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Western Hellenic subduction and Cephalonia Transform: local earthquakes and plate transport and strain

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Abstract

Focal parameters of local earthquakes in the region of the Ionian Islands of western Greece are constrained with a temporary dense array of three-component seismographs operated jointly offshore and onshore. Seismic deformation is documented to be confined to the east of the N20°E-striking steep continental slope west of Cephalonia island, the right-lateral Cephalonia Transform Fault, CTF, inferred from large earthquakes. The pre-Apulian continental material appears to be only deforming east of the transform fault, where it is in upper plate position to the Hellenic subduction. East of the transform fault, the transmission velocity tomography from local earthquakes, compared in depth-section with a previous marine reflection profile, provides evidence in support of a shallow landward dipping boundary around 12 km deep under the Ionian Islands along which they may override the lower plate. On either side of this interface local earthquakes occur with different focal mechanisms, in support with its interpretation as the interplate. Under Cephalonia island, reverse-faulting deforms the upper plate along NW–SE structures, which may also be affected by left-lateral bookshelf-faulting. Small earthquakes show normal faulting along the western coast of Cephalonia and its extension 20 km SSW, the trace of the CTF as inferred from the occurrence of the large strike-slip earthquakes. Another group of normal-fault earthquakes locates in the lower plate from under Cephalonia to Zante, just outboard of a possible change of interplate dip suggested from reflection seismics landward under the islands. These normal-fault earthquakes appear to coincide in position with that of the load imposed by the upper plate transported over them, rather than occurring in an outer rise, outboard the plate boundary and trench, as observed in other subductions and attributed to the control by the flexural bending of the lower plate under the pull of the sinking slab. Interpretation has to consider several peculiar features of plate interaction in western Greece with respect to a steady-state model for major subduction zones, in particular: a fast deformation of the upper plate in front of an orogenically overthickened crust and of the southwestward push of extruding Anatolia; its transport, which is the cause of the migration of the plate boundary rather than the roll-back of a slab which has been proposed to be detached; possibly a flat and ramp shape of the interplate; the geometrical complexity of the shear limit across the CTF between subduction and collision, and the nearby variation of the nature of the foreland crust. © 2000 Elsevier Science B.V. All rights reserved.

Keywords: Cephalonia Transform Fault; local earthquakes; plate transport

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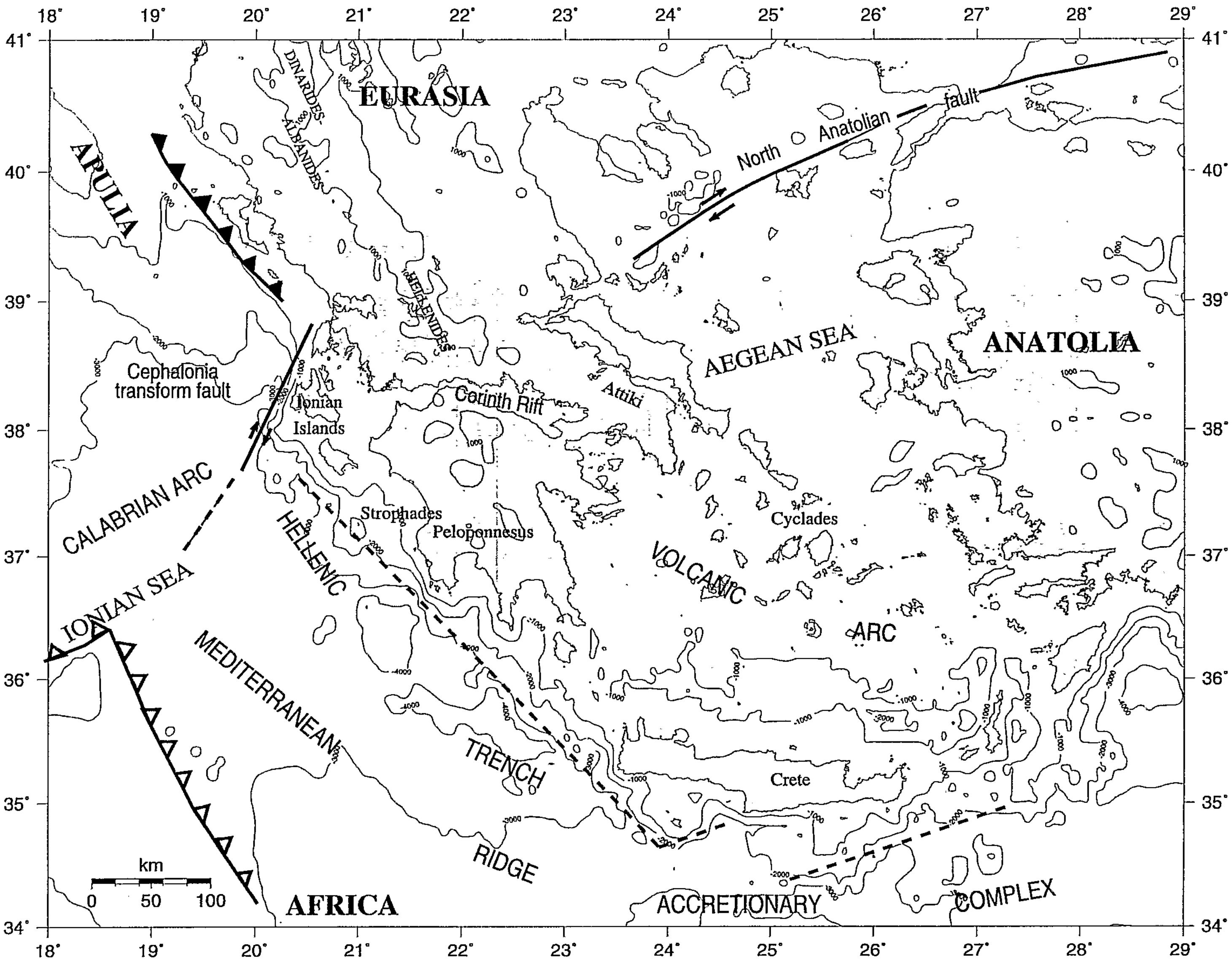

EURASIA
DINARIDES
ALBANIDES
HELLENIDES
APULIA
North Anatolian fault
AEGEAN SEA
ANATOLIA
Cephalonia transform fault
Ionian Islands
Corinth Rift
Attiki
CALABRIAN ARC
Strophades
Peloponnesus
Cyclades
HELLENIC
VOLCANIC
ARC
IONIAN SEA
MEDITERRANEAN
TRENCH
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1. Introduction

1.1. Geodynamical frame of the Hellenic subduction

The evolution of the Hellenic region is a subject of very active research that has recently been discussed and reviewed, e.g. by Jackson (1994), Le Pichon et al. (1995) and Armijo et al. (1996), and will be briefly summarized from the perspective of the results of the local earthquake survey reported in the present work.

The convergence of Africa and Europe since 80 Myr with an average rate of 10 mm/yr (Olivet et al., 1982; Argus et al., 1989) slowly consumes Mediterranean oceanic lithosphere. However, it is the more recent onset of intracontinental extension deforming the European upper plate that has caused boundaries across which, locally, a much faster rate for subduction of over 40 mm/yr has occurred since the early Pliocene, or earlier, in the Calabrian arc and the Hellenic Trench and Arc (Fig. 1). For the former one, no large interplate earthquakes are known and fast subduction appears to have recently halted (e.g. Hirn et al., 1997), but there is still activity in the Hellenic Arc, as attested by the distribution and focal mechanisms of earthquakes.

In the Hellenic Arc and Trench, subduction occurs in a domain that is spatially confined laterally, between the continental collision of the continental foreland of Apulia with the Albanides and Dinarides to the northwest, and in the east, by the fast westward lateral extrusion of Anatolia, under the northward push of Arabia at its eastern edge since the middle Miocene (e.g. Le Pichon et al., 1995). The latter mechanism contributes to the deformation of the Aegean domain, which is the upper plate in the convergence, as does the post-orogenic extensional evolution of this domain since its Alpine crustal thickening (e.g. Gautier and Brun, 1994; Jolivet et al., 1994) and its back-arc position with respect to the subduction, as attested by the Hellenic volcanic arc in the central Aegean, above the 150 km deep slab earthquakes.

