Aegean crustal thickness inferred from gravity inversion. Geodynamical implications

Céline Tirel,*, Frédéric Gueydan, Christel Tiberi, Jean-Pierre Brun

Abstract

Since Oligo–Miocene times, the Aegean domain has undergone regional extension due to the southward retreat of the Hellenic subduction zone. Boundary conditions have been more recently modified by the westward extrusion of Anatolia. A new map of the Aegean crustal thickness inferred from gravity inversion is proposed here to better constrain the variations in space and time of crustal thinning that has accumulated since Oligo–Miocene times. Moho topography is obtained by inversion of satellite marine gravity data. Data are first corrected for terrain anomalies and deep mantle effects (African subducting slab). They are then filtered between 50 and 300 km to avoid short wavelength intracrustal effects. Results are consistent with previous 2D geophysical studies (seismic refraction, receiver functions) and show that an overall regional isostatic compensation of the crust holds for the Aegean area, with a mean crustal thickness of 25 km. Three different provinces (North Aegean, Cyclades and Cretan Sea) can be identified. Thinner crust is observed both in the North Aegean region (NE–SW trending of thinning, with crustal thickness lower than 24 km) and in the Cretan Sea (crustal thickness of 22–23 km). Between these two regions, the Cyclades are marked by a rather flat Moho at 25 km. A two-stage model of the Aegean extension could well explain the observed crustal thickness variation. From Oligocene to middle Miocene, gravitational collapse of the Hellenides, due to the southward retreat of the African slab, reduced the Aegean continental crust from 50 km (by reference to continental Greece and Anatolia) to a mean value of 25 km at the scale of the whole Aegean. From upper Miocene to present, the westward extrusion of Anatolia modified the extension and the associated crustal thinning in the North Aegean domain. During this second episode, crustal thinning related to the southward retreat of the African slab tends to localize in the Cretan Sea. The Cyclades likely behave as a rigid block translated toward the south–west without significant deformation, in agreement with the GPS velocity field and the lack of major earthquakes.

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1. Introduction

Following continental collision and crustal thickening, the Aegean domain has undergone two successive stages of extension since Oligocene times. From Oligocene to middle Miocene, extension was first marked by the development of core complexes in the Cyclades, Menderes and Rhodope, with a dominantly N–S direction of stretching [1–4]. During this period, extension likely corresponds to a gravity collapse of the previously thickened and thermally softened lithosphere, controlled by the southward retreat of the south Hellenic subduction zone [5–9]. Since the late Miocene, the effects of the westward displacement of Turkey were superimposed on the previous kinematic pattern [4,10–14]. Particularly, the Cyclades underwent considerable stretching during the first stage and became rather inactive during the second one [10,15–20]. Deformation tends to localize within a restricted number of active faulting areas on the edges of the Aegean domain (continental Greece and Western Turkey, North Aegean Through, Cretan Sea) and the volcanic arc. Active normal faulting resulting from N–S stretching is especially well represented in the Gulf of Corinth and Evia rifts [13,15,21–23]. In the Peloponnesus and Crete, the present active extension is parallel to the subduction arc [24]. This is in good agreement with locations of major earthquakes and the present day displacement pattern demonstrated by GPS measurements [10,19,20]. The present day kinematics and strain patterns depict the superposition of a dominantly dextral shearing along the North Anatolian Fault and the North Aegean Through (Fig. 1a) to the north, and a N–S stretching due to the subduction retreat to the south.

In the North Aegean, the location of Plio-Quaternary sedimentary basins (Fig. 1d, [25]) is relatively well correlated with regions that have undergone important thinning during the second stage of deformation. However, a complete map of these recent basins is not available in the whole Aegean domain.

In the present paper, we propose a new map of the Aegean crustal thickness obtained from inversion of marine gravimetric data. To assess that the gravity signal only reflects the crustal thickness, a series of corrections were applied: water load and terrain corrections, effects of the subducting African slab and bandpass filtering of short wavelengths related to crustal heterogeneities smaller than 50 km. The results are then compared to available reflexion and refraction seismic profiles and receiver functions [26–30] that provide local estimates of crustal thickness. The variations of crustal thickness are finally discussed in terms of a two-stage deformation history of the Aegean since Oligo–Miocene times.

