman masonry and thus within the last 2000 a. Flemming found beachrock with pottery intrusions within Roman and Greek structures and quarries at Apollonia, Libya, in 1958, and at Phycus, Libya, in 1959. These observations have only been described previously in unpublished field reports. Higgins (1969) discusses the rate of formation of beachrock near archaeological sites in the Peloponnese and near Athens. He concluded that at no site is it possible to demonstrate that beachrock is forming now. Flemming *et al.* (1973 *a*, p. 13) described submerged walls of mediaeval origin at Trinisi (2) embedded in beachrock and in places almost covered by it.

In conclusion, beachrock provides interesting circumstantial evidence regarding the position of ancient shorelines, but full solution of the problem even at one site would need detailed coring and laboratory analysis requiring more time than was available during this study.

Unconsolidated coastal deposits

Coastal lagoons, deltas, infilled estuaries, beach ridges, dunes, salt flats, and marshes all provide possible means of identifying relative changes of sea level, and all these phenomena may overlap in some way with archaeological deposits or structures. Kraft (1972), Kraft, Rapp & Aschenbrenner (1975) and Kraft, Aschenbrenner & Rapp (1977) have demonstrated the value of boreholes and sedimentary analysis of recent coastal terrestrial deposits in the Peloponnese and eastern Greece, and have related their observations to archaeological remains in many cases. While the published work of other authors is taken into account during the present study, and field observations include gross data such as the growth of the Menderes delta (Snell 1963), the blocking of the lagoon at Patara (61), or the shallowness of the harbour at Salamis (163), detailed observations and coring were not attempted.

Probability of estimates

Taking into account all the foregoing notes on derivation of the best estimate of relative vertical displacement at a site, it is often preferable to express this estimate as a spread of

† Numbers after site names refer to table 1 and figures 3 and 4.

probabilities over a vertical range, rather than a single value. Since certain types of field evidence limit the range of estimates only on one side of the best estimate, the distribution of probability is often asymmetric. To avoid subjective weighting of the data for individual sites, a uniform weight of ten points was allocated, with a few exceptions where accuracy was very low. The ten weighting points were distributed to 25 cm classes on either side of the most probable value. Figure 2 shows the resulting histogram for one site, Karatoprak (51). Probability histograms were not constructed during the early studies in 1967–8. Table 1 lists the probability weightings for all sites where the method was applied. When computing best-fit polynomial surfaces to the rate of displacement of site over an area, all ten data points for each site were entered with equal weighting.

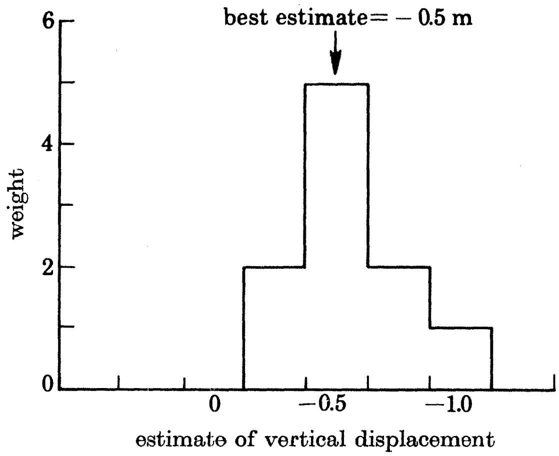

FIGURE 2. Histogram of the weighting of estimates of relative vertical displacement for a typical site, Karatoprak (51).

SUMMARY OF FIELD DATA, ARCHAEOLOGICAL

The location of all sites visited is shown in figures 3 and 4, and the results of measurements of vertical displacement relative to present sea level are shown in table 1. Sites in western Crete for which a solution notch level only has been obtained, without on-site correlation with archaeological remains, have been listed as single sites with the principal notch level recorded in table 1, although there may be up to nine levels observed on one profile (see figure 5). The correlation of sequences of solution notches is discussed separately later. Full archaeological data and references for sites in the Peloponnese and southwest Turkey have already been published (Flemming *et al.* 1973*a*), while maps and references for the remaining sites have already been drafted, and will be published shortly.

At most archaeological sites only one change of level was detected over a single time interval. At some sites, for example Ildir (34) and Plitra (3) (Flemming *et al.* 1973*a*, pp. 10, 24) two or more datable groups of structures were identified with different dates and displacements. In most cases these sites demonstrated a repetition of displacements in a consistent direction, even if the rate of displacement appeared to vary. At two sites the structures of different date appeared to produce internal contradiction, or reversals of direction.

At Matala (120) the slipway on the south side of the harbour appears to be adjusted to the correct level at present sea level, while numerous rock-cut tombs nearby are submerged by as much as 2.0 m. The simplest hypothesis is that the tombs pre-date the slipway, and that subsidence occurred between cutting the tombs and cutting the slipways. Archaeological evidence however indicates strongly that the tombs are Mediaeval, say A.D. 500–1000, while the slipway is Roman Imperial, about A.D. 100 (Blackman, personal communication). This suggests that Matala was uplifted by about 2 m after the construction of the slipway, then the tombs were cut, and then the site subsided again. Pirazzoli & Thommeret (1977) report that Loutron (127) may also have experienced reversals of displacement.

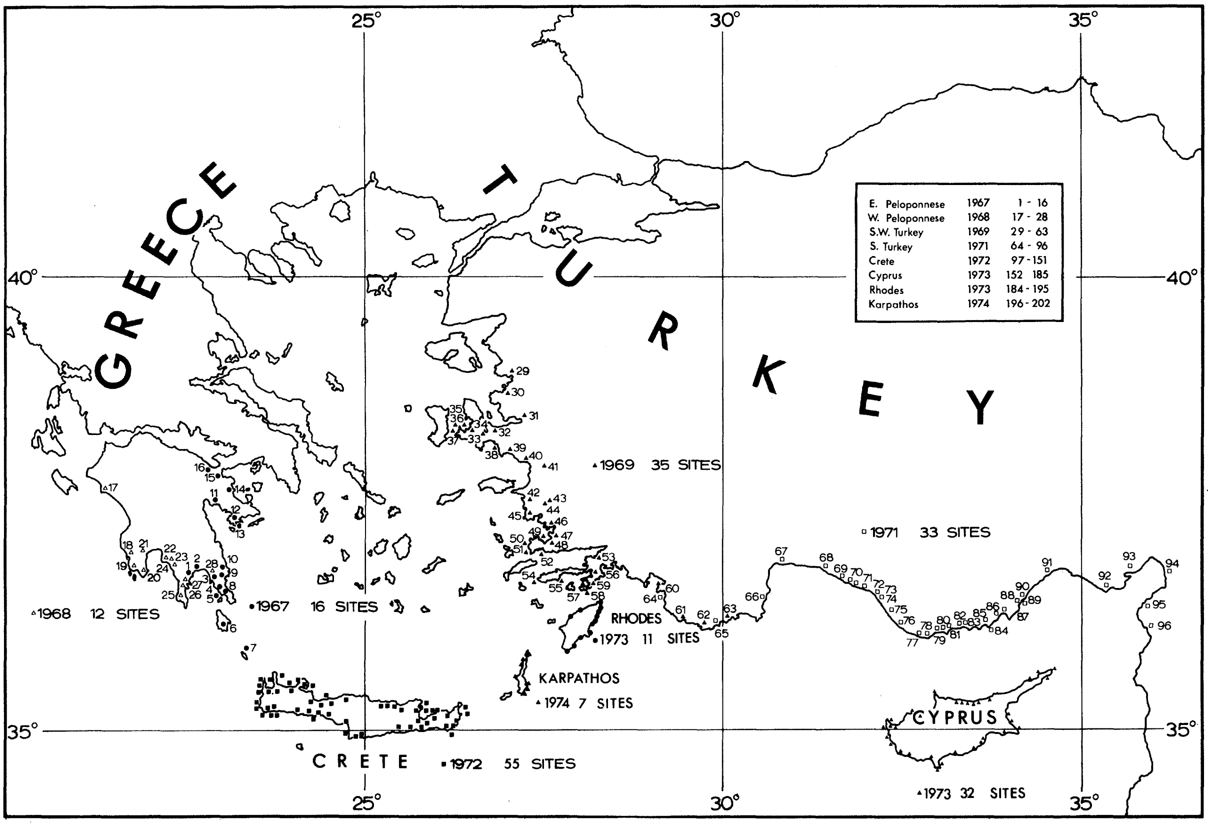

FIGURE 3. Distribution of sites for which field data were obtained. Each site number correlates with the sites listed in table 1, and represents an archaeological structure, or a single independent estimate of a solution notch level (see text). Details of Crete, Karpathos, Rhodes, and Cyprus are in figure 4.

At Kalithea (188) on Rhodes both uplift and subsidence were detected, and at Zimbule (187) subsidence was observed between two neighbouring uplifted sites. The sequence of movements here is even less clear than at Matala. However, these two sites are located at a hinge axis (see figure 22) as is Matala, and thus movements both up and down are not unlikely, though the reversal at Kalithea–Zimbule is more probably due to strain build-up between seismic events. The nature of both these hinge axes is discussed in appendix 1. (Copies of the appendix have been deposited in the archives of the Royal Society and the British Library, Lending Division.)† The fact that the only site with a reversed direction of movement of more than 50 cm is Matala, and that it lies on a major hinge axis, suggests internal consistency of the data.

† Copies of the appendix (including figures and plates therein) may be purchased from the British Library, Lending Division, Boston Spa, Wetherby, West Yorkshire LS23 7BQ, U.K. (reference SUP 10027).

PHILOSOPHICAL TRANSACTIONS OF THE ROYAL SOCIETY A MATHEMATICAL, PHYSICAL & ENGINEERING SCIENCES

Summary of field data, solution notches

Raised and submerged marine solution notches in calcareous rocks were measured on the islands of Antikythera, Crete, Karpathos, and Rhodes. The altitude, directions of movement, and chronology of these notches, and their relation to the archaeological sites, are discussed in detail in appendix 1. The multiple notches around southwest Crete are shown in figure 20, and plates 1–3; the notches on Karpathos are shown in figure 21 and plate 4; the notches on Rhodes are shown in figure 22 and plate 5. (Figures 20–22 and plates 1–5 are all contained in appendix 1.)

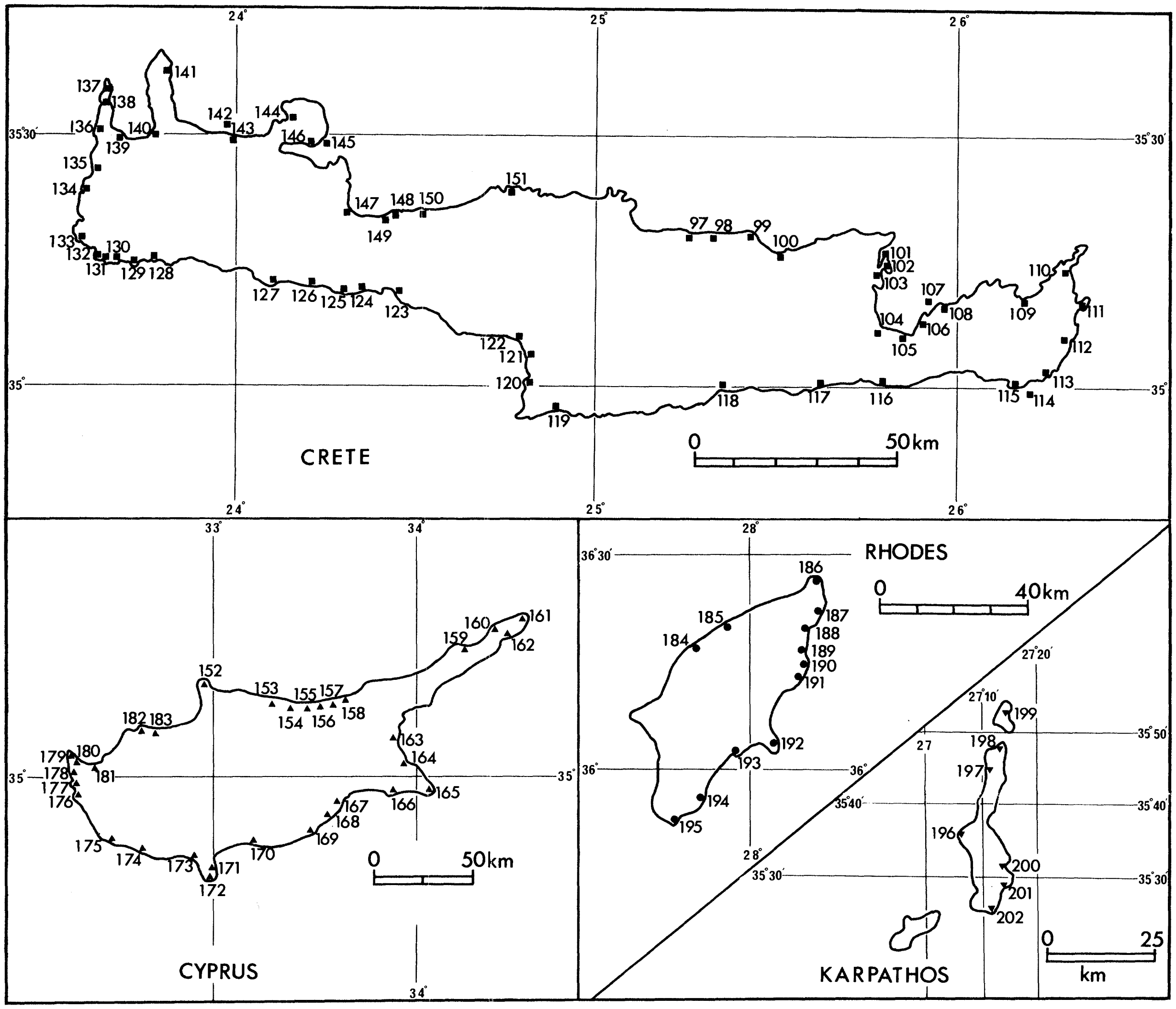

Figure 4. Site locations for Crete, Karpathos, Rhodes and Cyprus. See table 1 for age and displacement of each site.

In summary, the notches provide continuity of evidence between archaeological sites for synchronous movements over extended areas of 20–50 km. In addition, they show that these areas have moved as rigid blocks at periodic intervals, with a sequence of repeated vertical movements of approximately the same magnitude for each block. References to the literature are given in appendix 1.

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Statistical analysis of data

Introduction

The data presented in the previous section and summarized in table 1 and figures 3, 4, and 20 will be discussed in terms of their statistical significance and reliability, before relating the observations to the geology and tectonics of the region. The data immediately confirm extreme vertical instability of the coastal area and considerable regional variability. In these circumstances, simple means of ages, displacements, and rates of displacement, are of limited value. On the other hand, subjective grouping of the data into classes, or arbitrary definitions of boundaries to regions, would beg the question. Thus criteria for spatial or temporal grouping of the data must be based on preliminary deductions from the whole data set.

