

A model for the Hellenic subduction zone in the area of Crete based on seismological investigations

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Abstract: The island of Crete represents a horst structure located in the central forearc of the retreating Hellenic subduction zone. The structure and dynamics of the plate boundary in the area of Crete are investigated by receiver function, surface wave and microseismicity using temporary seismic networks. Here the results are summarized and implications for geodynamic models are discussed. The oceanic Moho of the subducted African plate is situated at a depth of about 50–60 km beneath Crete. The continental crust of the overriding Aegean lithosphere is about 35 km thick in eastern and central Crete, and typical crustal velocities are observed down to the upper surface of the downgoing slab beneath western Crete. A negative phase at about 4 s in receiver functions occurring in stripes parallel to the trend of the island points to low-velocity slices within the Aegean lithosphere. Interplate seismicity is spread out about 100 km up-dip from the southern coastline of Crete. To the south of western Crete, this seismically active zone corresponds to the inferred rupture plane of the magnitude 8 earthquake of AD 365. In contrast, interplate motion appears to be largely aseismic beneath the island. The coastline of Crete mimics the shape of a microseismically quiet realm in the Aegean lithosphere at 20–40 km depth, suggesting a relation between active processes at this depth range and uplift. The peculiar properties of the lithosphere and the plate interface beneath Crete are tentatively attributed to extrusion of material from a subduction channel, driving differential uplift of the island by several kilometres since about 4 Ma.

Since the Late Cretaceous, the tectonics of the Eastern Mediterranean region has been controlled by (1) convergence between Africa and Eurasia and (2) subduction of oceanic lithosphere of narrow small oceanic basins separating Gondwana-derived terranes (e.g. Dercourt *et al.* 1986; Gealey 1988; Stampfli & Borel 2004). Oceanic lithosphere is now almost completely subducted, with remaining remnants beneath the Ionian basin and in the Eastern Mediterranean south of western Turkey. The closure of the earlier subducted oceanic basins and the collision of the intervening terranes with the Eurasian active margin resulted in a number of distinct orogenic belts and accretion of continental crust. In the Aegean, rollback of the active continental margin has been important since at least 30 Ma (e.g. Angelier *et al.* 1982; Thomson *et al.* 1998; ten Veen & Postma 1999), collision between the African passive margin and the Aegean continental lithosphere is incipient (Mascle & Chaumillon 1997; Mascle *et al.* 1999; ten Veen & Kleinspehn 2003; Meier *et al.* 2004a).

The position of subducted oceanic lithosphere in the Hellenic subduction zone was imaged by seismic tomography in the upper mantle (Spakman *et al.* 1988; Ligdas *et al.* 1990; Papazachos *et al.* 1995;

Taymaz 1996; Alessandrini *et al.* 1997; Papazachos & Nolet 1997; Piromallo & Morelli 1997, 2003; Marone *et al.* 2004) and well into the lower mantle (Spakman *et al.* 1993; Bijward *et al.* 1998; Bijward & Spakman 2000; Karason & van der Hilst 2000; Faccenna *et al.* 2003), down to depths of about 1700 km. Only the upper part of the slab represents African lithosphere that was subducted at the present active continental margin south of Crete. Subduction south of Crete started at about 20–15 Ma, when the plate boundary shifted to the southern border of an accreted microcontinent building most of the continental crust of present Crete (Thomson *et al.* 1998). The active margin has since retreated by about 350–500 km in a southwestward direction (ten Veen & Meijer 1998; ten Veen & Kleinspehn 2003). The coeval convergence between the African and the Eurasian plates amounts to about 150 km (Dercourt *et al.* 1986; Gealey 1988; Jolivet & Faccenna 2000; Faccenna *et al.* 2003; Stampfli & Borel 2004). This means that a slab of about 500–650 km length entered the subduction zone since the plate boundary shift to the south of Crete (Meier *et al.* 2004a). In accordance with tectonic reconstructions (Dercourt *et al.* 1986; Gealey 1988; Stampfli &

Borel 2004), the slab of more than 2000 km length imaged by seismic tomography must be composed of lithosphere once underlying different oceanic basins that were successively subducted at different active margins. After closure of each oceanic basin and collision of the trailing continental terrane with the Eurasian margin, a new active margin developed to the south of the accreted terrane, and subduction of the next stripe of oceanic lithosphere commenced. To explain the continuous slab imaged by seismic tomography, Meier *et al.* (2004a) proposed that continental crust of the terranes was delaminated from the underlying continental mantle lithosphere, as proposed by Thomson *et al.* (1998, 1999) in the 'buoyant escape' scenario based on the geological record of Crete. If this is true, the slab imaged by seismic tomography would be composed of alternating sections of oceanic and continental mantle lithosphere (Meier *et al.* 2004a; van Hinsbergen *et al.* 2005) of variable width, corresponding to the oceanic basins and continental terranes depicted in the palaeogeographical reconstructions (Dercourt *et al.* 1986; Gealey 1988; Stampfli & Borel 2004). It is possible that the presence of sections of delaminated subcontinental mantle devoid of hydrated material is the reason for the missing deep seismicity in the present-day Hellenic subduction zone.

The recent kinematics of the Eastern Mediterranean are characterized by westward drift and counter-clockwise rotation of the Anatolian microplate and southward escape of the Aegean lithosphere (McKenzie 1970, 1972, 1978; Le Pichon *et al.* 1995; McClusky *et al.* 2000). Global positioning system (GPS) surveys demonstrate that the rate of SSW motion of the Aegean lithosphere with respect to stable Eurasia increases towards the south (McClusky *et al.* 2000). This implies extension in the overriding plate, which is attributed to rollback and retreat of the subduction zone (Le Pichon & Angelier 1979; Angelier *et al.* 1982; ten Veen & Meijer 1998; Armijo *et al.* 2003). Within the given kinematic framework, the high curvature of the active continental margin implies that subduction becomes increasingly oblique from west to east. In the western forearc, the vectors of motion are oriented normal to the continental margin. Towards the east, in the area of Rhodes, the angle between the motion vector and the active margin decreases to about 20°. The relative velocity between the subducting African plate and the overriding Aegean lithosphere is about 4.5 cm a^{-1} (McClusky *et al.* 2000) at the active margin.

Seismic activity in the forearc of the Hellenic subduction zone is considerable. In the last 2.5 ka several events reached a magnitude of eight (Papazachos *et al.* 2003). In AD 365, an event with an estimated magnitude of 8.3 occurred in the western forearc, to the SW of Crete. The late

Holocene uplift, of up to 9 m at the southwestern coast of Crete, is attributed to co-seismic deformation during this event (Pirazzoli *et al.* 1982; Stiros 2001). According to the global catalogue of Engdahl *et al.* (1998), eight events with a magnitude of six or above occurred in the Hellenic subduction zone in the second half of the 20th century.