2. Gravity inversion

2.1. Data processing

The complete Bouguer anomaly (CBA) of the Aegean area is used in this study to image the Moho variations. A complete Bouguer anomaly (CBA) is first compiled from the satellite derived free-air anomaly (FAA; Fig. 1b, [31,32]). The anomaly is then corrected from possible deep and crustal sources to only retain the crustal thickness information. Sandwell and Smith [31] and Smith and Sandwell [32] high-density satellite 2 min grid gravimetry and topography data set provide the free-air gravity anomaly (FAA) and bathymetry maps for the Aegean (Fig. 1b and a, respectively). The CBA (Fig. 2a) is then computed by removing the effects of the water load and terrain correction. These two corrections are computed with a 3D grid composed of elementary prisms of 2 min basal area and a thickness set to the bathymetry (Fig. 1a). The gravity signal of this 3D structure is computed with water and crustal densities of 1000 and 2670 kg m$^{-3}$, respectively. It is removed...
Fig. 2. (a) Complete Bouguer anomaly (CBA; isolines every 20 mGal). (b) Gravity anomaly due to the subducting African slab (isolines every 10 mGal). (c) Complete Bouguer anomaly without the subducted African slab effect (CBAS; isolines every 20 mGal), and (d) the same filtered with a bandpass 50–300 km (FCBAS; isolines every 10 mGal). This late filtered map is used to compute the Moho depth variation in Fig. 3a.
from the FAA to obtain the CBA (Fig. 2a). Note that a low value for crustal density is used to model the presence of sedimentary rocks near the water–crust interface.

The comparison between FAA and CBA (Figs. 1b and 2a) shows that the large positive signal observed in the south Aegean is amplified and more localized to reach a value of 160 mGal in the Cretan Sea. The increase of CBA (up to 80 mGal) in the North Aegean Trough emphasizes the strong influence of the water load in zones of high bathymetry.

The effect of deep low-frequency sources, particularly the African subducting slab in this region [33,34], must be taken into account. It is beyond the scope of this article to discuss in detail the shape of the subducting slab. We are particularly aware that this shape will control the gravity signature induced by the slab. However, the estimate of the slab gravity effect made by Tiberi et al. [34] presents the advantage of taking a non-ad hoc shape for the subducting lithosphere (unlike Tsokas and Hansen’s previous work [35]). Tiberi et al. [34] used tomographic data [36] and a linear relationship between P-wave velocity and density [37] to compute the slab effect. The modeled slab anomaly (Fig. 2b) is centred north of Crete with a maximum of 120 mGal and decreases radially to vanish in the North Aegean domain. The SW–NE decrease of the gravity signal is observed both in the computed slab anomaly and in the CBA (Fig. 2a,b). This good correlation first validates the proposed slab geometry used for the computation and also shows that the African slab is responsible for the major part of the CBA.

The complete Bouguer anomaly free from the African slab effect (CBAS) is shown in Fig. 2c. Compared to the CBA, the CBAS shows lower variations of the gravity signal in the Aegean Sea area. More specifically, the gravity signal in the Cretan Sea and the North Aegean Trough are now similar (+40 mGal; Fig. 2c), while part of the Cyclades is marked by a lower gravity anomaly. However, short wavelength variations of the CBAS are still present within the whole Aegean domain. These short wavelength variations are assumed to be related with intracrustal density variations, which have thus to be discarded to compute the Moho depth. To remove most of the intracrustal high-frequency signal, prior to the inversion, we filter the CBAS between 50 and 300 km with a bandpass taper through the Fourier domain using the Generic Mapping Tools (GMT) software [38]. A lower minimum value of the bandpass filter (30 km instead of 50 km for example) does not remove all the short wavelength variations. Setting the maximum of the bandpass filter to a larger value will decrease the amplitude of the CBAS signal. Indeed, a test with a bandpass filter of 50–1000 km shows a signal amplitude decrease of about 10 mGal and the same shape of anomaly signal than a 50–300 km filter. The choice of a bandpass filter between 50 and 300 km was thus found sufficient to remove intracrustal density variations and to preserve the amplitude of the CBAS signal. The filtered complete Bouguer anomaly without slab effect (FCBAS; Fig. 2d) only reflects the crustal thickness variations. The Cretan Sea and the North Aegean Trough correspond to two maxima (~40 mGal) in the FCBAS, suggesting a shallower Moho. Between these two regions, the FCBAS in the Cyclades reaches lower values. These observations are consistent with a thin crust in the Cretan Sea and the North Aegean Through and a thicker crust in the Cyclades, as previously mentioned [26–30,39].

