Geophys. J. Int. (1994) **117**, 834–849

The 1953 earthquake in Cephalonia (Western Hellenic Arc): coastal uplift and halotectonic faulting

S. C. Stiros,[1] P. A. Pirazzoli,[2] J. Laborel[3] and F. Laborel-Deguen[3]

[1] *Institute of Geology and Mineral Exploration (IGME), 70 Mesoghion St, Athens* 11527, *Greece*
[2] *CNRS-URA* 141, *Laboratoire de Géographie Physique,* 1 *Place Aristide Briand,* 92190 *Meudon-Bellevue, France*
[3] *Laboratoire de Biologie Marine, Université de Marseille II, Luminy, case* 901, 13288 *Marseille Cedex* 9, *France*

Accepted 1993 December 20. Received 1993 December 20; in original form 1992 December 18

SUMMARY

Geomorphological, marine biological and radiometric data in combination with earlier reports reveal that the $M_s = 7.2$, 1953 Cephalonia earthquake, the most destructive but least studied earthquake in Greece this century, was associated with a 0.3–0.7 m quasi-rigid-body uplift and westward tilting of the central part of the island. Another palaeoseismic event, around 1500 yr BP, associated with coastal uplifts was also identified.

Structural data indicate that the 1953 uplift is bounded by two subparallel, east-dipping major reverse faults and corresponds to a piston-like motion. This structurally unusual pattern of seismic deformation is detached from the deformation of the basement (conspicuously a thrust) and different from the long-term deformation pattern of the area; it is probably due to the particularities of salt tectonics: a ~1500 m thick salt layer acts as a regional décollement, while thinner layers are sandwiched between the carbonate thrust sheets that compose the overburden and reduce their friction during fault movement. This crustal anisotropy is responsible for the observed anomalous attenuation of seismic waves during the 1953 and historical earthquakes.

The 1953 seismic surface deformation mimics long-term halotectonic patterns, but is not directly indicative of the regional stress-field, for it reflects uplift-induced stresses only.

Key words: biological mean sea level, Cephalonia, crustal anisotropy, earthquake, palaeoseismic event, evaporites, raised beaches, salt tectonics, thrust sheets.

1 INTRODUCTION

A number of exposed fossil shorelines, associated with seismic events, have recently been identified in the Aegean and the Eastern Mediterranean (Crete: Pirazzoli 1986a; Rhodes: Pirazzoli *et al.* 1989; Euboea: Stiros *et al.* 1992; Gulf of Corinth: Papageorgiou *et al.* 1944; Pirazzoli *et al.* 1944; for review of results and ideas see Stiros *et al.* 1944). In all these cases, however, the coastal seismic event is inferred from biological and other data, and the only case for which an eyewitness account for a coastal seismic uplift exists is the 1953 earthquake of Cephalonia (Kefallenia) Island, in the Ionian Sea, at the west end of the Hellenic arc (Figs 1 and 2; Galanopoulos 1955; Mueller-Miny 1957). This earthquake was large ($M_s = 7.2$) and destructive, but one of the least-studied earthquakes in Greece in the last 100 years.

Another particularity of Cephalonia is that the 1953 earthquake affected the overburden of a thick evaporite layer (Fig. 3). While the long-term deformation of evaporites and the contribution of tectonics have been the matter of detailed investigations, especially in recent years (for references and results, see Trusheim 1960; Nely 1980; Jackson & Talbot 1986; Underhill 1988; Jenyon 1988; Mascle *et al.* 1988; Dardeau *et al.* 1990), to the best of our knowledge there have been no studies of the response of evaporites to earthquakes.

Thus, the 1953 earthquake provides an opportunity to study probably the first known case of seismic, halotectonic movements. The results of our study of the 1953 and older Holocene co-seismic uplifts of the coasts of the island, and their implications for the understanding of the mechanism of the 1953 earthquake are presented in this paper.

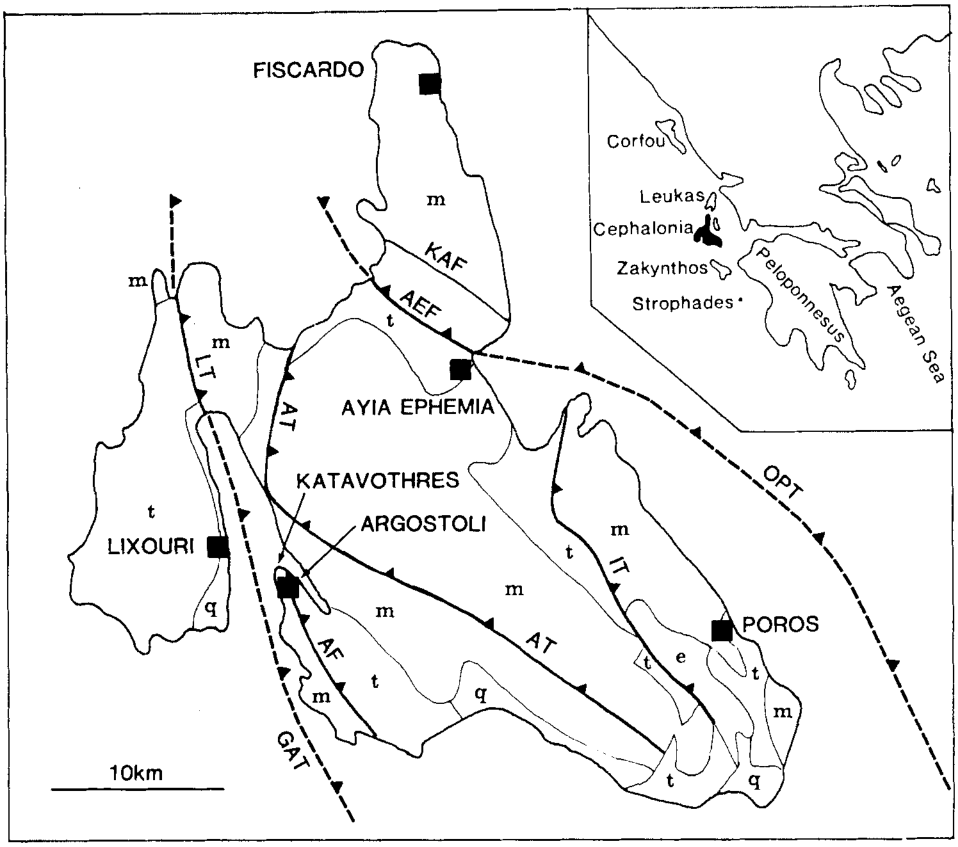

Figure 1. Simplified geology of Cephalonia Island (after Bornovas & Rondogianni 1983; Sorel 1976; Underhill 1989; this study): q, Quaternary; t, Tertiary; m, Cretaceous–Upper Triassic; e, Triassic evaporites. IT, Ionian Thrust; AF, Argostoli Fault; AT, Ainos Thrust; GAT, Gulf of Argostoli Thrust; AEF, Ayia Ephemia Fault; KAF, Kalon Anticline Fault; OPT, Offshore Poros Thrust; LT, Livadi Thrust. Inset: location map.

2 TECTONIC SETTING

2.1 Structural history

Cephalonia is located at the north-west edge of the Hellenic Arc in western Greece (Fig. 1). The tectonic fabric of the island and the wider area consists of east-dipping, NW to NNW striking thrust sheets (Fig. 4), fragments of a carbonate, epicontinental platform. During Mesozoic–Tertiary tectonic subsidence, ~5000 m of Upper Triassic and up to 500 m of Tertiary carbonates were deposited above a possibly 1.5 km thick (Monopolis & Bruneton 1982; Kamperis 1987) Triassic evaporite layer. In the channel between Cephalonia and the mainland, however, more than 5 km thick Neogene and Quaternary sediments have been identified (Kamperis 1987).

Tectonics were inverted in the Lower Pliocene, and the area is still under ENE–WSW compression. During three paroxysmic periods (Lower Pliocene, Lower Pleistocene and Middle Pleistocene–Holocene), separated by intervals of tectonic quiescence, the Tertiary (and probably Mesozoic) normal faults reactivated as reverse faults (or folded faults), new thrusts and folds formed, the carbonate rocks were uplifted and the present-day north-west part of the Aegean arc was formed (Bornovas 1964; IGRS & IFP 1966; BP Co. Ltd 1971; Sorel 1976; Mercier *et al.* 1979; Karakitsios 1992).

Lower Pliocene to Holocene contractional tectonics were associated with an important uplift: 300 m to more than 700 m during the compressional interval of the Lower Pliocene, 500 m to more than 700 m in the Lower Pleistocene, and ~100 m uplift since the Middle Pleistocene (Sorel 1976).

2.2 Active tectonics

Contractional tectonics are still active, as the intense seismicity (see Table 1), focal mechanisms of earthquakes (McKenzie 1972; Scordilis *et al.* 1985; Anderson & Jackson 1987) and microearthquakes (Hatzfeld *et al.* 1990), and tectonic (Sorel 1976; Mercier *et al.* 1979) data reveal. Uplift is still continuing, as coastal data (Fig. 7) indicate.

Some of the major faults are active (Fig. 4), while a number of minor faults cutting Quaternary sediments have been observed on the island (Sorel 1976; Underhill 1989). Also, 500 000 yr old folded marine sediments have been recognized (Sorel 1976).

2.3 Structural position and role of evaporites

Extensive evaporitic outcrops (limestone and dolomitic breccias and gypsum; see Fig. 3) occur in north-west Greece. In the past, their stratigraphic position has been something of an enigma, but detailed studies revealed that, except for some relicts of the Messinian salinity crisis of the

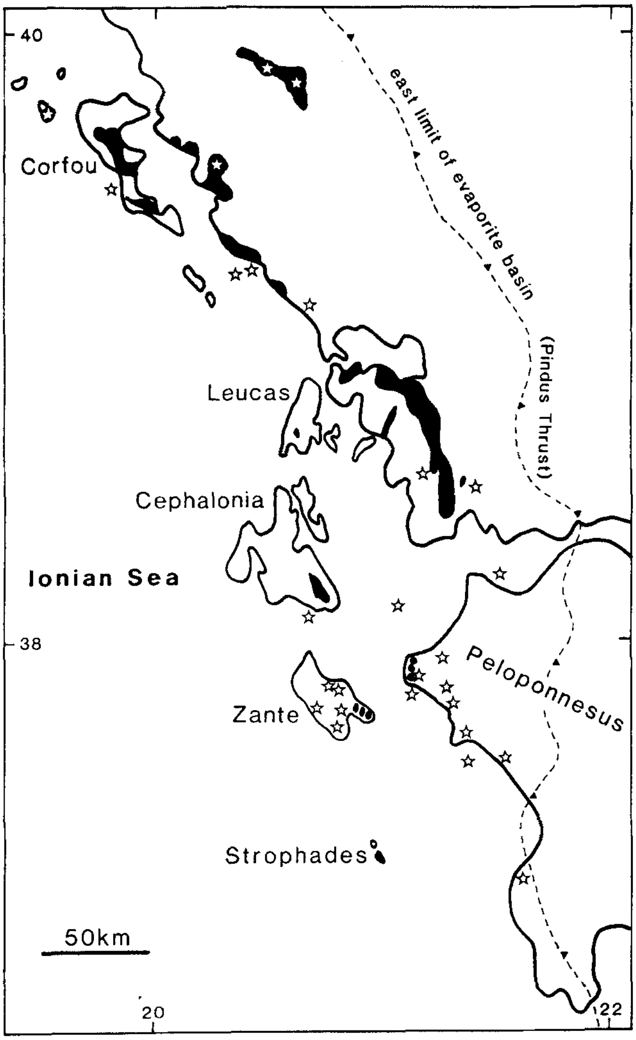

Figure 3. Evaporite surface exposures (black areas) and location of boreholes for petroleum investigation (open stars) in western Greece. Many of these boreholes, 3–4.5 km deep, crossed a ~1500 m thick evaporite layer or testify to large-scale diapiric movements. The Pindus Thrust, bounding the Triassic evaporite basin to the east is shown (after Nikolaou 1986, 1988, modified).