Parameters of the whole data set

The data from table 1 and figures 3 and 4 were combined so that every point representing an independent estimate of date and displacement was counted as a data point. Sites where two or three separate archaeological dates and displacements were found are classed as separate points; but where there are multiple notch systems and the dates for the lower notches are calculated by simply assuming a linear rate of uplift for the main notch, only one entry is made, that is, for the main notch. Some sites produced interesting data relevant to sea level determination, but not sufficient to give a numerical estimate at present. These sites are listed in table 1, and will be described in the forthcoming archaeological report, while sites which were not visited, but potentially contained useful data, are listed in table 2.

Table 3. Statistics based on treating the whole data set

	max value		min value			
	+ve	−ve	+ve	−ve	mean	standard deviation
age of sites/ka	5.0	n.a.	0.1	n.a.	1.96	1.07
magnitude of vertical displacement/m +ve = uplift	8.5	−5.0	0	0	0.01	2.14
modulus of vertical displacement/m	8.5	n.a.	0	n.a.	1.30	1.70
rate of vertical displacement/($m\ ka^{-1}$) +ve = uplift	3.9	−3.0	0	0	−0.14	1.05
modulus of rate of displacement/($m\ ka^{-1}$)	3.9	n.a.	0	n.a.	0.68	0.81

The total number of data points obtained in this way is 175 located at 158 different geographical positions. Of these 175 points, 100 indicate submergence, 43 indicate stability relative to present sea level, and 32 indicate uplift relative to present sea level. The simple statistical parameters derived from this set of data are shown in table 3. Apart from the consistently high scatter of the data, the following deductions can be made. The magnitude of uplift is greater than the magnitude of subsidence, although far more sites are submerged than uplifted. In spite of the fact that 132 out of 175 sites show vertical displacement of some magnitude, the mean displacement, taking into account the direction of movement, is only 0.01 m. The mean rate of displacement, taking into account direction of movement, is only −0.14 m/ka. (The change of sign is due to the fact that the uplifted sites are in general older than the submerged ones, and therefore contribute less to the average when corrected for age.) The relief of the coastal area is thus being increased through time while the mean altitude, relative to present sea level, is being altered very little.

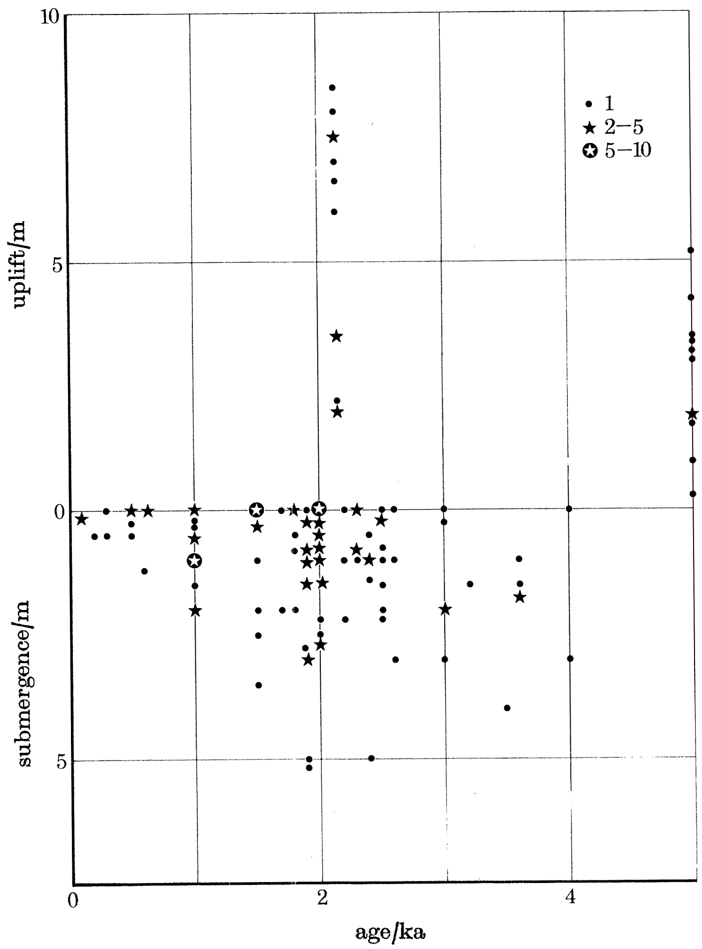

FIGURE 5. Relation between age and displacement for all sites. Black stars mark 2–5 points; hollow stars indicate more than five superimposed points.

If the displacement of sites relative to present sea level is plotted against age (figure 5) with the same data as above, the vertical scatter in each age group is seen to be extreme.

The frequency of occurrence of sites in each 25 cm ka^{-1} class of rate and direction of displacement is shown in figure 6. The distribution has some resemblance to randomness, but is heavily skewed towards a large number of sites slightly submerged, and a few sites uplifted by large amounts. The modal rate of displacement is the zero class, which can be compared with the mean displacement of 0.01 m in table 3. Both figures suggest minimal eustatic change. The paucity of field observations in the +25 cm ka^{-1} class is probably due to the difficulty of detecting uplift of the order of 20 cm. Though submergence of 20 cm may be apparent due to near-waterline structures becoming inundated, uplift of this magnitude is unlikely to produce any noticeable anomalies. Such uplifts are frequently allocated as possible in the spread of probabilities

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attributed to sites in table 1, but are not classified as the most probable estimate. Thus a small proportion of the zero class should probably be classified as slight uplift.

The skewness of figure 6 towards submergence, and the mean rate of submergence of 0.14 m/ka from table 3, combine to suggest the possibility of a net eustatic rise of sea level.

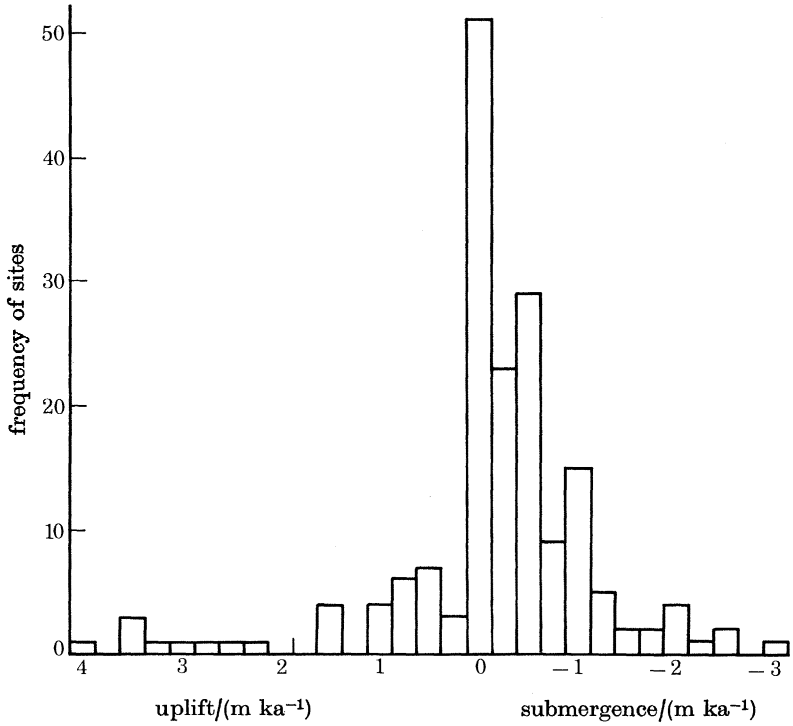

FIGURE 6. Frequency of occurrence of sites showing rate of uplift or submergence relative to present sea level, classed in intervals of 25 cm ka^{-1}. Class boundaries are 0.125–0.3749 etc.

Before a polynomial regression can be fitted to the data in terms of time and displacement, and regardless of whether the geographic-tectonic component is removed first, it is necessary to examine the data to check the reliability as a time series. Figure 7 shows the frequency of occurrence of sites in 250-year classes. The extreme peak in the class 1875–2125 B.P. (150 B.C.–A.D. 100) reflects the real prevalence of harbour construction and improvement in the early Roman Empire, combined with the numerous estimates of the notch level from western Crete. The broader concentration between 500 B.C.–A.D. 500 is also a reflexion of the high level of construction and maritime activity in classical Greece, the Hellenistic Empires, and the Roman and Byzantine Empires. The tendency for peaks to fall on whole and half millennia† is due to uncertainty in dating. The high weighting of the 2000 a class, and the very low values in many other classes means that the data can only be regarded as a reliable series if viewed in 500 a classes, and older than 2500 a even that is optimistic. This implies that the time series can only be regarded as a group of 5 or 6 independent samples, and consequently the highest degree polynomial which can be fitted in an attempt to obtain a eustatic curve is third degree.

The spatial distribution of sites may produce biased conclusions. There may be a sampling bias towards uplifted or submerged areas, and it is uncertain as to how far conclusions based on

† 1 millennium = 1 ka = 1000 years.

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coastal data can validly be applied onshore and offshore, or interpolated across channels and straits. Furthermore, it is important to determine objectively the degree of coherence between the tectonic displacement of neighbouring sites in order to define regions where polynomial regression surfaces may justifiably be fitted, and discontinuities where they may not.

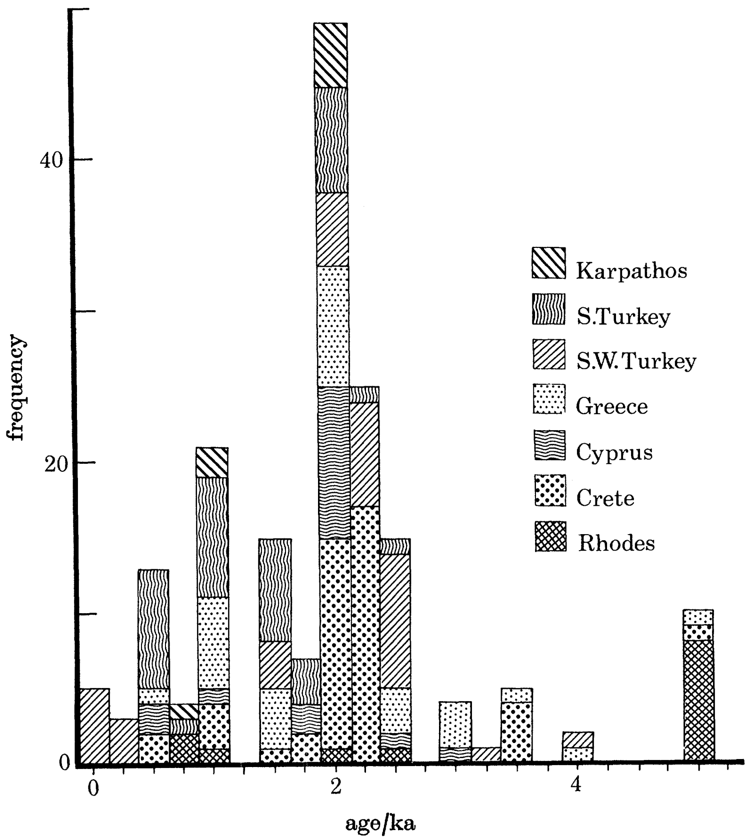

FIGURE 7. Frequency of occurrence of sites by age in class intervals of 250 a. Class boundaries, 1.875–2.1249 × 1000 a, etc., measured from A.D. 1975.

The mean spacing between sites measured in straight lines along the coast is 19 km. A 50 km grid was placed over the study area, and a convex boundary defined to include all the coastal areas studied. The number of sites per 50 km square was plotted as a frequency histogram. A test for randomness by using the Fisher dispersion index indicated a probability of randomness of less than 0.5 %, and that the sites were highly bunched. That is, no generalizations can be made about the area as a whole on the assumption that the data set is a random sample from the area, since there are too many empty squares and too many sites concentrated in just a few squares. Thus it must be accepted that conclusions from the present data are only relevant to a coastal strip on a scale of 50 km or less, and that generalizations more than 25 km onshore or offshore, or between data points separated by 50 km, are suspect.

To check for coherence of magnitude and direction of movement, the mean displacement of sites, regardless of age, was calculated for each 50 km square. This distribution is almost identical in form with that in figure 6, suggesting that on average, sites within each 50 km square tend to move coherently in the same direction. If the direction and magnitude of movement were random, each square would tend to average out with a net movement near the total mean net movement of 0.01 m. A further check was made on the distribution of gradients between adjacent sites. A network was constructed joining every site to its nearest site in a straight line. By using the data for rates of displacement at each site from table 1, the difference in rates of displacement in centimetres per thousand years was divided by the separation in kilometres, to obtain the gradient of relative movement between the sites in centimetres per kilometre per millennium.

The histogram of gradients was tested for randomness, and the Fisher dispersion index indicated that the distribution was extremely non-random and highly bunched. That is, there are far too many zero gradients and high gradients to be compatible with randomness. This is consistent with the hypothesis that the rate and direction of movement of sites is highly coherent, with a number of discrete zones of discontinuity or high deformation.

Against a background of low net eustatic sea level change, and probable coherence between adjacent sites, it is now possible to consider fitting geographical polynomial surfaces to describe regions of similar movement and their intervening zones of discontinuity.

In all surface-fitting tests two distinct trials were made, one plotting the absolute magnitude of movement, regardless of age, the other plotting the mean rate of movement obtained by dividing the magnitude by the age of the site. This was to check for the possible frequency or intermittency of earth movements in comparison with the 2–4 ka period of the whole study. It was assumed that if earth movements took place as large discrete movements widely spaced in time at intervals of 1000 a or more, each site would tend either to have moved once, or not at all, in the last 2000 a. Thus, a surface computed in terms of magnitude uncorrected for age would produce a good fit. If earth movements were smooth and continuous, or intermittent with a periodicity of less than a few hundred years, then a better fit for a polynomial of the same degree would be obtained by using the mean rates of displacement. The latter always proved to be the case, except in those areas where the data in any case were synchronous.

Peloponnese, Kythera and Antikythera

The work in this region, and the statistical treatment, has been discussed fully by Flemming (1968*a*, 1972) and Flemming *et al.* (1973*a*, *b*). Figure 8 is taken direct from Flemming (1973*b*). The contours suggest either doming of the Peloponnese, or anti-form folding along a northwest–southeast axis. There is a slightly depressed saddle-point in the region of sites 5–10 (figure 3), after which the axis rises again towards Antikythera. On either side of this there is marked depression of the Argolis peninsula and the Messenia peninsula, while the Mani appears as an isolated, faulted and depressed block. In view of the foregoing discussion, the continuity of the contours offshore is suspect, but it is at least consistent with the extremely deep depressions of the sea floor (figure 14).

Considered as an isolated set of data, the sites in this region show an overall depression averaging 0.97 m/ka, which might have been considered as evidence of a eustatic rise in sea level. This implication was not accepted because of the evidence from the western Mediterranean (Flemming 1969, p. 85) and the considerable vertical displacement between the sites of the same age. Since the data refer strictly to the coastal zone, there is no proof that the Peloponnese as a whole is subsiding. It is more probable that there is progressive folding and faulting, such that the centre of the Peloponnese, and possibly even the central axes of the peninsulae, are rising relative to present sea level. Kraft *et al.* (1975, 1977) find greater submergence for alluviated sites in the Gulfs of Lakonia and Argolis, suggesting compaction or possibly tectonic depression relative to the peninsulae.