In this narrow Hellenic convergence zone, the western part studied here is further complicated by two elements: the change in space on either side of the Cephalonia Transform Fault (CTF) from still oceanic subduction to continental collision, and the change in time of the position of the frontal thrust in the domain of the southern Ionian Islands (Fig. 2). In fact, the evolution here could be described in continuation of the Alpine orogenic evolution as a foreland-propagating fold and thrust belt of the Hellenides (e.g. Underhill, 1989), the front of which jumped in middle Miocene from its position at the Pindos thrust to a position west of the Ionian zone, at the Ionian Thrust (IT), and finally to a position west of Cephalonia and Zante islands, along which this southern part of the pre-Apulian continental domain now actively overthrusts the deep Ionian basin, probably since 4–5 Ma. Since the pre-Apulian domain does not extend further southeast, thrusting over the oceanic margin of Africa and subduction could have occurred earlier there.

From the plate tectonics point of view the region of the Ionian Islands is unique as a multiple junction, where all four types of plate boundary (collision, subduction, transform and spreading) connect at a 100 km scale in map view. The unraveling of the lithosphere structure, and deformation as retrievable from earthquake distribution and focal mechanisms, will be attempted here in

Fig. 1. Sketch map of Europe–Africa plate interaction in the region of the Hellenic subduction. Active volcanic arc in the Aegean sea. Frontal thrusts indicated: the line with full triangles for the collision with Eurasia of Apulia, the submerged northern continental margin of the Ionian sea being part of Africa; the line with open triangles for the front of the accretionary complexes, in the west for the Calabrian one, in the east for the Hellenic one and the Mediterranean Ridge. In this subduction zone, the Hellenic Trenches, between full and dashed lines, are interpreted as outer arc basins made of upper plate material of the Aegean extensional region as well as a continental backstop to the accretionary complex (Lallemant et al., 1994). Major active strike-slip faults are the North Anatolian fault, along which Anatolia is extruded, and the Cephalonia Transform Fault, the subject of the present study, which links collision with the western Hellenic subduction. The individual normal-faults of the continental extension of the Aegean domain, upper plate to the subduction, are not indicated, with the exception of the Corinth rift east of the study zone. Dinarides–Hellenides are alpine-age mountain ranges.

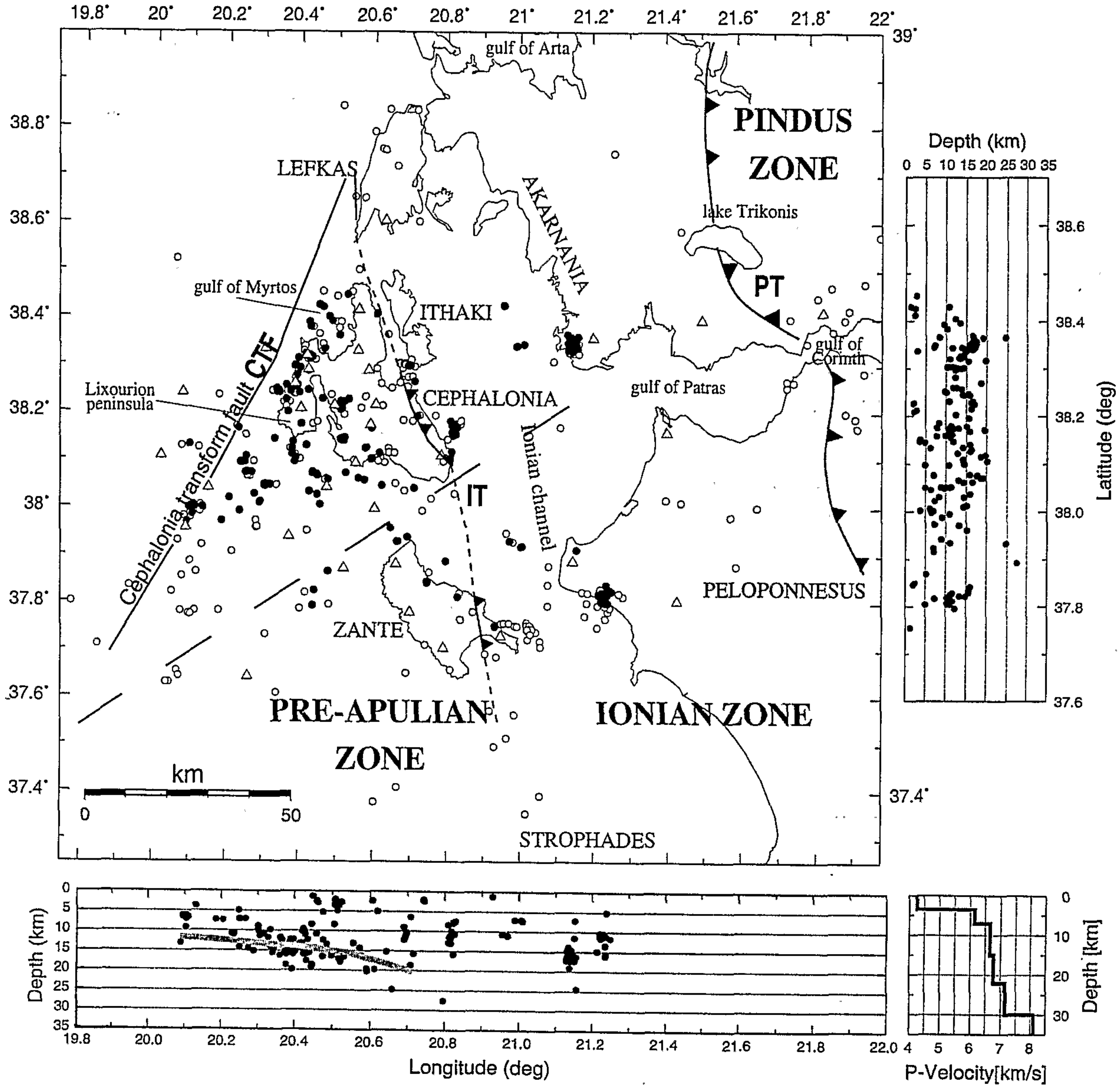


Fig. 2. (a) Western part of the foreland propagating fold and thrust belt of the Hellenides, with successive frontal thrusts: PT for the thrust of the Pindus over the Ionian zones, active before middle Miocene; IT for the thrust of the Ionian over the pre-Apulian zones, active until early Pliocene and replaced since in this region south of Lefkas by subduction outboard southwest, and by right-lateral shear along the CTF. Triangles show the locations of three-component seismographs of the temporary array at sea and on land. Full circles are the 200 local earthquakes best recorded, used in the inversion for a minimum 1D model (inset lower right) with which also the 300 lesser-quality open circle hypocenters have been relocated. Depth cross-sections display the E–W and N–S projections of the most reliable hypocenters. The position of the Ion-7 multiple-coverage reflection seismic profile is shown as a dashed line in plane view. In the E–W cross-section of seismicity a grey line is the projection of the main reflector imaged in this profile and suggested as the interplate (Hirn et al., 1996). Earthquakes are seen to locate on either side, but then differ in focal mechanisms (Fig. 4a and see text). (b) Epicenter map of earthquakes located with the world-wide network from 1963 to 1997 (ISC and NEIS sources). Magnitudes as indicated by symbols; points are smallest events.