2.2. Inversion procedure

The inversion used here is based on the direct formula of Parker [40]. It calculates the gravity signal \( \Delta g(x, y) \) of a layer having a density contrast \( \Delta \rho \) with its underlying semi-infinite space. The contact between the two domains is non-flat, and topography \( h(x, y) \) creates the gravity signal [41]. Oldenburg [42] described and solved the inverse problem within the frequency domain using the Fourier Transform. The topography of the contact between the two layers is obtained by iteratively solving the direct problem, assuming a constant density contrast \( \Delta \rho \). The following equation is then used:

\[
F(h(x,y)) = \frac{F(\Delta g(x,y))e^{jk|z_0}}{2\pi G \Delta \rho} = -\sum_{n=2}^{\infty} \frac{|k|^n \Gamma(n-1)}{n!} F(h^n(x,y))
\]

(1)
Fig. 3. (a) Moho depth from inversion of FCBAS (five iterations running, \( z_0 = 26 \) km, \( \Delta \rho = 0.4 \text{ g cm}^{-3} \); isolines every 0.5 km). (b) Crustal thickness from Moho depth and bathymetry filtered with a bandpass 50–300 km (isolines every 0.5 km) and (c) the residual (CBAS-computed gravity signal after gravity inversion, isolines every 10 mGal).
where $F$ represents the Fourier Transform, $G$ is the gravitational constant, $k$ is the wave number and $z_0$ is the reference depth from which the variations $h(x, y)$ are calculated. The absolute Moho depth is obtained using the following relation:

$$z(x, y) = z_0 + h(x, y)$$  \hspace{1cm} (2)

A value of 400 kg m$^{-3}$ is used for $\Delta \rho$ which reflects the mean density contrast between the crust (2800 kg m$^{-3}$, average density of the entire crust) and the mantle (3200 kg m$^{-3}$). The reference depth $z_0$ is calibrated with recent seismic data [26–30] and set to 26 km. It is worth noting here that the magnitude of the variations strongly depends both on the reference depth and the density contrast. Increasing $\Delta \rho$ or decreasing $z_0$ (with constant $z_0$ or $\Delta \rho$, respectively) both leads to a decrease in the magnitude of the variations.

The stability of the inversion is tested with respect to the variation of the parameters. The tests made on $z_0$ show that the magnitude of crustal thinning or thickening depth is about the same. When increasing the density contrast from 400 to 500 kg m$^{-3}$, the magnitude of the crustal thickness decreases about 0.5–1 km while the wavelength content remains remarkably stable.

For stabilization in the Fourier domain, the data are mirrored prior to the inversion and a low-pass filter is used (a cosine taper) to withdraw high-frequency anomalies arising from shallow crustal structures, if any. Five iterations were run, but in general, the convergence appeared after only two iterations, with a final root-mean-square (rms) of 2.2 mGal. The results are remarkably stable even when changing the cosine taper, which reflects a good wavelength coherence for the sources. A residual map is deduced from the inversion (Fig. 3c).

3. Aegean crustal thickness

3.1. Results

The Moho topography obtained by the inversion of the complete Bouguer anomaly corrected from the African slab and crustal density effects (FCBAS) is presented in Fig. 3a. The Aegean Moho appears quite flat for the whole region with variations of only +2 km (near Continental Greece and Anatolia) and −2 km (North Aegean Trough and Cretan Sea) around an average depth of about 25 km.