Crete

The field data for Crete have already been discussed. The raised notch system is analysed in appendix 1. In contouring the data for Crete there was the problem of what value to attribute as the net movement or rate of movement for Matala (120). Should the displacement of this site be considered as -2 m in view of the presently submerged tombs, or zero because of the correctly aligned slipway, or even some higher value to allow for the total movement in both directions? Consequently four different surfaces of displacement were fitted for Crete, with Matala entered as zero and -2 m, and for each case the surface contoured in terms of magnitude of displacement and rate of displacement (figure 9*a–d*).

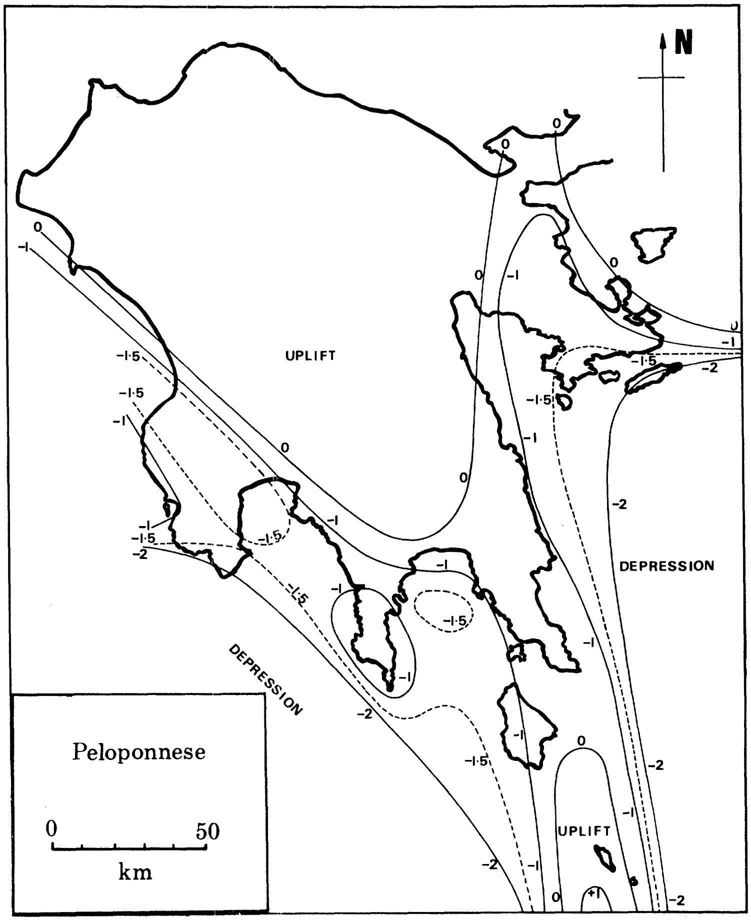

FIGURE 8. Contours of rate of vertical movement of the margin of the Peloponnese relative to present sea level in metres per millennium.

The quasi-planar uplift and tilt of western Crete is clear on all the diagrams, but when Matala is entered as -2 m (figure 9*b* and *c*) the contour lines in western Crete become much more sinuous, and residuals are increased. Since the sites in western Crete are of almost uniform age, this effect is almost identical in both diagrams. This test does not actually prove that the net displacement of Matala has been near zero over the last 2000 a, but it indicates a high

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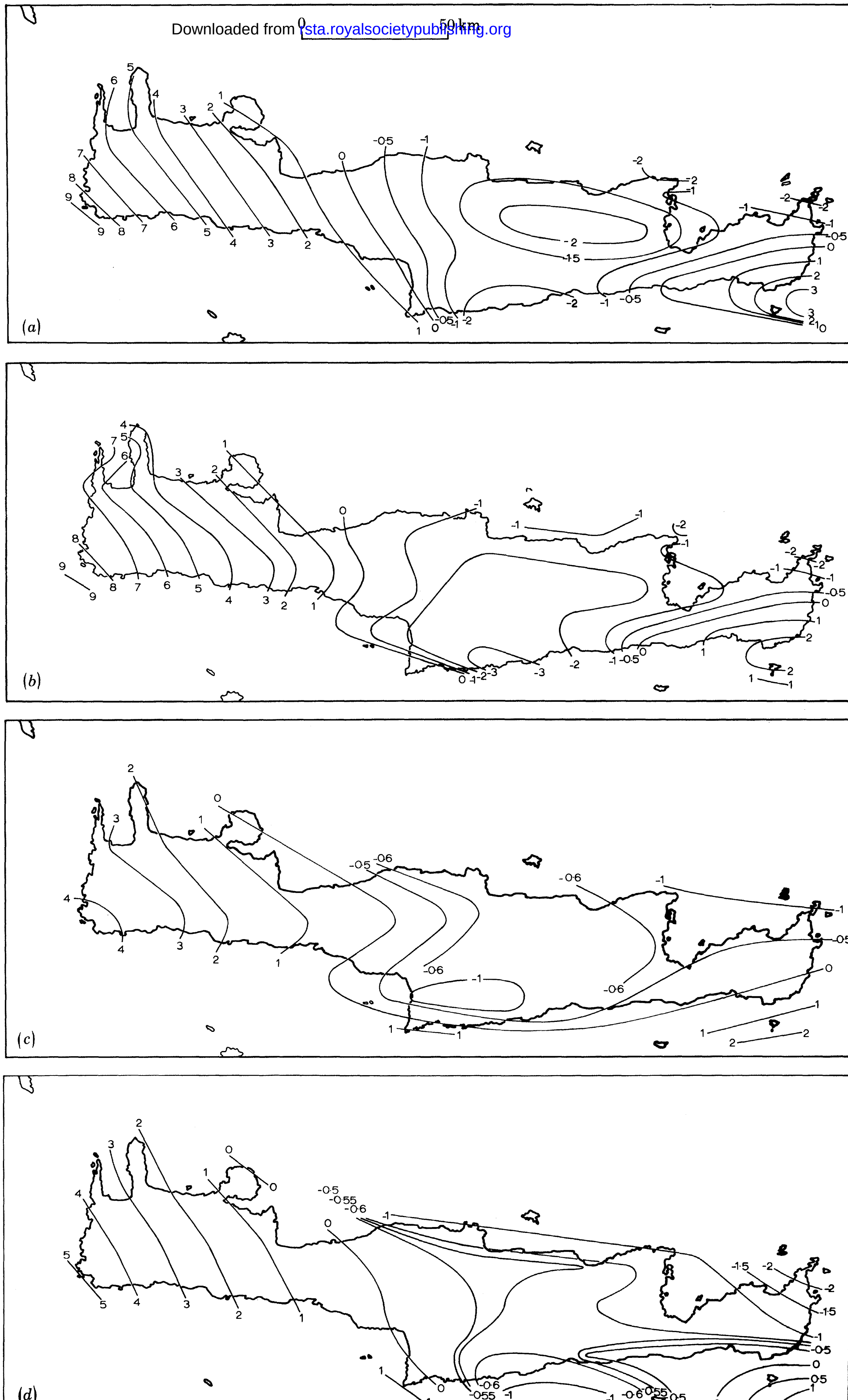

FIGURE 9 *a*, *b*, *c* and *d*. For legend see opposite.

probability. Thus, there is support for the archaeological suggestion that the site was first uplifted by about 2 m, and then subsided again.

All the contour patterns show a discontinuity at the transition from the broad central highlands of Crete to the narrow isthmus between Gournia (105) and Hierapetra (116). Eastern Crete indicates a tilting curved surface uplifted maximally in the southeast and submerged maximally in the northeast.

Karpathos

There are only seven data points from Karpathos, and they appear to indicate an almost horizontal subsidence. No statistical tests were made, but the nature of the subsidence seems completely discontinuous with either eastern Crete or with Rhodes (see appendix 1, figure 21).

Rhodes

The data from Rhodes is unsuitable for surface-fitting tests because of the almost total lack of control on the west coast, and the linear nature of the east coast.

On the partial evidence available the best interpretation of the overall form of the distortion of Rhodes is that it is composed of two plunging anticlinal or monoclinal forms with axes meeting at Afandou headland, and trending WSW and SW respectively. (See appendix 1, figure 22.)

Southwest Turkey

The data for southwest Turkey were gathered in 1969, and a first analysis published by Flemming (1972, p. 198) and Flemming *et al.* (1973 *a*, p. 58). The contours for rate of vertical displacement relative to present sea level with no eustatic correction (figure 10 *a*) show a progressive seaward subsidence of the Izmir-Cesme peninsula dipping towards the island of Khios. The area south of Ephesus (41) as far as Fethiye (60) shows undulating contours with no values outside the range 0–0.5 m/ka in the area which is actually constrained by the field data. To within the limits of the error of this study, the area is to all intents stable, but perhaps with a net average submergence of 20–30 cm, which may be attributable to eustatic rise of sea level. The demonstrable stability of this area in the last 2000 a is important in view of the large east–west faults associated with the Bodrum scarp and the Cnidos peninsula. These faults are clearly not active now and have not been active for at least 2000 a. East of Fethiye the contours show local subsidence.

The time-series for southwest Turkey (figure 7) is sufficiently evenly distributed to justify an attempt to separate a eustatic curve which is time-dependent but independent of geographical coordinates, from a tectonic component which depends on geographical coordinates, and some function of the age of the site. The method used is described in detail by Flemming (1972, pp. 194–195). It was assumed that the eustatic component could be defined by a third degree polynomial, while the vertical displacement of sites due to tectonic factors could be described by a fourth degree surface representing the rate of displacement as a function of arbitrary coordinates, and the rate of displacement was assumed to be linear in time for each site. The resulting polynomial in x and y (geographical coordinates), Z (vertical displacement) and T (time), was expanded into all its terms for each site, and the coefficients determined by a multiple

FIGURE 9. Contours of displacement of the coast of Crete. (*a*) Total vertical displacement with Matala set as zero. (*b*) Total vertical displacement with Matala set as -2.0 m. (*c*) Rate of vertical displacement in metres per thousand years with Matala set as -2.0 m. (*d*) Rate of displacement with Matala set as zero.

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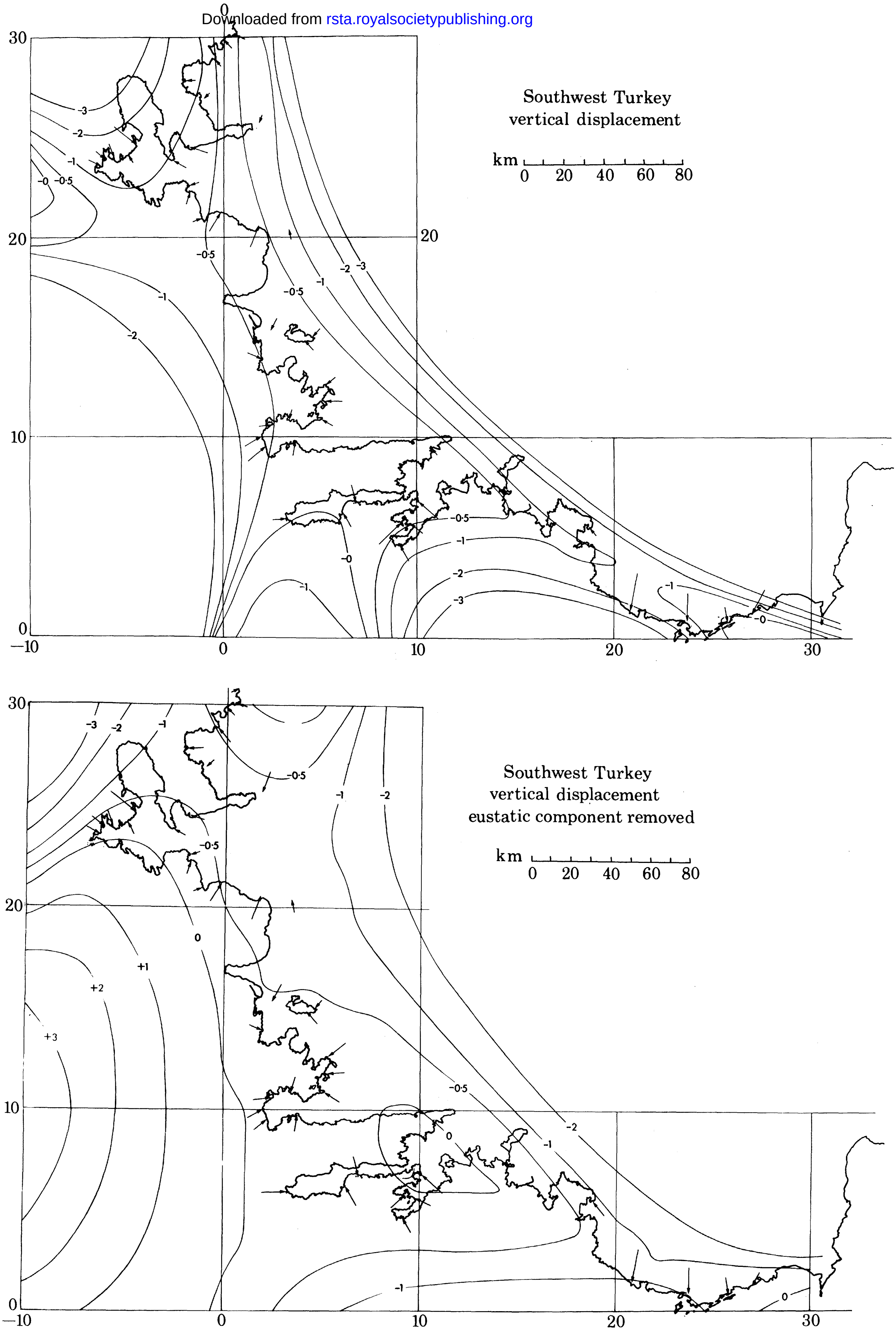

FIGURE 10. Contours of rate of displacement in metres per millennium for southwest Turkey. Arrows indicate data points. (*a*) Fourth degree contours fitted to field data. (*b*) Fourth degree contours of regional trend of rate of displacement after removal of the eustatic component.

step-wise regression. The ten values for each site were used in accordance with the probability histograms described above in figure 2 and table 1.

The equations obtained were as follows:

$$Z = 0.432 \times 10^{-1}\ T^3 - 0.317\ T^2 + 0.584T \tag{1}$$

$$\begin{aligned} Z/T = &-0.483 \times 10^{-5}x^4 + 0.224 \times 10^{-3}x^3y \\ &-0.191 \times 10^{-3}x^2y^2 \\ &-0.428 \times 10^{-4}xy^3 - 0.895 \times 10^{-5}y^4 \\ &-0.736 \times 10^{-4}x^3 \\ &-0.587 \times 10^{-2}x^2y + 0.897 \times 10^{-3}xy^2 \\ &+0.257 \times 10^{-3}y^3 \\ &+0.269 \times 10^{-2}y^2 + 0.151x - 0.661 \times 10^{-1}y \\ &+0.106, \end{aligned} \tag{2}$$

referring to the UTM grid on 1/0.5 M maps, series 1404, Turkey, with arbitrary grid origin at East 50, North 400, and x and y in units of 10 km.