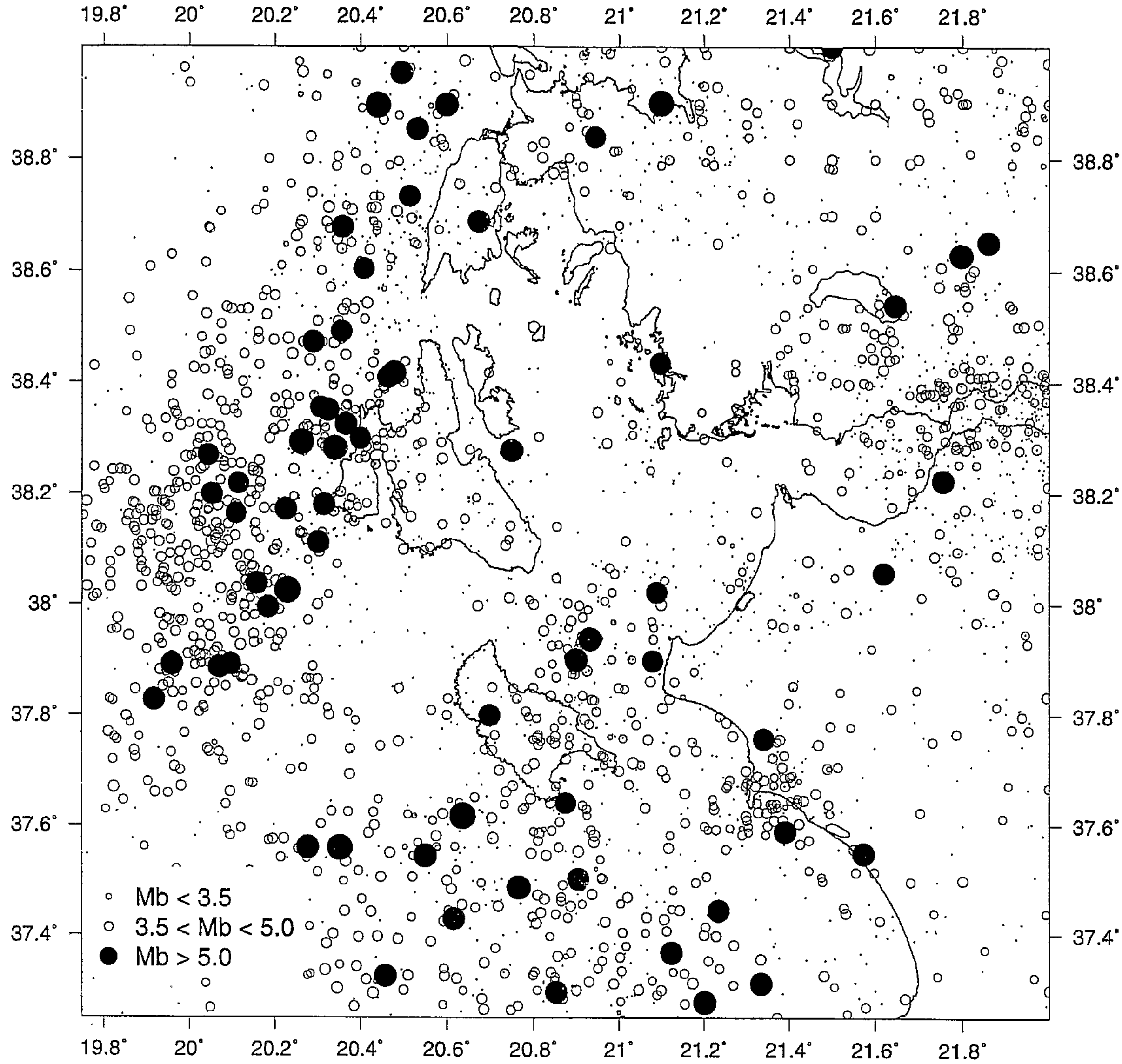


Fig. 2. (*continued*)

an effort to understand their relation by describing their interaction in three dimensions.

1.2. Seismic structure and activity in relation with subduction

The Hellenic Arc and Trench can be regarded as an active subduction zone, and its earthquake activity and plate and slab geometry compared with other examples. As a contribution to the framework of plate tectonics, Isacks et al. (1968) interpreted the world-wide distribution of large earthquakes along belts as marking the relative motion of lithospheric plates at their contact boundary, the major thrust-earthquake belts being those where convergence and subduction occur. Earthquakes at intermediate depths in subduction zones have been recognized to correspond to strain within the slab, such as downdip compression or extension, rather than its slip with respect to the surrounding mantle, and double seismic zones, with an upper level of compressional earthquakes and a lower one of extensional ones, being attributed to the slab unbending in the asthenosphere (e.g. Isacks and Molnar, 1969). In the outer rise, outboard of the front of upper plate overthrusting, normal-faulting identified for large events at shallow depth has been interpreted as a consequence of plate flexure (Hanks, 1971). Although it is most active at the European-scale and also singularly active in the upper plate extensional regime, the Hellenic region is, however, lacking a rate of large subduction-related earthquakes which would be comparable to that of the major global subduction zones. Earthquakes are restricted to above 150 km depth and their low magnitudes limit their location accuracy for a description of the slab geometry. From earthquake locations at dense temporary networks, Hatzfeld (1994) confirms that, in the western Hellenic Arc, the slab has a very shallow dip over the 200 km from the trench through the Peloponnesus, then a steeper one to 150 km depth under the Aegean Sea. Body-wave tomography results of Spakman et al. (1988) have revealed a high-velocity anomaly dipping down to at least 600 km, which they regard as the image of the slab of subducted oceanic lithosphere, prolongating well below its seismically active upper part, which leads to debate and revision of the age of onset and rate of the Hellenic subduction. In the western part of the arc, they suggest that the slab is now detached beneath 150 km depth. Only a few events in the interplate domain under Crete and the Ionian Islands have been large enough for focal mechanisms to be derived from world-wide observations (Taymaz et al., 1990; Papazachos et al., 1991). At greater depth, few mechanisms from large earthquakes have been interpreted as showing downdip compression and they have been supplemented by microearthquakes interpreted as showing reverse faulting at the top of the slab below 100 km (Hatzfeld, 1994). The image of the internal deformation of the slab and its interaction with the upper plate, however, remains poorly defined by the scarce database.

2. Local earthquake observations: experiment and results

2.1. Background and scope of the local earthquake survey

We are pursuing an effort at deciphering the structure and evolution of the western Hellenic Arc by a succession of field experiments. A first coincident reflection–refraction profile in 1992, from the Ionian basin through the channel between Cephalonia and Zante (Fig. 2a), has provided a two-dimensional (2D) reflectivity cross-section, with velocity control (Hirn et al., 1996). This reveals several features at depth, like, from west to east: an active block structure in the pre-Apulian Zone, the east-dipping thrust of the Ionian Zone over it; a west-dipping boundary in the upper crust, which may either be a backthrust or a normal fault. These features are all limited in depth above a conspicuously flat reflector slightly dipping landwards from a 12 km depth at 50 km offshore, then steeper under the islands. A test record of local earthquakes with three-component seismometers around this profile suggested that the location of seismic activity might be controlled at depth by this reflector (Sachpazi et al., 1996). The results presented here have been obtained from a third experiment, in August–September 1995, which focused on recording local earthquakes with 35 stations. A new effort with vertical reflection seismic and ocean-bottom seismograph (OBS) refraction in January 1997, will help in future to test how structures extend towards the south and west of the island of Zante.

The experimental set-up of the local earthquake survey was arranged to have some important improvements on those previously used to discuss the strain pattern in NW Greece (Hatzfeld et al., 1990, 1995). In our experiment, both OBSs and land stations were set up and operated jointly as an array in order to provide, for the first time, epicentral locations of local earthquakes without bias from the array being restricted in E–W extent to land, or in N–S extent to one side of the Cephalonia–Zante channel. In addition, the recording array was made up exclusively of three-component digital seismographs. Here, this was considered critical for constraining tightly the focal depth estimates from observations near the epicenter for earthquakes located anywhere in the study region. Achieving a resolution in the location similar to that obtained for the crustal structure from reflection seismics is essential to allow activity and structure to be discussed with respect to each other.