Mediterranean, they result from Plio-Quaternary and older diapirism of Triassic sediments. Evaporites in the wider area crop out at the leading edges of certain thrust sheets, are sandwiched between them (Figs 1, 4a and b) and played an important role in their emplacement (Bornovas 1960, 1964; BP Co. Ltd 1971; Monopolis & Bruneton 1982; Nikolaou 1986; Underhill 1988); for example, acting as lubricants along thrust planes (Bornovas 1964; Nikolaou 1986).

Gypsum is dominant and halite is absent (for an exception, see Nikolaou 1986, p. 35) in evaporite exposures in western Greece, but over 1500 m of halite have been

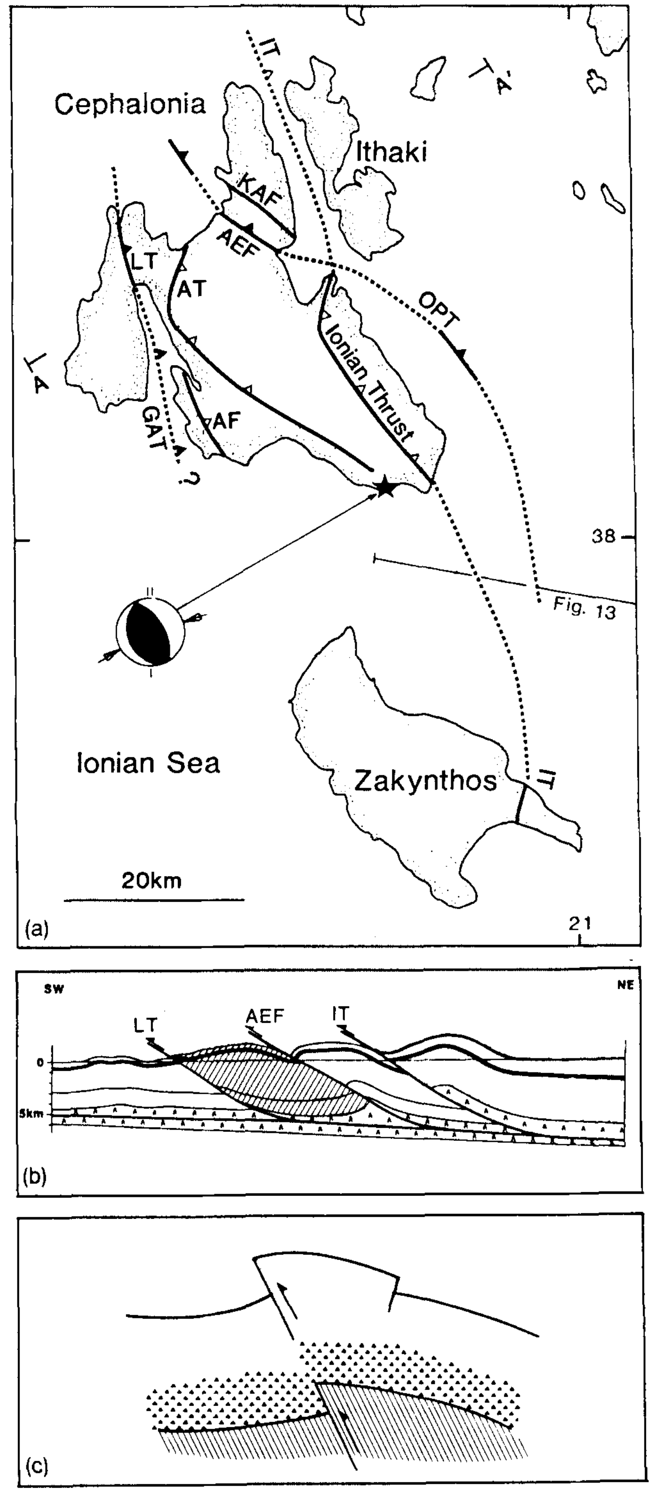

Figure 4. (a) Schematic structural pattern of the wider Cephalonia area, modified from Sorel (1976), Cushing (1985) and Underhill (1989). Continuous lines indicate observed thrusts and reverse faults, and dotted lines their inferred extension offshore (see text for details). Solid and open triangles indicate faults active and inactive in the Upper Pleistocene, respectively (after Sorel 1976). A star indicates the epicentre of the 1953 main shock, expected to be accurate to within 50 km; the focal mechanism by McKenzie (1972) is also shown. Abbreviations are as in Fig. 1. (b) Schematic cross-section along axis A–A′ shown in (a), based on the hypothesis of Triassic evaporites (layers marked with inverted v's) playing the role of a regional detachment zone. The block uplifted in 1953 is dashed (after Underhill 1989, modified). (c) Alternative model for basement control on thrusts and uplifts seen on the surface; a disharmony between basement and upper layers is caused by the quasi-ductile deformation of evaporites (layer with triangles; after IGRS & IFP 1966, modified).

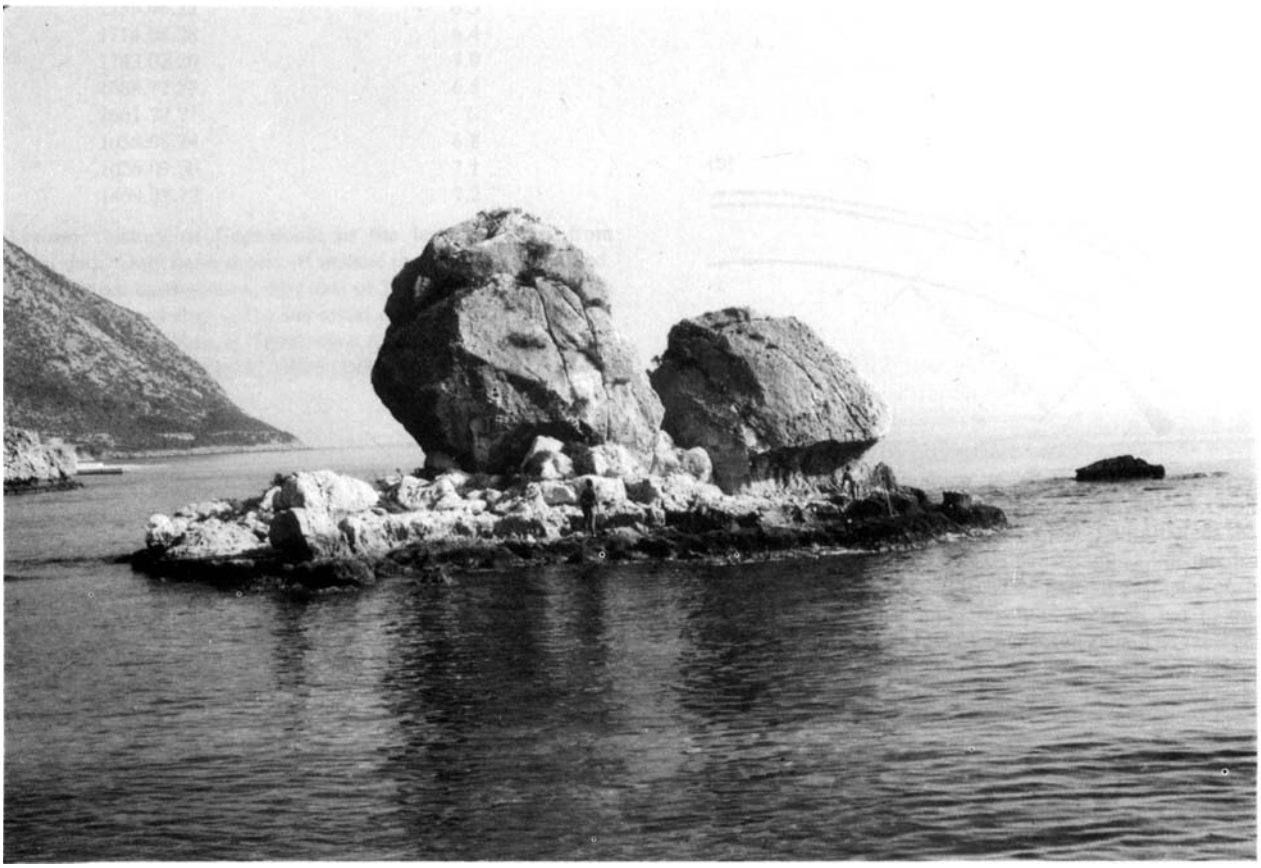

Figure 2. Photo of the mushroom-shaped rock at Poros, similar to that published by Galanopoulos (1955). Two fossil notches and benches are visible. The lower level corresponds to the 1953 co-seismic uplift, the upper one to a palaeoseismic event at around 1500 yr BP.

Table 1.

Date	Ms
1983.01.17	7.0
1972.09.17	6.3
1953.08.12	7.2
1939.09.20	6.3
1912.01.24	6.8
1867.02.04	7.2
1862.03.14	6.6
1767.07.22	7.2
1766.07.24	6.7
1759.06.22	6.5
1714.08.28	6.4
1743.02.20	7.0
1688.??.??	6.5
1661.??.??	?
1658.08.24	6.8
1636.09.30	7.1
1469.??.??	7.2

The seismic history of Cephalonia in the last 500 years from historical data. Only main shocks of seismic sequences are included. Among all these earthquakes, only that of 1953 was associated with coastal uplifts, according to the historical evidence and the coastal data discussed here. Source: Papazachos & Papazachou (1989); the 1661 earthquake after Vogt & Albini (1992).

identified in boreholes (BP Co. Ltd 1971; Fig. 3), a situation typical in many evaporite environments (Mascle *et al.* 1988), indicating that surface gypsum represents a relict of subsurface halite deposits (Bornovas 1964; Underhill 1988).

The role of evaporites was very important in the structural and geomorphological evolution of Cephalonia and the wider area, but remains a matter of debate. BP Co. Ltd (1971), Guzzetta (1982) and Underhill (1988) suggest that they acted as a regional detachment zone (Fig. 4b); according to IGRS & IFP (1966), no regional-scale décollements are likely to exist in north-west continental Greece (where the evaporite layer is, however, thinner) with basement faulting cutting through the ductile evaporite layer to the surface, though with a disharmonic deformation of uppermost strata relative to the basement (Fig. 4c).

2.4 Geometry of faults

Faults in Cephalonia and the wider area are either old normal faults that have been reactivated as reverse faults since the Lower Pliocene, or thrusts. In their majority, they are not simple planar faults, but flatten at the surface, with their hanging wall block edges folded and cut by minor low or high angle faults, subparallel to the trace of the main fault (Fig. 5; Sorel 1976; Cushing 1985; Underhill 1989).

This geometry is to a certain degree related to the variations with depth in the thickness and composition of intruded evaporites, producing variations in the friction coefficient between the fault blocks: thick halite layers and low-friction surfaces exist at certain depths, while the lack or scarcity of evaporites near the surface is responsible for 'locked' faults (Fig. 5).