One peculiarity of the Hellenic forearc is the geologically recent rapid uplift in the area of Crete (Meulenkamp *et al.* 1994; Lambeck 1995), of up to several kilometres since about 4 Ma with an average rate close to 1 mm a^{-1} . This Plio-Pleistocene uplift has produced a marked relief, with Crete rising to nearly 2.5 km above sea level, close to narrow furrows with water depths of more than 3 km to the south. These deep topographic furrows to the south of Crete do not mark the position of the plate boundary (Le Pichon & Angelier 1979; Angelier *et al.* 1982). The widely used term 'Hellenic trench', commonly used to refer to the distinct topographic feature marking the site where the subducted plate disappears beneath the forearc at a convergent plate boundary, may be misleading here. In the Hellenic subduction zone, the accretionary complex (e.g. Kastens 1991; Mascle & Chaumillon 1997) is located to the south of these 'trenches' beneath the Mediterranean ridge, and the trenches mark tectonic features within the forearc, and hence within the overriding Aegean lithosphere. The Cretan horst structure shows the following characteristic features: (1) it is narrow in the north-south direction; (2) it is bounded by normal faults on all sides; (3) uplift commenced abruptly in the early Pliocene; (4) uplift is contemporaneous with continuing forearc-parallel extension of the Aegean lithosphere. The question is: what drives Crete up?

In this paper, recent results of receiver function, surface-wave and microseismicity studies using temporary local networks are reviewed. The implications for the structure and dynamics of the Hellenic subduction zone are discussed and a model for the rapid localized uplift of Crete is proposed.

Slab segmentation in the Hellenic subduction zone

The seismicity of the Hellenic subduction zone between 1964 and 1998 according to the global relocated ISC catalogue (Engdahl *et al.* 1998) is shown in Figure 1. In this region the catalogue is complete for events with magnitudes greater than *c.* 4.7. Clusters of events are observed within the part of the forearc characterized by intense shallow seismicity. The trend of these clusters is oriented NE-SW (Fig. 1). The depth of hypocentres increases with distance from the plate boundary and reaches

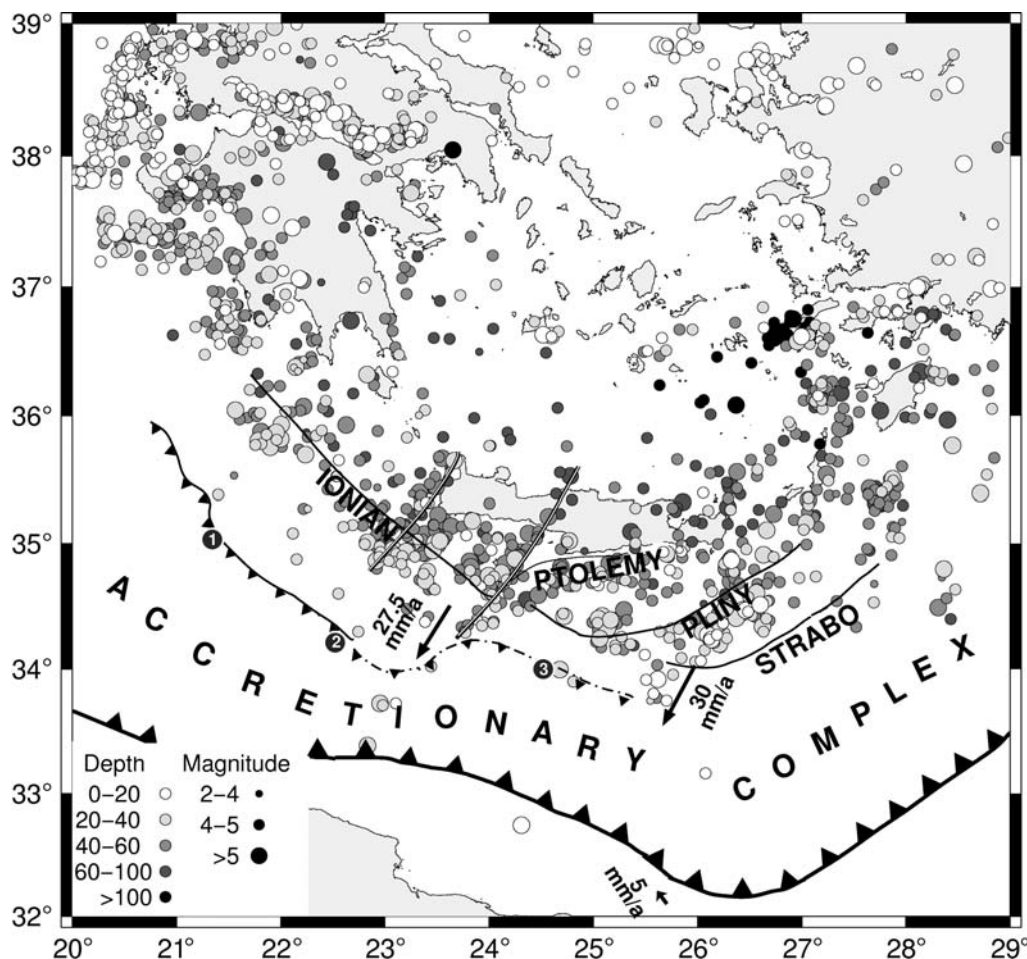


Fig. 1. Hypocentres of events in the Aegean region listed in the relocated ISC catalogue (Engdahl *et al.* 1998) for the years 1964–1998. The position of the deformation front of the Hellenic subduction zone (Le Pichon *et al.* 1995) and the border between the central and the inner units of the Mediterranean Ridge (continuous line with triangles, Lallemand *et al.* 1994; dashed–dotted line with triangles, Mascle *et al.* 1999) are indicated. Points labelled with numbers mark the southern border of the Aegean crust as determined by seismic studies (1, Lallemand *et al.* 1994; 2, Brönnert 2003; 3, Bohnhoff *et al.* 2001), corresponding to the transition between the inner and the central units of the Mediterranean Ridge. Relative horizontal velocities with respect to Eurasia (McClusky *et al.* 2000) are indicated by arrows. In addition, the deep furrows ('trenches') to the south of Crete and the NE–SW-striking linear patterns of seismicity in western and central Crete are emphasized.

about 100 km in the west and about 160 km in the east. These intermediate depth events define a strongly curved Benioff zone, with its lower termination approximately beneath the volcanic arc (Papazachos & Comninakis 1971; Comninakis & Papazachos 1980; Makropoulos & Burton 1984; Taymaz *et al.* 1990; Hatzfeld & Martin 1992; Knapmeyer 1999; Papazachos *et al.* 2000).

The hypocentres of the relocated ISC catalogue (Engdahl *et al.* 1998) are projected onto vertical cross-sections in Figure 2. The figure shows that

there is not much intermediate-depth Benioff zone seismicity compared with, for example, the western Pacific subduction zones. As is evident from Figure 2d and e, the Benioff zone can be traced to a depth of about 160 km, with a faint seismic gap between about 100 and 125 km (Papazachos *et al.* 2004). In the east, the Benioff zone dips more steeply than in the west. As is evident from Figure 2, data from local networks are required to image the seismogenic zones beneath the forearc in greater detail.