It was assumed that in minimizing the residuals between the sum of the two polynomials and the field data, the regression analysis would produce two polynomials which bore a resemblance to physical reality. This is not mathematically necessary, but is quite probable. The results are shown graphically in figure 10*b* and figure 11. Figure 10*b* shows the contoured surface of rates of displacement after removal of the 'eustatic' component. The area of near zero displacement in the centre of the map is significantly increased, tending to confirm the stability of this block. Otherwise the contour trends within the area of constraint by the field data are little altered. Outside the data constraint the surface after removal of eustatic factors suggests uplift of the sea floor of the Aegean, but this is probably a purely mathematical fiction.

Figure 11 illustrates the eustatic curve obtained from equation (1), as compared with the best fit third degree curve for the data as a whole for southwest Turkey without separation of the tectonic component. The eustatic curve after removal of the tectonic component suggests a minimum sea level of −30 cm at about A.D. 700; a level at about −23 cm at 2000 B.P., and a level within a centimetre or so of present sea level in the period 3500–4000 B.P. The small net eustatic sea level change found in this study, and by Flemming (1969, p. 85) as compared with the slight progressive rise found on the eastern seaboard of the U.S.A. is probably due to the different hydroisostatic corrections which should be applied to the two areas, as suggested by Walcott (1972, p. 3, figure 1).

Figure 10*a* and *b* suggest that the southwest of Turkey could be treated as several zones: an active zone of subsidence over the Cesme peninsula; a passive zone from Kusadasi to Bodrum; an active zone including Marmaris, the Rhodes Channel, and Rhodes island; and an area of rapid subsidence from Fethiye to Gelidonya. To define the nature of each of these zones more clearly, second and third degree geographical trend surfaces were fitted to the rate of displacement for the first three zones. The Rhodes area did not produce a good fit, indicating only that there was a discontinuity between the island and the mainland. The Fethiye–Gelidonya area does not contain enough data points to provide a significant trend. The plots for the other two areas are included on figure 13*a*.

South Turkey

South Turkey from Gemili (64) to Phaselis (66) shows subsidence. The south coast from Kas (62) to Kale (63) shows a uniform depression of almost exactly 1.0 m/ka. A general interpretation of the distortion of the block is made difficult by the apparent high rate of subsidence of Gemili (64) of 2.0 m/ka, but the dating of this site is very uncertain (see Flemming *et al.* 1973 *b*, p. 55).

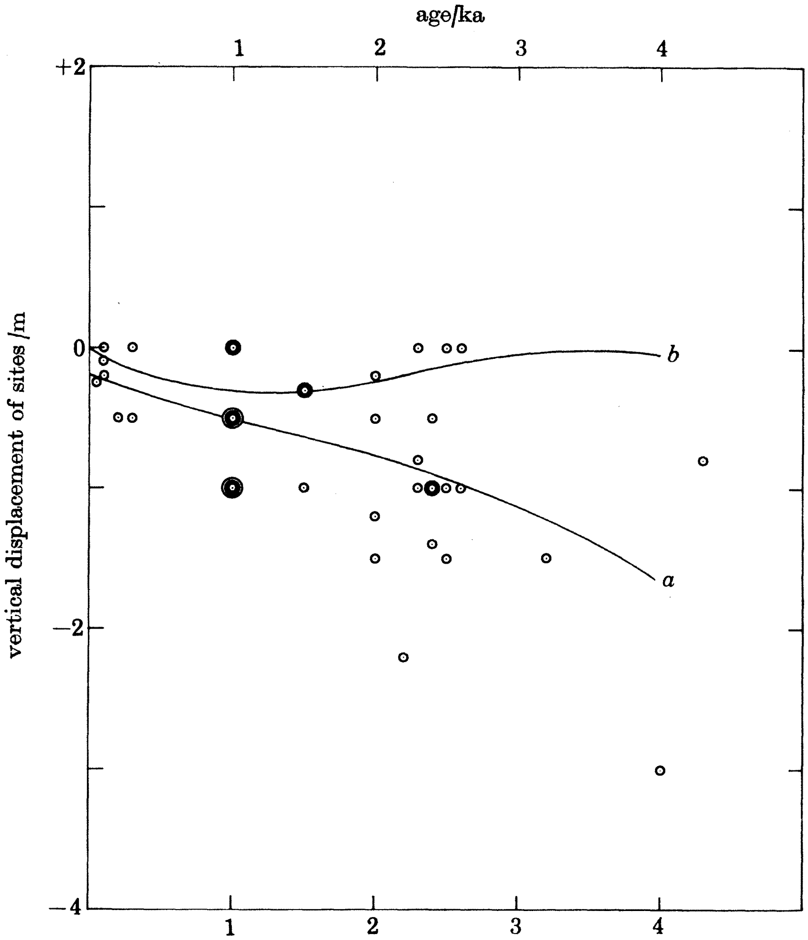

FIGURE 11. Comparison of (*a*) cubic trend fitted to total data of displacement and age of sites for southwest Turkey, with eustatic curve (*b*), obtained after removal of regional tectonic component.

The data set for south Turkey from Fethiye (60) to Seleucia Pieria (96) was contoured for rate of displacement, and the ten probability estimates for each site were used. East of 31° E the coast appears to be stable to within ±25 cm relative to present sea level. West of 31° E the contours perform a complex turn to accommodate the subsidence of the Fethiye–Gelidonya block.

Cyprus

The contoured surface of rate of displacement for Cyprus (figure 12) indicates clearly the stability of the north coast relative to present sea level. The contours for the rest of the coast

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form no simple pattern, and merely indicate general submergence of the margin of the island. The zero contour on the southeast coast is suspect, since there are no data control points between Amathus (170) and Dades (166).

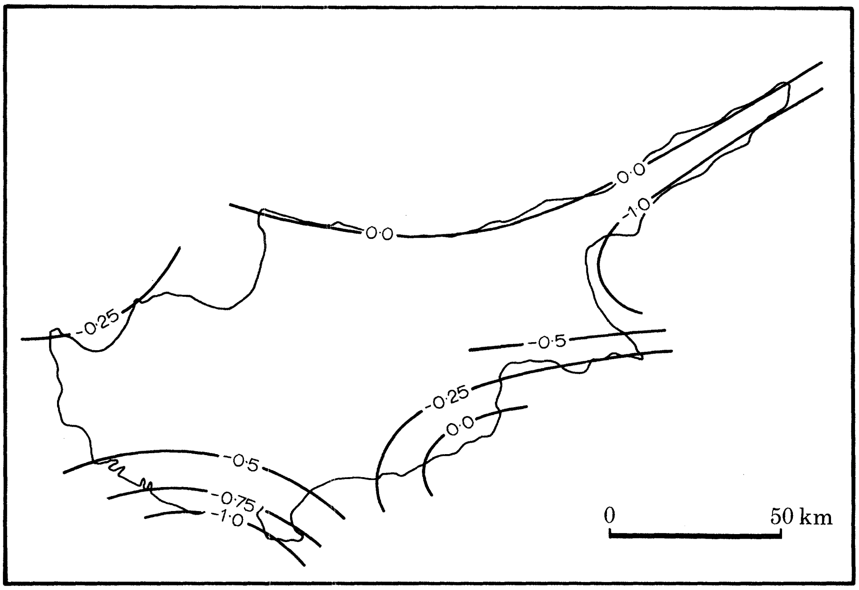

FIGURE 12. Contours of rate of vertical displacement of the coast of Cyprus in metres per millennium.

Conclusions

The regional best fit contoured surfaces of rate of displacement are plotted together in figure 13*a*, and the principal boundaries indicating discontinuity between blocks of distinct character are shown in figure 13*b*. Apart from the discontinuous junctions between blocks with internally coherent characteristics, a striking feature of figure 13 is the stability of southern Turkey and northern Cyprus. The distance between the two is 100 km, and hence the present statistical methods do not in anyway confirm the stability of the sea floor intervening. The geological implications of these results will be discussed in the next section.

DISCUSSION

Introduction

The summary of the data analysis provides a clear picture of the recent earth movements in the coastal zones. The nature of these movements, and the distribution of stable, tilting, and warped blocks, is sufficiently distinctive to warrant a unique explanation in terms of the geological history of the region, and the current stress pattern in the crust produced by the relative motions between the African and Eurasian plates. As a prefatory caution it should be noted that the movements detected in the present study refer only to the most superficial strata, and these movements are only a distant expression of the movements 10–100 km deep in the crust. However, the fact that some blocks of superficial strata behave coherently over 50–100 km suggests that the depth to which such blocks extend as undistorted rigid units is of a similar scale, and thus reasonable inferences can be drawn.

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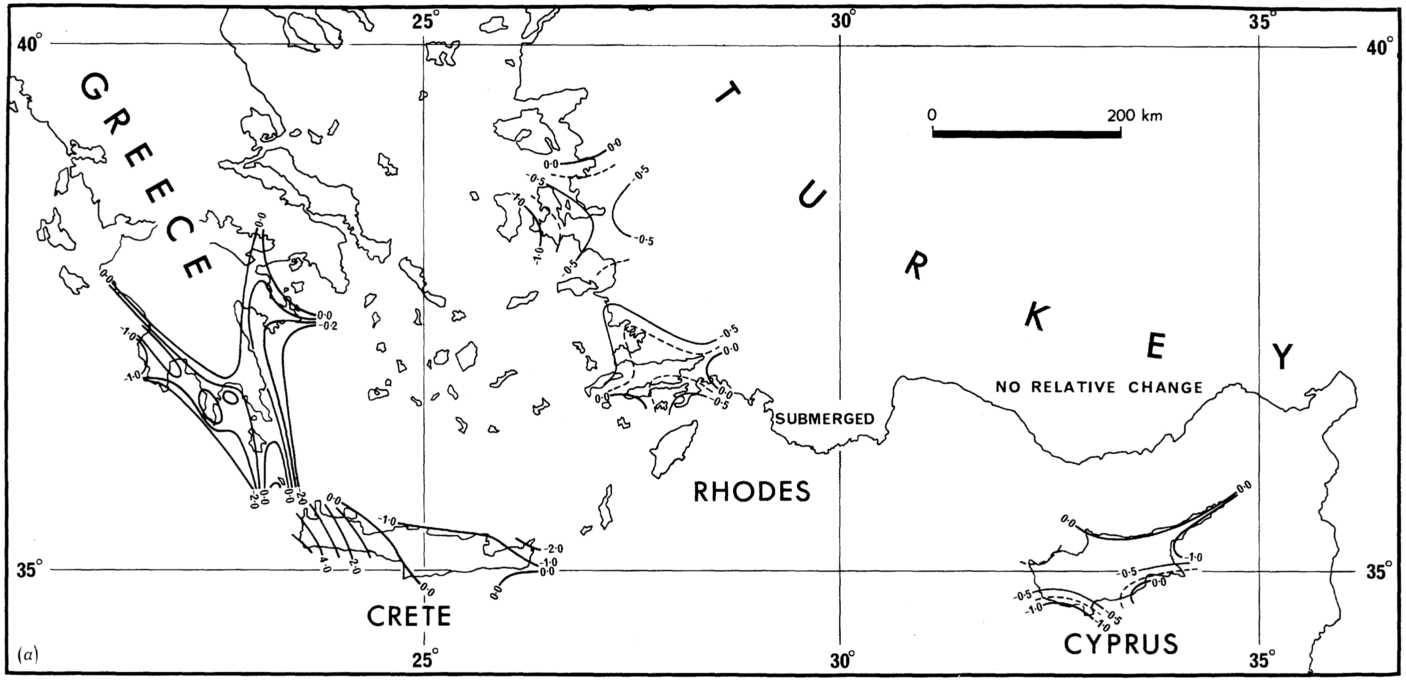

FIGURE 13. (*a*) Synthesis of regional contours of rate of vertical displacement in metres per millennium relative to present sea level. (*b*) Boundaries of coherent blocks and directions of tilt. The arrows in deep water south of the Hellenic Arc and Turkey are derived from Ryan *et al.* (1973) and Rabinowitz & Ryan (1970) and Woodside (written communication). Depths in metres.

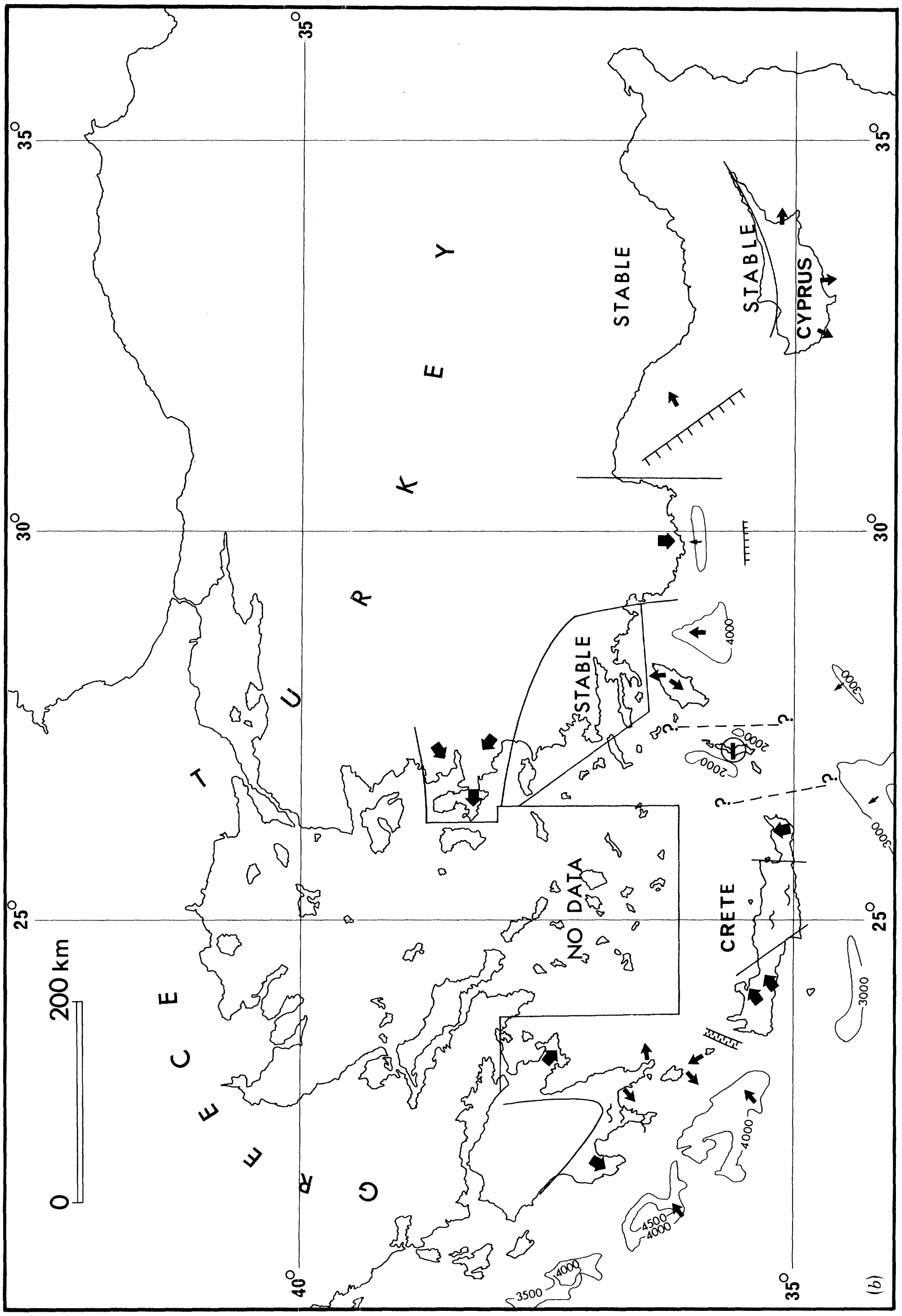
TURKEY
GREECE
STABLE
STABLE
STABLE
CYPRUS
CRETE
NO DATA
200 km
0
35°
30°
25°
40°
35°
2000
3000
3500
4000
4500
(b)

FIGURE (13b)

The present data refer to movements with a periodicity of 100–1000 a over the last 5000 a. The best correlation should be found with historical evidence for seismicity over the same period; less good correlation should occur with instrumental records of the last few decades; and only general correlation with movements deduced from geological structure, topography, and geophysical parameters such as gravity, crustal velocity profiles, and magnetism, which produce averages for millions of years. However, these general factors should be considered briefly in order to define independent boundary conditions to the present study.