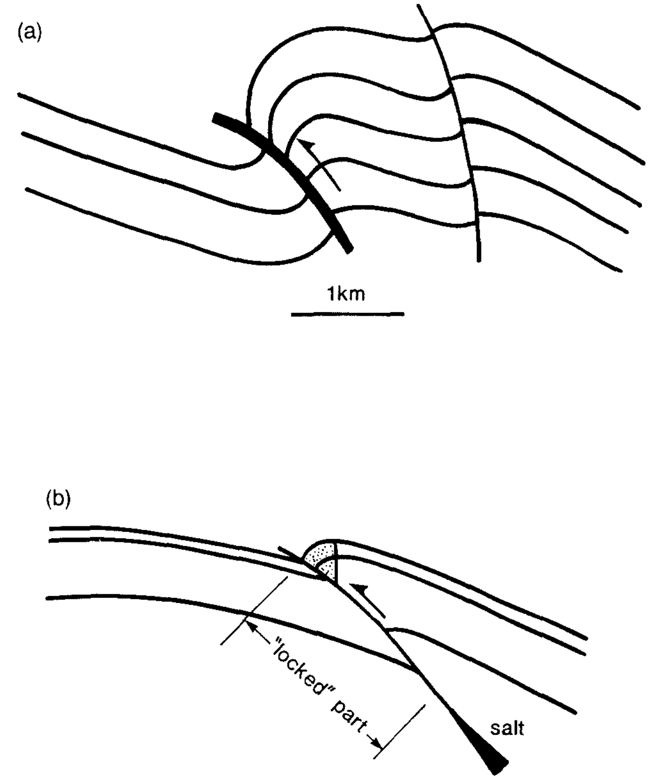

Figure 5. Pattern of faulting in Cephalonia. (a) Second-order deformation of the hanging wall of a typical fault at Cephalonia, simplified after Sorel (1976). The main fault flattens near the surface and is marked with a solid line. A faulted anticline is formed at the edge of the footwall. (b) The pattern of faults on a larger scale. The fault in the upper levels is 'locked', but in the lower levels, due to salt intrusions, friction is reduced. The fault geometry and variations of the friction coefficient are responsible for the deformation of the hanging wall edge (shaded area), which is reflected in relatively high seismic intensities and minor uplift.

3 SEISMIC HISTORY OF CEPHALONIA AND THE 1953 EARTHQUAKE

The $M_s = 7.2$, 1953 earthquake is the most recent major event to have affected the island. Although large earthquakes are not unusual in the area (see Table 1), this particular seismic event resulted in damage and a death toll of an unprecedented scale: more than 85 per cent of the buildings in Cephalonia (with the exception of the Fiscardo area where damage was limited—a point we shall discuss later), Ithaki (Ithaca) and Zakynthos (Zante) islands were completely destroyed. Inspite of the available detailed information on the seismic damage, there is no historical evidence that, with the exception of the 1953 event, any of the earthquakes listed in Table 1 were associated with coastal uplifts. This is confirmed by our results, for the second, upper notch shown in Figs 2 and 7(c) was associated with a probably co-seismic uplift dated around 1500 yr BP (see Section 5.2).

3.1 Seismological data

The $M_s = 7.2$, 1953 August 12 shock was preceded by two large foreshocks (1953 August 9, $M_s = 6.4$ and 1953 August

11, $M_s = 6.8$) and was followed by a number of aftershocks; the largest one (1953 August 12) of magnitude 6.3 (Papazachos & Papazachou 1989).

The epicentre of the main shock was located in the south-eastern part of the island, but Anderson & Jackson (1987) suggested that this location could be in error by as much as 50 km. Hypocentral depths of the earthquakes in the seismic sequence are also not well controlled, but are probably shallower than 50 km (Jackson, Fitch & McKenzie 1981). A focal mechanism for the 1953 mainshock by McKenzie (1972) shows thrust faulting with a NNW strike (Fig. 4), but is not well constrained and contains many inconsistencies (Jackson *et al.* 1981).

3.2 Anomalies in the distribution of seismic intensities

A particularity of the 1953 earthquake was that seismic intensities were of the order of IX–X+ (Modified Mercalli Scale) in Ithaki and Cephalonia, with the exception of the Fiscardo area, where the intensities were of the order of VI. Historical seismicity data (e.g. Papazachos & Papazachou 1989) reveal that such anomalies in the distribution of seismic intensities have been observed in a number of historical earthquakes as well. Stiros (1994) has suggested that the discontinuities in the seismic intensities may be identified with some major faults (Fig. 6) which are loci of evaporitic intrusions and responsible for the anomalous attenuation of seismic waves.

3.3 Earlier accounts of the 1953 uplift

According to Galanopoulos (1955) and Mueller-Miny (1957), the 1953 seismic sequence was associated with uplift of up to 60–70 cm at a number of sites in the south and south-east part of the island and the Argostoli area, but it is not known whether all, or which parts, of these uplifts should be attributed to the main shock. However, the much lower magnitude of aftershocks suggests that most of the seismic deformation was associated with the main shock.

At Poros, a mushroom-shaped (notched) rock, about 50 m offshore from the small harbour, was uplifted by about 60–70 cm, and a photograph of this rock was published by Galanopoulos (1955) (Fig. 2). At Katavothres, near Argostoli, the uplift prevented the regular flow of seawater to two water mills, and Mueller-Miny (1957) estimated that the amplitude of the co-seismic displacement was about 30 cm. An uplift of up to 50 cm was also reported at a number of sites along the southern coast of the island, south-east of Argostoli.

No uplift, however, was reported from Lixouri, some other harbours in the island and Ithaki, as is discussed in the following.

4 METHODOLOGY OF THE STUDY OF HOLOCENE SEA LEVEL CHANGES

During our 1990 field survey we examined all the coasts of Cephalonia and Ithaki Islands, either from land or by boat. We started to study the 1953 uplift from Poros, more specifically from the uplifted rock, the photograph of which was published by Galanopoulos (1955). At least two superimposed uplifted notches and benches, up to 1 m high,

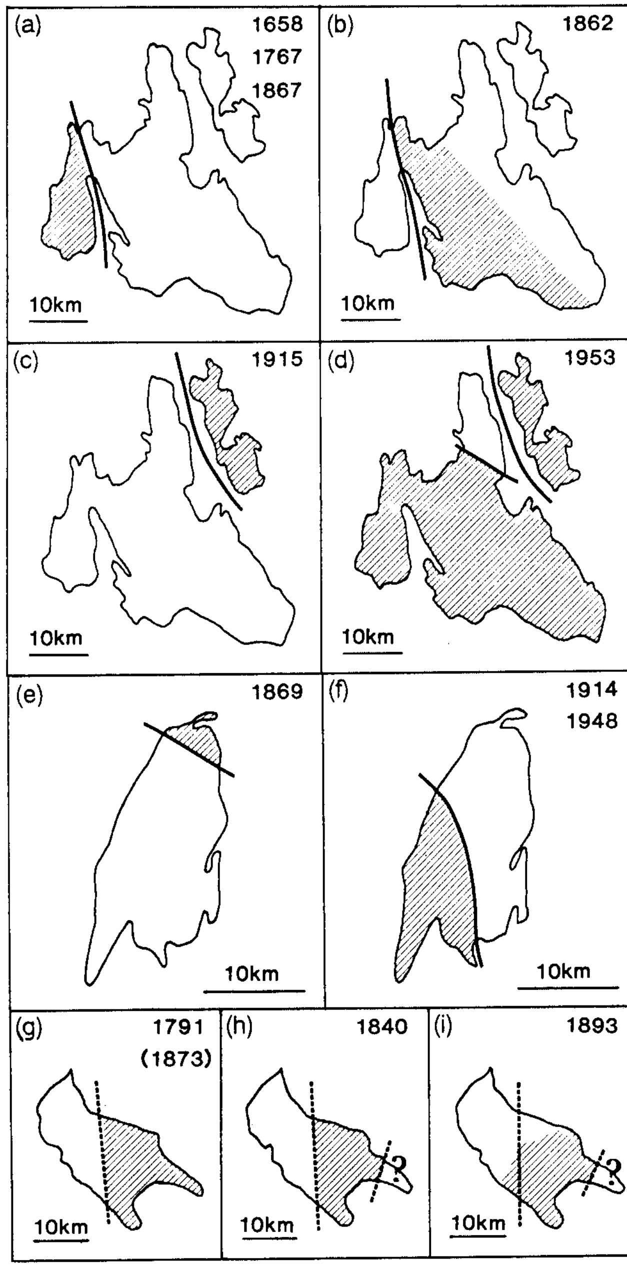

Figure 6. Major faults bound areas with high contrasts in the seismic intensities during major historical earthquakes. Shaded and blank areas correspond to intensities IX–X+ and IV–VI, respectively; parts of the meizoseismal areas only are shown. Salt intrusions along these faults are probably responsible for the anomalous attenuation of seismic waves (after Stiros 1994). (a)–(d): Cephalonia; (e)–(f): Levkas; (g)–(i): Zakynthos.

were recognized on this rock (Fig. 2). Such landforms are the product of the biological and mechanical erosion of coastal rocks during periods of relative sea-level stability and correspond to fossil shorelines (Pirazzoli 1986a).

Our systematic survey revealed that uplifted fossil shorelines exist along several parts of the island (Fig. 7). The lowest exposed fossil shoreline is very fresh and with many vermetid shells well preserved in growth position; it is clearly the youngest. Morphological and biological correlations and radiometric control (two dates are discussed below, while several other radiocarbon dates from samples

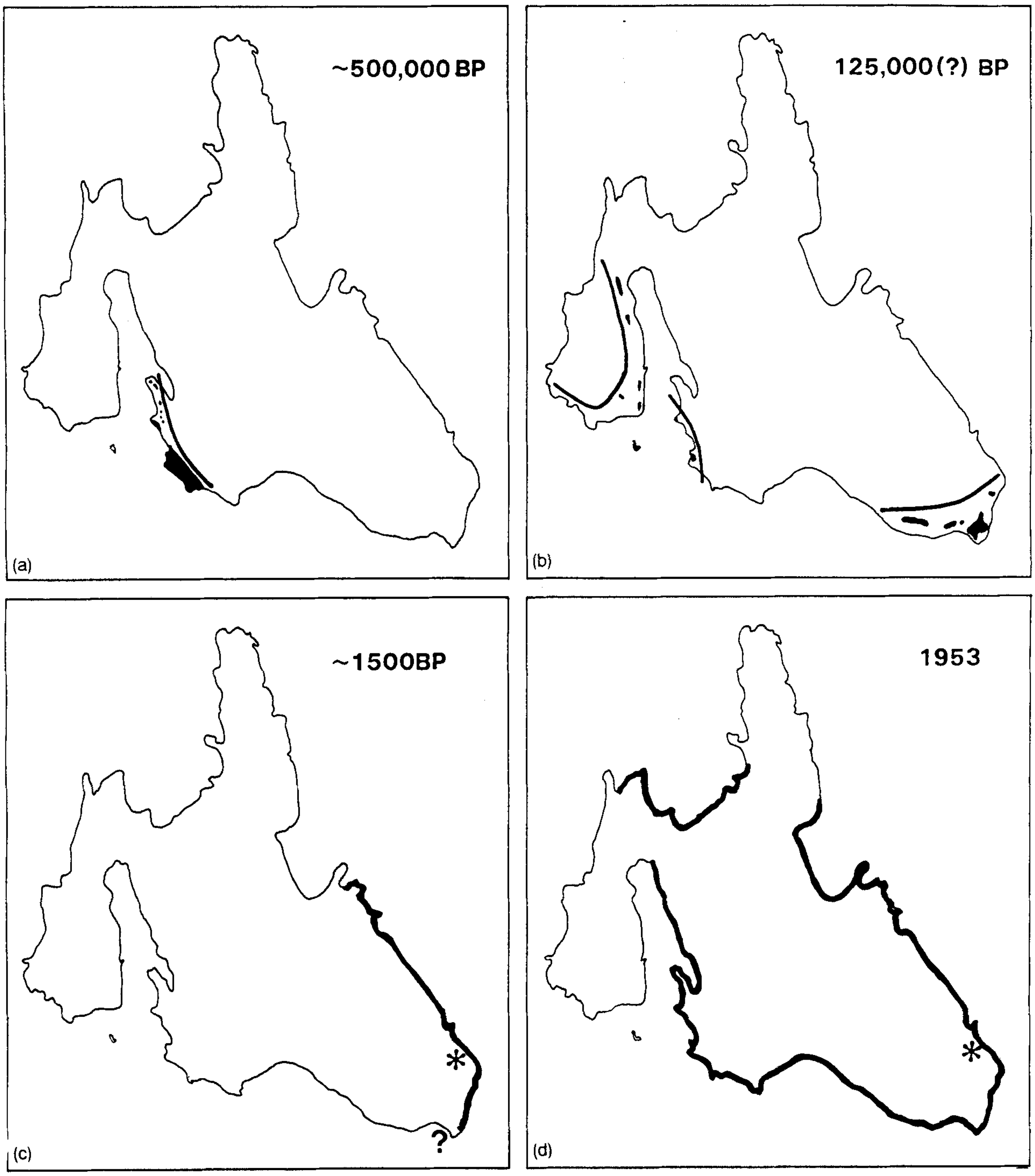

Figure 7. Upper Quaternary fossil shorelines in Cephalonia. Around (?)500 000 yr BP and (?)125 000 yr BP shorelines are drawn approximately from the present-day distribution of palaeo-Milazzian and Tyrrhenian sediments (shown as dots and black areas, after Sorel 1976; Braune 1973; BP Co. Ltd *et al.* 1985). The 1500 yr BP shoreline may extend to the west, along the south coast, where lithology does not permit preservation and identification of fossil Holocene shorelines. A star indicates the location of radiocarbon-dated samples of Table 2.

collected from the two exposed shorelines in various parts of the island are in progress) revealed that this shoreline was active just before the 1953 earthquake, as will be discussed more in detail elsewhere (Pirazzoli *et al.*, in preparation).