The generally good anticorrelation between the bathymetry (Fig. 1a) and the Moho shape strongly suggests an overall isostatic compensation of the Aegean topography. More specifically, regions of low bathymetry (less than 100 m), such as the Cyclades, are marked by a Moho depth of 25 km, whereas the Cretan Sea (2000 m depth bathymetry) is marked by a shallower Moho (22 km). However, a very simple calculation shows that there is no local (airy sense speaking) but regional isostasy in the present case.

The Aegean crustal thickness (Fig. 3b) is computed by subtracting a 50–300 km filtered bathymetry (computed from Fig. 1a) to the Moho depth (Fig. 3a). Similarly to the CBAS, the bathymetry was filtered to remove short wavelengths. Crustal thickness (Fig. 3b) seems to increase westwards from approximately 25 km (Turkish Coast) to about 27 km (Continental Greece), with a sharper gradient in the western edge of the Aegean Sea. Despite the poor constraint of marine gravity data near coastal areas, this smooth westward increase of the crustal thickness is in agreement with previous studies [7] and with the estimate of about 30 km for the continental crustal thickness near the Corinth Gulf obtained from inland gravity surveys [34]. Unfortunately, the lack of data precludes the same type of comparison for the Turkish side of the Aegean.

As previously seen from the filtered complete Bouguer map (FCBAS; Fig. 2d), the Cretan Sea and the North Aegean domains are marked by a thinner crust (22 and 23 km, respectively). In the North Aegean domain, the minimum crustal thickness defines an elongate zone trending in a NE–SW direction, and whose minimum is located beneath the North Aegean Trough (Figs. 1c and 3a). Mascle and Martin [25] already identified this orientation and
extended it through the Cyclades region. More recently, Goldsworthy et al. [43] proposed a similar trend from a study of fault systems, but that ends before the Cyclades, in Central Aegean along an Andros–Tinos line (see Fig. 1a).

In the Cyclades, between the two above quoted regions of minimum crustal thickness (North Aegean and Cretan Sea), the Moho has a rather uniform depth of c.a. 25 km. Note that the minimum SW–NE horizontal dimension of the Cyclades area is at least 100–150 km (Fig. 1a), which is much larger than the lower value of the bandpass filter used prior to the inversion. This rules out a flat Moho coming from any artefact of filtering.

In terms of crustal thickness, the Aegean Sea can thus be divided into three main regions: (1) the Cyclades with a flat Moho at 25 km depth, (2) the Cretan Sea and (3) the North Aegean domains, both with a thinner crust.

It is worth noting here that the variations of the Moho depth remain very similar and stable through all the tests performed for each modelling parameters ($\varepsilon_0$, $\Delta \rho$, cosine taper...). Thus, the relative variations of the crustal thickness described above cannot be numerical artefacts of the inversion. However, as previously mentioned, the absolute value of the Moho depth is dependent on the choice of the inversion parameters, and this trade-off induces nonuniqueness. For example, setting $\varepsilon_0$ to 30 km (instead of 26 km) yields to Moho depth variations from 27 to 33 km (instead of 23–28 km). At this stage, previous geophysical studies will help to solve this uncertainty.

### 3.2. Comparison with previous geophysical studies

The comparison of the above results with previous geophysical studies shows very similar overall relative variations of Moho depth (Table 1). In particular, in the Cyclades region, Makris and Vees [26], Vigner [29] and Li et al. [30] describe a regularly flat Moho at a depth of 25–26 km, in agreement with our calculations. However, some discrepancies of Moho depth estimates must be noted, particularly in the Cretan Sea, where our estimate of 22 km is larger than values of 15–20 km proposed by Makris and Vees [26], Makris [27] and Bohnhoff et al. [28] (Table 1). This apparent inconsistency could be explained by the presence of sedimentary basins, which were not taken into account for in our inversion.

