Geodetic constraints on the tectonic evolution of the Aegean region and strain accumulation along the Hellenic subduction zone

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A B S T R A C T

We present evidence that GPS velocity estimates of plate motions and fault slip rates agree to within uncertainties with geologic estimates during the most recent phase of the geologic evolution of the E Mediterranean region (post-Late Miocene). On this basis, we use the GPS differential velocities to estimate the timing of initiation of the principal structures in NW Turkey, the N Aegean Sea, and central Greece, including, the Marmara Sea, the Gulfs of Evia (GoE) and Corinth (GoC), and the Kephallonia Transform fault (KTF). We interpret these ages to indicate that the North Anatolian fault propagated across the N Aegean, opening the GoE and GoC and initiating the KTF, during the past 1–4 Ma. We further suggest that Aegean extension that was earlier more distributed across the Aegean Basin became focused on this new fault system allowing the southern Aegean and Peloponnisos to translate SW with little internal deformation, as observed today with GPS. This change in tectonic configuration may account for the clear geologic evidence for crustal thinning throughout the S Aegean in apparent contradiction with low present-day strain rates. We further show that the low present-day strain rate along the southern edge of the Aegean micro-plate requires substantial aseismic slip along the plate interface below Crete, consistent with the low level of historic, subduction-type earthquakes along this segment of the subduction zone.

1. Introduction

Understanding the relationship between geodetically determined surface motions and longer-term motions estimated from geologic observations is a prerequisite to using highly precise, present-day geodetic motion estimates to constrain the Earth’s recent geologic evolution. Since the earliest geodetic studies of plate motions, earth scientists have demonstrated good agreement between recent geologic (−2–3 Ma) and geodetic (~10 yr) plate motion estimates (e.g., Smith et al., 1990; Larson et al., 1997; Sella et al., 2002; McClusky et al., 2003; Kreemer et al., 2003). However, in the more complex zones of deformation within plate interiors and along plate boundaries where geologic estimates of motion rates are more difficult to determine, the relationship between geodetic and geologic motions is less clear.

Geodetic observations in and around the Aegean Sea indicate that at present, the southern Aegean Sea and Peloponnisos Peninsula are translating towards the SW with little internal deformation (McClusky et al., 2000; Kahle et al., 2000; Reilinger et al., 2006). This present-day motion appears inconsistent with geophysical and geological evidence for widespread post-Pliocene extension throughout the region (e.g., Makris, 1978; Armijo et al., 1996). This apparent inconsistency is at odds with evidence for agreement, within uncertainties, between GPS and geologic estimates of plate motions and fault slip rates within the E Mediterranean region (McClusky et al., 2000, 2003; Allen et al., 2004; Reilinger et al., 2006).

Here, we present evidence for a tectonic scenario to reconcile this apparent contradiction. We show that the present tectonic structure of the Marmara Sea, the Gulfs of Evia and Corinth in central Greece, and the Kephallonia Transform fault would have developed in the past few million years at present-day, geodetic rates. We suggest that these structures result from the propagation of the NAF across the N Aegean during the past few Ma (Armijo et al., 1999; McClusky et al., 2000) and that this tectonic event concentrated Aegean extension in the N Aegean resulting in apparent contradiction. We show that the present tectonic structure of the Marmara Sea, the Gulfs of Evia and Corinth in central Greece, and the Kephallonia Transform fault would have developed in the past few million years at present-day, geodetic rates. We suggest that these structures result from the propagation of the NAF across the N Aegean during the past few Ma (Armijo et al., 1999; McClusky et al., 2000) and that this tectonic event concentrated Aegean extension in the N Aegean...
earthquakes along the Hellenic Arc (Jackson and McKenzie, 1988; Shaw et al., 2008).

1.1. Tectonics of the Aegean region

The present-day tectonic framework of the Aegean region is illustrated schematically in Fig. 1A, and the GPS-derived velocity field in Fig. 1B. The tectonic history of the region is complex involving a long history of north-dipping subduction, an early (Oligocene?) stage of continental collision (Meso-continental blocks), followed by recent (Miocene–Pliocene?) extension (see Jackson, 1994 for review). The tectonics are thought to be in large part a result of subduction-related processes, with continental collision being ascribed to the “rafting in” of continental fragments, and extension being related to slab roll-back (Le Pichon and Angelier, 1979;
Aegean tectonics are also thought to be intimately related to regional tectonic processes including the collision of Arabia with Eurasia, and presumed extrusion of the Anatolian plate (Sengor et al., 1984), although the relationship between these different tectonic events/processes is subject to debate (Gautier et al., 1999). We refer the reader to Dixon and Robertson (1984) for a more complete discussion of the broad geologic evolution of the Aegean region.