4.1 Biological mean sea-level and fossil shorelines

The amplitude of the 1953 co-seismic uplift was estimated on the basis of marine biology observations, and more explicitly on the concept of the *biological mean sea-level* (BMSL). The latter is defined as the limit (usually a sharp horizontal line) between the infralittoral and midlittoral biological zones. The upper limit of the infralittoral zone is defined as the upper limit of vermetids *Dendropoma petraeum*, *Vermetus triqueteur*, *Lithophaga* bivalves, Clionid sponges, *Paracentrotus*, etc., and as the lower limit of *Patella, Balanus*, *Cyanobacteria*, etc. (Fig. 8). In favourable conditions, a few years can be sufficient to the

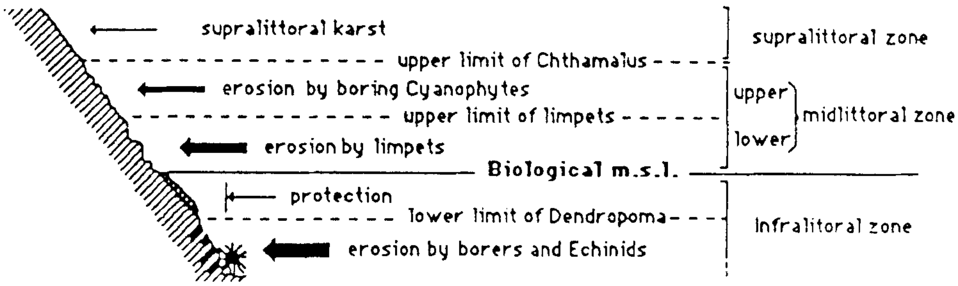

Figure 8. Biological zonation and definition of the mean biological sea-level (MBSL), not necessarily coinciding with that deduced from tide-gauge records (after Peres & Picard 1964). When biological remains of an uplifted shoreline exist, the amplitude of the relative sea-level change (uplift) equals the difference in height between fossil and active biological mean sea-level. The scale of the figure depends on various factors (exposure to winds, etc.) and differs from place to place, but the BMSL is independent of small-scale (mainly seasonal) fluctuations of the sea-level, and is defined with an accuracy sometimes better than 10 cm (Laborel 1979/80; Pirazzoli *et al.* 1991; Stiros *et al.* 1992; Laborel & Laborel-Deguen 1994).

development of a new BMSL on a bare hard limestone rock. The degree of this development is not the same everywhere, however, varying with the nature of the substrate and with the local ecological and physical conditions. The (absolute) height of the BMSL (and the scale of Fig. 8) depends on various factors (exposure to winds, etc.) and differs from place to place, but is independent of small-scale (mainly seasonal) fluctuations of the sea-level. If after a (relative) land uplift, some characteristic fossils are preserved (mainly *Dendropoma, Vermetus trigueteur, Lithophaga*, barnaches, etc.), a fossil BMSL can also be determined; the difference in elevation between fossil and active BMSL can give a reliable estimate of the amplitude of an uplift sometimes with an accuracy of ~10 cm. (Laborel 1979/80, 1987; Pirazzoli *et al.* 1991; Stiros *et al.* 1992; Laborel & Laborel-Deguen 1994).

4.2 Sea-level indicators in Cephalonia

Indeed, the best sea-level indicator in Cephalonia is *Dendropoma* (especially frequent along the eastern coast of the island and near Argostoli), which develops as a veneer over hard substrates, with a very narrow vertical zonation (generally less than 10 cm). The upper surface of this veneer is more or less dry at low tide (the tidal amplitude in Cephalonia is of the order of 10–20 cm) or during very calm weather. The occurrence of *Dendropoma* shells exposed in their growth position above sea-level is therefore sufficient to identify a former sea-level position with an accuracy of the order of ±10 cm. Consequently, an emergence of the order of 30 cm can be determined unambiguously, without overlaps between active and fossil BMSL.

Since *Lithophaga* may live as deep as several tens of metres, the presence of fossil perforations is generally considered to be an inaccurate sea-level indicator. However, as the upper limit of *Lithophaga* is the BMSL, the upper limit of elevated *Lithophaga* burrows can provide very reliable information on sea-level. This was the case, in particular, in the Myrtos Gulf, where energy conditions are generally insufficient for development of *Dendropoma*. In addition, preservation of *Lithophaga* shells inside fossil burrows, as in some places of the Myrtos Gulf, is generally an indication of co-seismic change in sea-level (Stiros *et al.* 1992; Laborel & Laborel-Deguen 1994).

5 FOSSIL EXPOSED COASTLINES

5.1 The 1953 coastline

Our observations of the 1953 co-seismic uplift are summarized in Fig. 9. The bulk of observations come from the south-east coast of Cephalonia, where the lithological conditions were favourable (compact limestones outcropping on the coast), and only a few measurements were possible in other sites.

At Poros, Galanopoulos' (1955) report of an uplift of 60–70 cm of the mushroom-shaped rock in Fig. 2 is consistent with our estimate of 50–70 cm in the wider area.

Some 3 km south of Poros, well-preserved *Dendropoma* shells, in their growth position at 0.5 m above the living counterpart, are clearly associated to the lowest exposed fossil shoreline, which is also marked by a clear notch (Fig. 10). In the same place, a second notch indicates a higher shoreline at about +1.2 m (see below).

Preliminary results of radiocarbon dating with the accelerator mass spectrography method, carried out by M. Arnold (Centre des Faibles Radioactivities, Gif-sur-Yvette, France) on a *Dendropoma* shell collected at +0.5 m (Table 2; Figs 7 and 10) indicate a conventional age (without reservoir correction) of 450 ± 60 yr BP, which is of the same order of magnitude with the apparent age of seawater before the beginning, in 1956, of the 'bomb effect'. This confirms that the lowest exposed shoreline was active just before the 1953 earthquake.

At Katavothres, near Argostoli, our estimation, using exposed vermetid evidence, of a 1953 uplift of 30–35 cm, is in close agreement with the reports of Mueller-Miny (1957) for an uplift of 30 cm and with the fact that the mill seawater wheels (one of which is still preserved) could not be used after the sudden uplift.

Along the southern coast the outcropping soft Neogene and Quaternary sediments do not preserve marks of exposed fossil Holocene shorelines; however, according to Galanopoulos (1955) and Mueller-Miny (1957) this coast was uplifted about 50 cm in 1953; from our experience from the other sites we suggest that this estimate is reliable and precise.

Lithological conditions were not favourable for similar observations along the coasts of the western (Lixouri)

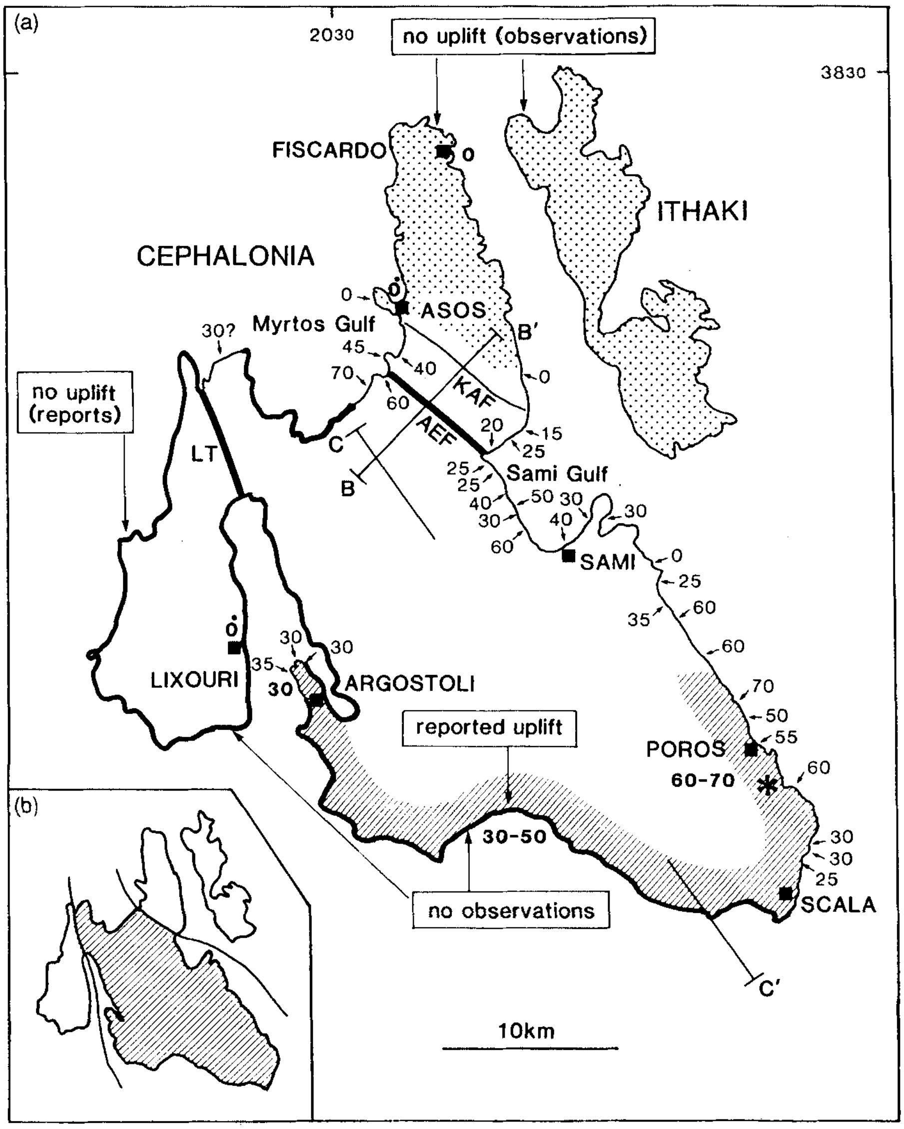

Figure 9. (a) Summary of observations of the 1953 uplift (in centimetres). Numbers with arrows: our observations; bold numbers: estimations of co-seismic uplift from reports of Galanopoulos (1955) and Mueller-Miny (1957); bold number with star: our estimate for a land movement in certain harbours, based on various 1953 reports. A solid coastline indicates lithology not permitting observations of Holocene raised beaches. The hatched coast is that reported to have been uplifted in 1953; the dotted area is where local lithology and morphology reveal that no Holocene raised beaches exist. The trace of the Livadi Thrust (LT), Ayia Ephemia Fault (AEF) and Kalon Anticline Fault (KAF) are also shown. A star indicates the location of radiocarbon dated samples of Table 2. B–B′ and C–C′ are the axes discussed in Figs 13 and 15. (b) Summary map of the 1953 uplift (hatched area).

peninsula, for which no co-seismic motion in 1953 was reported. We believe that no co-seismic uplift took place there, for it would have been easily identified in the various harbour installations; however, the possibility of a small (up to a few centimetres) subsidence in this peninsula cannot be excluded.