Two main differences with the previous gravity study of Tsokas and Hansen [35] must be pointed out. The first difference occurs along the Evvia profile (Fig. 1a). Makris and Vees [26] show an increasing Moho depth along this profile from 32 km depth at Mandouthi, in North Evia, to 26 km at Amorgos (East Cyclades). This is confirmed by Vigner [29] and our study. However, the same profile using Tsokas and Hansen’s data [35] shows a strong gradient of down-dipping Moho between Andros and Amorgos. The second discrepancy concerns the Cretan Sea, where an elongate zone of shallower Moho (15 to 20 km) runs parallel to the north of Crete. This tendency is exemplified in most studies, including ours. However, in Tsokas and Hansen [35], the Moho depth variations display a rather different pattern with no elongate minimum of crustal thickness. These differences should come either from their modelling method (multiple-source Werner deconvolution) or, more probably, from the fact that they do not consider the African slab gravity effect. Based on tomographic imaging of the African slab [36], and following Tiberi et al. [34], we have estimated more precisely and fairly the shape and thus the gravity effect of the subducting lithosphere. This has a strong effect on the complete Bouguer anomaly (CBAS; Fig. 2c), as shown for

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**Table 1**

Comparison between the estimates of Moho depth of the present study and previous 2D local geophysical studies

<table>
<thead>
<tr>
<th>Authors</th>
<th>Method</th>
<th>Crete</th>
<th>Cretan Sea minimum</th>
<th>Cyclades</th>
<th>North Aegean</th>
</tr>
</thead>
<tbody>
<tr>
<td>Makris and Vees [26,27]</td>
<td>Refraction</td>
<td>30–32</td>
<td>20</td>
<td>26</td>
<td>–</td>
</tr>
<tr>
<td>Bohnhoff et al. [28]</td>
<td>Reflection refraction</td>
<td>24–32.5</td>
<td>15</td>
<td>–</td>
<td>–</td>
</tr>
<tr>
<td>This study</td>
<td>Gravity inversion</td>
<td>28–31</td>
<td>23</td>
<td>25</td>
<td>24–26</td>
</tr>
</tbody>
</table>
example by the positive free-air gravity anomaly centred in South Aegean (Fig. 1b) that disappears after removing the African slab effect (Fig. 2a).

3.3. Role of sedimentary basins

The gravity inversion tool used in this study to compute Moho depth and crustal thickness did not take into account sedimentary basins. The negative gravity signal due to sedimentary basins (negative intracrustal density variation) has thus been disregarded, leading to an overestimate of the Moho depth and crustal thickness in regions where basins are present. For a maximum estimation, we calculate that a sedimentary layer at the surface with a thickness of 3 km yields a negative anomaly of 40 mGal. This could be misinterpreted as a crustal thickening if one is not aware of the presence of sediments in the area.

The presence of low density sediments in the Cretan Sea can in particular explain our overestimate of the Moho depth (22 km) compared to that proposed by Makris and Vees [26], Makris [27] and Bohnhoff et al. [28] (15–20 km; Table 1). Sedimentary thickness in the Cretan Sea might reach values close to 3 km [26], leading to shallower Moho depths than those mapped in Fig. 3a. Similarly, in the North Aegean, sediment thicknesses of about 5–6 km in the Orfanos Gulf and in the Sporades basin have been measured [29,39]. We estimate the Moho depth to be 22 km after a basin correction, instead of the 24 km previously estimated. Consequently, we expect the North Aegean region to be marked by a more pronounced NE–SW trending of crustal thinning because of the numerous sedimentary basins reported in the region (Fig. 1d).

Because a complete map of the basement depth is not available for the whole Aegean Sea, we can only give some insights on the change of Moho depth induced by the presence of large basins in regions where the basement depths is well documented as in the North Aegean Trough.

4. Geodynamical implication

As extension must lead to significant crustal thinning, the above results bring information on the stretching accumulated since the beginning of extension and its regional variations. The Aegean domain has undergone two successive stages of extension since Oligocene times. The first stage of Oligo–Miocene age is related to the southward migration of the African slab [5–9], and the second stage (Pli–Pleistocene) is related to the combined effects of the still active migration of the African slab and of the westward extrusion of Anatolia [4,10–14]. In this work, we identify in the Aegean Sea three regions (Fig. 4) that suffered either only the first stage of extension (the Cyclades) or the two successive stages (the North Aegean and the Cretan Sea). Because we cannot have access to the crustal thickness prior to extension, we assume, following McKenzie [44] and Gautier et al. [4], the amount of crustal thickness prior to the extension to be c.a. 50 km by reference to continental Greece and Anatolia [26,45].