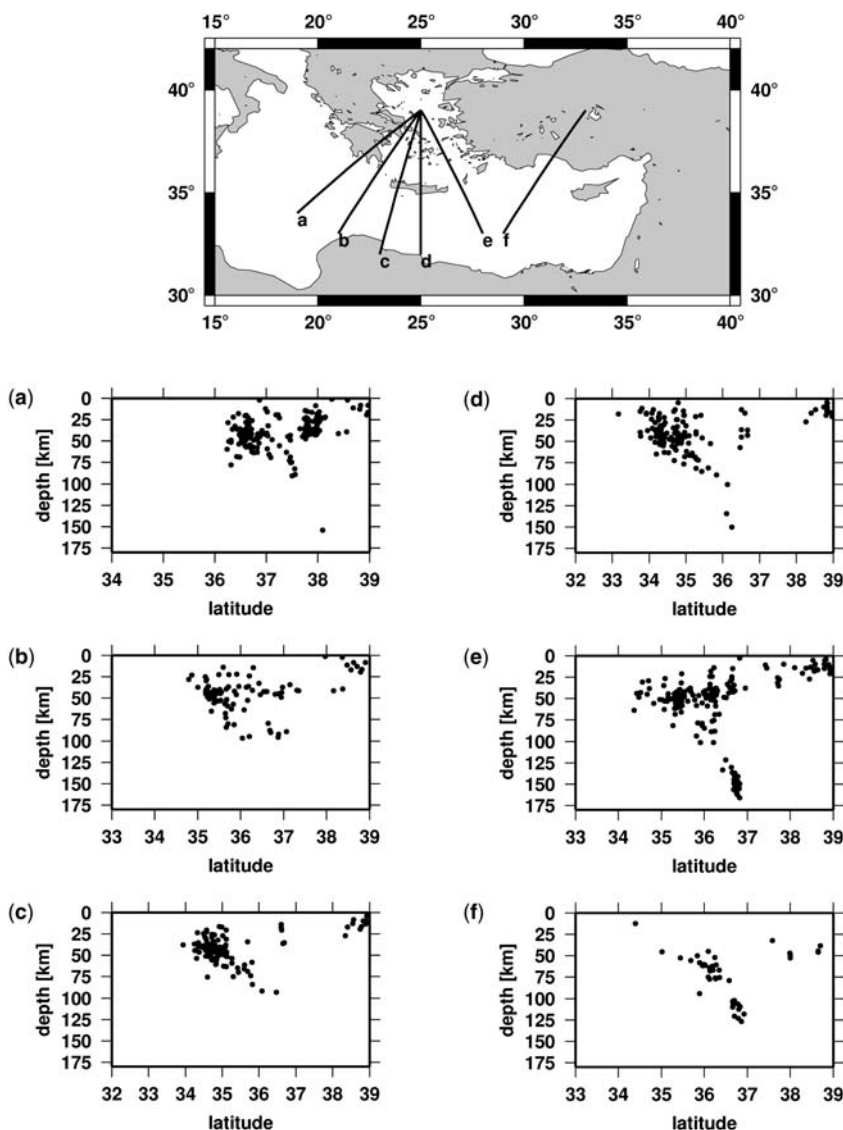


Fig. 2. Hypocentres from the global catalogue by Engdahl *et al.* (1998) projected onto vertical planes along the profile lines indicated in the map.

In the western Cyprus arc (section f in Fig. 2), the hypocentres define a Benioff zone with subduction towards the NE beneath Anatolia. As the slab in the eastern part of the Hellenic subduction zone is inclined towards the NW, a slab window is expected to exist beneath western Turkey, which probably affects deformation and magmatism of this region. The schematic sketch in Figure 3 illustrates the situation, as suggested by intermediate-depth seismicity (e.g. Engdahl *et al.* 1998).

In the western segment of the Hellenic subduction zone, the dip of the slab increases at a depth

of about 75 km (Hatzfeld 1994; Papazachos *et al.* 1995, 2000; Taymaz *et al.* 1990; Papazachos & Nolet 1997). Based on a regional tomographic study, Papazachos & Nolet (1997) concluded that at about 200 km depth the western and eastern slab segments meet along a rather sharp edge with a right angle. Wortel & Spakman (2000) assumed a tear within the slab, propagating horizontally from the Ionian Sea along the forearc, which causes incipient slab detachment. Apart from the strong curvature enhanced by rollback (e.g. ten Veen & Kleinspehn 2003), the segmentation of

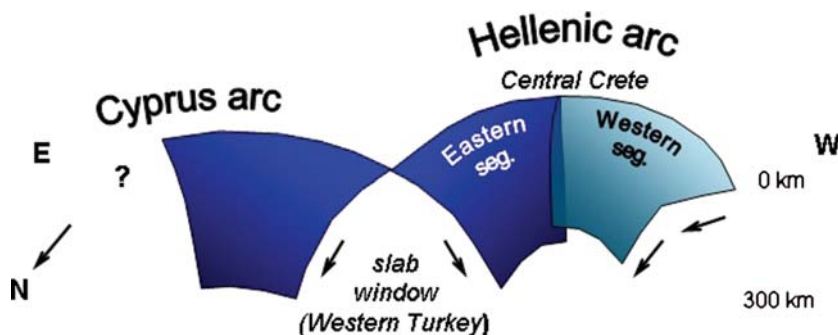


Fig. 3. Schematic illustration of the first-order segments of the African slab in the Eastern Mediterranean down to about 300 km (view from the north).

the subducting slab may result from lateral heterogeneity of the African lithosphere, which is indicated by body-wave and surface-wave studies for the Libyan Sea south of Crete (Piromallo & Morelli 1997; Bijward & Spakman 2000; Marone *et al.* 2004). To the south of western Crete, seismic velocities in the mantle lithosphere are lower than those typically observed for oceanic lithosphere (Marone *et al.* 2004; Meier *et al.* 2004a). This indicates that the passive margin of northern Africa is entering the Hellenic subduction zone to the south of western Crete (Meier *et al.* 2004a). In contrast, to the south of eastern Crete, according to tomographic studies (Bijward & Spakman 2000; Marone *et al.* 2004), remnants of oceanic lithosphere seem still to be present.

Topography and tectonics in the area of Crete

Crete is situated in the central forearc of the Hellenic subduction zone. Figure 4 depicts the topography of the Hellenic convergent margin along a north–south section at 25.12° E, from the Cyclades over Crete to northern Africa. The volcanic arc is underlain by continental crust of the Cyclades, which has undergone considerable stretching since mid-Tertiary times and is characterized by metamorphic core complexes. The average water depth in the volcanic arc region is a few hundred metres. It increases to more than 1 km in the Cretan Sea north of Crete, where crustal thickness is reduced to well below 20 km (Bohnhoff *et al.* 2001). As is evident from Figure 4, Crete forms a narrow spike in topography, abruptly rising on both sides.

The Ionian trench is a NW–SE-trending furrow barely 30 km from the southwestern corner of Crete (Fig. 1). According to wide-angle seismic data, the southern border of the Aegean lithosphere is located

about 150 km to the SSW of this deep furrow (Truffert *et al.* 1993; Lallemand *et al.* 1994; Bohnhoff *et al.* 2001). Consequently, the Hellenic trench system does not mark the plate boundary but constitutes tectonic features within the forearc. Further to the east, to the south of central Crete, the Ionian trench system bifurcates into three branches trending NE–SW, referred as the Ptolemy, Pliny and Strabo trenches (Fig. 1). These furrows are at an angle of about 30° to the direction of relative motion between the Aegean and the subducted African plate. The Ptolemy and Pliny trenches are therefore interpreted to represent trans-tensional structures within the Aegean lithosphere (Huchon *et al.* 1982; Huguen *et al.* 2001; Brönnert 2003; ten Veen & Kleinspehn 2003). South of eastern Crete, the Strabo trench marks the position of the backstop, where sediments of the Mediterranean Ridge accretionary complex are backthrust onto Aegean continental crust (Huguen *et al.* 2001). According to Mascle & Chaumillon (1997), the southern border of the Aegean lithosphere appears to be offset by about 50 km to the south of central Crete (Fig. 1).