Geological and tectonic background

Geological maps have been consulted as follows: Greece, Renz, Liatsikas & Paraskevaidis (1954); Crete, sheets of the 1/50000 series of the geological map of Greece published by the Institute for Geology and Subsurface Research; Turkey, sheets of the 1/500000 series edited by Pamir & Erentöz (1963); Rhodes (Desio 1931); Cyprus, 1/250000, the Geological Survey of Cyprus (1963). A general tectonic map of the land areas was prepared by Bogdanoff, Mouratov & Schatsky (1964), and this has been modified with the later results of marine seismic findings by Biju-Duval (1974). The most comprehensive field work and synthesis of the geological and tectonic data for Greece and the islands of the Hellenic Arc has been carried out by Aubouin and his co-workers (Aubouin 1958, 1965; Aubouin *et al.* 1963; Aubouin & Dercourt 1965, 1970). Brunn *et al.* (1971) provided a tectonic analysis of southern and southwest Turkey. The bulk of this research was completed before 1963, and hence is interpreted in terms of a geosynclinal sequence rather than the more recent concepts of plate tectonics. However, Kay (1951) and Dewey & Bird (1970) show that for each subdivision of the geosynclinal terminology there is an equivalent dynamic stage or process in the evolution of continental margins according to plate tectonic theory. Thus there is no inherent contradiction in drawing equally from work carried out within the two different frames of reference.

Aubouin (1958, 1965) describes the evolution of the geosyncline couple since the Triassic solely in terms of crustal shortening, but Smith (1971) reports evidence from central Greece indicating crustal extension and the formation of ophiolites followed by shortening. This confirms the possibility that the geosynclines and nappes described by Aubouin and others may be explained in terms of continental rifting, mid-ocean spreading in the area which is now the Mediterranean, embryo-oceans, island arc formation, back-arc spreading, trenches, and finally continent-continent collisions. In a very general way this has already been attempted by Dewey *et al.* (1973). Dewey *et al.* (1973) propose that 24 small blocks or plates in the Tethys–Mediterranean area were in continuous motion relative to each other, and to the African and Eurasian plates, during the varying rotation and movements of the African plate relative to Eurasia over the last 180 Ma. Each of these small plates were at various times bounded by spreading zones, transform faults, and zones of convergence or overthrusting. Angelier (1976, 1977) combines the geosynclinal sequence proposed by Aubouin with plate movements and fault analysis to construct a Plio-Quaternary model of the south Aegean involving phases of compression and extension, with an overall southward extension or radial expansion of the Hellenic Arc. Brunn *et al.* (1971, p. 251) envisage that in the middle Miocene a phase of compression in the SE–NW direction acted on the previously gently curved arc system so as to fold the Hellenic Arc back on itself, while having less effect on the eastern Taurus range. This is consistent with a drastic decrease in the radius of curvature eastwards along the Hellenic Arc producing a 'knot' in the region of Isparta. The 'knot' may then have become sutured to the main Turkish plate along the Antalya–Isparta–Izmir line. This can be compared with the Marmara region (Crampin & Ucer 1975, pp. 270–271) in northwest Turkey. Magnetic and profiling data (see below) indicate a quasi-linear anomaly extending from the Antalya–Gelidonya region directly southeast towards western Cyprus and it has been suggested by Biju-Duval *et al.* (1974) that the ophiolite belt is actually continuous with the Troodos complex in Cyprus. This is unlikely, since the Troodos Mountains are dated as Upper Cretaceous (Vine *et al.* 1973, p. 38) as opposed to the Triassic or Upper Jurassic age for the Antalya ophiolites (Dewey *et al.* 1973, fig. 5). It is more probable that some other discontinuity is responsible for the anomalies, and other authors (McKenzie 1970; Lort & Gray 1974) indicate that the convergence zone of the present-day plate boundary between Africa and Turkey must occupy more or less this line.

The origin and mode of emplacement of the Troodos ophiolite series is a matter for much discussion (Gass & Masson-Smith 1963; Gass 1968; Miyashiro 1973; Vine *et al.* 1973). Dewey *et al.* (1973, p. 3166, fig. 15) invoked a short spreading zone with a SW–NE axis to produce oceanic crust just in the region of Cyprus during the Late Cretaceous. This has the virtue of separating the Cyprus ophiolites from the rest of the Hellenides, and indeed the same type of process could be suggested as the origin of the Antalya ophiolites outside the main geosynclinal-island arc system.

Topography and Quaternary–Holocene faulting

The bathymetry of the northeast Mediterranean and the main tectonic trends are shown in figure 14. The Hellenic Arc has been referred to as an 'island arc' (Allan *et al.* 1964; Ninkovich & Heezen 1965; Nicholls 1971; Caputo *et al.* 1972; McKenzie 1972; Galanopoulos 1973; Papazachos 1973) and the combination of a concave seismically active belt with active volcanicity on the concave side, certainly fits the principal criteria. However, the most striking topographic characteristic of the Aegean is not smooth curving arcs, but rather the linearity and angularity of all the features in the area, including the so-called Hellenic Arc (Angelier 1976, 1977).

The Hellenic Trough extends in a NW–SE trend from the Ionian Sea to a point about 100 km south of the mid-point of Crete in an almost perfect straight line. If any curvature can be detected, the radius must be several thousand kilometres rather than the few hundred suggested for the Hellenic Arc as a unit. The deepest axis of the Trough is broken in several places by sub-rectangular offsets and blocks, and south of Crete this develops into a double trench system, the Pliny Trench and Strabo Trench, separated by the Ptolemy Mountains. Two rhomboid banks appear south of Crete bearing the islands Gavdhos and Nisos Khrisi. From this point the double trough trends due northwest to the Rhodes Abyssal Plain and the Gulf of Fethiye. Where the Hellenic Arc joins southwest Turkey all semblance of arcuate structure ends abruptly against the digitated peninsulae of Cnidos and Marmaris, and the highlands of the nappe system between Fethiye and Cape Gelidonya. The coast from Gelidonya north to Antalya is exceptionally straight and steep, and is one of the most striking topographic features of the region, both on the map and on the ground.

The islands of the Hellenic Arc are themselves bounded by linear features, and internally divided by rhomboid conjugate fault patterns (Aubouin & Dercourt 1965, 1970; Galanopoulos 1967; Angelier 1977). The coast of Crete is dominated by long east–west coastal segments interspersed with abrupt north–south offsets, and long north–south peninsulae. Karpathos trends directly north–south, with extensive rhomboid conjugate faulting on a NNW–SSE and NNE–

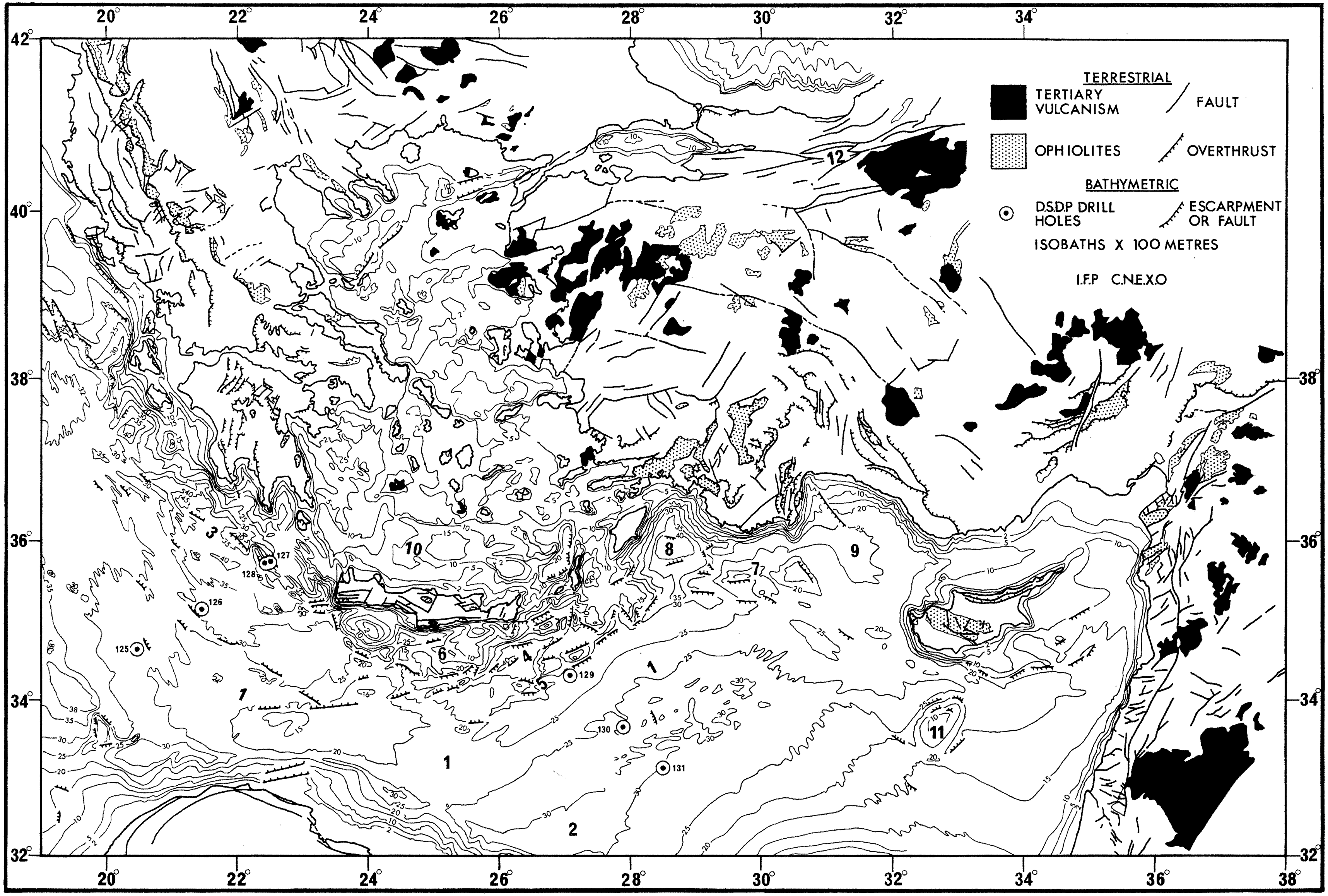

FIGURE 14. Bathymetry and tectonic trends of the Eastern Mediterranean after Biju-Duval (1974). 1, Mediterranean Ridge; 2, Herodotus Abyssal Plain; 3, Hellenic Trough; 4, Strabo Trench; 5, Pliny Trench; 6, Ptolemy Mountains; 7, Anaximander Mountains; 8, Rhodes Abyssal Plain; 9, Antalya Abyssal Plain; 10, Cretan Furrow; 11, Eratosthenes Seamount. DSDP Leg 13 drilling sites are shown with the hole numbers.

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SSW pattern. Rhodes continues the NE–SW trend, and borders the Rhodes Abyssal Basin with its remarkable triangular form.

Aubouin & Dercourt (1970, p. 467) describe the massifs of Rhodes as horsts isolated from the surrounding deposits by faults; Crete is described as a horst (Aubouin & Dercourt 1970, p. 455), while the location of Karpathos as an isolated, almost stable block bordered by deep basins is similarly horst-like. In all these cases the intensive fault patterns within the horst blocks are seen to be currently non-active, and the present data show that the blocks are now tilting or moving as rigid units. Aubouin & Dercourt (1970, p. 470) conclude that the currently active fault pattern observed in the south Aegean is completely independent of the previous tectonic stresses which created the nappes. The post-Miocene pattern cuts across the previous structural trends, especially in Crete, and the near coincidence with the structural trend of the two limbs of the Hellenic Arc is a coincidence. The present fault system is seen as consistent with N–S extension of the Aegean (Aubouin & Dercourt 1970, p. 470), although E–W extension may also be present. Zarudzki (unpublished seismic reflexion profile) has detected a graben-form in the Antikythera channel which could be associated with both the southward movement of Crete relative to Antikythera, and the less obvious E–W dilation of the centre of the Hellenic Arc.

Outside the Hellenic Arc, Rabinowitz & Ryan (1970, p. 604) describe a series of folds and overthrusts with a NE–SW strike, parallel to the eastern limb of the Hellenic Trough. Rabinowitz & Ryan (1970, p. 606) interpret the various compressive features of the intrabasinal deformation in the east Mediterranean and in Cyprus, as nappes in the process of formation. In order to provide estimates of the probable directions of movement and tilting of the areas immediately adjacent to the islands data were collated from several sources (Giermann 1966; Wong, Zarudzki, Philips & Gierman 1971; Sancho *et al.* 1973; Lort & Gray 1974; Stride, Belderson & Kenyon 1977; Woodside, personal written communication and unpublished reflexion profiles), and plotted qualitatively as shown in figure 13*b*.

The sediments between Cyprus and Turkey are only very slightly foreshortened and folded, but with massive thickening towards the Turkish coast, especially into the Gulf of Iskenderun (Wong & Zarudzki 1969) (Woodside, written communication).

Bore-holes 127–129 of the Joint Oceanographic Institutions Deep Earth Sampling (JOIDES) program are shown on figure 14. Seismic profiling in preparation for Hole 127 showed sediment conformations in the Hellenic Trough indicating tilting of the sediments in the basin in a NE direction at a rate of 1°/Ma (Ryan *et al.* 1973, p. 243, and pp. 270–272).

Gravity and magnetic fields

Gravity anomaly maps for the Aegean have been published by Allan & Morelli (1971), for the eastern Mediterranean by Rabinowitz & Ryan (1970), and by Cambridge University for the Cyprus–Levant area (in the press). The implications for the interpretation of the structure south of the Hellenic Arc are summarized by Rabinowitz & Ryan (1970). The Hellenic Trough is associated with a pronounced gravity low, and the linear trend with a right-angle bend south of Crete is emphasized by the northern margin of the anomaly. The eastern flank shows a double low anomaly, correlating with the double trough. The double anomaly continues eastwards, across the area of the Anaximander Mountains, and towards southwest Cyprus. This is consistent with the previous discussion of the complex form of the discontinuity in this region (Matsuda & Uyeda 1971). Woodside & Bowin (1970) and Harrison (1955) interpret the inverse Bouger anomaly–topography relation in the south Aegean as due to uplifting of the mantle and thinning of the crust. The gravity highs on Cyprus and other topographic highs are interpreted by Rabinowitz & Ryan (1970) as due to mantle material being incorporated within uplifted blocks during nappe formation.