1.2. Geodetic and geologic estimates of active deformation in the Aegean

Fig. 2 shows the updated GPS velocity field for the Aegean and surrounding areas in an “Aegean-fixed” reference frame (tabulated in Auxiliary Table 1; velocity field modified from Reilinger et al., 2006). Data processing strategy and error estimation are as described in Reilinger et al. (2006). The Aegean reference frame is determined by minimizing, in a least squares sense, the relative motions between selected GPS sites within the Aegean region (selected sites are shown in red in Fig. 2). The low level of deformation in the southern and central Aegean and the Peloponnisos is well defined. This result was unexpected given the clear evidence for geologically recent deformation of the entire Aegean region (e.g., Jackson, 1994; Armijo et al., 1996; Reilinger et al., 1997).

Well-determined geodetic and geologic estimates are available for Arabia plate motion as well as for slip rates on major faults in the eastern Mediterranean region. Chu and Gordon (1998) report the geologic estimates of the Arabian plate motion based on magnetic anomalies in the Red Sea covering the past 3 Ma. The 3-Ma Euler vector (Arabia–Nubia) is consistent with the GPS Euler vector in location and rotation rate at the 1 standard deviation level (McClusky et al., 2003).

Fig. 3 shows a comparison of geodetic and geologic estimates of fault slip rates in the greater E Mediterranean region (modified from Reilinger et al., 2006). Geodetic slip rates are determined from the elastic block model described by Reilinger et al. (2006). This agreement provides the basis for using GPS motion estimates to interpret regional geologic evolution.
comparison indicates agreement within uncertainties for almost all structures, including the Gulf of Corinth (Armijo et al., 1996) that we examine in more detail below. Only the Main Recent fault (MRF) in Iran shows what appears to be a significant difference — this may be due to a real change in slip rate on the fault or insufficient geologic and/or geodetic observations. We conclude, in agreement with some prior studies (e.g., Allen et al., 2004), that relative plate motions and slip rates on major fault systems indicated by GPS in the Eastern Mediterranean region are equal within uncertainties (±10%) to the geologic rates averaged over the past few million years.

Fig. 4. A. Red lines showing our estimates of total right-lateral offset across the Kephelonia Transform fault (55–70 km) and oblique extension across the Marmara basin (i.e., in the orientation of GPS differential velocities across the basin), and our estimates of the ages of these structures (to nearest 0.5 Ma) based on present-day GPS velocities (insets). For the Marmara Sea, we also estimate basin age based on the GPS-derived NAF slip rate (25±2 mm/yr, Reilinger et al., 2006), and our estimate of the along strike length of the basin (115–150 km) assuming it is a pure, right-lateral, pull-apart basin. B. Red lines showing our estimates of extension across the western (two estimates are shown) and eastern Gulf of Corinth (10–15 km; ~20 km, respectively) and the Gulf of Evia (10–15 km) and the differential GPS velocities across these structures used to estimate the time these basins were initiated (insets). C. As B for the North Aegean Trough. Multiple widths are estimated to indicate uncertainty. Tick marks on width estimates indicate alternate interpretations for estimating basin width. D. Summary of the ages of major structures (to nearest 1 Ma) traversing the N Aegean and central Greece. We interpret these results to indicate that deformation associated with the NAF traversed the N Aegean and central Greece during the past 1–4 Ma. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
1.3. Propagation of the North Anatolian fault across the North Aegean and central Greece

Fig. 4A shows our estimate of right-lateral offset across the KTF (55–70 km) and the oblique opening of the Marmara Sea (80–90 km). GPS velocities for sites on opposite sides of the Kephelonia fault, and outside the zone of elastic strain accumulation indicate a current differential rate of $30 \pm 1$ mm/yr (GPS sites KRTS–XRIS, Fig. 2; see also Hollenstein et al., 2008). If the rate of motion has been constant, it would require 1.8–2.3 Ma to develop the total present-day offset across the Kephelonia Fault.

Similarly, the differential GPS velocity oblique to the Marmara Sea is $18 \pm 2$ mm/yr (Fig. 4A). This rate and the estimated offset projected to this orientation (80–90 km) suggest that the Marmara pull-apart basin developed during the past 4–5.6 Ma. Alternately, we estimate the age of the Marmara basin from the observed GPS slip rate on the NAF and the along-fault length of the basin, assuming the basin is a classic pull-apart on the NAF (e.g., Barka and Kadinsky-Cade, 1988; Flerit et al., 2003). We estimate a total length for the basin of 115–150 km (Fig. 4A; see also Le Pichon et al., 2003), and use the $25 \pm 2$ mm/yr NAF GPS slip rate (Reilinger et al., 2006) to deduce a total age for the Marmara basin of 4.3–6.5 Ma. This age for the Marmara basin, and presumably for the present configuration of the NAF in NW Turkey, compares well with the estimated age for the NAF from offsets on the western NAF (Armijo et al., 1999) and more regional estimates from offsets along the length of the fault (Westaway, 1994).

