Moho, crustal architecture and deep deformation under the North Marmara Trough, from the SEISMARMARA Leg 1 offshore–onshore reflection–refraction survey

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Abstract
Understanding further the nature and evolution at lithospheric scale of the Sea of Marmara on the North Anatolian Fault needs constraints on the deep crustal and Moho spatial variation. This has been probed here with offshore–onshore and OBS, Ocean Bottom Seismometer refraction seismics, in addition to coincident MCS, marine multichannel reflection seismic profiles over the whole North Marmara Trough. The diverse strikes of MCS profiles in a dense grid allow to avoid misinterpretation of late echoes in the deep basin as Moho reflections and attribute them to sidesweeps. Moho is instead positively identified from reversed observations of first-arrival head and refracted waves at the top of the mantle obtained at large offset by land stations. A significant and sharp reduction in its depth, on the order of 5 km occurs beneath both the eastern and western rims of the North Marmara Trough, with a more progressive crustal thinning from the south. The wide-angle reflections on OBS and land stations document in addition to Moho the top of a lower crustal reflective layer, which is also sampled by MCS profiles, and appears to follow Moho topography. The dense grid of MCS profiles along the southwestern margin of the North Marmara Trough reveals a dipping reflector through the upper crust with tilted basement blocks on top. This low-angle fault is suggested as a normal sense detachment extending in depth towards the reflective lower crust. The upwarp of the Moho and lower crustal layer towards the North Marmara Trough suggests that crustal thinning occurs mostly in the upper crustal part, with lateral transport of the material towards WSW in the footwall of the detachment, and possibly other features to the south, in the motion of Anatolia with respect to stable Eurasia oblique to the North Marmara Trough. Thinning can be accommodated in an asymmetric partitioning of the displacement on several branching faults at lithospheric scale.

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1. Introduction

1.1. Study area and methods

The North Anatolian Fault (NAF) is a continental transform boundary, which cuts across Turkey over more than 1500 km. It is accommodating about 25 mm/yr (McClusky et al., 2000) of right-lateral motion between the Anatolian and the Eurasian Plates (Fig. 1a inset). The surface expression of the North Anatolian Fault is a relatively narrow and simple fault zone over its entire morphological trace except in its western part as it becomes more complex before entering the Marmara Sea, where it is partitioned into several branches.

The Sea of Marmara has been interpreted as a large-scale transtensive region above an extensional jog on a right-lateral step-over (Armijo et al., 1999) or a releasing bend (Flerit et al., 2006) between the two North Anatolian strike-slip fault segments marked by recent large strike-slip earthquakes, the Izmit segment in the east and the Ganos one in the west. At a larger scale, it connects to the west with the Aegean region of extensional deformation.

Offshore and onshore scientific investigations have been recently conducted in the Marmara Sea region, including geological and GPS studies (Reilinger et al., 1997; Straub et al., 1997; McClusky et al., 2000; Flerit et al., 2003), high-resolution bathymetry (Armijo et al., 2002,
coring and high-resolution seismic profiles (Ergun and Ozel, 1995; Smith et al., 1995; Wong et al., 1995; Aksu et al., 1999; Okay et al., 1999, 2000; Imren et al., 2001; Le Pichon et al., 2001; Parke et al., 2002; Demirbag et al., 2003).

The French–Turkish seismic survey, “SEISMARMARA-Leg1” was carried out after the 1999 earthquakes of Izmit and Düzce from July to October 2001, as a multi-method approach to investigate the seismic structure and activity of the northern Sea of Marmara, the North Marmara Trough, NMT (Hirn et al., 2002, 2003; Bécel et al., 2004) (Fig. 1). The aims of the programme were to shed light on the regional tectonics and recent evolution at crustal scale. The crustal-scale architecture of the NMT is revealed by its dense grid of Multi-Channel Marine Seismic (MCS) profiles that have an unprecedented depth of penetration. Before this survey, MCS data had been collected but only with a short recording streamer and a modest strength of the source, limiting the penetration to the sea bottom multiple.

Selected MCS profiles outlining the general architecture and lateral heterogeneity in the North Marmara Trough have been presented in Laigle et al. (2008). They revealed the supra-crustal structure of the deep Cinarcik and Central Basins as well as elements of the intra-crustal and deep structure on the southern shelf of the trough.