2.2. *Earthquake epicenter distribution and plate boundaries at outcrop*

Usual earthquake location, with a one-dimensional (1D) velocity–depth model, can be seriously in error where the assumption of lateral homogeneity is not satisfied. This is the case in subduction zones because of the high-velocity slab and low-velocity back-arc basin that alter downgoing rays to non-local stations, depending on azimuth. In the specific case of the Hellenic subduction zone in Peloponnesus and western Crete, Papadopoulos et al. (1988) showed that adding few local stations shifted computed earthquake locations by several tens of kilometers, from an African intraplate position to a better constrained and more reasonable location at the Aegean plate boundary. Their station network, however, was confined in both cases to the landward half-space, and locations were obtained in a 1D velocity-model. In northwestern Greece local earthquakes have also been studied previously with temporary arrays, but which were looser and were confined either to the north of the Gulf of Patras or the channel between Cephalonia and Zante (Amorese, 1993; Hatzfeld et al., 1995), or to its south (Hatzfeld et al., 1990). In order to have a more definitive proof on the true location of earthquakes, we obviated previous limitations in our survey across the Ionian Islands. Firstly, we used an array of OBSs in order to detect and enclose seismicity at sea as well as on land. Secondly, the complete offshore–onshore array was arranged with an aperture and density ensuring that first-arrival waves used for location were generally not downgoing rays interfering with deep heterogeneity, and that there was a dense enough cross-propagation in the source–receiver domain to explicitly take into account heterogeneity by a joint tomographic inversion for the three-dimensional (3D) heterogeneity and hypocenter location.

In the region studied, Fig. 2b displays the epicenters reported from 1963 to 1997 by international agencies [International Seismological Centre (ISC), UK; National Earthquake Information Service (NEIS), USA], using data of seismological observatories worldwide. These epicenters distribute as a cloud, the center of which would define the Cephalonia fault. Seismicity located by temporary arrays operated previously in the region (Hatzfeld et al., 1995) has been suggested to lie more landward than apparent in this image of the larger magnitude seismicity over 35 years. A main result of the experimental effort of the present survey, which controlled the seaward extent of seismicity with OBSs, appears in the map of local earthquake epicenters displayed as Fig. 2a. The earthquakes appear to be confined to the east of a sharply defined line that follows the western edge of the island of Cephalonia. This has already been obtained in a 1D velocity–depth model encompassing the sources, which is indicative of the fact that the array geometry is adequate. These well-constrained epicenters hence do not appear distributed like those in Fig. 2b. Earthquakes are here all confined to the eastern side of the CTF, although the basement on either side, down to the southern end of the Lixourion peninsula, is considered to be of the same nature: continental platform of Apulia, submerged on one side and outcropping on the islands on the other. The difference in seismic activity may then be related to the situation with respect to the converging plate boundary, since, to the east of the CTF, pre-Apulian material is in the upper plate position, whereas on the western side the aseismic zone is in the lower plate position far in the foreland to the collision. Further south, the geodynamical and the lithological nature of these entities are contrasted: oceanic for the Ionian basin, thinned continental for the Apulian platform, thickened Alpine for the Western Greek mainland.

To the east of the islands, the Ionian Channel has already been noted in previous temporary deployments to have a reduced seismic activity. In the present case, however, if large parts of this area have no earthquakes, there is the notable exception of two clusters of earthquakes at the eastern edge of the channel with the mainland and Peloponnesus. These also occurred clustered in time, as did the events at the southeastern tips of both Cephalonia and Zante. Not all of those regions appeared active in previous temporary surveys, and would not have here for shorter periods. This conversely poses the question of how many other regions would have appeared active in a longer deployment, and is a reminder of the usual problem of trade-off between the increased accuracy but reduced representativeness of temporary observations.

2.3. Seismic structure and activity in section: the upper and lower plates

A first step in the analysis is to define a 1D velocity model and corresponding station corrections. The region sampled spans structures as diverse as oceanic crust offshore southwest, thinned continental to the west, and thickened continental under the Hellenides to the east. There is, however, a need to invert the data in order to define a single 1D model before attempting a 3D inversion, in order to reduce in the latter the effect of the non-linearity of the inverse problems in having an optimally small deviation of the ray paths in 3D with respect to their initial geometry in a tabular 1D velocity model. Since the aperture of the array has been kept small enough, first-arrival mantle refractions are uncommon and the expected principal lateral variation, which is that of the crustal thickness, does not invalidate the search of a minimum 1D model for velocity within the crust. We follow the approach of Kissling (1988) to derive a tomographic image.

Among the 500 locatable earthquakes recorded, a selection is made considering their residuals, azimuthal coverage (gap smaller than 180°) and number of observations (more than six stations with both P and S) resulting in about 200 high-quality events used for further processing. A minimum 1D velocity model, with corresponding station corrections, results from a simultaneous inversion for both hypocenter parameters and model. It is designed to locate these events with the smallest possible uniform location error (Kissling, 1988). The computation of a minimum 1D model is a trial-and-error procedure for different initial velocity models, initial hypocenter locations and damping and control parameters for the coupled inverse problem, using the VELEST code (Kissling et al., 1994). An initial model derived from reflection–refraction seismics is used, with respect to which the minimum 1D model shows, in general, an increased velocity under the sediments, but the average velocity across the shallower 13 km is consistent with that resulting from velocity analysis above a major reflector of the multiple coverage reflection seismics profile in the Cephalonia–Zante channel (Sachpazi et al., 1996). In the process of determining the 1D minimum model, station delays are estimated. They are useful to absorb artifacts, like systematic picking biases or local delays due to near-surface conditions, but they may also partially account for the three-dimensionality of the velocity field, in particular the inhomogeneous near-surface structure. In the present case these station corrections are not randomly scattered, as would be expected if they were noise, but they are consistent among groups of neighboring stations. Hence they contain information on regional variation of structure in space, though probably in the shallow layers. In fact, their first-order distribution outlines the outcrop of the hard and high-velocity limestone cover of the pre-Apulian zone in parts of the emerged islands. Since these station delays contain information on the upper layers, the 3D inversion is performed in damping them significantly, in order that this information goes into constraining the heterogeneity in structure. The 3D inversion is made for P-velocity, but S-arrivals are jointly used for constraining source locations. The 3D inversion for the 205 events, with about 2300 P arrivals, results in reducing the data variance to 0.06 s, half its initial value.