No sign of uplifted fossil shorelines exists along the northernmost coasts of the island (Fiscardo area), where compact limestones that can preserve signs of fossil shorelines outcrop; since there were no reports of a co-seismic uplift (especially in harbour installations) and the seismic intensity and damage were considerably less than elsewhere in the island, we conclude that no seismic uplift took place in this area. Yet, a subsidence of up to a few centimetres may have occurred.

Emerged shorelines are completely absent from the island of Ithaki, as our survey by boat indicated.

The similarity of the amplitude of the co-seismic displacements deduced from our geomorphological and biological observations and those reported just after the earthquake, indicates that in the 40 years following the 1953 earthquake no significant post-seismic displacement occurred. The long recurrence intervals for 1953-type earthquakes may provide an explanation for that.

5.2 Upper Quaternary shorelines

A nearly continuous, up to 1.2 m high, apparently older shoreline with very scarce (already eroded) fauna remains, similar to that of the 1953 shoreline, is observed along the

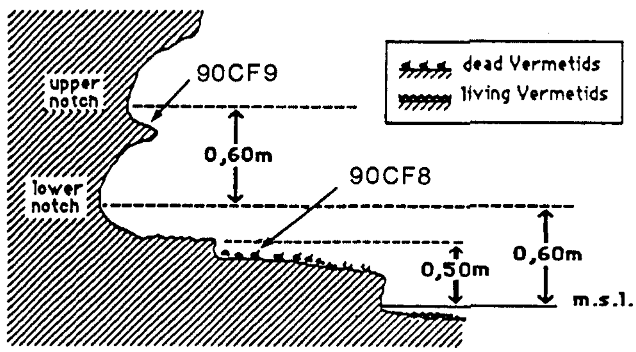

Figure 10. Coastal profile at 3 km south of Poros, showing two superimposed raised shorelines. The lower notch corresponds to the shoreline active just before the 1953 earthquake. The position of samples 90CF8 and 90CF9 (collected a few tens of metres away) are shown. The location of the section is marked with a star in Figs 7 and 9.

south-east coast of the island (Figs 2 and 7). No emerged signs of it exist in the other parts of the island, even where the lithological conditions are favourable for the preservation of its geomorphological and biological expression, for example in the Katavothres area. However, it is possible that it extended along the south coast of the island, south-east of Argostoli.

In the same place where sample 90CF8 was collected (Figs 7 and 10), an isolated *Vermetus triqueteur* shell was found in its growth position on the walls of a former pool at the height of 1.1 m, just below the level of the base of the higher notch of Fig. 10. Preliminary results of AMS radiocarbon dating of this sample by M. Arnold (Gif-sur-Yvette, France) indicate a conventional age of 1820 ± 70 radiocarbon years (Table 2), suggesting that the higher notch was uplifted probably between the fourth and sixth century AD. For the reasons discussed earlier (preservation of fauna of the former infralittoral zone), this uplift was probably episodic (conspicuously co-seismic).

In the literature there is also evidence of fossil Pleistocene shorelines. At different sites along the southern coast of the island (Fig. 7b), at heights varying between 6.5 and 21.6 m, Braune (1973) and BP Co. Ltd *et al.* (1985) reported marine sediments, locally forming terraces, which were assigned a Tyrrhenian age (125 Ka). Although the characteristic fossil *Strombus bubonius* was not found, a Tyrrhenian age may be assumed for these sediments, in analogy with those of the opposite Peloponnesian coast (Keraudren 1971).

Table 2.

Sample	AMS ^{14}C conventional age (yr BP)	Height (m)	Dated fossil
90CF8	450 ± 60	0.5	*Dendropoma*
90CF9	1820 ± 70	1.1	*Vermetus triqueteur*

Preliminary radiocarbon dating by M. Arnold (CNRS-CEA, Gif-sur-Yvette, France), without reservoir correction. For location see Figs 5 and 7. Additional dates and details are in progress and will be reported by Pirazzoli *et al.* (in preparation).

In various parts of the island, Braune (1973) also marked four other higher fossil shorelines, ranging up to the height of 200 m. Sorel (1976), however, showed that in the area south-east of Argostoli, marine sediments above 100 m are of Pliocene age, while the lower ones are palaeo-Milazzian (about 500 000 yr BP), and their variations in height result from folding. Furthermore, from our survey we have not been able to confirm the existence of fossil Pleistocene shorelines in locations other than those indicated in Fig. 7. From these data, approximate shorelines of the corresponding periods have been drawn (Fig. 7).

6 PATTERN OF THE 1953 UPLIFT

The area that was uplifted in 1953 is confined to the southern and central part of the island (Fig. 9b). In the following, we shall present evidence that (1) the uplifted area is bounded by two subparallel and homothetic, segmented thrusts, (2) seismic uplift reflects reactivation of the thrusts, and (3) the pattern of the 1953 uplift is different from the long-term deformation pattern of the island.

6.1 The western boundary of seismic uplift

To the north-west edge of the island, the uplift seems to be bordered to the west by the east-dipping Livadi Thrust (LT), the only major fault to be active in the Upper Quaternary in the area (Figs 4a and 9; Sorel 1976). The following lines of evidence indicate that this thrust extends southwards, and can be identified with the Gulf of Argostoli Thrust (GAT); this modifies previous views (e.g. Sorel 1976) that the Livadhi Thrust abuts to the Argostoli Fault (AF in Figs 1 and 4a).

(1) The landscape of the Ionian Islands is young and definitely fault-controlled, and the depression of the Argostoli Gulf is unlikely to be an exception; a major fault along the whole length of the gulf (a faulted syncline according to Nikolaou & Armoutidis 1979) is therefore expected.

(2) The discontinuity in the seismic intensities (Fig. 6) implies a major structural discontinuity along the gulf.

(3) The geology and the geomorphology around the gulf (Figs 1 and 11) are consistent with the hypothesis of a major thrust: older formations and uplift are observed to the east (hanging wall), younger formations, tilting and subsidence to the west (footwall; Fig. 11c). Furthermore, minor faulting, indicative of internal deformation of the hanging wall is expected, and in fact has been observed in the Argostoli area (Sorel 1976; Underhill 1989), among them the Argostoli Fault (AF in Figs 1 and 4a). This last fault cannot be the major feature in the vicinity of the Argostoli Gulf, as previous workers suggest, for the following reasons: first, it was inactive from middle Pleistocene (Sorel 1976), and consequently not related to the late Quaternary uplift in the Argostoli area; second, the observed uplift at its footwall (Fig. 7) is not explained; and third, the wavelength of elevation changes in this last area is very small relative to the other parts of the island (Fig. 11), indicative of a secondary deformation.

(4) An escarpment 20 m high and 15 km long is seen along the east coast of the Argostoli Gulf in all bathymetric

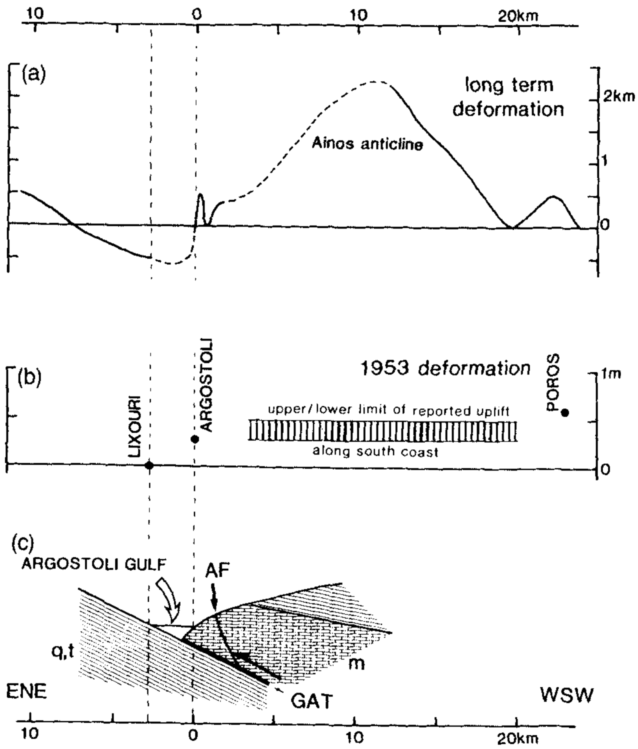

Figure 11. (a) Variation of the amplitude of vertical motions in Cephalonia since Lower Pliocene, along an axis perpendicular to the average strike of the LT–GAT. A dashed line indicates uncertain or inferred data. This deformation is likely to reflect folding associated with the Gulf of Argostoli thrust (GAT, see (c); compare for example with Mansinha & Smylie 1971). Small amplitude variations of the uplift curve near Argostoli reflect secondary faulting (AF in (c), among others). Based on data from Sorel (1976). (b) Seismic motion along the same axis, probably indicating uplift and tilting of a quasi-rigid block. The difference in the style from the long-term deformation is evident. (c) Schematic geology and geomorphology along the Argostoli Gulf. (Partly based on data from Galanopoulos 1955; Voutetakis 1954; Braune 1973; Sorel 1976; Bornovas & Rondogianni 1983.) Differences in the dip of Quaternary (q) and Late Tertiary (t) sediments west and east of the Argostoli gulf are schematically shown.

(echosounder) profiles collected in this gulf (Fig. 12, profiles VIII–XIV). The morphology of the scarp, according to Braune (1973), is indicative of very young faulting bringing into tectonic contact the Mesozoic rocks of the Argostoli area and the eastward-dipping Tertiary and Quaternary formations exposed on the Lixouri peninsula; apparently the surface trace of a thrust. This escarpment is unlikely to reflect older events: after ~100 m differential uplift since the Mindel-Riss period and 600 m since Early Pleistocene documented for this area (Sorel 1976), an old scarp of this size along this closed gulf would have long since been obliterated and buried, especially since the rainfall average in the area is significant, $\sim 1\,\mathrm{m\,yr^{-1}}$ (Bornovas 1964), and the western peninsula is built of easily eroding Tertiary sediments, forming badlands.

Since on the other hand, the seismic uplift is confined to the hanging wall of the LT-GAT, it can be concluded that these thrusts reactivated in 1953.

6.2 The eastern boundary of seismic uplift

The north-east limit of the 1953 uplift broadly correlates with the Ayia Ephemia Fault (AEF), the only major and active fault in the area (Sorel 1976). Some minor uplift continues attenuated farther north, both along the Myrtos and Sami Gulfs (Fig. 13), up to the trace of the Kalon Anticline Fault (KAF), a satellite fault in the hanging wall of the main fault (Figs 1, 4 and 9). This minor uplift reflects a secondary deformation; its geometry and kinematics are explained in Figs 5 and 13.

Marine reflection data (seismic section MEDOR 19 in Cushing 1985) indicate that AEF extends to the north-west as a prominent east-dipping thrust (Fig. 4). There is evidence that AEF extends (segmented) to the south-east as well.

(1) The discontinuity in seismic intensities along the straits between south-west Ithaki and Cephalonia (Fig. 6) implies a major structural discontinuity along these straits.

(2) The available bathymetric data offshore the uplifted south-east coast of Cephalonia indicate a 300 m steep bathymetric escarpment (Fig. 12a), probably a thrust ramp (thereafter the Offshore Poros Thrust, OPT), which is clearly seen in a seismic line 6 km offshore Poros (Fig. 4; line PT132 in Nikolaou & Armoutidis 1979). Farther south-east, this thrust is likely to abut to one of the thrusts identified in the reflection profiles of Brooks & Ferentinos (1984) farther south; these last thrusts are also sites of diapirism (Figs 4 and 14).