Magnetic anomalies for the northeast Mediterranean have been published by Vogt & Higgs (1969). The Mediterranean Ridge and the trench system of the Hellenic Trough are magnetically quiet, and therefore presumably associated with sedimentary rocks. The Cretan Furrow is fairly quiet, with one linear positive anomaly trending east of north, and terminating at the active volcanic island of Santorini. Fisher & Hess (1963) suggest that crustal fissures open on the concave side of island arcs, normal to the trend of the arc, and lead to volcanic activity. Vogt & Higgs (1969, p. 440) indicate that the magnetic anomaly trends in the north Aegean are probably due to a continuation of the Mesozoic granites of northwest Turkey under that part of the sea. In any case, there is a characteristic difference between the north and south Aegean basins.

The positive magnetic anomaly over Cyprus continues clearly towards the northern corner of the Bay of Antalya. This trend lies on the northeast margin of the previously discussed gravity and profiling anomalies. Biju-Duval *et al.* (1974, p. 714) interpret the magnetic anomalies as a direct continuation of the Cyprus ophiolite series, with possible continuity with the Antalya ophiolites described by Brunn *et al.* (1971). Caution about this has already been expressed.

Correlation of the present data with recent seismicity, volcanicity and fault mechanisms

The previous sections have outlined the geological complexity of the area, so that it is possible to correlate the observed vertical earth movements and historical seismicity, volcanicity, and present fault mechanisms. Figure 13*b* summarizes the principal tectonic units defined by the present study, with their directions of tilt. Figure 15 illustrates in approximate form the intermittent rates of movement for the four units which have been demonstrated to be moving vertically in a stick-slip manner. Since the observations of Reid in 1906 (cited by Benioff 1964) intermittent fault movements have been analysed in terms of aseismic strain accumulation alternating with periods of rapid slip releasing seismic energy (see for example, Burridge & Knopoff 1967; Okada 1970; Fitch & Scholz 1971; Lensen 1974). Scholz (1972, p. 203) suggests a model for underthrusting at plate boundaries on this basis, and goes on to indicate that the recurrence of seismic slippage will have a frequency of the order of 100 a. The periods of movement in figure 15 are 714, 333, 217, and 157 a with an average of 355 a. This is longer than that suggested by Scholz (1972), but is heavily weighted by the data for Rhodes, which appears to have an exceptionally long period.

The magnitudes of movement in metres per millennium for the various blocks can be converted to rates of tilt in seconds of arc per millennium as follows: West Crete, 10.6″/ka; East Crete, 10.3; South Rhodes, 3.6; North Rhodes, 9.0. These rates are the averages over 2–5 ka, and do not take into account intermittent accelerations or reversals. Lensen (1974) finds that tilt rates on the south coast of Shikoku, Japan, are of the order of 3″ over the last 100 a (equivalent to 30″/ka) but average only 9″/ka over the last 400 a, and over the last 110 ka. Lensen (1974, table 3) summarizes the work of several authors to show that major seismic earth movements in various regions of the circum-Pacific belt and Alpine chain occur at periodicities of the order of 500–1000 a within each region. He concludes that while available data are mostly for regions in horizontal shear, the periodicity in tensional regions may be as low as a few hundred years, and in compressional regions as long as 100 ka.

Karnik (1972) reports that for shocks greater than $m = 6.25$, activity at the beginning of the twentieth century A.D. was concentrated at depths of 150–200 km near Crete, and then in 1925–6 moved out to the east and west limbs of the arc. From 1935–50 activity was concentrated at depths of 60–100 km, and after 1950 there was an increase in shallow crustal activity. Periodic seismic movement of different islands of the Hellenic Arc could therefore be associated with migration of dislocations either up and down the subducted slab, or along the limbs of the arc.

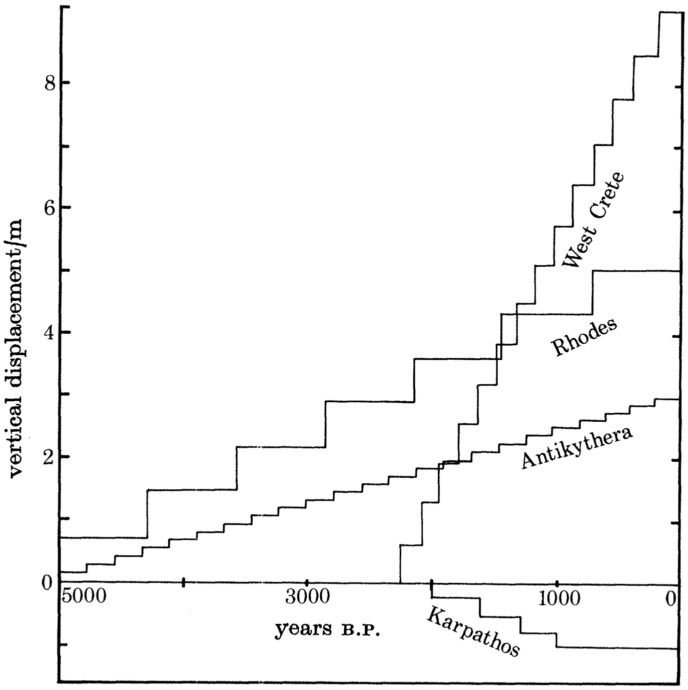

FIGURE 15. Simplified diagram for intermittent vertical movement of points of most rapid displacement in the islands of the Hellenic Arc. Displacement of each site is plotted in steps of vertical magnitude and time duration equal to the mean obtained by dividing the total displacement and age into equal steps. The number of steps at each site is determined from solution notches. See text for age estimates, appendix 1.

Ambraseys (1971) describes historically recorded earthquakes at Istanbul during the last 2000 a. Major earthquakes occurred at intervals of approximately 200 a, varying from 200 to 350 a. This relates to right lateral movement on the Anatolian Fault, which manifests a generally periodic peak of seismic activity at intervals of 150 a (Ambraseys 1970). Ambraseys further compared the overall seismicity of the lateral movements on the Anatolian Fault and the so-called Border Zone (between the Dead Sea Rift and the eastern end of the Anatolian Fault). The comparison is shown in the inset in figure 16. Ambraseys (1971, p. 379) comments that for the first five centuries the Border Zone is quiet, while the Anatolian Fault is active; then from A.D. 500–1100 the Border Zone is active and the Anatolian Fault quiet; from A.D. 1100–1300 there is activity on both fronts; after A.D. 1300 the Border Zone is quiet, while the Anatolian Fault continues to be active. Periodicity on this scale seems to vary from 500 to 800 a.

The periodicity of uplift and tilting of the islands in the Hellenic Arc is intermediate between the simple stick-slip periodicity predicted by Scholz (1972), and the implied periodic movement of Turkey suggested by Ambraseys (1971). The size of the coherent tilting blocks in the present

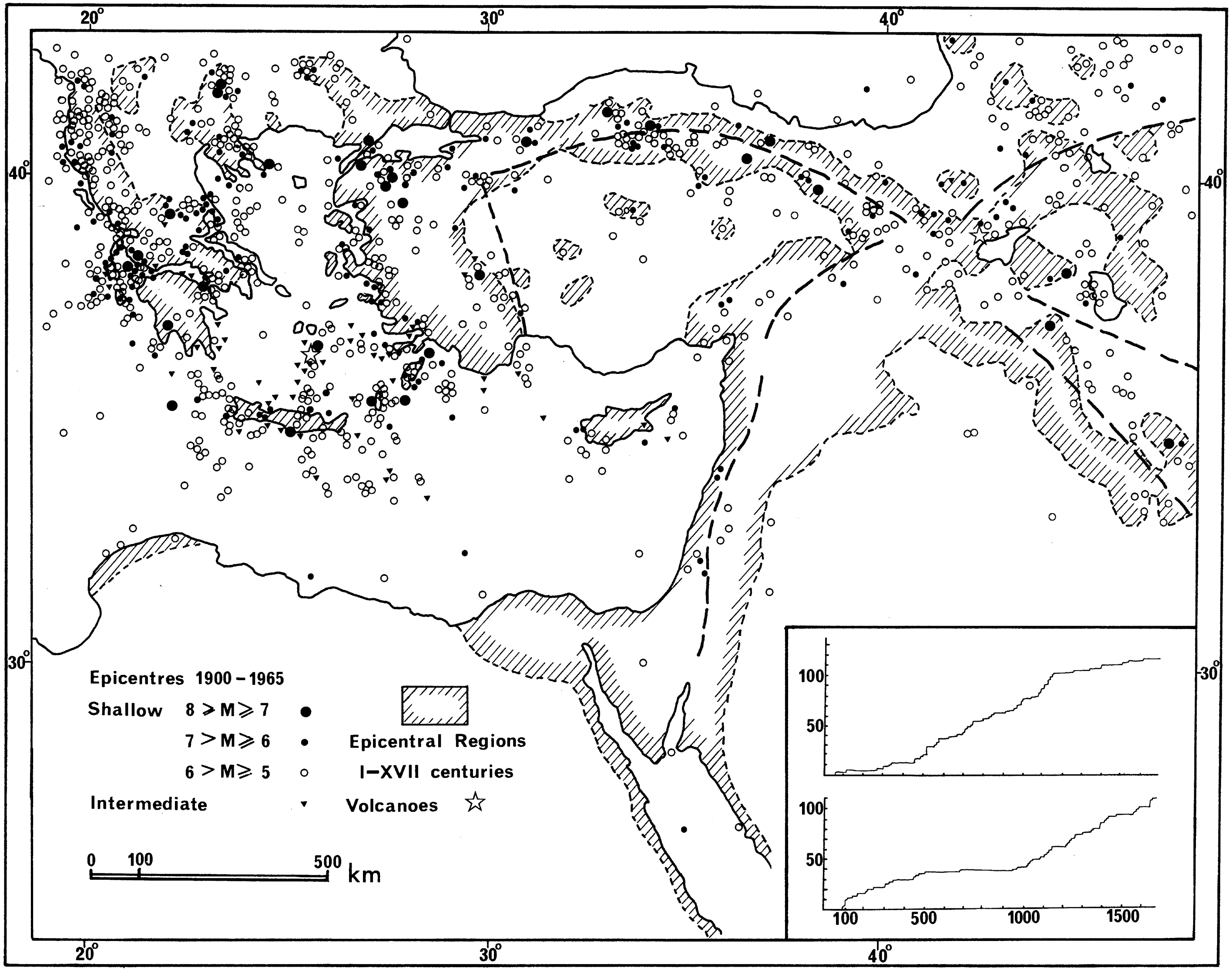

FIGURE 16. Historical and recent seismicity of the Eastern Mediterranean and Middle East, from Ambraseys (1971). The heavy dashed lines indicate principal trends of fault zones presumed to be plate boundaries. The north–south line in western Turkey is hypothetical. Inset, top: number of damaging earthquakes on the Border Zone; bottom: number of damaging earthquakes on the Anatolian Fault Zone.

study is consistent with thickness of the order of 50–100 km, and therefore it can be assumed that their movement is a fairly close reflexion of subduction processes of the crust below the Hellenic Arc.

Ambraseys (1971) shows that in terms of broad areas of seismicity the modern records of the last 50 years correspond well with historical data of the last 2000 a, except for the fact that the modern records show low values in the Red Sea–Dead Sea Rift area (Ben-Menahem, Abodi, Vered & Kovach 1977). Comparison of figure 13*b* and figure 16 shows that the stability of the south Turkish coast is confirmed both for the historical period of the last 2000 a, and for recent instrumental records of seismicity. The shaded area in figure 16 is so generalized elsewhere that no direct correlations can be made with the boundaries suggested in figure 13*b*. However, Ambraseys (1970, fig. 2) plots a narrow belt of historic epicentres from Cape Gelidonya, up towards Isparta, and then curving across to Izmir.

The relative stability of North Cyprus (figure 16) is confirmed historically by Ambraseys (1961). An arc of recent seismicity curves from a point seawards of Gelidonya, through Antalya, and then northwest of Izmir to join the Anatolian Fault near the Bosphorus. This line is east and north of the line suggested by Ambraseys (1970, fig. 2) for historic seismicity, but almost parallel to it. The large number of shallow, low magnitude epicentres around south-west Crete correlates with the frequent movement in that area. An oddity of the recent seismicity (figure 16) is the two radial lines of high seismicity, one stretching west of south from Santorini towards central Crete, the other almost due east from just north of Santorini towards Cnidos. The southerly lineation coincides almost exactly with the magnetic anomaly described by Vogt & Higgs (1969). There is a suggestion here of radial weakness of the crust in the back-arc area.

Profiles of hypocentre distribution in the Aegean area have been compiled by Caputo *et al.* (1970), and in more detail by Papazachos (1973) and Galanopoulos (1973). The following discussion shows the extreme difficulty of applying simple concepts of plate tectonics, Benioff zone, subduction, etc. to the south Aegean area. Data from the present study does not fit at all well with the suggestion that there is a single subduction zone dipping northeast as suggested by Caputo *et al.* (1970) and Nicholls (1971, p. 379). Galanopoulos (1973) assumes the plate boundaries suggested by McKenzie (1972) and admits that the primary problem, given the enormous scatter of hypocentres, is to establish the correct plane and dip of the Benioff zone. He concludes that the zone is dipping 35° in a direction north–58° east, but states clearly that this conclusion is speculative. He states that the data are extremely confusing, and that the dipping slab may be torn into two segments as it descends (Galanopoulos 1973, p. 102).

The tearing of descending crustal slabs has been discussed by Isacks & Molnar (1971, p. 123) and Abe (1972) in connection with the Aleutian island arc, and in general principle by Frank (1968). Frank (1968) pointed out that, given an island arc whose radius subtends an angle θ at the centre of the Earth, the angle of descent of the crust which permits the crust to be neither torn nor compressed is necessarily 2θ. Steeper angles create tearing, shallower angles compression. Since the radius of the Hellenic Arc, considered as a true island arc, would be somewhere between 200 and 500 km, the dip of the Benioff zone to maintain a constant area of the dipping crust would be from 5° to 9°. This is obviously absurd. Although the angle and location of the subduction zone is in dispute, the range of angles considered has always been between 20° and 35°. Clearly, any crustal slabs subducted around the perimeter of the Hellenic Arc must be torn very considerably.