Fig. 4B shows our estimates of the extension across the Gulf of Corinth (GoC) and the Gulf of Evia (GoE) and the differential GPS velocities across these structures. For the W GoC we estimate a total extension of 10–15 km based on the present-day width of the rift, and the differential GPS rate across the fault of $10 \pm 1$ mm/yr (LIDO–KOUN), yielding an age of 0.9–1.7 Ma. This age for the GoC is consistent within uncertainties with the age estimated from more detailed geologic observations along this segment of the Gulf (Armijo et al., 1996). For the eastern GoC we estimate a total extension of ~20 km and a present-day differential velocity of $10 \pm 2$ mm/yr yielding an age of ~2 Ma. Similarly, for the GoE we estimate a total extension of 10–15 km, and a differential GPS velocity of $4.5 \pm 1$ mm/yr (Fig. 4B), yielding an age in the broad range of 1.8–4.3 Ma.

Total extension across the North Aegean Trough (NAT) (Fig. 4C) is more difficult to estimate. Basin geometry is considerably more complex than the other structures we consider. In addition, it seems likely that the North Aegean was subject to multiple episodes of extension since the late Oligocene (e.g., Dixon and Robertson, 1984). In any case, we use the broad range of offsets across the deepest basins shown in Fig. 4C to estimate a total age for the well-defined basins of the NAT of 1–6 Ma.

Uncertainties on our estimated ages are large due to geometric complexities of the basin bounding faults (Armijo et al., 1996), as well as a lack of direct constraints on the time required to establish the presumed new fault configuration (i.e., NAF transecting the NAT). Any time delay required for the system to develop will cause an underestimate in the time of initiation of the present fault geometry. Extension outboard of the main bounding faults will also cause a underestimate of the age of the basin. Converesely, less than 100% extension within the basin will cause an overestimate. We do not attempt to estimate these effects in our analysis — our expectation is that active extension outboard of the main basin bounding faults will tend to offset any overestimates of extension within the basins, at least to some extent. Furthermore, the general agreement ($\pm 10\%$) between geodetic and

Fig. 5. Schematic illustration of present-day Aegean tectonics showing coherent motion of the central Aegean and Peloponnisos (pink dotted area) with strain concentrated in the N Aegean and Gulfs of Corinth and Evia. The gray dotted area shows the extent of the Anatolian Plate, the blue dotted areas show deformation zones in the SE Aegean (trenchward motion relative to the Aegean micro-plate) and between the north and south branches of the North Anatolian fault system in the N Aegean (see Fig. 2). Blue arrows show the sense of motion across major tectonic structures, and red arrows, rates of micro-plate motion relative to Eurasia. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
Fig. 6. Perspective view of the model for the African–Aegean plate interface along the southern Hellenic Arc. The interface extends to the outer trench (black dots) and dips at an average angle of 20° below Crete [Meier et al., 2004]. The 12, 20, and 40 km depth contours of the plate interface are marked. The coupling of the plate interface (\(\phi\)) can vary gradually between 0 (free slipping) and 1 (fully locked). In the case represented here, the fully locked part of the zone of plate interaction lies below Crete (from 20–40 km depth) and is indicated by red (see profile in Fig. 7D). The N–S components of the GPS motion rates and 1-sigma uncertainties in the Aegean are projected onto the profile shown for comparison to the model results in Fig. 7 (dashed lines indicate width of the profile). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Fig. 7. N–S Aegean GPS velocity profile crossing Crete and theoretical deformation due to locking different segments of the plate interface. A. no-locking, B. locking between 0 and 20 km depth (outer blue segment in Fig. 6), C. locking of the interface between 0 and 40 km (outer blue segment and red segment in Fig. 6), D. locking between 20 and 40 km depth (red segment in Fig. 6), showing full (solid grey) and 20% locking (dashed line), E. topography and bathymetry along the profile, and F. earthquake hypocenters along the profile (Engdahl and Villaseñor, 2002). These results indicate little (<20%), if any, locking on the plate interface, and consequently a reduced probability of large inter-plate earthquakes on this segment of the subduction zone.

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geologic fault slip rates (Fig. 3) suggests that the time required to establish the new fault geometry is short in comparison to fault ages (i.e., \(-10%\) of the total age of the faults or \(-10^3\) yrs. in central Greece). We are encouraged by the consistency of the results from our simple analysis and those from more detailed structural studies of the Marmara basin and NAF (e.g., Westaway, 1994; Armijio et al., 1999), and the GoC (Armijo et al., 1996). In addition, we expect that errors in our estimates are roughly equivalent for the different structures we consider, resulting in more accurate relative than absolute ages that are more relevant for the present study.