The southern deformation front of the Hellenic convergent margin is located to the south of the Mediterranean Ridge (e.g. Le Pichon *et al.* 1995; Mascle & Chaumillon 1997; Mascle *et al.* 1999). To the south of western Crete, the deformation front bends northward (Fig. 1). This is attributed to incipient collision of the accretionary complex with the passive continental margin of Africa (LePichon *et al.* 1995; Mascle & Chaumillon 1997; Mascle *et al.* 1999). The arc-normal contraction in the western forearc, indicated by focal mechanism solutions for events down to about 20 km depth (Lyon-Caen *et al.* 1988; Taymaz *et al.* 1990; Hatzfeld *et al.* 1993a; Baker *et al.* 1997; Benetatos *et al.* 2004; Bohnhoff *et al.* 2005), is consistent with incipient collision.

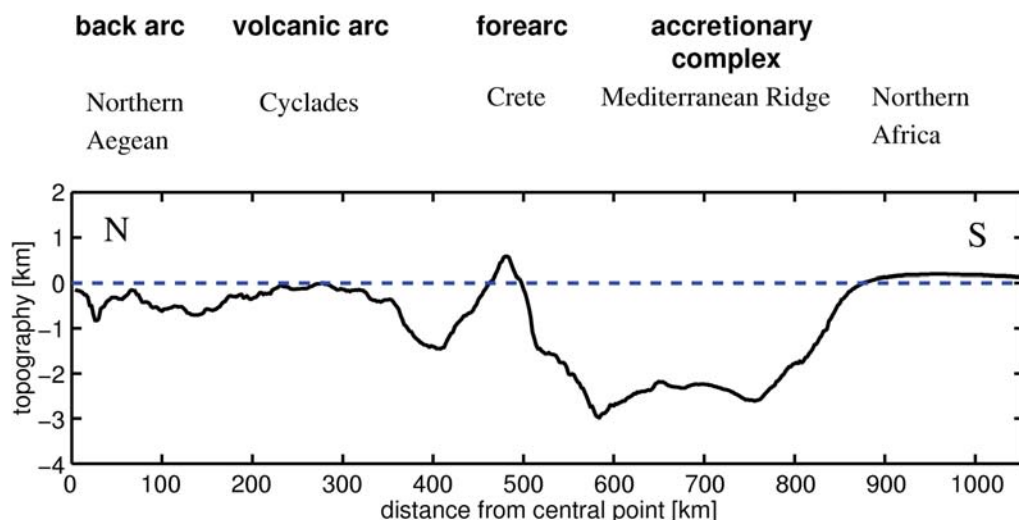


Fig. 4. Average topography along a profile from the Cyclades to northern Africa along 25.12° E from 39.5° N to 29° N.

Temporary seismic networks

To gain more detailed information on the structure and seismicity of the convergent margin, several temporary local seismic networks were installed in the Aegean in recent decades. In 1988, microseismic activity was investigated using a temporary seismic network covering large parts of the southern Aegean (Hatzfeld *et al.* 1993a). Regionally focused studies of microseismic activity were dedicated to the Peloponnese (Hatzfeld *et al.* 1989; Sachpazi *et al.* 2000; Makris *et al.* 2004) and to the eastern part of the volcanic arc (Makris *et al.* 2000). In 2002–2004, a temporary seismic network provided insight into microseismic activity in the Cyclades (Bohnhoff *et al.* 2004). The island of Crete was the base for a temporary local network of seismic stations, first installed in eastern Crete and then displaced to the western part of the island for a second month of observation (de Chabaliér *et al.* 1992; Hatzfeld *et al.* 1993b). This study was complemented by a further campaign in central Crete in 1995 (Delibasis *et al.* 1999). Since 1996, a more extensive database was obtained by operating five temporary local onshore networks on Crete with up to 47 stations and recording intervals of up to 18 months (Becker 2000; Meier *et al.* 2004b). Data from these five networks are considered in this paper. Recently, these onshore networks were complemented by an offshore ocean-bottom seismograph (OBS) network in the Libyan Sea south of central Crete, with the aim of gaining information on microseismicity near the Ptolemy and Pliny trenches and localizing interplate microseismicity

with higher accuracy. The location of the temporary stations deployed between 1996 and 2004 is shown in Figure 5. Moreover, nine permanent broadband stations are currently operated in the region of Crete by the GEOFON network (Hanka & Kind 1994), the National Observatory of Athens and MEDNET (Boschi & Morelli 1994). This dense station coverage yields an extensive database for receiver function, surface-wave and microseismicity studies.

Structure of the subduction zone in the area of Crete

The most important of the structural features of the subduction zone is the position of the subducted African lithosphere. To obtain this first-order information, Knapmeyer & Harjes (2000) used receiver functions (Fig. 6) for short-period stations in western Crete, and Li *et al.* (2003) studied receiver functions of broadband stations on the island. The results indicate that the crust–mantle discontinuity within the subducting slab is at about 55 km depth beneath Crete. This was confirmed by Meier *et al.* (2004a), who studied surface-wave propagation between two broadband stations on Crete and found the African Moho at a similar depth of 50 km. A WNW–ESE-oriented cross-section through migrated receiver functions of short-period stations in western and central Crete (Endrun *et al.* 2004) is shown in Figure 6a, and clearly depicts the discontinuity interpreted to represent the African Moho within the slab at about 50–55 km depth.

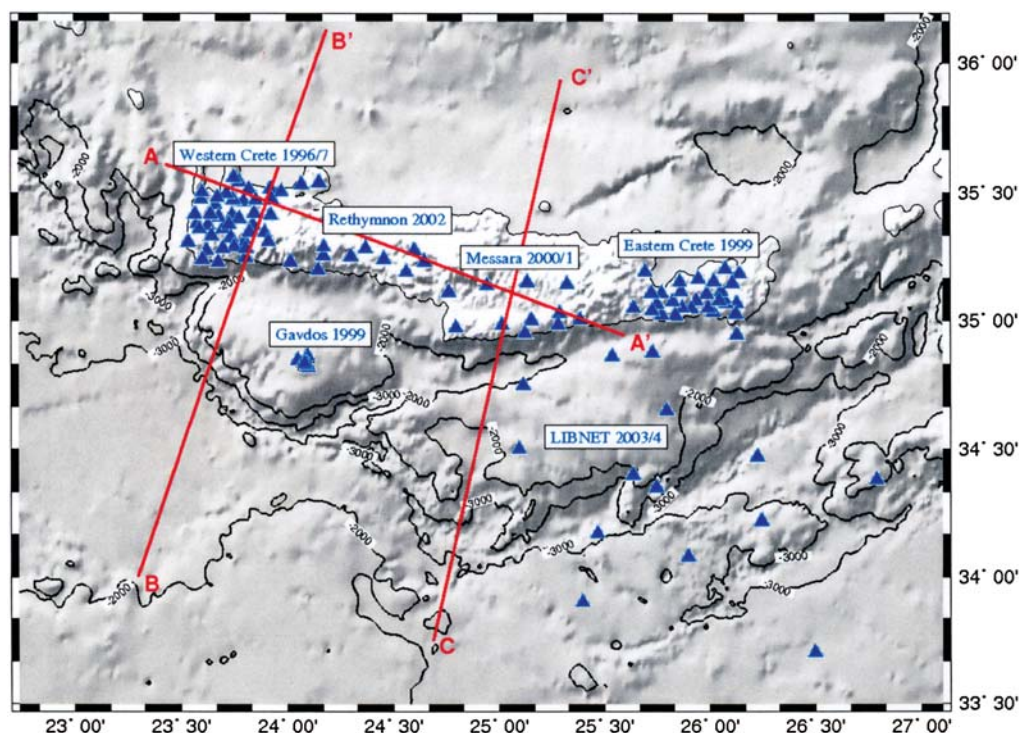


Fig. 5. Temporary short-period networks deployed in the area of Crete between 1996 and 2004. Western Crete: six stations in May–August 1996 and 47 stations in September–December 1997; Gavdos: 12 stations in May–August 1999; Messara: eight stations in April 2000–December 2001; Rethymnon: seven stations in May–November 2002; LIBNET: up to eight OBS deployed in four 2 month campaigns in 2003–2004; eastern Crete: 27 stations in January–March 1999. Location of receiver function profiles shown in Figure 6 are indicated by red lines.