Assuming that the northern boundary of the African plate is indeed the deepest part of the

Hellenic Trough, continuing eastward via the Anaximander Mountains to Cyprus, let us consider the tearing and/or compression which must result from a slab dipping at an angle of 30° along this strike line. As pointed out above, the Hellenic Arc and the Hellenic Trough are dominated by linear and angular features rather than curvature. If the subducted crustal slab were a rigid continuous unit except at the most obvious 90° bends, i.e. south of Crete and near the Anaximander Mountains, then major breaks would occur at these two points only. There would be a divergence of the descending slab of about 10° below Crete, and a similar convergence under the Anaximander Mountains. The line of seismicity and magnetic anomaly from west–central Crete to Santorini lends some weight to this suggestion, but a divergence of 10° would produce a tear 35 km wide for a subducted slab length of 200 km. It is unlikely that a discontinuity of this magnitude occurs in one place. Similarly, convergence of this magnitude would produce enormous compressive forces even if, as suggested by Ambraseys (personal communication), the slab effectively thickens as it descends. It is also unlikely that a structure based on three massive planar slabs at angles of 90° could have arisen from the east–west foreshortening of an island arc.

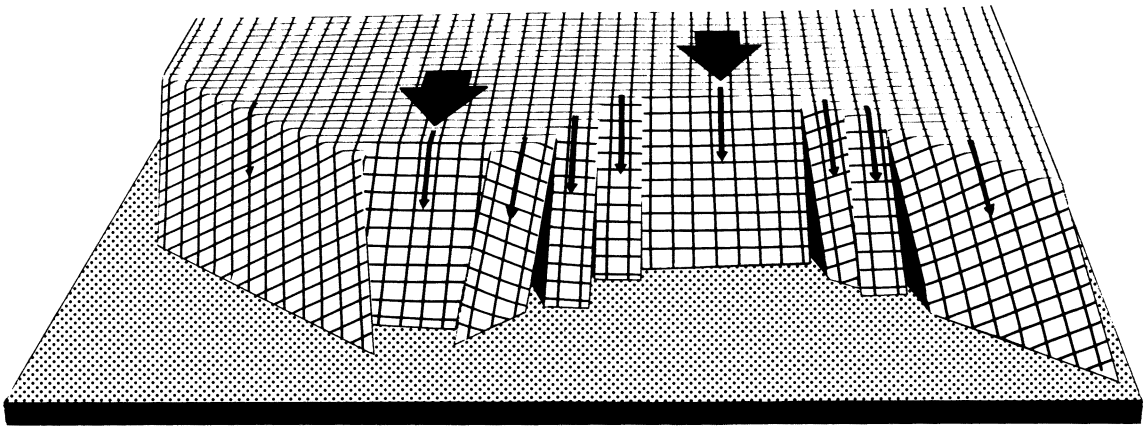

FIGURE 17. Schematic model of possible mode of subduction of the African plate beneath the Aegean.

Both objections are partly overcome by suggesting a short segment of subducted crust with a due east–west strike across the angle of each of the major corners. In the case of the divergence area this results in two narrower divergence zones, each of about 5°. In the convergence zone the problem of converging slabs of crust 50–100 km thick is not resolved unless a short tear fault is also introduced at each end of the east–west segment. An offset north or south would prevent immediate collision of the descending slabs. Thus it is possible that the 'arc' is made up of short segments of dipping crustal slabs with strikes of E–W and NW–SE on the NW–SE trending limbs of the arc, and strikes of E–W and SW–NE on the SW–NE limb. Such segments must necessarily be interspersed by tear faults where the change in trend produces a relative curvature towards the south, and may be interspersed by tear faults, but not necessarily, when the change of trend produces a relative curvature to the north. The final effect, in model form, is shown in figure 17. It should be stressed that the model is intended to illustrate that such a sequence of small subduction zones can produce a total structure extremely similar to the Hellenic–Cyprus Arc, but there is no intention to suggest that the actual position of faults in the model represents the actual structure of the arc. It is implied by the model that fingers of crustal slab, sometimes no more than 50 km wide, are subducted beneath the south Aegean. Given the varied regional stresses arising from changing relative motions of Africa, Turkey, and the south Aegean, it is unlikely that such fingers would remain intact to a depth greater than 2–3 times their width or thickness. This is consistent with the absence of hypocentres deeper than 160 km; see figure 18 (Galanopoulos 1973).

Since it is probable that such finger-slabs crack or break at various depths, it is not surprising that the distribution of hypocentres does not plot on a simple Benioff zone, even for one limb of the 'arc'. Such broken slabs will tend to produce increased resistance to further subduction at that point, and a short segment of the arc will tend to be offset southwards by 50–100 km (Papazachos 1973). This adds to the angular disjointed appearance of the arc. Figure 18 shows the profiles of hypothetical subducted slabs proposed by Galanopoulos (1973) superimposed on the recorded locations of hypocentres. The distribution of hypocentres is more consistent with the proposition that the 'arc' consists of two separate slab systems, one striking NW–SE and the other SW–NE.

East of Rhodes the model proposes large tear faults on either side of Fethiye and Gelidonya. The eastern fault, if it is a fault, coincides with the most dramatic linear feature of the whole Turkish coastline. The subduction zone east of Gelidonya, seen as a single linear feature, would then trend southeast from Antalya towards western Cyprus. This correlates with previous discussion, with a qualification that the whole subduction zone east of Crete may be a double feature, with the Pliny and Strabo Troughs continued eastwards, representing the arc system being forced southwards.

Papazachos (1973) assumed that the earthquake hypocentres of the Hellenic Arc lie on an 'amphitheatrical or conical surface' at a fixed distance from the line through the back-arc volcanoes. The inner edge of the seismic zone plotted in terms of distance from the volcanic axis does show a clear trend with a dip of 35°, but it is impossible for the lithospheric slab to avoid tearing on such a steep dip. Papazachos' profiles also show hypocentres of intermediate depth beneath the Hellenic Trough, outside the island arc. This appears to be associated with the double trench topography of the Strabo and Pliny trenches, and may be due to the breaking-up of subducted fingers of slab, and out-stepping of short segments of the arc, as suggested by the model (figure 17). The vertical motions observed (figure 13*b*) correlate fairly well with the model (figure 17). The size of blocks which are shown to be moving as rigid units, typically 100×50 km, is consistent with the idea of blocks thrust outwards from the Aegean, each being tilted differently as the result of underthrusting by narrow subducted segments of the African plate.

Karig (1970, 1971, 1974*a*, *b*) and Oxburgh & Turcotte (1971) suggest that the back-arc areas of the island arcs of the Pacific are spreading, since the subducted material behind the island arc produces a net accretion of metamorphic material and magma which then rises in the back-arc area to produce extension of the crust. Karig (1971, p. 340) states that in the case of the Mariana island arc system there is an axial topographic high in the back-arc area, where most of the extension takes place, but that there are minor extension areas parallel to the main axis, between it and the island arc, and further extension areas normal to the main axis and to the island arc within the back-arc area. Packham & Falvey (1971) also propose a quasi-linear mode of spreading for the Pacific back-arc seas, but without axial symmetry. Hart, Glassley & Karig (1972) studied the basalts from the Marianas back-arc area and concluded that there was a great deal of similarity to basalts from slow-spreading mid-ocean ridges. The Mariana back-

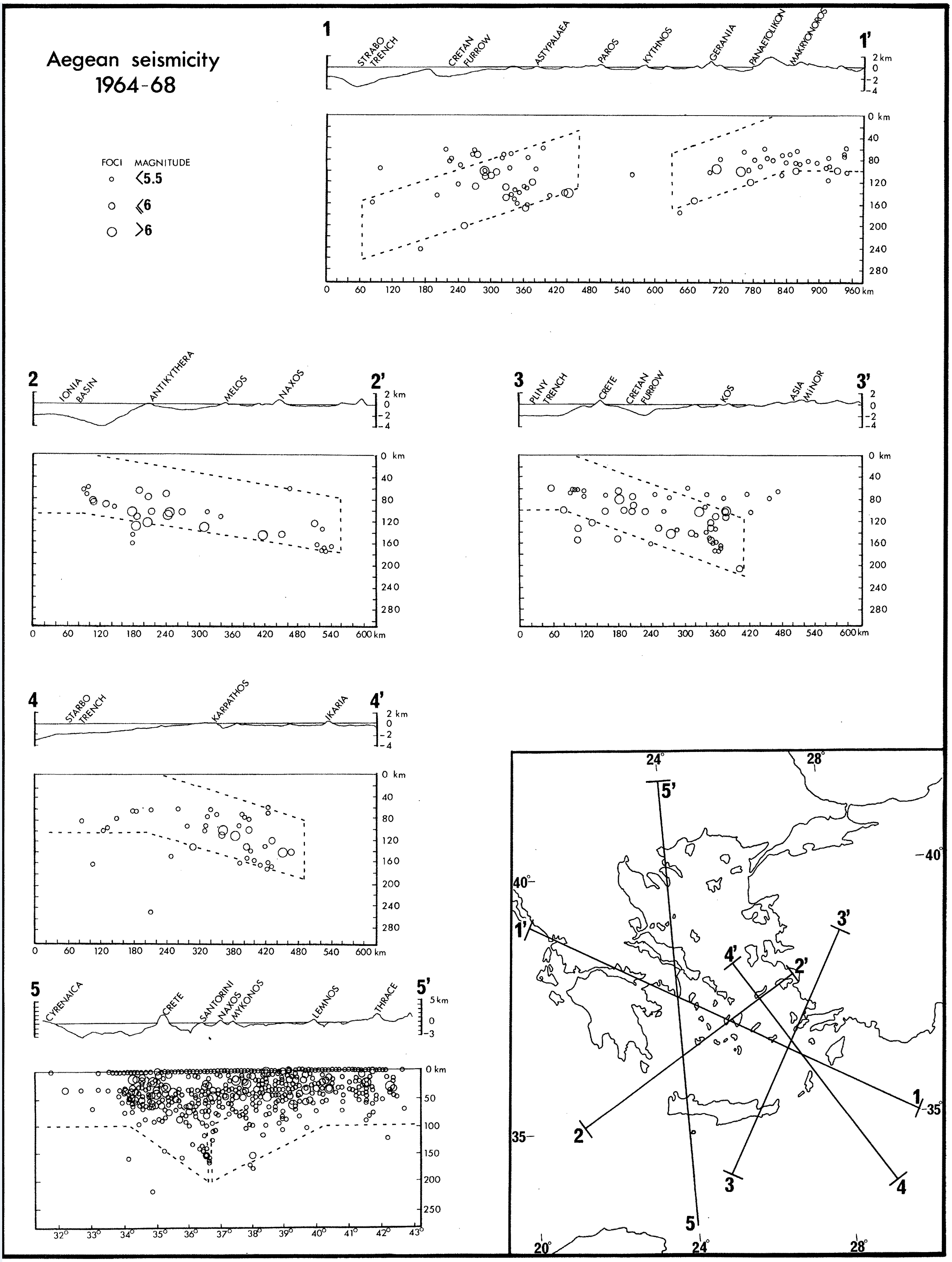

Figure 18. Seismic hypocentre profiles adapted from Galanopoulos (1973). Profiles 1–4 are derived from zones 200 km wide. Profile 5 is derived from a zone 500 km wide. All profiles are plotted with the presumed African plate on the left, and the presumed Eurasian or Aegean plate on the right. Note that this has required reversing the diagrams for profiles 1 and 4 as compared with the original figures in Galanopoulos (1973). The dotted boundaries outline the presumed profile of the subducted slab as depicted by Galanopoulos (1973).

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arc basalts have a more alkaline affinity than most mid-ocean ridge basalts, and could be associated with a spreading rate equivalent to 1 cm/a. Barker (1972) develops the oceanic type spreading hypothesis for the special conditions of the Scotia Arc.

Dewey & Bird (1971, p. 3187) support Karig's model (1970, 1971) to the effect that 'thermal diapirs' form above the descending slab, and that these would be alkalic if formed at depth, and tholeiitic if formed at shallow levels. They propose that this is compatible with the complex magnetic anomaly patterns in back-arc areas, and the high heat flow. Matsuda & Uyeda (1971, p. 21) also support the hypothesis that magma ascends from the descending lithosphere 'almost everywhere', so that magnetic lineations are not generally produced behind island arcs. Most of the magmas would solidify before reaching the surface, and there may be short-lived micro-spreading centres. Matsuda & Uyeda (1971, p. 22) also observe that back-arc spreading may be associated with longitudinal constraints on the arc, so that curvature becomes more acute, while the ends of the arc make sharper angles of intersection with adjacent arcs. Makris (1973) and Makris *et al.* (1973) conclude from deep seismic sounding cross-sections of the Aegean and Ionian Seas that a hot plume of low seismic velocity rises through the lithosphere across much of the width of the south Aegean.

Nicholls (1971) attempts to correlate all the volcanics of the Aegean with the concept of a rigid plate moving SW over a Benioff zone dipping NE under the Cretan Arc, but Vilminot & Robert (1974) show that the volcanicity of the NE Aegean cannot be caused by subduction south of Crete, since, it is 500 km from Crete. Vilminot & Robert (1974) conclude that the volcanicity of the NE Aegean is continuous with Turkey itself, and support the concept that the north Aegean floor is part of the 'paleo-plate' of Turkey.

The Plio-Quaternary volcanicity of the south Aegean is further discussed by Nicholls (1971), Vilminot & Robert (1974) and Hedervari (1973). Vilminot & Robert (1974) discuss the eight volcanoes, from west to east, Krommyonia, Methana, Aegina, Poros, Milos, Antimilos, Santorini and Nisyros. Hedervari (1973) omits Methana, Poros, and Antimilos, and adds Vromotopos and Chaikutes on the Turkish coast near Bodrum. Thus there appear to be ten significant volcanoes in an 'arc' about 200 km north of the Hellenic Trough. In practice, this arc amounts to two almost straight lines radiating from Santorini. Vilminot & Robert (1974) conclude that the products of these volcanoes are typically calc-alkaline with predominance of andesites. Nicholls (1973) stresses that Santorini is in fact the only Aegean volcano to have been active in historic times. This may be due to the proposed intrusive belt trending from west-central Crete as implied by the magnetic data of Vogt & Higgs (1969) and the seismicity (Ambraseys 1971; Papazachos 1973).

Hedervari (1973) states that shallow seismic events do not occur close to, or immediately beneath volcanic vents, and calculated from the nearest hypocentres that there is a conical-cylindrical column, or 'translithospheric vent' beneath each of the volcanoes of the south Aegean. This column is supposed to have such low rigidity that it is incapable of accumulating strain which might later be released seismically. The vent immediately below the volcanoes has an approximate radius of 10–20 km at a depth of about 50 km, but narrows to less than 1 km near the surface. Deeper than 50 km the low rigidity column continues to a depth of about 100 km with a radius of at least 25 km. This picture correlates with the concept of thermal diapirs summarized by Dewey & Bird (1971), from the data of Karig (1970, 1971), but, in the case of the Aegean, the multiplicity of vents may be due to the splitting and divergence of the descending lithosphere.