The present paper has a specific focus on the deep structure, with the different and additional dataset of wide-angle reflection and refraction (WARR) on both OBS and land stations. The WARR modeling reveals the deep structure under the North Marmara Trough itself,

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**Fig. 1.** a) The North Anatolian Fault (NAF) in the Marmara Sea region. Multi-beam bathymetric image of the northern Sea of Marmara collected in 2000 with R/V Le Suroît with observed and interpreted faults superimposed from Armijo et al. (2002). Elements of the SEISMARMARA-Leg1 survey: Black lines indicate the MCS profiles acquired with the 8100 cu. in. source, dashed black lines indicate the MCS profiles acquired with the 2900 cu. in. source. Black circles show the land station locations and the black squares show those of the ocean bottom seismometers (OBS). Dashed grey line indicates the Intra-Pontide suture zone and the part in light-grey the limits of the Thrace Basin (after Görür and Okay, 1996). Box indicates area enlarged in b. Inset shows tectonic framework (faults from Armijo et al., 2002). b) Location map showing the interconnected profiles in the southwestern part of the North Marmara Trough with identification of profile discussed in black dash-lines.
which can not be approached by the MCS data. It thus allows us to extend the discussion on crustal thinning under the central NMT and quantify the crustal deformation in space. We illustrate on several MCS profiles diverse aspects of the intra-crustal structure, also documenting further those suggested in Laigle et al. (2008), which give us a clue to the mechanism of deformation. We also deal here with the problem of reflection response of the Moho under the deepest marine basins with an evaluation and interpretation of a tight grid of MCS profiles.

1.2. Seismic data

1.2.1. Data acquisition

The SEISMARMARA-Leg1 program consisted of marine multi-channel reflection seismic (MCS), ocean bottom seismometers (OBS) and land stations recording of wide-angle reflection–reflection from the same source. It was the first time that marine wide angle–reflection reflection seismic data were recorded by OBS in the Sea of Marmara, thus allowing coincident reflection and reflection recording and joint modeling.

The French N/O Le Nadir acquired about 2000 km of MCS profiles (Fig. 1) in the northern Sea of Marmara during the first leg. The MCS profiles have an unprecedented depth penetration due to the 4.5 km length of the 360 channel digital streamer and also to the strength of the sources of 8100 or 2900 cu. in. (Laigle et al., 2008) provided by a 12-airgun array in single-bubble mode (Avedik et al., 1995, 1996).

Navigation safety was provided by a vessel of the Turkish Coast Guards (Sahil Güvenlik). At sea-bottom, 37 OBS with 3-component sensors and continuous recording over 1 to 2-month, were deployed and collected by the Turkish ship MTA Sismik-1. On land, a similar number of temporary stations, with 3-component 2 Hz sensors were installed.

During this Leg 1, 4 E–W lines and 30 N–S lines crossing the whole Marmara Trough have been acquired. The survey was designed to have crossing profiles in diverse azimuths (Fig. 1) with OBS at nodes and land stations in the continuity of profiles. This geometry has been designed in order to estimate the dip of reflectors. This proved essential to discriminate between primary reflections and signal-generated noise such as multiples and also point- or line-diffractions in the section plane from beneath or broadside. This marine survey took place in a narrow, deep bathymetric trough, with complex shallow structures, a rather adverse response of the Moho under the crustal thinning at its base, have already shown for MCS profiles of SEISMARMARA Leg 1 that are located on the margins of the deep sedimentary basins (Hirn et al., 2002, 2003; Bécel et al., 2004; Laigle et al., 2008). Concerning the deep sedimentary basins themselves, we could observe a flat seismic event with a strong energy at about 9 s right on the on-board processed brute MCS stack section of the first profile shot. This was the regional E–W MCS profile SM-1, also labeled SM-23 in the location map of Fig. 2. This line has been shot successively with the 8100 cu.in. source and 15-fold coverage as profile SM-1, and as profile SM-23 with the 2900 cu. in. source which has a smaller shot interval distance and a higher coverage-fold of 45.