The notable differences in the structure between the eastern and western parts of section A (Fig. 6a) suggest a significant lateral heterogeneity within the overriding Aegean lithosphere. A discontinuity observed at about 35 km beneath central Crete is interpreted as the crust–mantle discontinuity of the Aegean lithosphere. This interpretation is in accordance with wide-angle seismic data (Bohnhoff *et al.* 2000) and surface-wave studies (Endrun *et al.* 2004; Meier *et al.* 2004a). According to wide-angle seismic data, the depth of the Aegean Moho strongly decreases northwards and is located at about 17 km depth beneath the Sea of Crete (Bohnhoff *et al.* 2000), probably as a result of extensional thinning of the crust in the recent geological past. A negative phase in the receiver function of a broadband station in northern central Crete (Li *et al.* 2003) and in the receiver function image along section C (Fig. 6c) may be caused by the strong Moho topography (Endrun *et al.* 2005). In contrast, in western Crete the receiver function image shown in Figure 6a does not reveal a positive discontinuity at about 35 km depth that could be

attributed to a transition from crustal to mantle velocities. Wide-angle seismic data reveal high reflectivity and a complex structure of the Aegean lithosphere at this depth (Bohnhoff *et al.* 2000). Li *et al.* (2003) found a negative phase in the receiver function of a broadband station in western Crete at about 4 s, corresponding to about 35 km depth. The detailed analysis of the spatial distribution of this negative phase (Endrun *et al.* 2004) revealed a marked lateral heterogeneity at 35 km depth. Arc-parallel stripes with a negative receiver function phase at 4 s delay time in western Crete point to slices with varying velocity including distinct low-velocity bodies. Finally, Figure 6b shows a NNE–SSW-oriented section of migrated receiver functions in western Crete. Negative phases are present at 20–40 km depth beneath northwestern Crete. Analysis of surface-wave propagation within the short-period network in western Crete indicates typical crustal velocities for the Aegean lithosphere down to the plate contact at about 45 km depth (Endrun *et al.* 2004). If so, the crustal thickness in western Crete would be

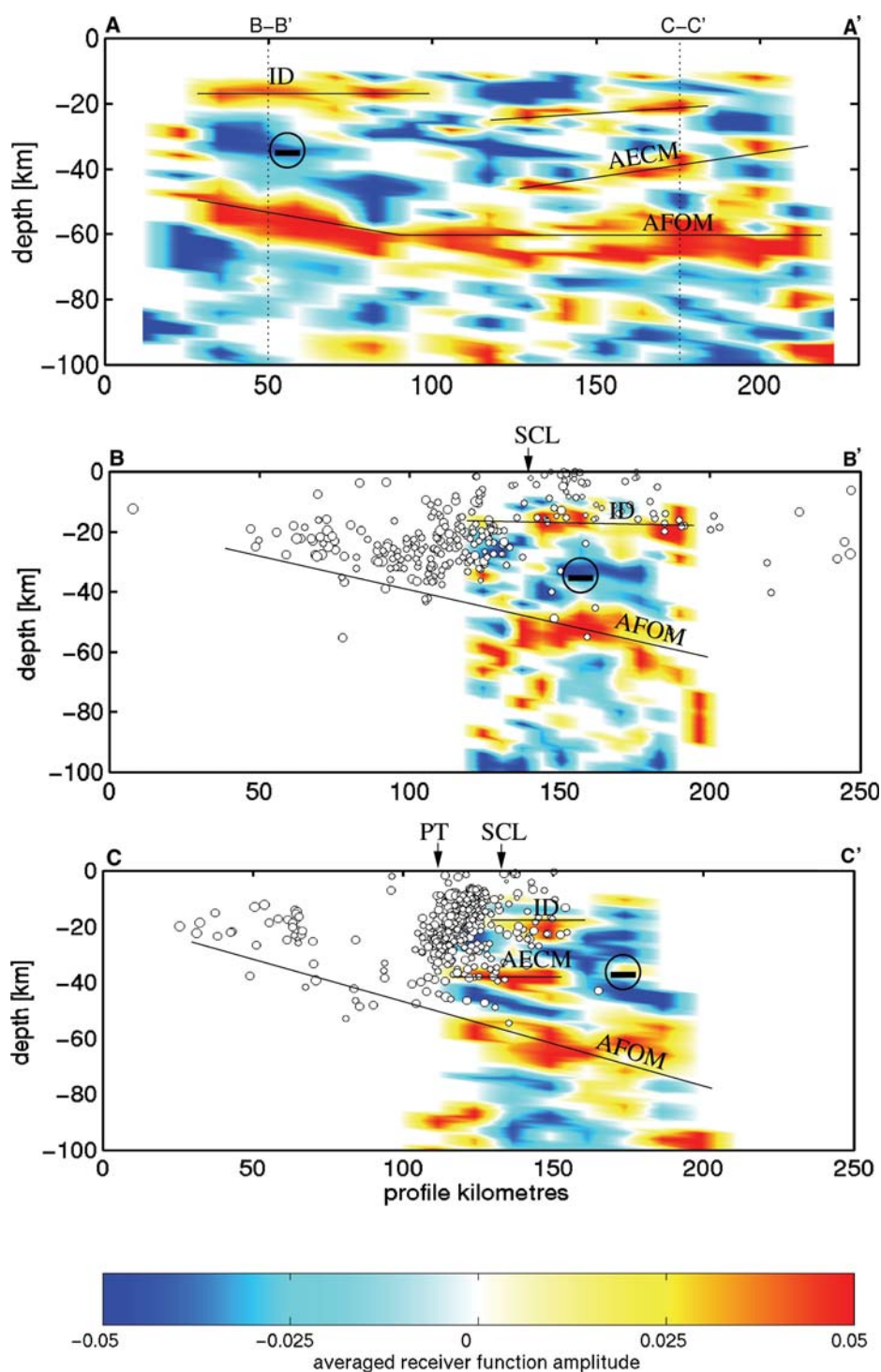


Fig. 6. Cross-sections through migrated receiver functions of short-period stations (location of the profiles shown by red lines in Fig. 5). The hypocentres of microseismicity are projected onto arc-normal cross-sections BB' and CC'. AECM, Aegean continental Moho; AFOM, African oceanic Moho; ID, intracrustal discontinuity; PT, Ptolemy trench; SCL, southern coastline of Crete.

comparable with that beneath the Peloponnese. The persistent low-velocity layer at mid-crustal depths identified by Papazachos *et al.* (1995) beneath the Peloponnese seems to be absent beneath western Crete, however (Endrun *et al.* 2004). Instead, low-velocity zones indicated by negative receiver function phases seem to represent individual slices of limited lateral extent within the Aegean lithosphere.