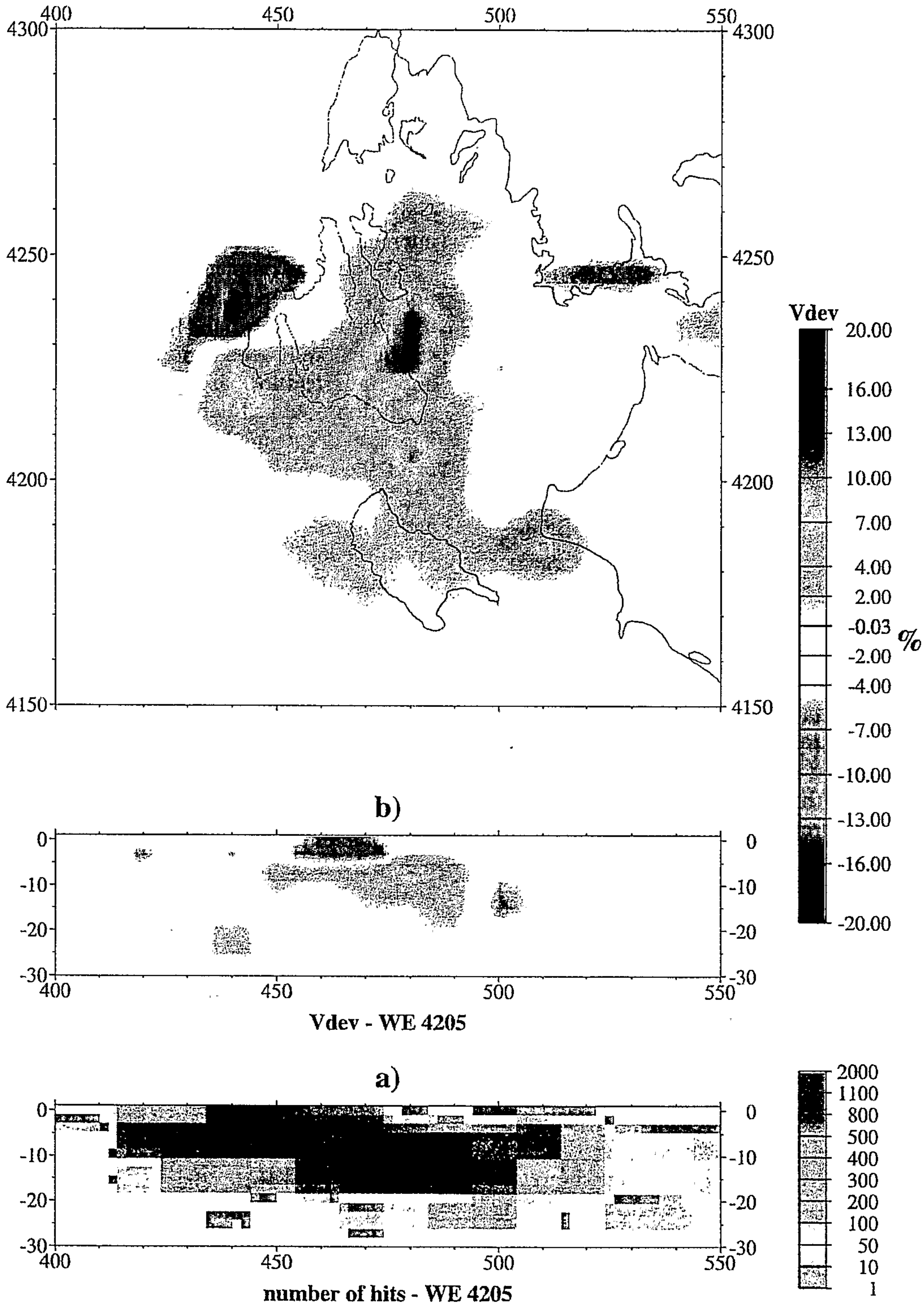
The next step is to derive the structural heterogeneity by local earthquake tomography and compare it with reflection–refraction results. The main aim of the 3D inversion is to provide improved locations, as well as take-off angles taking into account a 3D model as realistic as possible. The central part of the structure beneath Cephalonia is well sampled by the database from the present array, as shown by the distribution of hit counts in Fig. 3a. The display of the spatial deviation of P-velocity from the minimum 1D model in Fig. 3b can then be regarded as representing a good approximation of the structural heterogeneity obtainable in a cross-section along the southern coast of Cephalonia. In this well-resolved part of the survey, a high-velocity deviation in the depth range of 7–11 km is underlain by a null or slightly negative deviation. The retrieval of the signs of velocity deviations is robust; the reliability of their magnitudes is the subject of further synthetic tests. In the present case, these anomalies with opposite signs superposed in depth are in two layers of only slightly different initial velocities, and hence result in a net velocity inversion with depth. The top of the corresponding low-velocity layer is found at a depth of 11 km, where it is shallowest west of Cephalonia; it is dipping eastward, first slightly, then much more steeply and then out of the domain sampled. This section is largely along a previous marine multiple coverage reflection profile (Hirn et al., 1996). This only provides a unique opportunity of comparing the structure obtained from completely independent data sets. Also, ray geometry and sources are fundamentally different artificial sources at the surface and waves reflected on interfaces on the one hand, and, on the other hand, natural sources at depth with waves continuously refracted in upward transmission. From the reflection seismic image, which revealed a very shallowly dipping reflector, we suggested that the pre-Apulian crust of the Ionian Islands was overriding along that reflector, the Ionian oceanic crust. In this lower plate, its sedimentary cover appeared to be included, in keeping with a model in which it was not the lower plate that was actively pushing under the upper, possibly scraping off its sedimentary cover, but the upper plate was instead deforming and accommodating transport over it. With the present ray-coverage, however, we take the indication of the high velocity reaching deeper to the east only as a suggestion; this, however, is consistent with the change to a steeper dip of the interface imaged by vertical reflection seismics, suggesting a ramp and flat geometry for the interplate.

The addition of local earthquake tomography to this knowledge is to provide an idea of the extent in plane-view of the structural elements distinguished in the depth-section. The P-velocity deviation displayed in Fig. 3c for the cells in the layer from 5 to 12 km depth shows spatial variation consistent with the largely NW–SE extent of the high-velocity domain, orthogonal to and interrupted by the CTF.

Although a complete discussion of the distribution of earthquakes with respect to structure in three dimensions would require a more complete sampling, and hence more earthquakes, a few important features are apparent already at the resolution obtained here. First, in Fig. 2a, there is a general deepening landward of the cut-off depth, or of the average depth of seismicity, that resembles the dip of the suggested interplate reflector and of the deeper limit of the high-velocity anomaly of the local earthquake tomography of Fig. 3b. However, the depth distribution of earthquakes appears to span both sides of the depth range of the interface interpreted on the reflection section as the basal thrust of the upper plate. Hence seismic deformation appears to occur both in the upper and lower plates. This statement would, however, remain weak if the focal mechanisms evidence, described hereafter, were not to provide independent support of it. In the lower plate the earthquakes tend to occur in the upper few kilometers, but not as interplate events at its top, rather at the base of the low-velocity layer, hence at the top of the basement of the lower plate.

2.4. Focal mechanisms and the nature of seismic deformation with respect to plate subduction

The focal mechanisms presented in Fig. 4a are obtained from an average 15 first-arrival polarities. They are considered as a representative sample of the data recorded. When compared with focal mechanisms of large earthquakes displayed in Fig. 4b, the present sample of mechanisms of local earthquakes does not simply mimic them in spatial distribution or type. The two sets of observations will be discussed together from the perspective of regional geodynamics, with the possible implica-



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tions of their differences for the seismic cycle in this region being deferred to a further study.

Focal mechanisms can be discussed with respect to deformation within plates and transport of plates with respect to each other. A wide diversity of focal mechanisms has been reported for western Greece, for different large earthquakes, at the 100 km scale of the region at which true spatial variation is likely, and also for the same earthquake when analyzed with different techniques (Sachpazi et al., 1996). Microearthquake surveys also yielded a variety of mechanisms. Geodynamical interpretations have considered focal mechanisms from diverse viewpoints, discussing attributes as different as either the population of horizontal projections of P and T axes (Hatzfeld et al., 1990, 1995), or the principal stress directions derived assuming a regionally uniform stress and a slip on pre-existing faults (Amorese, 1993), or the population of horizontal projections of the slip directions resulting from a choice made for the fault-plane among the two nodal planes (Baker et al., 1997).

For our specific survey region, the focal mechanisms derived in a previous microearthquake survey (Amorese, 1993; Hatzfeld et al., 1995) correspond to events that clustered geographically under the Gulf of Myrtos, NE of the Lixourion peninsula of Cephalonia, but showed a great variety with respect to focal depth. Apart from those, three events under the center of the island, at a large depth of over 15 km had normal-fault mechanisms. An interpretation of seismicity by deformation due to a regional stress-field (Hatzfeld et al., 1995) does not use the information contained in the large degree of variation in focal mechanisms, and these deep normal-fault events remain unexplained. Here we suggest instead to discuss the full diversity of focal mechanisms by analyzing individually their high-resolution location with respect to the features of internal structure resulting from reflection–refraction and local earthquake tomography.

The experimental effort of the present survey resulted in an array that encompasses all the Ionian Islands at a time, as well as the neighboring mainland and Peloponnesus, and extends far into the sea, thus offering a geometry adequate for constraining in a reliable and tight way the epicenter locations, depths and focal mechanisms. With the tightly constrained take-off angles resulting from the high-resolution location and with the large number of first motions for the same event due to the dense array, the short-scale spatial variation of focal mechanisms can be resolved as significant. This variation of mechanisms can then be described with respect to that of structure in space, offering an opportunity to approach the complexity of kinematics. Mechanisms are obtained over a wide region of 50 km dimension centered on Cephalonia (Fig. 4a). They apparently are not consistent with fault-slip due to a unique regional stress-field; for instance, under Cephalonia earthquakes with same epicenters exhibit reverse and normal fault solutions that cannot be reconciled with such an assumption. Based on the resolution of a difference in their focal depths, which locates them either side of an interface imaged by reflection seismics slightly further south, previously suggested as the basal thrust along which the Ionian Islands are transported to the SW (Hirn et al., 1996), they can be readily understood as revealing different states of stress in the upper and lower plates of the convergence zone.