Profile SM-1/23 is a strike-line with respect to the known structures, which form the North Marmara Trough, i.e. parallel to the NMT margin, and is thus prone to record side-swipes. The seismic acquisition system, with receiver groups along the streamer, as well at the stacking processing is designed to enhance vertical reflection from horizontal interfaces at depth. Indeed what is enhanced is the signal coming in orthogonally to the streamer, hence as well the signal returned by objects located broadside to the streamer direction. Thus we did not interpret this event as a deep reflector like the Moho. This late energy on the western part of profile SM-1/23 of our Leg 1 survey has been however proposed as corresponding to the Moho, crust–mantle boundary (Carton, 2004).

This is contradicted by the crossing profiles as discussed in detail in the following sections which establish by modelling that the energy corresponds to echoes from out of the plane of the vertical section of the E–W profile. In short, this is because on those crossing profiles which are N–S transects of the NMT, the corresponding wave “move-out” on shot-gathers, that is the travel-time difference of the wave between the two ends of the streamer, is very much larger than the NMO, normal move-out for a Moho reflection that would be less than a tenth of a second. This indicates that the corresponding raypath difference is about the length of the streamer and also that velocity is low, documenting thus that the bouncing point from which the wave is returned is shallow and aligned with the streamer. This being true for several N–S transects, indicates a line-diffractor along the southern edge of the NMT as the object giving rise to this late amplitude, and not the Moho. As shown later, the Moho under the basins of the NMT can only be identified by the WARR P-waves arrivals recorded far away onshore, since its response remains hidden behind side-echoes in MCS profiles in the basins.

2.2. Observational evidence provided by crossing profiles

On this regional E–W profile SM 1/23 in the basins rather flat seismic energy at about 9–10 s is observed also in its higher-fold stack section profile SM-23 (Fig. 2b and c). This flat late energy on this E–W
The north-dipping event seen on the crossing profile SM-2 (Fig. 2a), which is striking N22°W and is nearly a dip-line with respect to the NMT. Hence, this event recorded on profiles SM-1/23 as being flat does not correspond to Moho reflections at the vertical beneath this E–W profile, but comes from out of the plane of section. The N–S profile SM-2 shows that the north-dipping event already exists as early as 3 s beneath the southern margin of the basin. Whatever causes this echo, diffraction or dipping reflector, its source is thus at shallow depth or even outcropping near this southern basin margin, and thus cannot be the Moho.

2.3. Modeling the response of a 2D multi-layered model to dip-line stacks

In order to discuss the origin of the steepest event on the N–S profile SM-2 that exhibits a strong energy down to about 10 s at the cross-point with profile SM-1/23 (Fig. 2b) we have computed the zero-offset response of 2D models (Sartoretti, 2003). This 2D model (Fig. 2d) is based on the interpretation of the brute stack (Fig. 2a), and shows northeastwards dipping sedimentary reflectors down to 4 s, and the “basement” reflector (labeled B) corresponding to the bottom of the stratified sedimentary sequence. Other dipping events beneath the top of the basement can also be seen. Some of them are easily identifiable: this is the case of the event (D), which we will discuss later in Section 4 and which has been interpreted as a detachment. Deeper, steeper events include “X”, the one seen flat at a late time on the strike-line, which origin we want to constrain. For designing the model, the main reflectors have been picked and converted to depth using realistic layer velocities above.

We have then tested two different hypotheses for the origin of this deeper event “X”, which are either a diffracting point, or a steep reflecting interface, which geometry can be tested and constrained. In the case of the steep reflector, we find that it has to be so steep that its echo, although spreading widely through the seismic section of profile...

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**Fig. 2.** a) MCS profile SM-2 with the calculated reflection travel-times in the model in (d) superimposed. b) MCS profile SM-23 — strike line with respect to the NMT. B: Basement, D: detachment, X: steep seismic event. c) Location map of the EW regional profile SM-1/23 (black line), the NE-SW profile SM-2 (black line) and the profile N–S SM-44 (dashed black line). d) Velocity depth model from the 2 dimensional reflection travel-times modeling: Case of a steep reflector. e) MCS profile SM 44 (f) MCS profile SM-23. Vertical exaggeration of 3 with respect to the sea bottom.