McClusky *et al.* (2000) demonstrated that the southeastern part of the overriding Aegean lithosphere is moving relative to the southwestern part. Segmentation of the subducted slab and lateral heterogeneity of the overriding Aegean lithosphere in the area of Crete are indicated by several features: (1) the change in the dip and the ultimate depth of the Benioff zone; (2) the NE–SW-oriented cluster of seismic events in central Crete (Fig. 1); (3) the lateral heterogeneity of the Aegean lithosphere beneath central Crete found by receiver function, surface-wave and seismicity studies; (4) changes in the structure of the Hellenic trench system along the forearc; (5) changes in the position of the southern border of the Aegean lithosphere south of central Crete (Fig. 1).

Microseismicity

The catalogue of global seismicity published by Engdahl *et al.* (1998) covers a timespan of about 40 years and an earthquake magnitude range of about 4–6.5. Localization uncertainties can be as high as 20 km in the Aegean. Temporary onshore local networks reduce these uncertainties to about 5–10 km (Meier *et al.* 2004b). Localization accuracy of microseismicity south of Crete will be further improved when data from offshore networks become available. The observation periods of temporary networks are much smaller, however, than that of the global catalogue by Engdahl *et al.* (1998). For the networks shown in Figure 5 observation times vary between 3 and 18 months. Nevertheless, about 5000 events with magnitudes between -1 and 4.5 were detected and localized. The persistent high level of microseismic activity in the forearc allowed investigation of the spatial distribution of seismically active zones despite the relatively short observation times inherent in the temporary networks. Clearly, an assessment of temporal variations in the seismic activity would require longer observation periods. The occurrence of swarm-like clusters of microseismicity in the forearc was detected by Becker *et al.* (2006).

Within the observation period, microseismic activity is observed between about 20 and 40 km depth (Fig. 6b and c) along the inferred plate contact. The current width of the interplate

seismogenic zone is about 100 km to the south of western Crete, close to 50 km to the south of central Crete, and again up to about 100 km in the east. At the southern border of the overriding Aegean lithosphere, the plate contact is characterized by a microseismically quiet zone about 50 km wide (Meier *et al.* 2004b). This zone may represent a décollement horizon with aseismic deformation localized in unconsolidated sediments.

Particularly intense microseismicity at the plate contact is observed to the south of central Crete (Fig. 6c). The northern border of this seismically active zone along the plate interface coincides with the southern coastline of Crete (Meier *et al.* 2004b). The microseismically active zone along the plate interface to the south of western Crete is currently about 100 km wide (Meier *et al.* 2004b) and coincides with the fault plane proposed by Papazachos *et al.* (1999) for the historical magnitude 8.3 event that occurred on 25 July AD 365. The length of the fault plane activated in this event is estimated to be about 200 km, based on palaeoseismological studies (Pirazzoli *et al.* 1982; Stiros 2001). If the microseismically active zone corresponds to the part of the fault activated in the AD 365 event, the ruptured area would be about 200 km $\times$ 100 km, roughly compatible with the expected rupture area of a magnitude 8 event. Laigle *et al.* (2004) proposed that the rupture area exceeded the width of the currently microseismically active zone in a down-dip direction.

Beneath Crete, the plate contact is almost devoid of microseismicity. In this area, microseismicity is restricted to the upper c. 20 km of the Aegean lithosphere, with a low level of microseismic activity down to the inferred upper interface of the subducted African plate (Fig. 6b and c). In addition, a vertical zone of localized seismic activity follows the Ptolemy trench and cuts through the entire Aegean lithosphere, from the plate interface through the uppermost crust (Meier *et al.* 2004a).

Microseismic activity in the area of Crete is concentrated in a certain depth range, which is not uniform over the area. Owing to complete coverage of Crete by the successively operated local networks, the median of the vertical depth distributions of the events could be derived for the entire island and its surroundings; the pattern is shown in Figure 7. To the south of Crete the median depth of the microseismicity corresponds approximately to the inferred depth of the plate contact. The shift towards a shallower median depth of microseismicity correlates with the southern coastline of Crete (Fig. 7). Seismic activity at the inferred level of the plate interface is almost absent beneath Crete. The coastline of Crete delineates a seismically quiet zone at a depth between about 20 and 40 km beneath Crete. This coincidence suggests that the

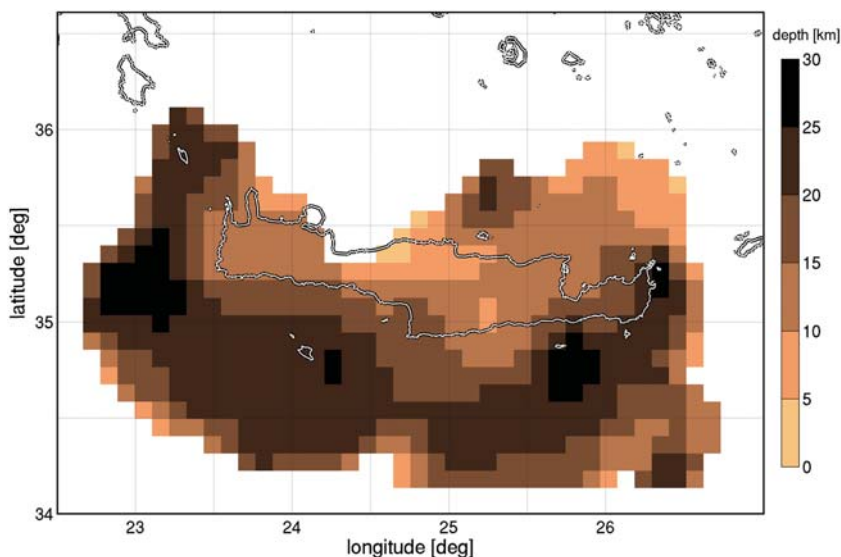


Fig. 7. Map view of average hypocentre depths of microseismicity. Hypocentres of c. 5000 microseismic events were detected and localized using data from temporary stations shown in Figure 5.

recent uplift of Crete may be related to processes currently active at these depths.

Implications for geodynamic models of the Hellenic subduction zone around Crete

The inferred structure of the Hellenic subduction zone in the area of Crete, and the distribution of seismicity, is shown by the schemes in Figure 8. Figure 8a displays the situation in western Crete, viewing to the NW, and Figure 8b that in eastern Crete, viewing to the NE. As shown in Figure 8, there are some differences in the structure and distribution of seismicity in the two areas. In western Crete, the leading edge of the passive African continental margin has entered the subduction zone and collision has commenced, whereas oceanic lithosphere is still subducted south of eastern Crete with rollback being effective (ten Veen & Meijer 1998; ten Veen & Postma 1999; ten Veen & Kleinspehn 2003). Beneath eastern Crete, the slab dips at an angle of about 19° compared with about 15° beneath western Crete.

As emphasized in Figure 8, interplate microseismicity is intense to the south of Crete. In contrast, the plate interface appears to be microseismically inactive further to the north, where the interface is at depths greater than c. 40 km, with a sharp cutoff beneath the southern coastline of Crete. This finding is similar to observations in other subduction zones (e.g. Tichelaar & Ruff 1993). Fault-plane

solutions (Taymaz *et al.* 1990; Bohnhoff *et al.* 2005) show that the seismically active upper part of the Aegean lithosphere beneath western Crete undergoes contraction normal to and extension parallel to the plate boundary. A similar situation holds for the forearc beneath the Peloponnese (Lyon-Caen *et al.* 1988; Hatzfeld *et al.* 1993a; Baker *et al.* 1997; Benetatos *et al.* 2004), indicating coupling between the plates in the frontal part of the forearc. The current overall extensional tectonics observed at the surface (e.g. Angelier *et al.* 1982) is in apparent contrast to the forearc-normal compressional regime at upper crustal depths indicated by focal mechanisms. Focal mechanisms of intermediate-depth events point to forearc-parallel compression and down-dip extension (Papazachos *et al.* 2000). These intermediate-depth events are supposed to occur within the subducting plate and suggest decoupling between the plates below about 50 km depth.