Except for the inferred tectonic control of the seismic uplift, which is confined to the footwall of the (segmented) thrust, there is additional evidence for its reactivation in 1953. For weeks after the earthquakes a cloud of white dust emerged from the sea in Myrtos Gulf. Sorel (1976) explained this white dust as result of the rise of gas mixed with the product of friction of the carbonate rocks due to seismic movements, obviously along this fault. Such fine-grained mylonitization of near-surface limestones is not unusual both on the island and the Greece mainland.

It must also be noted that because of the internal deformation of the hanging wall of the faults that are inferred to have reactivated in 1953, no important surface faulting is expected to have occurred, nor was any reported. Certainly, there are some vague reports of 'seismic cracks' in various parts of the island, which may in fact testify to seismic surface faulting; however, these reports are without any tectonic value.

6.3 Seismic versus long-term deformation

While the long-term deformation of the central part of the island is associated with folding (Fig. 11), from Figs 9, 11 and 15 it is likely that the uplifted fault-bounded block in 1953 was tilted to the west without important internal deformation; especially, there is no evidence of folding similar to the long-term one. This result may be extrapolated for the whole Upper Quaternary period. There is some evidence that the tectonic event at around 1500 yr BP discussed above was also associated with a westward tilting of the central part of the island

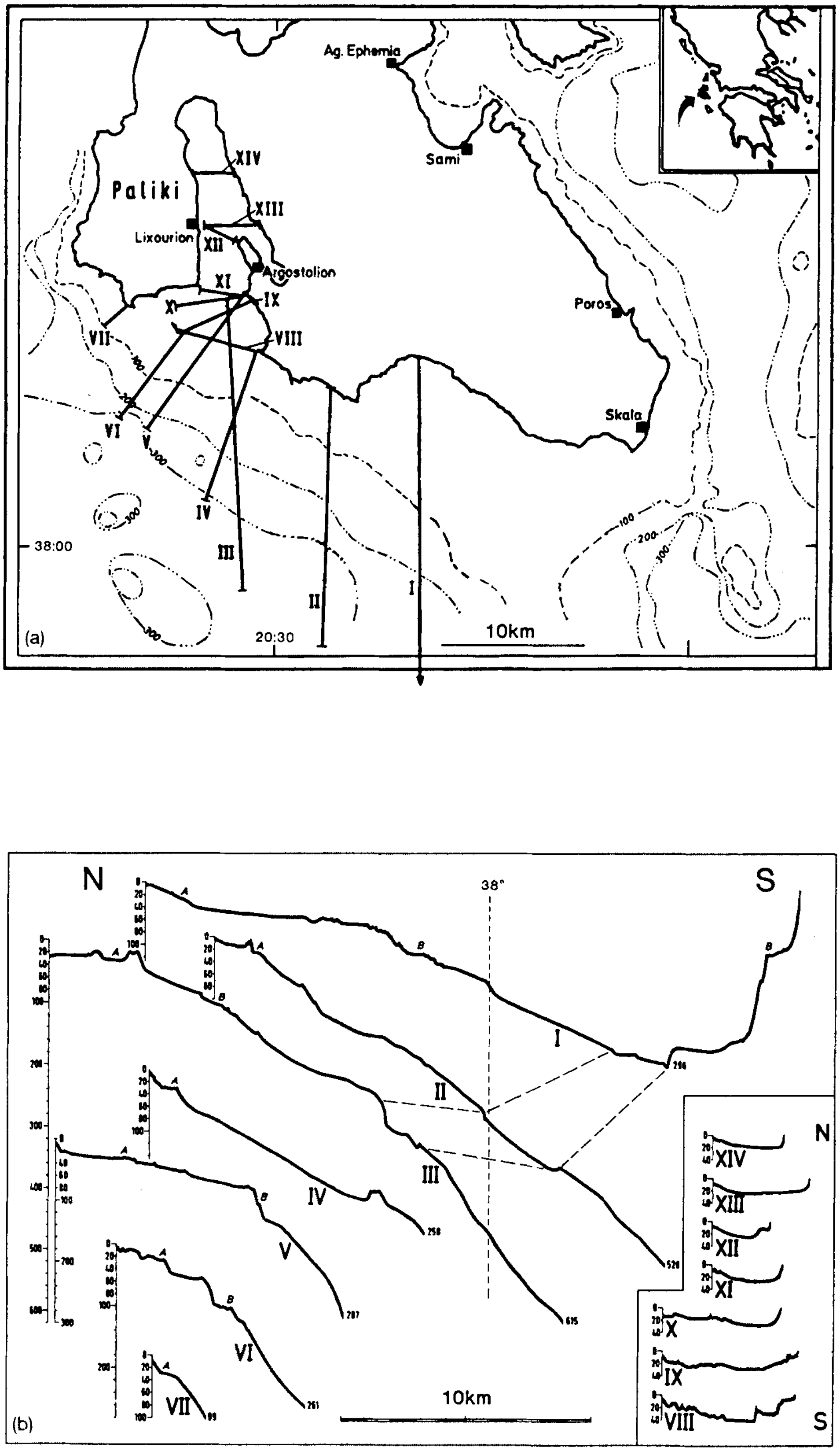

Figure 12. (a) Location map of (b) bathymetric profiles in the Gulf of Argostoli and the south-west Cephalonia shelf (after Braune 1973, simplified).

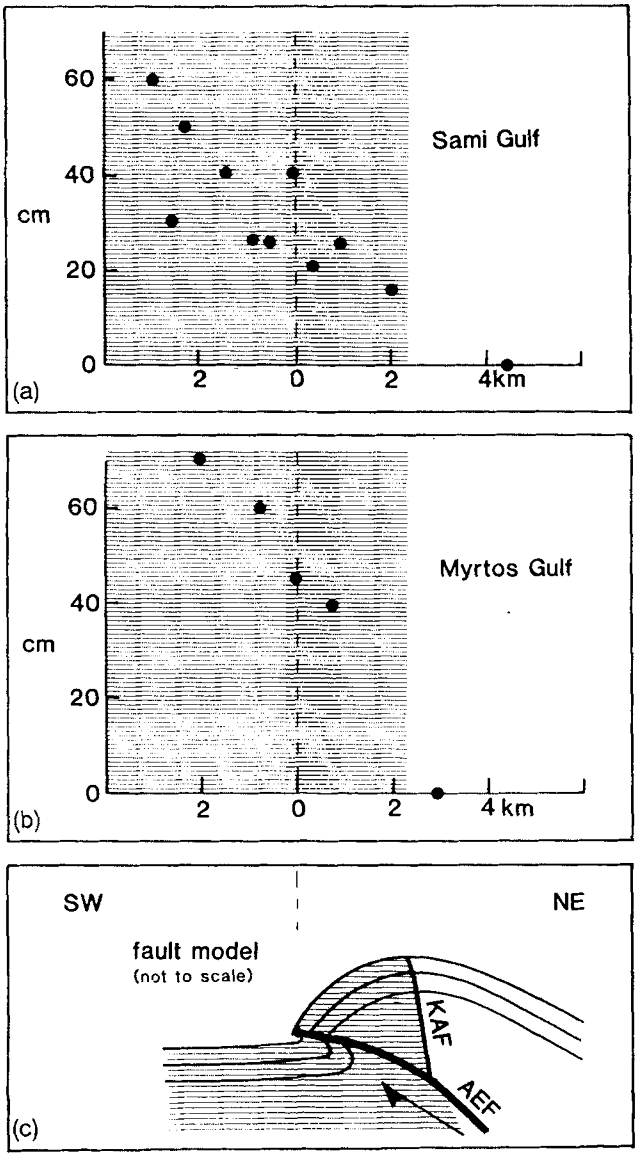

Figure 13. Uplift observed along the Sami Gulf (a) and Myrtos Gulf (b) versus distance from the surface trace of Ayia Ephemia Fault (AEF). Values projected along axis B–B′ in Fig. 9. The geometry (after Sorel 1976, modified) and the suggested direction of movement of AEF is shown in (c). The uplifted area is shaded and extends up to the Kalon Anticline Fault (KAF). Compare with Fig. 5.

(Pirazzoli *et al.*, in preparation), while the distribution of Upper Quaternary marine sediments (Fig. 7) does not provide evidence of large-scale folding.

7 DISCUSSION

The active tectonic style of the wider area, and the available, though poor, focal mechanism of the 1953 earthquake (see above) indicate that the seismic deformation was probably associated with a thrust in the basement. However, our study reveals that the surface deformation is different, and may be described as a *piston-like uplift of a nearly rigid block*, bounded by two subparallel and homothetic segmented faults that moved as a thrust and a normal (separation) fault, respectively (Figs 16a and b).

Parallel thrust and normal faults have been observed in several areas (e.g. Dalmayrac & Molnar 1981), including the Ionian Sea (Cephalonia, Nikolaou 1986; Leukas, Bornovas 1964; Strophades, Fig. 16c). Yet it is difficult to understand how a thrust and a normal fault with the same strike can be reactivated by the same earthquake. Or in other words, it is difficult to imagine a stress field that can produce such a structural peculiarity.

This problem, however, refers to nearly elastic, isotropic and homogeneous media. But is this the case with Cephalonia? We suggest that the answer is probably no, for the crust in the study area is highly anisotropic and inhomogeneous, with ductile layers not only subhorizontal (as is the case with other areas, where the sedimentary cover is partly (Jackson 1980; Jackson & White 1989) or totally (Makarov 1982) decoupled from the basement), but also oblique to the surface (Figs 4, 5, 13 and 16b); these last layers (salt intrusions along main thrusts) are responsible for the observed anomalies in the propagation of seismic waves (Fig. 6).

It is therefore reasonable to suggest that (at least in certain cases) the deformation of such media cannot be approached by standard, elastic dislocation uniform slip models, and alternative explanations based on the theory of salt tectonics are required.

According to this theory, basement shortening and earthquakes tend to mobilize salt bodies (Trusheim 1960; Jackson & Talbot 1986), whose movement and deformation is conveyed to the overburden. The deformation of the latter is directly related to the evolving morphology of salt structures and is characterized by a complex pattern of extensional, compressional and gravitational features (Jen-

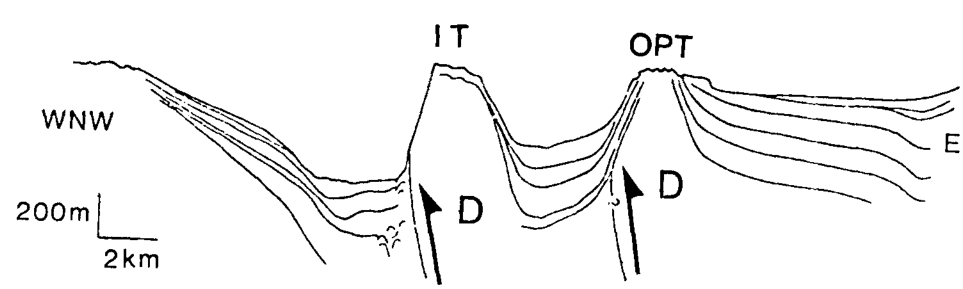

Figure 14. Interpreted single-channel reflection profile south-east of Cephalonia that indicates two east-dipping thrusts associated with uplifted evaporites (D). The two thrusts correlate with the Offshore Poros Thrust (OPT) and the Ionian Thrust (IT) respectively. After Brooks & Ferentinos (1984), modified. For location, see Fig. 4(a).