The model proposed by Hedervari (1973) requires that at least seven zones of low rigidity, each about 50 km across, exist in a band between Corinth and Bodrum. That is, of 500 km of crust, 350 km are of low rigidity at a depth of 50–100 km below the surface. Quite apart from the possible occurrence of closely spaced faults and decoupled blocks which might reduce the general rigidity of the crust in the Aegean area, the conclusion of Hedervari (1973) goes some way to explain the low seismicity of the Hellenic Arc, as noted by Isacks & Molnar (1971, p. 159) when compared with the island arcs of the Pacific. North (1974) finds an order of magnitude discrepancy between the proposed rates of slip on plate boundaries (McKenzie 1972) and the seismic moments calculated from earthquake magnitudes and the presumed surface areas of fault surfaces. North (1974) concludes that the observed low seismicity for the Mediterranean region as a whole might be explained if much of the deformation took place as viscoelastic creep. Somewhat confusingly, the three 'plate boundary' segments for which North (1974) finds least bad agreement between observed seismicity and proposed slip rates, are the north edge of the Aegean, the Anatolian Fault, and the proposed north–south boundary through western Turkey (see figure 19*b*). This probably arises because the plate boundary length required by the model (McKenzie 1972) is considerably shorter than the real boundary length (see below) and hence the seismicity per unit length appears high.

The seismic and volcanic data therefore suggest a very low rigidity for the south Aegean, and this is compatible with the occurrence of small scale units of crust which may move relatively independently, thus correlating with the varied stick-slip periodicities found for the different islands of the Hellenic Arc. Such variation in periodicity and magnitude of movement would be less compatible with the concept of a rigid Aegean plate.

Fault plane solutions for the Aegean have been published by Papazachos & Delibasis (1969), Isacks & Molnar (1971), McKenzie (1972), and Crampin & Ucer (1975). Papazachos & Delibasis (1969, p. 253, fig. 2) conclude that the acting stress in all parts of the Aegean area is tensional, with the exception of the north Aegean, which is compressional, as are Albania and the Ionian islands. It is suggested (Papazachos & Delibasis 1969, p. 254) that the stress field is equivalent to a thick beam bent around the stable mass of the central and northern Aegean. The faulting near the Turkish coast is generally normal, and dextral. Faulting in other areas is generally sinistral (Papazachos & Delibasis 1969, fig. 3).

McKenzie (1972, p. 139) notes the relative aseismicity of the central Aegean, and appears to assume implicitly that this is due to plate rigidity, whereas the present discussion indicates that it is more probably due to extreme lack of rigidity. McKenzie's discussion (McKenzie 1972, pp. 137–158) of the Aegean is stimulating and valuable in that a number of anomalies and inconsistencies mitigating against the acceptance of a simple rigid Aegean plate are honestly pointed out and highlighted. In particular, there is the difficulty of accounting for thrust faults in the north Aegean, and for the presumed relative motion between Turkey and the Aegean plate by means of a set of horst and graben normal structures in western Turkey. The total absence of a suitable structure showing the required 100 km offset across central Greece from Thessaloniki to the Ionian Sea is also stressed as an anomaly (McKenzie 1972, p. 155). McKenzie (1972) suggests that the Benioff zone under the Hellenic Arc dips north near Crete, but northeast under the Peloponnese.

McKenzie (1972, caption to fig. 2*b*) stresses that his sketched plate boundaries in the region of western Turkey are only intended to show the general nature of relative motions, and not actual plate boundary features on the ground. This, in conjunction with the discussion

(McKenzie 1972, pp. 137–158), suggests that the diagrams shown in figure 19 (*a*, *b*) have been taken a bit too literally by subsequent authors, especially by Nicholls (1971, based on McKenzie 1970), Galanopoulos (1973) and Vilminot & Robert (1974).

A plate tectonic model

McKenzie (1972, p. 179) proposes that the Aegean plate was derived by fracturing of a single plate extending from Greece to Turkey with the two present land masses further apart in an east–west direction. Smith (1971) does not mention an Aegean plate. Dewey *et al.* (1973) discuss the plate tectonic history of the region in terms of three plates denoted as Turkey, Rhodope and

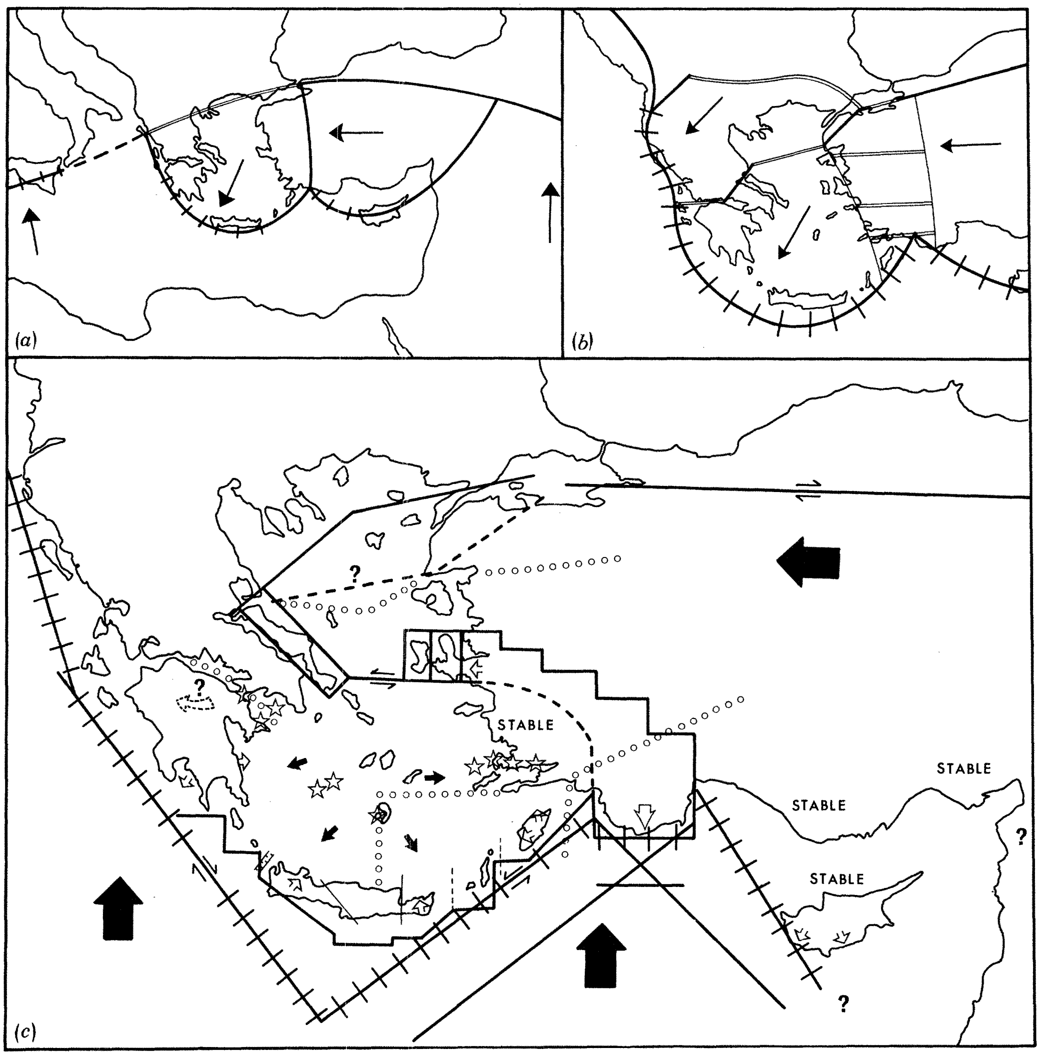

FIGURE 19. Proposed plate boundaries. Cross-hatched lines, subduction zones; double lines, spreading axes; black arrows, direction of plate movement. (*a*), McKenzie (1970); (*b*) McKenzie (1972); (*c*) this paper. In (*c*) the open arrows, direction of tilt of coastal blocks; lines of circles, seismic lineations from Papazachos (1973) and Crampin & Ucer (1975); hollow stars, volcanoes. The dotted arrow on the Peloponnese indicates possible sense of rotation. Areas of superimposed symbols indicate concurrent, but not precisely defined, processes in one zone: thus, the eastern limb of the Hellenic Arc is a zone of net subduction and crustal consumption, with a component of sinistral strike slip, tear fault offsets, and a possible complete double structure.

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Apulia, but do not suggest the existence of an Aegean plate before about 5.5 Ma ago (Dewey *et al.* 1973, p. 3170). Aubouin (1973) states that the Aegean plate must be a 'neoplaque', and that it cuts across older structural trends. Galanopoulos (1973) and Vilminot & Robert (1974) concentrate on the problem of how the plate boundaries and motion of the supposed Aegean plate (McKenzie 1970, 1972) could have arisen from the earlier boundaries and motions of the Turkish, Apulian, and African plates.

The evidence of the present research is that the rigid Aegan plate suggested with careful qualifications by McKenzie (1970, 1972) does not exist. The northern Aegean appears to be a continuous prolongation of the Turkish block, while the south Aegean is probably a non-rigid zone of radial and circumferential crustal extension. The problem remains however of defining the present position of the boundaries of the discrete rigid blocks bordering the Aegean, if they exist.

The smallest blocks considered by Dewey *et al.* (1973, fig. 8*b*) have dimensions less than 100 km, though the majority of units termed microplates have dimensions of about 500 km. Since all 24 units move independently as a result of the stresses between them, produced ultimately by the relative movement of Africa and Eurasia, it is a determined attempt to utilize the concepts of rigid blocks and strain-rate discontinuities with a high resolution over a large total area. However, Papazachos (1973) and Crampin & Ucer (1975) report observations indicating that plates of the size of Turkey subject to extreme lateral stresses do experience internal deformation.

Crampin & Ucer (1975) have recomputed epicentres for northwest Turkey and the Marmara Sea area to show that no simple pattern of plate boundaries can be detected. They conclude (Crampin & Ucer 1975, p. 285) that there is no overall pattern of extension as would be implied by McKenzie (1972) (see figure 19*b*). Crampin & Ucer (1975, pp. 286–287) adopt the concept put forward by Papazachos (1973) of plotting lineations of seismicity, rather than fault plane solutions, and demonstrate that the North Aegean is bounded by a series of small block structures. They observe (1975, pp. 270–271) that tectonic units of the Pontid mountains, showing evidence of pre-Alpine and post-Alpine orogeny, occur on the southeast margin of the Sea of Marmara. This accretion of mountainous blocks on the advancing edge of the Turkish plate is similar to the process which has been suggested in the present paper for southwest Turkey.

The planes of discontinuity and movement revealed by the present study have been combined with the evidence from gravity anomalies, magnetic anomalies, seismicity, volcanicity, and fault-plane solutions to generate the plate boundary diagram shown in figure 19*c*.

The model proposed in figure 19*c* accounts for the following observations:

1. Compression stress field in the north Aegean (Papazachos & Delibasis 1969; McKenzie 1972).
2. Absence of a lateral fault across central Greece (McKenzie 1972).
3. Absence of a north–south lateral fault on the Aegean coast of Turkey (McKenzie 1972).
4. Continuation of magnetic anomaly patterns from Turkey across the north Aegean (Vogt & Higgs 1969).
5. Seismic belt from Antalya to Izmir (Ambraseys 1970; present data).
6. Normal and strike-slip faulting in south-west Turkey (McKenzie 1972).
7. Low rigidity and low seismicity of the south Aegean (Galanopoulos 1967; Isacks & Molnar 1971; Hedervari 1973; North 1974).
8. Radial cracking of the crust in south Aegean (Vogt & Higgs 1969; Fisher & Hess 1963).

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9. Radial lines of seismicity extending from Santorini (Ambraseys 1971; Papazachos 1973).

10. Graben extension between the islands of the Hellenic Arc (Zarudzki, unpublished data; present data).

11. Circumferential extension of the Cretan Furrow (bathymetry).

12. Linear radial pattern of volcanicity extending from Santorini (Hedervari 1973; Vilminot & Robert 1974).

13. Decoupled tilting blocks on a 50–100 km scale in the Hellenic Arc (present data).

14. Linear, angular, offset double structure of most of the Hellenic Trough (Rabinowitz & Ryan 1970; Woodside & Bowin 1970; Papazachos 1973; Biju-Duval *et al.* 1974).

15. North–south coastal lineations at Fethiye and Gelidonya–Antalya (present data).

16. Stability of south Turkey east of Antalya and north Cyprus (Ambraseys, 1971; Woodside, written communication; Biju-Duval *et al.* 1974; present data).

17. Gravity high in south Aegean and extreme gravity low in Rhodes basin (Rabinowitz & Ryan 1970; Matsuda & Uyeda 1970).

18. Deduced nature of hypocentre distribution below Hellenic Arc (Galanopoulos 1973; Papazachos 1973).

19. Stick-slip periodicity of vertical movements in Hellenic Arc (present data).

20. There is little strike-slip component in the subduction under the Hellenic Arc, even on the outer limbs of the Arc (McKenzie 1972). Several further comments and deductions can be made from the model, hypothetical as it is.

21. The radial plus circumferential expansion of the south Aegean is consistent with observed normal faults and asymmetrical grabens cutting across Greece (Brunn 1956; Heezen, Ewing & Johnson 1966) of which the Gulf of Corinth is by far the biggest. These grabens are not associated with any east–west offset. The trench along the east coast of the Peloponnese, and the fact that the Gulf of Corinth and the Saronic Gulf widen towards the east, suggest that the Peloponnese may be rotating slightly clockwise.

22. The break-up of Euboea from the coast of central Greece results from the oblique westward pressure of the Turkey plate.

23. There is no confirmation from the present data for an active plate boundary east of Cyprus trending directly into the Gulf of Iskenderun. If there is such a boundary it would meet the Dead Sea Rift at a triple point a few kilometres inshore from the Gulf of Iskenderun (McKenzie 1972, figure 2*b*) which seems incompatible with the present data, even if the border zone is atypically quiet at the moment (Ambraseys 1971). The absence of vertical movement in the last 2 ka as far south as the mouth of the Orontes indicates that the plate boundary south of Cyprus must pass south of the Orontes, although the possible uplift of Seleucia Pieria (figure 3 and table 1, site 96) suggests that the boundary may be close by.

24. The exact structures of the plate boundaries in this region are probably all at least as complex as that proposed for the Hellenic Arc. Figure 19*c* must be regarded as approximate and simplified at all points.

Conclusions

The present study demonstrates that it is possible to detect and identify rigid crustal blocks on a scale of 50–100 km in a region of intense deformation, and to describe the relative motions in the zones of high strain between the blocks. This is achieved by combining data on strike-slip motion on boundary faults, and subduction rates on trenches, with vertical rates of movement from the present data. The central Aegean itself has been treated as a more or less viscoelastic non-rigid body, but even this area would probably turn out to be resolvable into block-slip graben patterns if studied on a scale of about 20 km and 200 a time intervals.

The model of the Hellenic Arc subduction zone (figure 17) and the plate boundaries in figure 19*c* should be considered as strictly hypothetical, but the model does provide a lot of specific subsidiary hypotheses which are open to test in the field.

The final model indicates that the present state of the Aegean consists of a reduced island arc and back-arc area in a late phase of annihilation. As Turkey encroaches from the east, and Africa from the south, the contortion of the Hellenic Arc subduction zone will become even more acute, and the fragmentation of the Peloponnese, Crete, and possibly southwest Turkey will become more chaotic. The thinning of the Aegean crust will continue, and oceanic crust may be exposed, as suggested by McKenzie in a general diagram (1972, fig. 38). It is salutary to consider that when the Aegean is finally eliminated, the tectonic structures which result will bear very little relation to the north–south compression which caused the event.

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