To the south of Crete, the NW–SE-trending Hellenic trench system bifurcates eastwards into WSW–ENE-trending transtensional graben structures (Fig. 1). For the Ptolemy trench (Fig. 6), to the south of eastern Crete, microseismic activity indicates a nearly vertical fault system that penetrates the entire Aegean lithosphere. Therefore, the Ptolemy trench is interpreted as a boundary with an independently moving forearc sliver (Fig. 8b) driven by oblique subduction (ten Veen & Kleinspehn 2003). The sense of motion along the Ptolemy trench is sinistral. The same probably holds for the Pliny trench, a transtensional structure

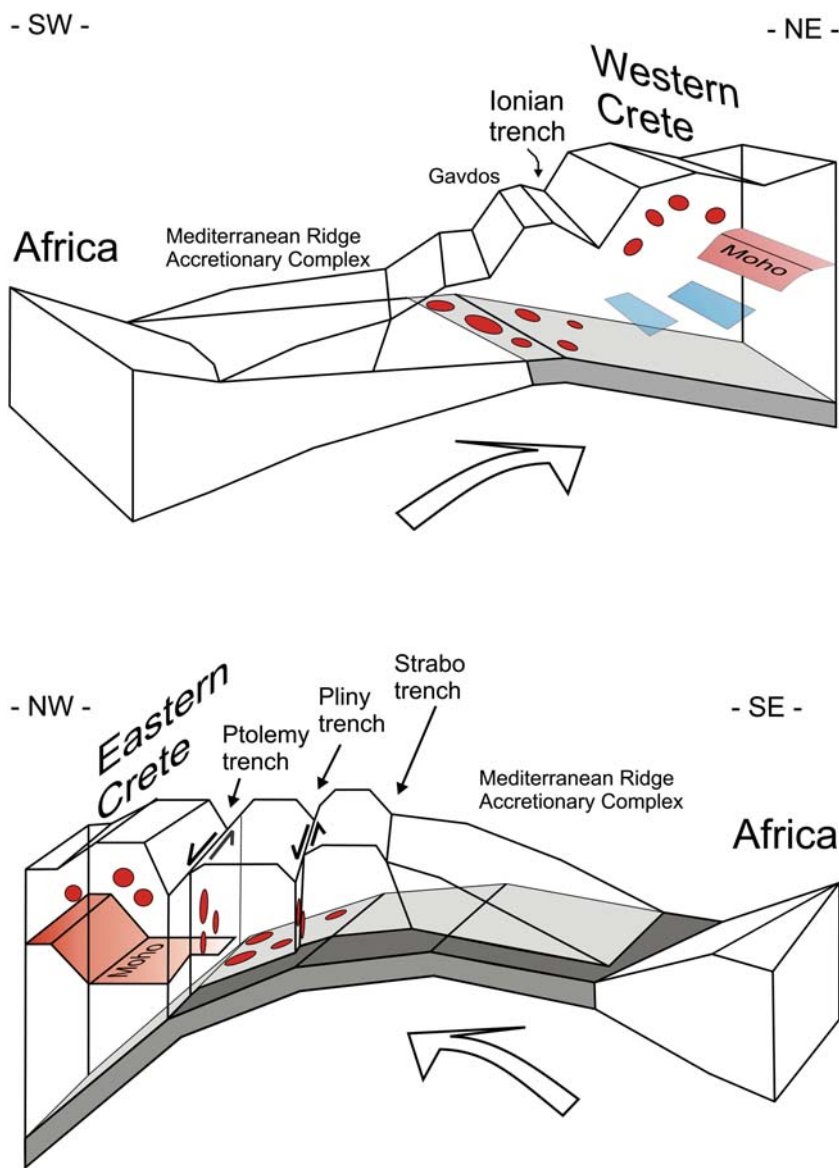


Fig. 8. Interpretation of the seismological results, shown in 3D sketches (not to scale) for (a) western Crete and (b) eastern Crete. The location of seismic events in the observation period is schematically indicated by red spots. The inferred position of the Aegean Moho based on receiver functions is marked by a red surface. Discontinuous blue surfaces schematically depict the negative phases observed in receiver functions beneath western Crete, suggesting low-velocity bodies in the overriding lithosphere (see Fig. 6).

paralleling the Ptolemy trench to the SE. Based on the observed microseismicity, motion between the forearc slivers is active. Sinistral displacement between these slivers created a basin south of central Crete, filled by a thick sedimentary sequence (Casten & Snopek 2005). The position of this basin is discernible from the offset in the front of the

overriding Aegean lithosphere (Masclé & Chaumillon 1997) and the distribution of interplate seismicity (Fig. 1).

Integration of the tectonic structure of the forearc and the information on present-day kinematics is hampered by the fact that a marked change in the mechanical state of the system must have occurred

a few million years ago in the Pliocene, when the rapid localized uplift of the narrow Cretan horst structure commenced. The onset of collision in the area of western Crete is expected to have caused further modifications. Notably, there is ample evidence for significant crustal thinning by arc-normal extension in the area of the Sea of Crete, but GPS surveys demonstrate no active extension in this part of the forearc (e.g. Le Pichon *et al.* 1995; McClusky *et al.* 2000). Recent extensional structures observed at the surface on Crete could therefore be the result of gravitational spreading of a mountain belt undergoing rapid localized uplift.

Several possible causes for this localized uplift have been proposed. Angelier *et al.* (1982) assumed that the uplift is caused by underplating of subducted sediments. Alternatively, buckling of the forearc as a result of coupling of the plates has been proposed (e.g. Savage 1983). Giunchi *et al.* (1996) introduced an additional arc-normal force in their numerical models to achieve the observed uplift rates. The buckling models seem to be in conflict with the narrow and discontinuous character of the horst structure, being bounded by normal faults. Forearc buckling by arc-normal compression would be expected to result in a continuous bulge of longer wavelength. As an alternative, based on the results summarized in the present paper, the driving

mechanism for the localized uplift of Crete is suspected to be extrusion of material from a subduction channel (Cloos & Shreve 1988*a, b*). This hypothesis is schematically depicted in Figure 9.