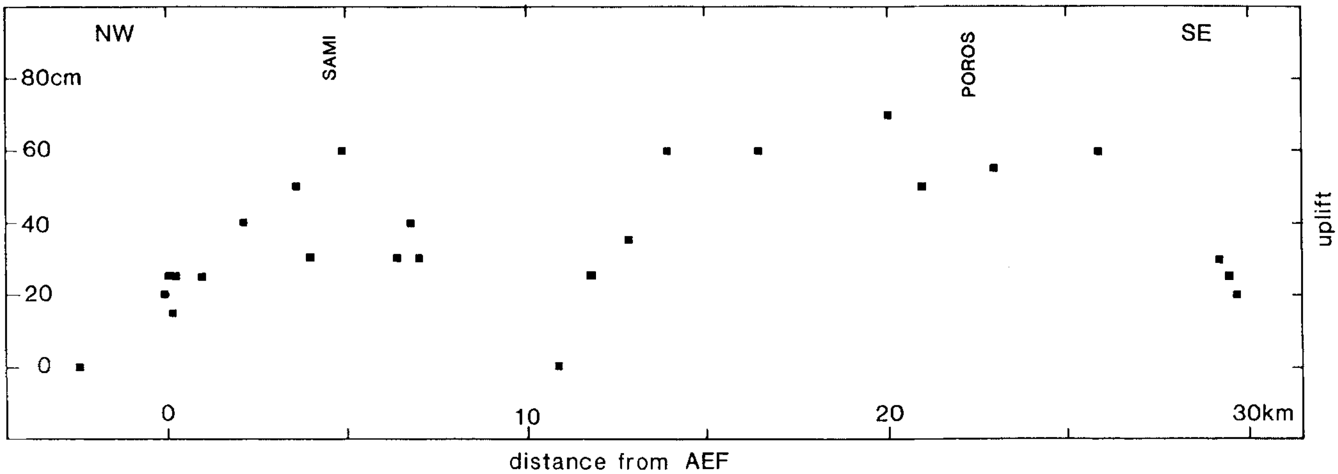

Figure 15. Variation of uplift observed along the south-east coast of Cephalonia; values projected along axis C–C′ of Fig. 9.

yon 1988)—a 'structural peculiarity' to cite Trusheim (1960). Obviously, such *halotectonic* features do not provide *direct* information on the regional stress field.

The inferred structural peculiarities of the surface seismic deformation of the overburden of an evaporite basin lead us to assume that salt had a double, key role in this seismic thin-skinned style deformation. First, salt, usually flowing at low strain rates, responded as a quasi-rigid body to the high strain rates during the earthquake (compare with the hypothesis of Jackson & White 1989 for the ductile transitional layer between the upper and lower crust) and transmitted the basement deformation to a fault-bounded block of the overburden. Second, evaporite intrusions along the thrust sheets act as lubricants, facilitating the piston-like movement (uplift) of a certain block, and preventing its internal deformation. This pattern is schematically shown in Fig. 16(b) and is compatible with both the models of Figs 4(a) and (b).

There is still another point to clarify. The available data (e.g. Fig. 11) reveal that the long-term deformation of Cephalonia can be approached by models of dislocations in elastic media, but not the seismic one. A probable explanation is that most of the relief of the island was formed in the Pliocene and Lower Pleistocene (Sorel 1976), when diapirism was less mature, and the crust more elastic and isotropic.

8 ALTERNATIVE MODELS

The proposed model is not unique, but is simple, fits well the available data, is compatible with the tectonic structure of the island and the surrounding region, as well as with the theory of evaporite migration, and explains the observed anomalous attenuation of seismic waves during the 1953 and historical earthquakes.

Furthermore, its parallels are found in everyday life; for example, the uplift of paving tiles in sidewalks: roots of trees push the tiles upwards as rigid blocks, overwhelming the strength of the cement of the tile joints.

Yet, some alternative mechanisms for the surface, seismic deformation must be considered. A possibility is that the uplifted block moved in response to lateral contraction. Low-friction faulting is the key point in this hypothesis also, which requires that the eastern fault has a higher dip; this is not unreasonable (see Section 2.4), but it is difficult to explain why the observed uplift to the west is smaller than to the east, and what the mechanism for the shortening of the overburden is, if the focus of the earthquake was in the basement.

In addition to the above models which emphasize the role of evaporites, alternative models can be proposed: (1) folding above the GAT–LT or a somewhat similar, blind thrust, (2) development of a positive flower structure and (3) a system of a fault and a transfer fault.

The first of these models, based on the elastic-dislocation uniform-slip concepts (e.g. Mansinha & Smylie 1971) predicts folding of the central part of the island, and subsidence in the Argostoli Gulf area. This pattern is in excellent agreement with the long-term deformation of the island (Fig. 11), but is questionable if it can describe the 1953 co-seismic deformation: our observations give no evidence of folding, no significant subsidence is observed in the Lixouri area, while uplift is much larger in the Poros than in the Argostoli area; furthermore, the strong correlation between the boundaries of uplift and faulting is not explained.

An alternative model, inspired from small-scale structures seen in the field (e.g. Fig. 37 in Sorel 1976), is that the eastern boundary of the seismic uplift line corresponds to a reverse, west-dipping fault (a positive flower structure model); yet, this hypothesis is not consistent with the tectonic fabric of the island (Fig. 4).

A third possibility, again inspired from the modelling of nearby faults by Sorel (1976) is that the LT–GAT can be regarded as a faulted fold, or as a thrust, with the AEF as a transfer fault which accommodates a certain amount of shortening because it is oblique to the master (LT–GAT)

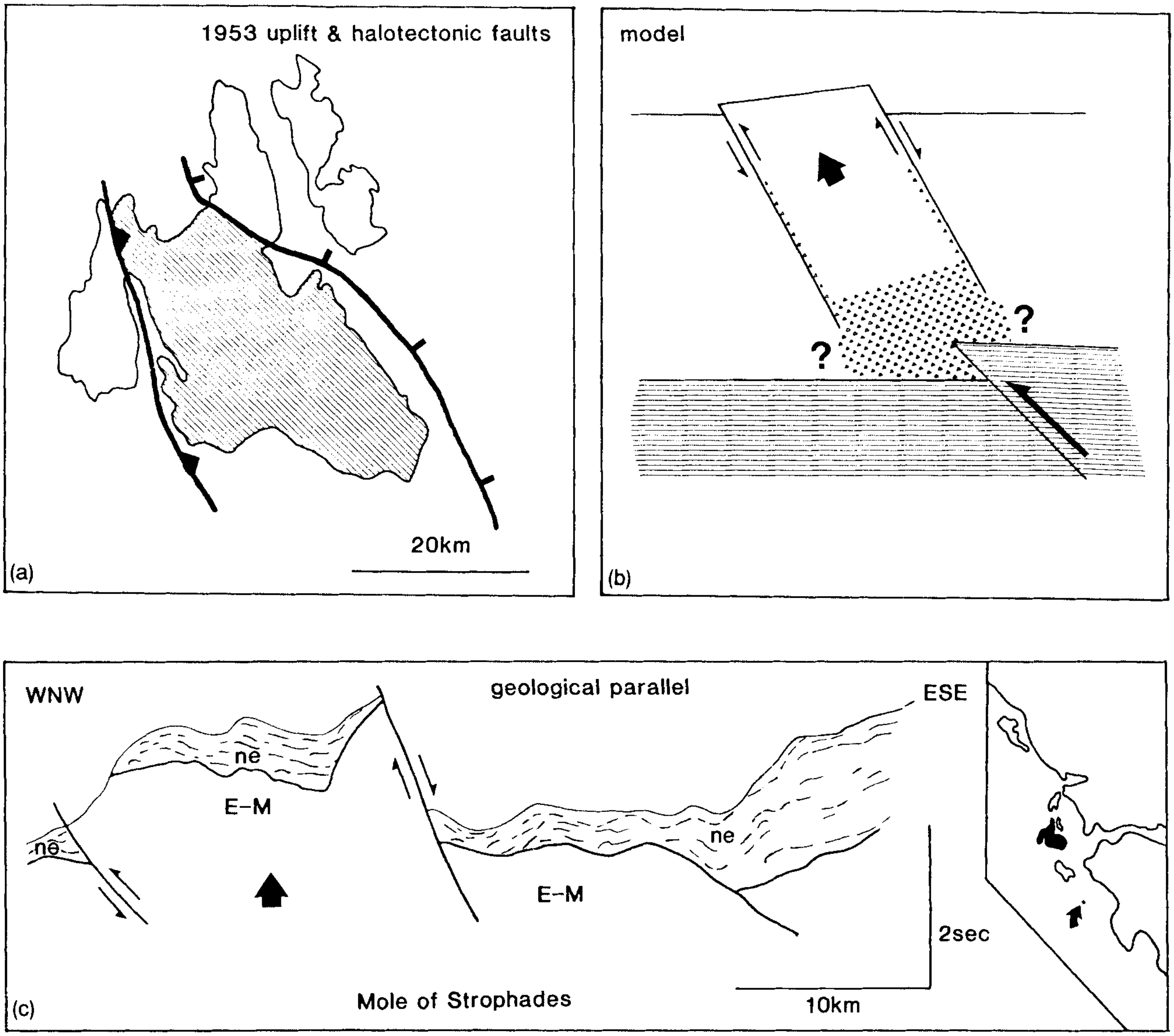

Figure 16. (a) Geometry of surface halotectonic seismic faults (triangles indicate thrusts, ticks normal faults) and uplifted block (hatched area) during the 1953 earthquake. (b) A sketch to explain our model for the mechanism of the 1953 earthquake and probably at least one other Holocene event. Faulting in the basement (hatched area) mobilized a salt body (marked by small triangles) that was uplifted, pushing upwards the overburden as a rather rigid block; the uplift was accommodated by two pre-existing homothetic thrusts which reactivated as halotectonic faults. For simplicity, faulting is shown planar to the surface, but in the uppermost levels the deformation pattern is complicated (see Fig. 5). Geometry of faulting in the basement is taken to conform with the mechanism of McKenzie (1972). The amplitude of uplift is 30–70 cm, the thickness of the overburden and the halite body up to 5 km and up to 1.5 km, respectively. (c) Geological interpretation of an E–W striking Sparker profile just north of the Strophades Isles, after Lyberis & Bizon (1981) and Nikolaou (1988), simplified. ne, Neogene; E–M, Triassic evaporites and Mesozoic basement. Mark the similarity with the proposed pattern of seismic deformaton in (b). This structural pattern is not unusual in halite environments. In the inset, an arrow points to the Strophades Isles. Cephalonia is shown in Black.

fault (see Cushing 1985; for an extensional parallel, see Ambraseys & Tchalenko 1972).

The major difficulty with this model is that uplift is much larger along the south-east coast of the island than in the Argostoli area, while there is no evidence of right-lateral movement along the AEF, as the pattern of streams in the 1:5000 scale topographic maps reveals.

All these last three models, however, ignore the décollement and the 2-D crustal anisotropy produced by salt bodies.

9 CONCLUSION

Geomorphological and marine biological data, in combination with eye-witness reports of the seismic uplift and radiocarbon datings, indicate that the 1953 earthquakes were responsible for a ~30–70 cm uplift of the central part of Cephalonia. The uplifted part is bounded by two subparallel and homothetic major thrusts. Obviously, this deformation pattern is structurally peculiar, and can be better explained if we assume that the surface deformation reflects a halotectonic deformation. This hypothesis is in excellent agreement with our knowledge of the geologic structure of the area, and indicates that the seismic deformation of the uppermost crustal strata in the area may mimic the style of the long-term halotectonic deformation.

ACKNOWLEDGMENTS

We thank M. Arnold (Centre des Faibles Radioactivities, Gif-sur-Yvette, France) for ^{14}C AMS dating. The text has benefited from the careful reviewing by J. A. Jackson, D. B. Slemmons and an anonymous reviewer. S.C.S. acknowledges stimulating discussions with J. Jackson, J. Bornovas, V. Makarov, M. Loucogiannakis and G. Valensise.

This is a contribution to the Active Geodynamics Project of IGME (DE-8661704-IGME), to the DGXII Project of the CEC 'Climate change, sea-level rise and associated impacts in Europe' (contract EPOC-CP-90-0015), to the Franco-Hellenic Scientific Cooperation Programme, to the task group 'Paleoseismicity of the Late Holocene' of the Inter-Union Commission of the Lithosphere, and to the IGCP Project No. 367 'Late Quaternary coastal records of rapid change'.