The Aegean Moho appears rather flat for the whole region, with variations of only +2 km (near Continental Greece and Anatolia) and −2 km (North Anatolian Trough and Cretan Sea), around an average depth of about 25 km (Fig. 3a). Such a regional-scale flat Moho is commonly observed in domains of wide rifting like the Basin and Range, in particular, beneath core complexes domains [46,47]. This requires a lower crust viscosity, low enough to flow rapidly, allowing the surface and the Moho to remain relatively flat during dome rise and continuing extension. Crustal thickening during Cretaceous and Eocene times created high thermal conditions (Moho temperature higher than 700 °C) suitable for such a mode of extension in the Aegean lithosphere [48,49]. Gravitational collapse has likely been triggered by the southward retreat of the Hellenic subduction zone [2,4], as represented by $F_{sr}$ in Fig. 4. Following the above arguments, it is therefore likely that after the first stage of Aegean extension, characterized by the development of metamorphic core complexes, the Moho should have had a rather flat geometry at a mean depth of around 25 km at the scale of the whole Aegean, from the Rhodope to Crete. Crustal thinning during this first stage of extension is thus approximately 100% (reduction of crustal thickness from ~50 km to 25 km), an estimate consistent with previous studies [4,44].

Variations of Moho depths around the mean value of 25 km obtained from gravity modelling (Fig. 3a) should therefore represent variations in crustal thinning related to more recent extension. The westward
extrusion of Anatolia, which started around 5 Ma ago [4,10–14], modified the kinematics of extension in the North Aegean. The North Aegean domain (NW–SE trending zone of thinning) and the Cretan Sea show crustal thicknesses smaller than 25 km. These two regions are also marked by Plio-Quaternary basins (Fig. 1d), such as the North Aegean Trough, the Edremit Trough, the Lesbos–Psara Trough [25,29,39] for the North Aegean domain and the Mirthes Basin and the Iraklion Central basin in the South Aegean [25]. Each of these basins shows evidences of post-Messian deformation [29], whereas no major Plio-Quaternary basins are present in the Cyclades. This second stage of deformation induces an additional thinning of around 10% (reduction of crustal thickness from 25–26 to 22 km) in restricted regions (the North Aegean and the Cretan Sea). Crustal thinning during this second stage of deformation is mostly governed by the extrusion of Anatolia in the North Aegean domain \( (F_{\text{ext}}) \) and by the still active slab retreat in the Cretan Sea \( (F_{sr}) \) [4,9,10,24, 50,51]. This interpretation is confirmed by the follow-

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Fig. 4. Schematic map showing the three main regions of the Aegean domain inferred by our gravity inversion and marked by different crustal thickness: (1) The Cyclades with a flat Moho at 25 km, (2) the Cretan Sea and (3) the North Aegean domain, both with shallower Moho. Since Oligo–Miocene, the Hellenic slab retreat triggers a gravitational collapse, which led to a ~25 km thinned Aegean crust. More recently, since 5 Ma, the extrusion of Anatolia \( (F_{\text{ext}}) \) changes the pattern of deformation in the North Aegean, giving NE–SW trending narrow zones of crustal thinning. The strongly thinned crust of Cretan Sea is mostly due to the slab retreat \( (F_{sr}) \). Thick lines correspond to the volcanic arc and the southwestern limit of the NE–SW trending narrow zones of thinning.
ing geophysical features in the three main regions. In the North Aegean domain, three major active strike-slip zones (North Aegean Trough, Edremit Trough and Lesbos–Psara Trough) have been recognized by a strongly localized seismicity [15,18] (Fig. 1c). The observed NE–SW trend of thinning (Fig. 3a) is also well marked by these active strike-slip zones, indicating a significant extensional component along these fault zones. The present day extensional strain axes calculated by Kalhe et al.[19] are indeed oblique to the strike-slip fault trend, in agreement with the elongate zones of thinning observed below the strike slip fault systems (Fig. 1c). Note that this obliquity of stretching axes indicates that the strain is not only controlled by the NE–SW strike-slip faults but also by submeridian stretching related to the southward retreat of the subduction zone. In the Cretan Sea, no significant seismicity is recorded, which apparently contradicts the observed strong and recent crustal thinning. However, strain rate deduced from GPS velocity [19,20] reveals a significant stretching rate within the Cretan Sea and, more specifically, near the Peloponnesus and Rhodes, which is consistent with the presence of normal faults in this region (Fig. 1c). However, more recent strain estimates [52] do not seem to indicate major deformation in the Cretan Sea. This discrepancy between strain rate estimates of these studies in the Cretan Sea has to be related with a poor constrain on GPS measurement (less than three points of measurement). The only evidence of post-Messinian deformation in the Cretan Sea is therefore given by the presence of Plio-Quaternary basins and normal faults. Moreover, Plio-Quaternary normal faults in the Cretan Sea indicate a recent extensional deformation despite the lack of seismicity. Opposite to the Cretan Sea and North Aegean Trough situation, strain rates and GPS velocities in the Cyclades are low [10,19,52] and seismicity is scarce and scattered [18]. Located between the two recently thinned regions of North Aegean and Cretan Sea, the Cyclades domain is thus likely translated as a rigid block towards the South, as already suggested [4,10,53,54]. As quoted by Walcott and White [54], the Cyclades block seems to be isolated from the rest of the Aegean domain when the Anatolian extrusion started (late Miocene–early Pliocene). The Cyclades do not undergo major extensional deformation since Oligo–Miocene times. The reasons for the absence of recent deformation in the Cyclades are however still a matter of debate [53,55,56] that is beyond the scope of the present paper.