Rapid return flow of material along the hanging wall of the subducted slab has been proposed to explain the rapid exhumation (e.g. Rubatto & Hermann 2001) of high- and ultrahigh-pressure metamorphic rocks. Numerical simulations (Gerya *et al.* 2002; Gerya & Stöckhert 2007) support the feasibility of this concept, predicted particle trajectories and pressure–temperature–time paths being in accordance with the natural record in collisional belts. In these simulations, material extruded from the subduction channel is shown to spread out beneath the forearc. Such material, including hydrated mantle material, may build up part of the lower Aegean lithosphere (Fig. 9), which is characterized by typical crustal velocities, a complex internal structure, and the absence of seismicity. If so, the sudden onset of uplift in the Pliocene (e.g. Le Pichon *et al.* 1995) is tentatively interpreted to be the surface expression of the onset of extrusion from the subduction channel, after a period without return flow. This period, during which return flow and extrusion were not effective, could be envisaged to be a consequence of the mid-Miocene accretion of the microcontinent (Thomson *et al.* 1999) now building up most of the continental

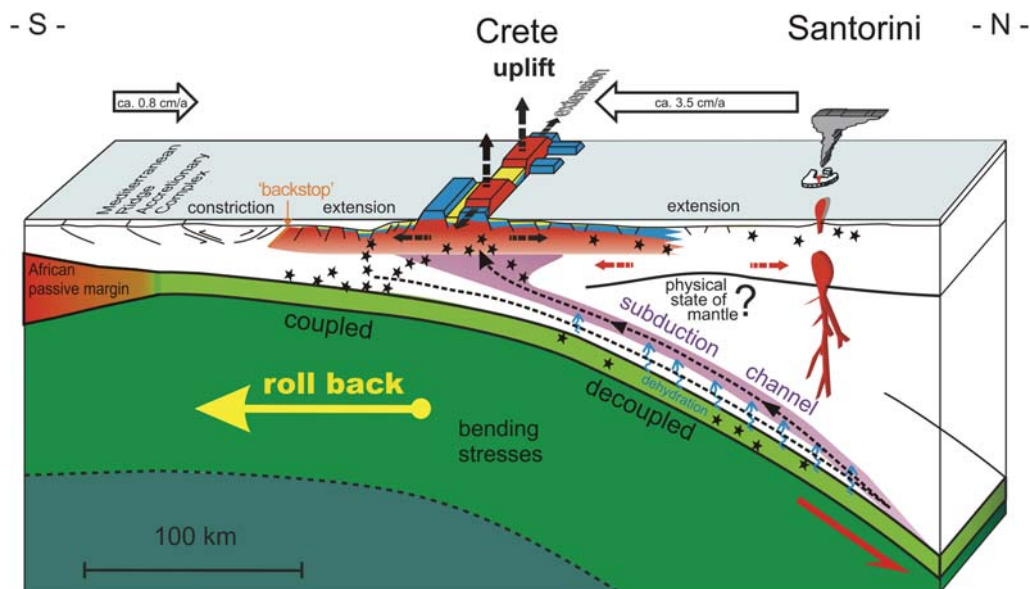


Fig. 9. Three-dimensional schematic illustration of the principal features of the Hellenic subduction zone in the area of Crete, and the proposed model for uplift of the island. The location of seismic activity within the observation period is shown schematically by stars. The bold black dashed arrows and labels 'extension' and 'constriction' refer to the geological record since about 4 Ma, when the uplift of Crete commenced. The proposed realm of backflow in the upper portion of a subduction channel is shown in purple.

crust beneath Crete. The simulations by Gerya *et al.* (2002) have demonstrated the importance of hydration of the mantle wedge atop the subducting slab for return flow. Subduction of dry subcontinental lithosphere, from which the buoyant continental crust of the accreted microcontinent had been detached (Thomson *et al.* 1999), is probably not favourable for return flow in a subduction channel. It is possible that a sufficient length of partially hydrated oceanic lithosphere of the African plate, trailing the continental lithosphere of the microcontinent, had to be subducted before return flow could be established. Assuming convergence and rollback with a constant rate close to 3 cm a^{-1} , about 400 km of oceanic lithosphere have been subducted between the time of accretion of the microcontinent at around 20 Ma and the onset of rapid uplift at about 4 Ma. A timespan of some 10 Ma would be sufficient to circulate material to a depth of about 100 km and back into the lower crust by return flow with plate velocity. It is therefore suspected that subduction of barren subcontinental mantle may have caused a temporary absence and subsequent re-establishment of return flow in a subduction channel, with the sudden onset of Cretan uplift as the surface expression of extrusion into the deeper crust becoming effective.

The recent extensional structures at the surface on Crete would then be a response of the upper crust to uplift, whereas contraction normal to the forearc, as indicated by the focal mechanisms, could persist in the middle crust because of coupling between the plates (Fig. 9). Notably, the addition of material extruded from the subduction channel would lead to growth of the forearc. This growth could partially compensate rollback and may have rendered lithospheric extension beneath the Sea of Crete ineffective since about 4 Ma, simultaneous with the sudden onset of uplift of Crete.

Summary and conclusions

Using data from short-period temporary networks, the lithospheric structure in the area of Crete was imaged by migrated receiver functions. A pronounced discontinuity at 50–60 km depth beneath Crete is interpreted to represent the African Moho within the subducting slab. Assuming a thickness of the African oceanic crust of about 5–10 km, the interplate contact is estimated to be at a depth of about 40–55 km beneath Crete. Within the overriding Aegean lithosphere, receiver function images display a marked forearc-parallel heterogeneity. In central and eastern Crete the Aegean continental crust is about 35 km thick and underlain by a *c.* 15–20 km thick Aegean mantle wedge. Beneath western Crete, a crust–mantle boundary in the

overriding Aegean could not be identified. A study of Rayleigh wave propagation across western Crete revealed typical crustal velocities down to the plate interface. Although a negative phase in the receiver functions could be explained by Moho topography in central Crete, this feature probably indicates the presence of low-velocity bodies in the lithosphere beneath western Crete. The lateral distribution of the negative receiver function phase points to considerable heterogeneity within the thickened crust of western Crete, with low-velocity bodies that elongate parallel to the forearc. The low velocities possibly reflect serpentinization of mantle material (e.g. Oleskevich *et al.* 1999). In the area of western Crete, the passive continental margin of Africa has entered the subduction zone (Meier *et al.* 2004a), and the plate contact dips at a lower angle than in eastern Crete (Meier *et al.* 2004b).

Tomographic studies (e.g. Papazachos & Nolet 1997) and the distribution of intermediate-depth seismicity point to a segmentation of the subducting slab in the Hellenic subduction zone, and to a corresponding segmentation of the overriding Aegean lithosphere. This east–west segmentation is indicated by (1) changes in the structure of the Hellenic trench system, (2) an offset in the southern border of the Aegean lithosphere, and (3) marked lateral heterogeneity within the Aegean lithosphere. The tectonics in the area of eastern Crete is governed by oblique subduction, causing left-lateral movement between forearc slivers, separated by nearly vertical microseismically active faults systems, which cut through the entire Aegean lithosphere.

To the south of Crete intense seismicity along the plate interface is observed in a zone up to 100 km wide. In contrast, little microseismicity is observed at the level of the plate interface beneath Crete, where the interface is located at a depth greater than *c.* 40 km. A change in the orientation of the stress field at *c.* 40 km, indicated by focal-plane solutions of intermediate-depth seismicity (Papazachos *et al.* 2000), points to seismic coupling between the plates in the shallower part of the subduction zone south of Crete, and decoupling of the plates at deeper levels. This decoupling is consistent with the concept of a subduction channel as shown in Figure 9.

The level of microseismicity is very low at a depth of *c.* 20–40 km beneath the island of Crete. The coastline of Crete follows roughly the boundaries of this seismically quiet realm within the Aegean lithosphere, where typical crustal velocities are observed down to the plate interface beneath western Crete. These observations point to a relation between the uplift of Crete and the physical state of the lithosphere in this depth range. It is tentatively proposed that the uplift of Crete, which commenced at about 4 Ma, is driven by return

flow in the upper portion of a subduction channel (Fig. 9) and by extrusion of material into the deeper crust.

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