REFERENCES

Ambraseys, N. & Tchalenko, J., 1972. Seismotectonic aspects of the Gediz, Turkey, earthquake of March 1970, *Geophys. J. R. astr. Soc.*, **30**, 229–252.

Anderson, H. & Jackson, J., 1987. Active tectonics of the Adriatic region, *Geophys. J. R. astr. Soc.*, **91**, 937–983.

BP Co. Ltd, 1971. *The Geological Results of Petroleum Exploration in Western Greece*, Publication No. 10 Institute of Geology and Subsurface Research, Athens.

BP Co. Ltd, Bergmann, H., Braune, K., Dremel, G., Hatzopoulos, E., Hug, F., Uliczny, E. & Migiros, G., 1985. *Geological Map of Greece*, 1:50 000 scale, Cephalonia Island (Northern and Southern parts), Institute of Geology and Mineral Exploration, Athens.

Braune, K., 1973. Die rezenten und pleistozaenen Sedimente des Sublitorals von Kephallinia (Ionische Inseln), *Senckenbergiana Maritima*, **5**, 99–133, Frankfurt.

Bornovas, J., 1960. Observations nouvelles sur la géologie des zones préapulienne et ionienne (Grèce occidentale), *Bull. Soc. Géol. France*, ser. 7, **2**, 410–413.

Bornovas, J., 1964. Geological study of Leukas Island, *Geol. Geophys. Res., IGSR*, 10/1 (in Greek with extended abstract in French).

Bornovas, J. & Rondogianni, T., 1983. *Geological Map of Greece*, 1:500 000 scale, Institute of Geology and Mineral Exploration, Athens.

Brooks, M. & Ferentinos, G., 1984. Tectonics and sedimentation in the gulf of Corinth and the Zakynthos and Kefallinia channels, *Tectonophysics*, **101**, 25–54.

Cushing, E., 1985. Evolution structurale de la marge nord-ouest Héllénique dans l'île Levkas et ses environs (Grèce Nord-Occidentale), *PhD thesis*, Université de Paris-Sud, Centre d'Orsay.

Dalmayrac, B. & Molnar, P., 1981. Parallel thrust and normal faulting in Peru and constraints on the state of stress, *Earth planet. Sci. Lett.*, **55**, 473–481.

Dardeau, G., Fortwengler, D., de Graciansky, P.-C., Jacquin, T., Marchand, D. & Martinod, J., 1990. Halocinèse et jeu de blocks dans les Baronnies: Diapirs de Propiac, Montaulieu, Condorcet (Département de la Drôme, France), *Bull. Centres Explo. Prod. Elf-Aquitaine*, **14**, 111–151.

Galanopoulos, A., 1955. Seismic geography of Greece, *Ann. Géol. des pays Hélléniques*, **6**, 83–121.

Guzzetta, G., 1982. Thin-skinned style of deformation in Epirus, in *Proceedings, Int. Symp. on the Hellenic Arc and Trench (HEAT)*, vol. 1, pp. 151–175, National Technical University, Athens.

Hatzfeld, D., Pedotti, G. Hatzidimitriou, P. & Makropoulos, K., 1990. The strain pattern in the western Hellenic arc deduced from a microearthquake survey, *Geophys. J. Int.*, **101**, 181–202.

IGRS (Institute de Géologie et des Recherches du Sous-sol) & IFP (Institut Français du Pétrole), 1966. *Etude Géologique de l'Epire*, Editions Technip, Paris.

Jackson, J., 1980. Reactivation of basement faults and crustal shortening in orogenic belts, *Nature*, **283**, 343–346.

Jackson, P.A. & Talbot, J., 1986. External shapes, strain rates, and dynamics of salt structures, *Geol. Soc. Am. Bull.*, **97**, 305–323.

Jackson, J. & White, N., 1989. Normal faulting in the upper continental curst: observations from regions of active extension, *J. struct. Geol.*, **11**, 15–36.

Jackson, J., Fitch, T. & McKenzie, D., 1981. Active thrusting and the evolution of the Zagros fold belt, in *Thrust and Nappe Tectonics*, pp. 371–379, eds McClay, K. & Price, N., The Geological Society of London Special Publication **9**, Blackwell Scientific Publications, Oxford.

Jenyon, M., 1988. Overburden deformation related to the pre-piercement development of salt structures in the North Sea, *J. geol. Soc. Lond.*, **145**, 445–454.

Kamperis, E., 1987. Geological and petroleum geology study of NW Peloponnesus, *PhD thesis*, Dept of Geology, National Technical University, Athens (in Greek).

Karakitsios, V., 1992. Ouverture et inversion tectonique du bassin Ionien (Epire, Grèce), *Ann. Géol. pays Hélléniques*, **35**, 185–318.

Keraudren, B., 1971. Les formations quaternaires marines de la Grèce- Deuxième partie, *Bull. Musée d'Anthropologie Préhistorique de Monaco*, **17**, 87–169.

Laborel, J., 1979/80. Les gasteropodes vermetidés: Leur utilisation comme marqueurs biologiques de rivages fossiles, *Oceanis*, **5**, (Fasc. hors série), 221–239.

Laborel, J., 1987. Marine biogenic constructions in the Mediterranean: A review, *Sci. Rep. Port-Cross Natl Park (France)*, **13**, 97–126.

Laborel, J. & Laborel-Deguen, F., 1994. Biological indicators of relative sea-level variations and of co-seismic displacements in the Mediterranean region, *J. coastal Res.*, in press.

Lyberis, N. & Bizon, G., 1981. Signification structurale des îles Strophades dans la marge Héllénique, *Mar. Geol.*, **39**, M57–M69.

Makarov, V., 1982. Structural and dynamic dysharmony of the lithosphere by neotectonic and geophysic data, Tien-Shan and adjacent regions (in Russian), in Tectonic layering of the lithosphere of neotectonic active belts, pp. 8–25, eds Makarov, V., Trifonov, V., Shchukin, J. & Kuchai, V., *Trans., Acad. Sci. USSR*, **359**, Nauka, Moscow.

Mansinha, L. & Smylie, D., 1971. The displacement fields of inclined faults, *Bull. seism. Soc. Am.*, **61**, 1433–1440.

Mascle, G., Arnaud, H., Dardeau, G., Debelmas, J., Delpech, P.-Y., Dubois, P., Gidon, M., de Graciansky, P.-C., Keckore, C. & Lemoine, M., 1988. Salt tectonics, Tethyan Rifling and Alpine folding in the French Alps, *Bull. Soc. Géol. France*, **8**, 747–758.

McKenzie, D., 1972. Active tectonics of the Mediterranean Region, *Geophys. J. R. astr. Soc.*, **30**, 109–185.

Mercier, J.-L., Delibassis, N., Gauthier, A., Jarrige, J., Lemeille, F., Philip, H., Sebrier, M. & Sorel, D., 1979. La néotectonique de l'arc égéen, *Rév. géol. dyn. géogr. phys.*, **21**(1), 67–92.

Monopolis, D. & Bruneton, A., 1982. Ionian Sea (western Greece):

its structural outline deduced from drilling and geophysical data, *Tectonophysics*, **83**, 227–242.

Mueller-Miny, H., 1957. Beitrage zur morphologie der mittleren jonischen Inseln, *Ann. Geol. pays Helleniques*, **8**, 1–28 and Tables I–XXV.

Nely, G., 1980. Faciès et morphologie sismiques des évaporites, *Bull. Centre Explor. Prod. Elf-Aquitaine*, **4**, 395–410.

Nikolaou, K., 1986. Contribution to the study of the Neogene and the geology and boundaries of Ionian and pre-Apulian isopic zines in relation to petroleum geology observations mainly in the Islands Strophades, Zante, Cephalonia, *PhD thesis*, University of Athens (in Greek).

Nikolaou, K., 1988. Noe-diapiric movements of Triassic evaporites in Zakynthos and Strophades, *Bull. geol. Soc. Greece*, **20**, 83–99.

Nikolaou, K. & Armoutidis, C., 1979. *Geological study of South Cephalonia*, Unpublished Report 060, DEP (National Petroleum Corporation), Athens.

Papageorgiou, S., Arnold, M., Laborel, J. & Stiros, S., 1994. Seismic uplift of the harbour of ancient Aigeria (central Greece), *Int. J. Naut. Archaeo.*, in press.

Papazachos, B. & Papazachou, K., 1989. *The Earthquakes of Greece*, Zitis, Thessaloniki (in Greek).

Peres, J. & Picard, J., 1964. Nouveau manuel de bionomie benthique en Mediterranée, *Rec. Trav. Station marine Endoume*, **31**, 47, 1-131.

Pirazzoli, P., 1986a. Marine notches, in *Sea-level research, A manual for the collection and evaluation of data*, pp. 361–400, ed. de Plaasche, O., Geobooks, Norwich.

Pirazzoli, P., 1986b. The early-Byzantine tectonic paroxysm, *Z. Geomorph. Suppl.*, **62**, 31–49.

Pirazzoli, P., Laborel, J., Saliège, J., Erol, O., Kayan, I. & Person, A., 1991. Holocene raised shorelines on the Hatay coasts (Turkey): palaeocological and tectonic implications, *Mar. Geol.*, **96**, 295–311.

Pirazzoli, P., Montaggioni, L., Saliege, J., Segonzac, G., Thommeret, Y. & Vergnaud-Grazzini, C., 1989. Crustal block movements from Holocene shorelines: Rhodes Island (Greece), *Tectonophysics*, **170**, 89–114.

Pirazzoli, P., Stiros, S., Arnold, M., Laborel, J., Laborel-Deguen, F. & Papageorgiou, S., 1994. Episodic uplift deduced from Holocene shorelines in the Perachora peninsula (Corinth area, Greece), *Tectonophysics*, **229**, 201–209.

Scordilis, E., Karakaisis, G., Panagiotopoulos, D. & Papazachos, B., 1985. Evidence for transform faulting in the Ionian Sea: The Cephalonia Island earthquake sequence of 1983, *Pageoph.*, **123**, 387–397.

Sorel, D., 1976. Etude néotectonique des îles Ioniennes de Céphalonie et de Zanthe et du l'Elide occidentale (Grèce), *Thèse 3e cycle*, Université Paris Sud, Orsay.

Stiros, S., 1994. Anomalous attenuation of seismic waves in the Ionian Islands (Greece), *EOS, Trans. Am. geophys. Un.*, in press.

Stiros, S., Arnold, M., Pirazzoli, P., Laborel, J., Laborel, F. & Papageorgiou, S., 1992. Historical co-seismic uplift on Euboea island, Greece, *Earth planet. Sci. Lett.*, **108**, 109–117.

Stiros, S., Pirazzoli, P., Laborel, J., Laborel, F., Arnold, M. & Papageoriou, S., 1994. Holocene sea-level changes in Euboea, *Bull. geol. Soc. Greece*, in press.

Trusheim, F., 1960. Mechanism of salt migration in Northern Germany, *Bull. Am. Ass. petrol. Geol.*, **44**, 1519–1540.

Underhill, J., 1988. Triassic evaporites and Plio-Quaternary diapirism in western Greece, *J. geol. Soc.*, **145**, 269–282.

Underhill, J., 1989. Late Cenozoic deformation of the Hellenide foreland, western Greece, *Geol. Soc. Am. Bull.*, **101**, 613–634.

Vogt, J. & Albini, P., 1992. An outstanding Ionian testimony, in *Abstracts, Regional Workshop on Archeoseismicity in the Mediterranean Region*, Damascus.

Voutetakis, S., 1954. *Subsurface exploration of the towns of Argostoli, Lixouri and Zakynthos*, Unpublished report, Inst. Geol. Subsurface Res., Athens (in Greek).