The CTF has been inferred from large earthquakes showing right-lateral strike-slip. At first glance, the focal mechanisms of local earthquakes displayed in Fig. 4a are not dominated by strike-slip near the western edge of Cephalonia. In an

Fig. 3. (a) East–west cross-section, at 4205 km northing of UTM coordinates system, of the distribution of hit counts of rays in cells of the 3D inversion, to show sampling of P-velocity anomalies displayed in (b). (b) and (c) East–west cross-section, at 4205 km northing and map at 5–12 km depth, of the distribution of P-velocity anomalies in cells of the 3D inversion, with respect to minimum 1D model. Note that the base of the high velocity anomaly corresponds to the lower part of the upper plate along the interplate suggested from a reflection seismic profile (Fig. 2a) and is the top of a layer of absolute low-velocity consistent with a sedimentary cover of the lower plate.

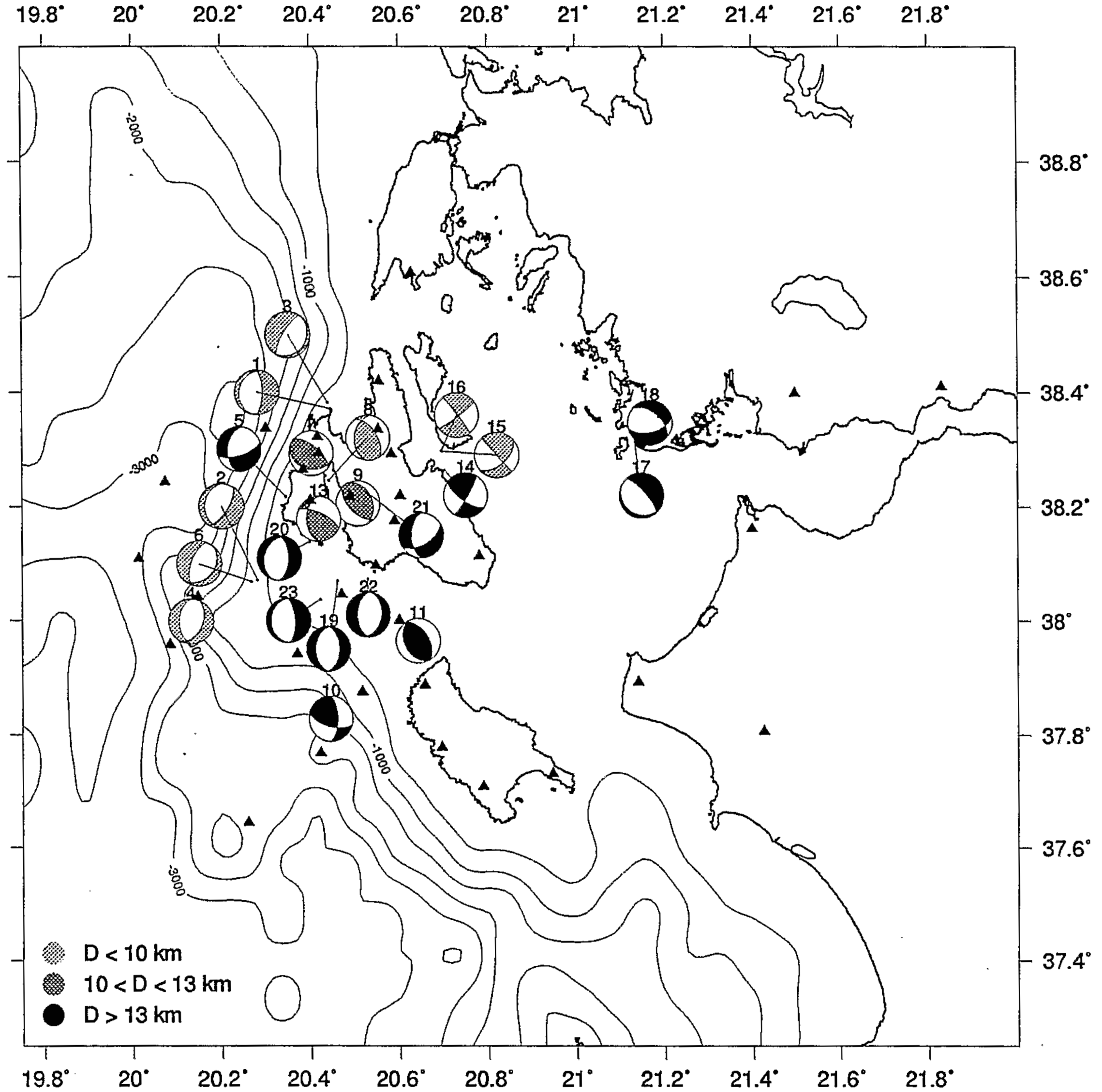

Fig. 4. (a) Focal mechanisms of local earthquakes with take-off angles computed in the 3D heterogeneous structure resulting from the inversion, an average 14 first-arrival polarities, lower hemisphere. Compressional quadrants shading as a function of depth: black for events deeper than 13 km (under Cephalonia these are suggested lower-plate events, mostly normal faults); dark gray between 10 and 13 km; light gray above 10 km. (b) Focal mechanisms of large earthquakes compiled from diverse sources as indicated (Harvard is for Dziewonski et al., 1983; Scordiles et al., 1985; Anderson and Jackson, 1987).

early study of large earthquakes, McKenzie (1978) wondered about finding no transform-faulting events consistent with the trend of the bathymetry along Cephalonia and its location linking two segments of the converging plate boundary. Since then, other earthquakes have occurred; nevertheless, the interpretations of their nature has only progressively converged to recognize that strike-slip faulting occurred along the CTF. A compilation displayed in Fig. 4b illustrates that major earthquakes correspond to right-lateral strike-slip occurring along a fault striking slightly clockwise

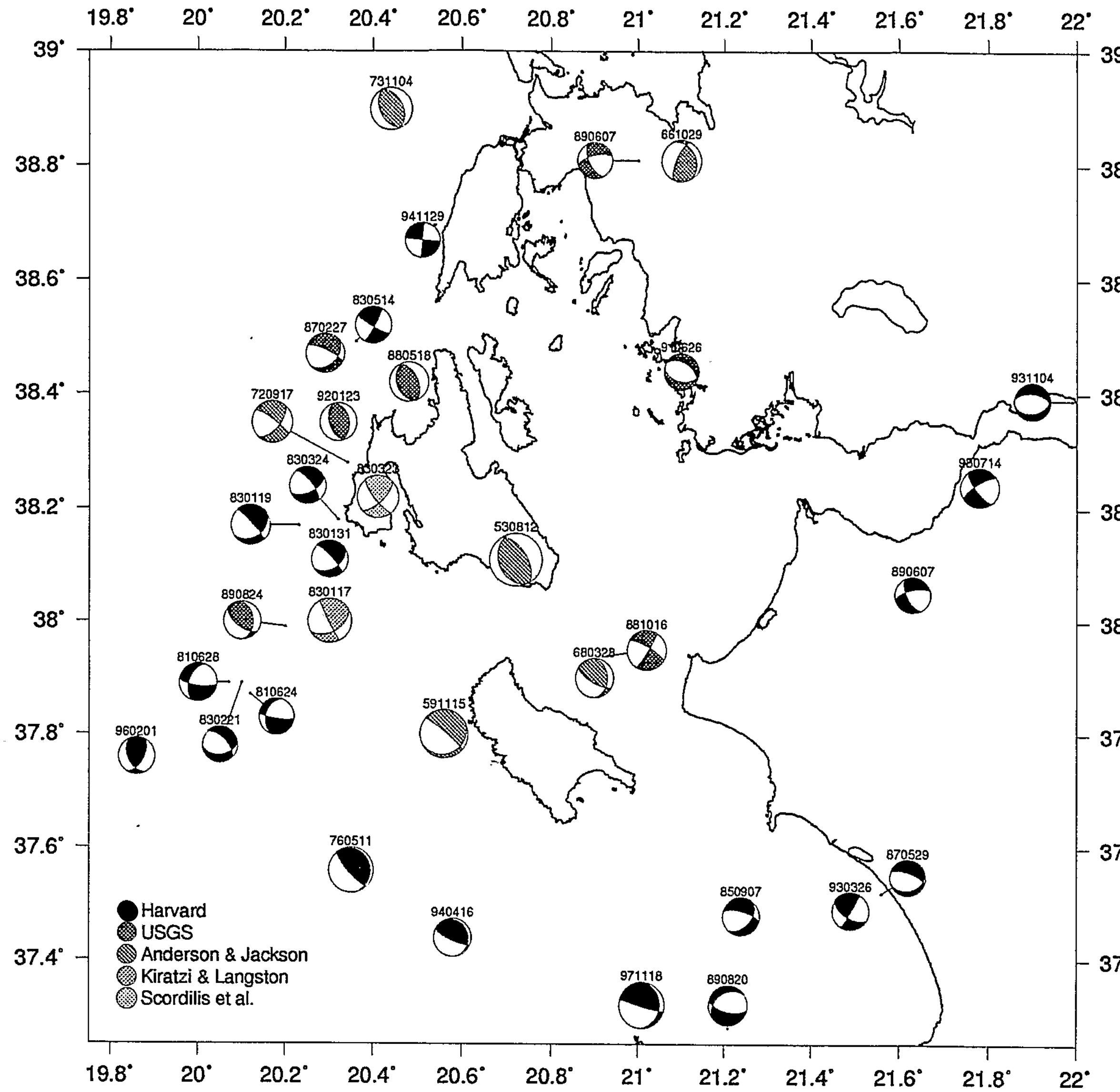

Fig. 4. (*continued*)