Overall, the relationships between present-day, GPS motion estimates and basin structures and total fault offsets (Fig. 4D) are consistent with the deformation having developed sequentially from east to west during the past \(1\rightarrow4\) Ma.

1.4. A simple, conceptual model for the recent evolution of the Aegean region

McClusky et al. (2000) recognized the apparent discrepancy between the GPS evidence for coherent motion of the SW Aegean and Peloponnissos Peninsula and geological and geophysical evidence for recent extension. They proposed a model where the entire Aegean underwent extension during the initial period of African slab rollback; during this period the Aegean crust thinned and subsided. At some later time, the North Anatolian fault (NAF) developed, apparently in response to the extrusion of the Anatolian micro-plate (Sengor et al., 1985), and propagated across N Turkey and the Marmara region developing the Marmara pull-apart basin (see also Armijo et al., 1999). The NAF continued to propagate westward crossing the Northern Aegean Sea and Central Greece, developing the Gulf of Corinth and Kephelonia Transform fault and completely decoupling the southern and central Aegean and Peloponnissos from the Eurasian plate. At this time the south and central Aegean and Peloponnissos began translating SW with little internal stretching. This model, schematically illustrated in Fig. 5, allows initial extension in the Aegean that developed the current extended crust, and subsequent translation with little internal deformation, as observed today by GPS, thereby reconciling present-day and geologic evidence for recent deformation of the Aegean region.

1.5. Strain accumulation along the Hellenic Arc

In addition to the implications for the tectonic evolution of the Aegean region, the observed low GPS strain rates in the southern Aegean provide constraints on the nature of the interaction between the African and Aegean plates along this subduction zone plate boundary. The location of the plate interface has been estimated from seismic observations (Bohnhoff et al., 2001; Meier et al., 2004). There is a substantial, apparently aseismic, accretionary prism south of Crete as shown in Fig. 6. The main plate interface has been estimated to lie south of and beneath Crete, dipping north at an angle of about 20° (red section in Fig. 6). The elastic lithospheric block rotations and interseismic elastic strains associated with active subduction are modeled using DEFNODE (McCaffrey, 2002). Fig. 7 shows the resulting horizontal deformation rate along a N–S profile (see Fig. 6 for location of the profile) for locking of different sections of the plate interface. Fig. 7A and B show that the GPS measurements are insensitive to the interseismic elastic strain resulting from the locking of the shallower part of the plate interface south of Crete (0–20 km depth). However, it seems unlikely that the poorly consolidated sediments within the accretionary prism are capable of maintaining high stresses (Mascle and Chamillion, 1998). GPS measurements in the Aegean are mainly sensitive to the coupling of the plate interface located between 20 and 40 km depth that lies below Crete and corresponds to the area of highest seismic activity. However, the low strain rate along the leading edge of the Aegean plate precludes substantial (i.e., greater than \(-2\%\)) locking of the interface below Crete (see Fig. 7D). Based on this analysis, and the absence of significant historic earthquakes that can be associated with the plate interface, we concur with previous interpretations of historic seismic observations (Jackson and McKenzie, 1988) that plate convergence is occurring primarily aseismically, possibly due to the introduction of poorly consolidated sediments into the trench and/or the effects of slab rollback on the dynamics of the plate interface. This conclusion is consistent with the low level of historic seismic activity along the arc, in spite of the high rate of convergence and with a recent study indicating that the 365 AD Crete earthquake occurred off the plate interface (Shaw et al., 2008).

2. Conclusions

Available evidence indicates that present-day, geodetic deformation in the E Mediterranean is generally equal within uncertainties (\(\pm\)10%) to geologic deformation since initiation of the present plate configuration in the late Miocene/Pliocene. We use this relationship and young basin structures and fault offsets to estimate the age of initiation of the main structures in the NW Turkey, the N Aegean, and central Greece. We show that these structures, including the Marmara Sea pull-apart basin, the Gulfs of Corinth and Evia rift structures, and the right-lateral Kephelonia Fault developed roughly sequentially from east to west. We suggest that this sequential development resulted from the propagation of deformation associated with the NAF across the N Aegean during the past 1–4 Ma, and that this change in tectonic configuration resulted in the concentration of extensional deformation associated with Africa slab retreat in the N Aegean allowing the S Aegean and Peloponnissos to translate SW with little internal deformation. This scenario reconciles the apparent discrepancy between geologically determined present-day coherent motion of the Aegean and Peloponnissos with geologic evidence for post Late Miocene, distributed extension.

We further show that the low level of present-day strain within the S Aegean precludes substantial strain accumulation along the Aegean–Africa plate interface below Crete. This observation suggests a low likelihood of subduction-type earthquakes along this segment of the plate boundary, consistent with historic seismic observations and with the well-documented 365 AD Crete earthquake having occurred off the plate interface (Shaw et al., 2008).

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at doi:10.1016/j.tecto.2009.05.027.

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