In summary, a two-stage model for the Aegean extension could well explain the observed crustal thickness variation within the whole Aegean region. First, during Oligo–Miocene, the southward migration of the South Hellenic subduction zone triggers the gravitational collapse of a previously thickened crust, leading to an overall crustal stretching of the whole Aegean domain by a factor of two [4,44] and to a flat Moho geometry at regional scale. Second, in addition to the still active southward migration of the Hellenic subduction zone, the Anatolian westward extrusion has recently changed the pattern of extensional deformation in the North Aegean domain. This second phase of extension, which probably began about 5 Ma ago [4,10–14], is responsible for ~10% additional thinning mostly located in South and North Aegean. The Cretan Sea thinning is mainly controlled by the back-arc extension, while the North Aegean extension is due to the combined effects of the extrusion of Anatolia and back-arc extension. Between these two regions, the Cyclades likely behave as a rigid plateau.

5. Conclusions

A simple gravity inversion was used in this paper to compute a new map of the Aegean crustal thickness. The Aegean crust thickness is homogeneous and relatively thin within the whole region, with variations of only +2 km (near Continental Greece and Anatolia) and −2 km (North Anatolian Trough and Cretan Sea) around an average depth of about 25 km. Our results are consistent with local 2D geophysical studies and show a regional isostatic compensation.

In this study, we emphasize the potential of gravity analysis in the understanding of the extensional processes in the Aegean region. An important result is the identification of three domains of different crustal thickness at a regional scale (Cyclades, North Aegean Trough, Cretan Sea), each of them being related to the two-stage evolution of Aegean extension. The first stage took place in Oligo–Miocene times, when gravitational collapse of a previously thickened Aegean crust led to a crustal stretch of about 100%. During the second stage, the westward
extrusion of Anatolia modified the kinematics of extension and led to more localized deformation within specific areas. The North Aegean Trough and the Cretan Sea show maximum thinning with a crustal thickness of 24 and 22 km, respectively. In the North Aegean Trough, thinning likely results from the combined effect of the extrusion of Anatolia and back-arc extension, whereas the Cretan Sea is mainly controlled by the back-arc extension. The Cyclades domain has an average crustal thickness of 25 km and seems to have not accommodated any additional extension since the late Miocene times.

Our study takes only into account the long wavelength part of the gravity signal. It is worth noting that the residuals of the inversion show a NW–SE trend in the Cyclades (Fig. 3c). These short wavelength patterns could be due to remaining crustal density contrasts related to tectonic and/or geologic features. A 3D inversion of this gravity component, together with a study of the sources depth, could be thus further considered to investigate the reason for the rigid block-type behaviour of the Cyclades during post-Miocene times.

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