from N–S, from event 941129 (year, month, day) alongside Lefkas, down to event 830514, offshore the northern tip of Cephalonia. The slip directions in those major earthquakes have commonly been taken along the N35°E strike, since this is not far from the average N20°E strike of the bathymetric scarp along the west coast of Cephalonia. An important feature, which caused alternative interpretations depending on waves and methods used to retrieve focal mechanisms (Sachpazi et al., 1996), is that the strike-slip plane chosen as having an azimuth similar to that of the bathymetry is not vertical but has only a dip of 60°. A thrust component has also been proposed (e.g. Kiratzi and Langston, 1991), but is debated (Baker et al., 1997). The local earthquake focal mechanisms may accommodate some strike-slip component but do not clearly document it by themselves.

The proportion of normal-fault mechanisms in Fig. 4a is unexpected with respect to the general transform-fault and subduction context recognized for this area. In the original study of focal mechanisms of McKenzie (1972), the two large events then available were displayed as normal-faults, with the mention that very shallow-dipping thrusts were equally acceptable, which was the option chosen later (McKenzie, 1978) and commonly followed in the analysis of solutions for more recent events displayed in Fig. 4b. The normal-faults obtained in the present study for the local earthquakes do not, however, have a nodal-plane so near to the vertical that they could have flipped to apparent normal faults from true flat thrusts as a result of lack of accuracy.

The control by OBS data of the locations and the sampling of the focal sphere allows one to distinguish a westernmost domain of shallow, 6–10 km deep normal-faulting with N20°E strike along the western coast of Cephalonia. Noteworthy is that this strike is also that of the bathymetric scarp that limits the narrow continental slope of Cephalonia towards the 3000 m deep basin to the west. The earthquakes, however, locate between this slope and the coastline, and along the same direction, in continuation to 20 km southwest of the coast. Normal faults have been reported from surface geology as en échelon structures related to the internal deformation in the flexed top of the oblique thrust on the strike-slip CTF. To consider these normal-fault mechanism events as due to surface bending of the upper plate, their depth would be too large and their strike too near to that of the strike-slip as derived from large earthquakes, instead of intermediate with the direction of transport of the thrust sheet. Alternatively, normal faults in the upper plate have been related to inferred changes of dip in the interplate, for instance in northern Chile by Armijo and Thiele (1990). There is, however, no reflection–refraction data at the edge of Cephalonia that could help to locate independently the depth of the interplate with respect to those hypocenters. As discussed before, although the CTF is considered a strike-slip, the mechanisms of large earthquakes there exhibit a focal plane that is not vertical but rather is shallowly dipping to southeast. It is hence possible that the normal-fault local earthquakes, though rather shallow, could occur under the inclined plane of interplate strike-slip, over which a thrust component has been suggested. One of the events locates deeper and could definitely be considered of this kind. In this frame, these events would correspond to the local yielding of the western plate along the boundary where the Ionian Islands slide along it but also tend to overlie it obliquely. If these events were in the lower plate, the extensional style would then extend in continuity to the region of deeper normal-fault events further southeast, discussed in Section 2.3.

The question of flat interplate thrusts and steep reverse-faulting, and of their bearing on the deformation within the upper plate or motion at the interplate, may be discussed briefly. For the Cephalonia earthquake of 1953 (530812), the major earthquake of $M = 7.4$ of this century in Greece, McKenzie (1972) derived a thrust mechanism which he regarded as evidence that Hellenic subduction reached as far north as the Ionian Islands. None of the nodal planes, however, was flatly dipping as expected for interplate events. Several earthquakes could be added since, which, as discussed previously, could be either vertical normal or very flat reverse faults, as displayed in Fig. 4b. They are commonly considered as shallow-dipping thrust events forming a belt of interplate earthquakes (e.g. Baker et al., 1997), which begins 30 km SW of Zante with the 760511 event and spans over 60 km to the SE to the Strophades island with the recent 971118 event.

Local earthquake data are not yet complete enough for this southernmost region from our sole array, but in the more central region the focal solutions are well-constrained and exhibit reverse-faults striking NW–SE, which have similar values of rather steep dip for the two nodal planes. The best-constrained reverse-faulting events are located under the Lixourion peninsula and further east under Cephalonia (Fig. 4a). Their focal depths would place them just above the interplate, if this can be extrapolated from its position inferred on the reflection section; hence, they would indicate internal strain in the upper plate under Cephalonia. Among larger earthquakes, the moderate ones of 880518 and 920123 show similar focal mechanisms

(Fig. 4b). Their locations to the northwest near offshore of Cephalonia may not be accurate; the second, for instance, has an aftershock zone extending well through Cephalonia (Tselentis et al., 1997), which suggests that they may be considered together with our better-located reverse-fault local earthquakes. The depths of these larger events are not well enough constrained to attribute them either to the interplate or to the upper plate, like the local earthquakes we interpret here as due to upper plate deformation.

Further south, a moderately well constrained thrust mechanism is obtained for a local earthquake that is located in the channel between Cephalonia and Zante. It is at a significantly larger depth than the ones under Cephalonia, but it could be already far enough to the northeast, on the steeper dipping part of the interplate reflector, that, in spite of its 16 km depth, it may still either mark upper plate deformation or be an interplate event. The location is not very far from the largest earthquake of 530812, which has a similar steep reverse-fault. These events have a distinctly more downdip position in the convergence and a steeper dip of their thrust-fault, with respect to the other large events south of Zante, which are instead typical of very flat interplate thrusting. For the major 530812 earthquake, Stiros et al. (1994) reported coseismic uplift from which geometry they inferred a piston-like block motion at the surface, above a decoupling level provided by the Triassic evaporites ubiquitous in the Hellenic domain, but invoke, however, a necessary deep thrust-fault as the primary cause of the earthquake. If further examples supported the steeper and deeper thrust events, this would add to the evidence of the change of dip of the interplate suggested by reflection seismics, in favor of a ramp and flat geometry of the interplate thrust.

From local earthquakes, strike-slip and normal faulting appears to occur far from the edge of the upper plate at the surface. Along the eastern coast of Cephalonia the activity occurred mostly in two clusters east of the surface trace of the Ionian thrust (Fig. 2a) boundary where convergence was localized until 5 Myr ago. Focal mechanisms are presently well-constrained only for the northern cluster between Cephalonia and Ithaki. Almost pure strike-slip on vertical faults can be fitted to the data. One nodal plane is NE–SW, along which right-lateral shear would have some resemblance with that of large earthquakes in the significantly more western position of the transform-fault along the other, western coast of Cephalonia. Alternatively, taking as the focal plane the NW–SE-striking one would allow one to take into account the inherited heterogeneity of this former boundary between the Ionian and pre-Apulian zones, as well as connect with the other seismic cluster at the SE tip of Cephalonia, and with the large earthquake of 881016, between Cephalonia and Zante, on this same boundary and with the same mechanism. Such left-lateral slip of NW–SE strike could be understood in the regional context by invoking bookshelf-faulting (McKenzie and Jackson, 1983) of pre-existing faults rotating together with the elongated blocks they bound. NW–SE-oriented features would produce left-lateral strike-slip, as observed, in a total dextral shear over the region; this would be consistent with the spatial variation resolved for the GPS displacement vectors of Kahle et al. (1995): increase in magnitude and clockwise rotation from north to south, through Lefkas, Cephalonia, Zante and the western Peloponnesus.

At the edge of mainland Greece, normal faulting appears to dominate the earthquakes of the cluster. As such, they look similar to the mechanism of the larger earthquake of 910626. Their depths range to 20 km. So far inland they are likely in the upper plate and show that its extensional regime extends over the Aegean domain.

The most striking result in Fig. 4a is the occurrence of normal faulting at a depth resolved as being in excess of 13 km, well-constrained and consistent in several events. The location of these events is 2–3 km beneath the interplate as interpreted from reflection seismics, if this can be extrapolated sideways over 20 km, and in correspondence with the base of the high-velocity anomaly of the upper plate in the local earthquake tomography. They would hence likely affect the top of the crust of the lower plate, probably under a sedimentary cover. In fact, on the reflection section, a reflector is seen almost parallel to the one attributed to the interplate, about 1 s two-way time later, corresponding to 2–3 km deeper. The location in the plane of these normal-fault events is just seaward of the change in dip of the interplate imaged by reflection seismics. In the shallow part of subduction zones, normal-fault lower plate earthquakes are found in an outer rise seaward of the plate boundary trench, and are interpreted as the result of lower plate yielding under the bending moment due to the pull of the sinking slab. Here the normal-fault events are also in the lower plate, and seaward of a steepening of its topography; this could similarly mark a yield to a bending, but in a different position, being here under the load of the upper plate, which increases in thickness landward of a ramp.

3. Conclusions and discussion

A dense seismic array deployed temporarily allows one to constrain reliably the focal parameters of local earthquakes in the region of the Ionian Islands, by the use of three-component seismographs jointly offshore and onshore. Seismic deformation is documented to be confined to the east of the N20°E-striking steep continental slope west of Cephalonia island, marking the surface outcrop of the right-lateral CTF. With respect to the general convergence, the material west of the CTF, which is not deforming, is in a lower plate position of the convergence with respect to its collision segment further northwest. The same material is instead deforming east of the transform-fault, where it is in the upper plate position to the subduction. The local earthquake transmission-velocity tomography, compared in depth-section with a previous marine reflection profile east of the transform-fault, provides evidence in support of a shallow landward dipping boundary around 12 km deep under the Ionian Islands along which these may override the lower plate. The earthquake hypocenters, which locate east of the transform, occur both above and below this interface suggested as the interplate thrust boundary. Diverse types of focal mechanism can be resolved and discussed with respect to structural features. They do not mimic in a simple way the image suggested from large earthquakes. Reverse-faults deform the upper plate under Cephalonia along NW–SE structures. Strike-slip occurs at its eastern edge, which would be left-lateral if occurring along this same NW–SE structural direction of previous boundaries. This could be interpreted as left-lateral bookshelf-faulting in a zone where the magnitude of SW-directed transport increased from NW to SE, as suggested from space geodesy. At the edge of mainland Greece, normal faulting at depth, but presumably in the upper plate, may indicate the spatial transition to the generalized Aegean extension far from the domain of transport and thrusting above the neighboring plate. Under the Ionian Islands, normal-fault earthquakes locate in the lower plate seaward of a change of interplate dip suggested from reflection seismics landward under the islands. Their position could hence appear related directly to that of the load imposed by the upper plate, and is not in an outer rise outboard the plate boundary and trench, as observed in other subductions where a flexural bending of the lower plate is attributed to the pull of the sinking slab. Normal faulting is also documented with the strike of the CTF, in a narrow belt along the western coast of Cephalonia and extending 20 km SSW.

The interpretation of the location and the type of slip of earthquakes, both the small ones studied with the temporary array and the large ones, is not straightforward in a regular model of major subduction zones; one has to consider several features peculiar to plate interaction in western Greece. Seismic activity and mechanisms have been summarized (e.g. Hatzfeld, 1994) as indicating that the seismically active part of the slab of the western Hellenic subduction, which is not well defined in its deeper part, remains flat and shallow over 200 km under the Peloponnesus, then dips steeper, to reach 150 km depth. A flatly dipping slab has also been derived from the study of the foreland basins (e.g. Doglioni, 1993). According to results of seismic tomography (Spakman et al., 1988), the velocity anomaly interpreted as a large and long slab which continues to depth in the seismically active part of the subduction north of Crete would be presently detached from the shallower part of the western subduction zone studied here. The slab-pull in this western region is then likely to be locally small and slab roll-back can hardly contribute to the retreat of the convergence boundary and to a back-arc spreading type of upper plate extension.

However, the recent and present relative motion across the converging boundary is much larger in the region of the Ionian Islands, as derived from paleomagnetism since the Pliocene and Pleistocene (Laj et al., 1982) and measured by GPS (Kahle et al., 1993, 1995), than the general motion of Europe with respect to Africa. This implies that, instead of being controlled by a slab retreat, the convergence boundary moves in the direction of the lower plate by the active overriding of the upper plate which is transported at a rate uncommon in other subduction zones. Hence, whereas the lower plate does not yield by bending and forming a flexural outer rise under a slab-pull, since the slab is short and flat, it is increasingly loaded by the advection on top of it, of the material which overrides at a fast rate and deforms internally. Internal deformation of the upper plate during transport, and of the lower plate under loading by this transported material, provides a frame for understanding the local earthquakes presented here, in addition to the slip at an interplate with a possible ramp and flat geometry, rather than describing the situation in terms of a regional compressional stress regime.

The very large deformation and transport of the upper plate has been the subject of much attention. Diverse causes and their evolution in both time and space have been considered and a complete review is beyond our present scope. Among the features or mechanisms active since 5 Myr, or the onset of which has modified the deformation and transport more recently, the following have been particularly emphasized, covering sometimes the same reality from different viewpoints: propagation of rifting ahead of the North Anatolian fault to the Corinth Rift (Armijo et al., 1996); capture of the Peloponnesus in the counterclockwise rotation due to the extrusion of Anatolia and differential motion with northern Greece across the Gulf of Corinth and the Ionian Islands (Le Pichon et al., 1995); slab-detachment of the western subduction (Spakman et al., 1988; Meijer and Wortel, 1996); spreading caused by body forces stored in the alpine overthickening of the crust onshore of deep basins (Davies et al., 1997); east–west Holocene extension (Lyon-Caen et al., 1988); collision of the Hellenic accretionary wedge with the African continental margin (Le Pichon et al., 1995).

The lateral variation from subduction to collision and the onset of the latter and of the CTF as a feature cutting through the extremity of the Apulian domain are less documented and discussed. The geometry and heterogeneity of the lower plate in the subduction–transform system likely also plays a significant, but less well understood, role. For instance, a Moho and basement highstand in the Ionian oceanic basin have been evidenced in the westward continuation of the seismic profile discussed here (Cernobori et al., 1996). They may be coupled with the localization of the Cephalonia transform, since the southward continuation of the latter could be regarded as separating the opposite-directed flexures of the Ionian oceanic crust respectively under the Calabrian and the Hellenic accretionary wedges (Le Pichon, 1996). Another poorly documented, but possibly important, structural variation is that from the Ionian oceanic basin to its Apulian continental margin, which can be seen in S–N direction west of the CTF, and the possible burial of which further to the east would likely have influenced structure and activity.

This region is characterized by such a strong variation of conditions, both in space at the scale of a 100 km and time at the scale of less than a million years, that it cannot be considered an example of a steady-state subduction zone. The structure corresponding to the sum of past evolution may not be directly indicative of present processes. The upper plate deformation, fast motion of the plate boundary and loading of the lower plate are expressed in features of the seismic activity discussed here. They characterize this convergence region, the peculiarity of which can be related both to its spatial confinement, with the northward motion of Arabia pushing it sideways towards the converging boundary, and to its geological evolution, in continuity with an alpine orogenic phase which thickened the crust, building up present body forces with respect to the adjacent marine domain.

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