**Fig. 3.** a) Location map of the EW regional OBS profile SM-1/23 — black squares onshore indicate the land station, black squares offshore the ocean bottom seismometers. b) Final P-waves velocity depth model from wide angle reflection–refraction data (WARR) with color scale at the bottom. Black arrows indicate the coincident multichannel seismic profile SM-1/23. Vertical exaggeration of 2.
SM-2 (Fig. 2d) only documents it as a short segment from 2 to 7 km depth near the southern rim of the marine basin. Calculated reflection times through the model are superimposed to the stack section of the profile SM-2. For this steep reflector, a 70° dip is found in the plane of section, which is a lower bound to the true dip. The latter could be even steeper if the profile SM-2 were not exactly a dip-line as it is likely. The top of this steep reflector would be located around the southern rim of the NMT, NE of Marmara Island. We explored the other type of models, of diffracting points rather than a reflector and found an alternative one, where the observed steepest energy can also be associated to diffraction points, localized at about 2 km depth under the southern rim of the trough.

It thus appears from the existence of dipping events on the two N–S transects SM-2 and SM-44 and the flat attitude of the event on the E–W profile that there is an alignment of sources of reflection or diffraction along the southern edge of the NMT. Both hypotheses for the source, as being a line-diffractor or a steep shallow segment of a reflector along the southern rim of the NMT have similar geological meanings. The line-diffractor can be regarded as the linear edge of an interface or layer that is interrupted by a fault and the steep reflector as the outcropping or buried fault-scarp itself. The later dipping event “X” on the dip-line, can thus be modeled to have a shallow few kilometers depth origin, at 15 km south of the strike-line: diffractions from a buried shallow edge of the NMT or reflections from edge fault itself, that they define at most from 2 to 7 km depth. In the following, this dipping event “X” will be labeled “LE”, for lateral echo, on the strike-lines.

The imaging of the Moho boundary under narrow marine basins by reflection seismic appears thus to be plagued by echoes of lateral basin-bounding features. This inability to image deep is not due to the source strength and penetration. With a stronger source the energy of the lateral echoes will be stronger too. On strike-lines lateral echoes are recorded with an enhanced energy that may lead to their misinterpretation. In any case a significant number of crossing profiles with different azimuth is required to avoid pitfalls.

3. Large offset refraction seismic identification of Moho

3.1. The Moho under the NMT: evidence and constraints from Pn refractions and wide-angle reflections (line SM-1/23)

In order to constrain the Moho geometry and to avoid the difficulty in MCS imaging described in the previous section, wide-angle reflection and refraction recording by land stations had been incorporated in the survey planning. In this special geometry the Moho may be sampled by undershooting the upper crustal complexity.

3.1.1. Approach: designing of the starting velocity–depth model

A 2D interface and velocity model has been developed in several steps for the main regional line SM-1/23 striking East–West along the NMT (Fig. 3). This line is 150 km long and has 13 OBS. The OBS data under the NMT sample mainly the basement by refracted waves. With these basement first arrivals we can easily constrain its velocity and deep geometry beneath the profile if we have a reasonable model for the sedimentary supra-crustal structure. This can be obtained here from the coincident multi-channel seismic profiles, which have the finer resolution for the strong variations in velocities expected here. The interface depths and layer velocities for the sedimentary part in the refraction modeling have been derived from the interpretation of the velocity analysis of the multi-channel data done for stacking the CMP gathers data of line SM-1/23 (45 fold-coverage).

This has been critical for the OBS which are at large depth in the marine basin, where the sea bottom is down to 1300 m depth. Indeed, the shots being near the sea surface, refracted waves sampling sediment velocities under deep basins cannot be observed for long distances, or not at all, as first arrivals before the direct water-wave. Even with a closer spacing than the 10 km between OBS here, a detailed velocity depth model of the sediments could not have been derived since the corresponding first-arrival refractions are covered by the direct water wave because of the large water depth. The MCS data provided a 2D
Fig. 5. a) For the land stations ELM (Elma), PAL (Pala) located west of the Sea of Marmara and KOR (Koru) located in the east, raypaths of the Pn waves through the final structural depth model of Fig. 3b extended beneath both sides according to onshore recordings. b) Seismic section from the land stations Elma, Pala and Koru (top), same seismic sections with overlain Pn calculated travel times (bottom).
velocity model down to the basement which topography has been tuned to fit with the wide-angle reflection arrivals observed on the OBS sections. All the calculated travel times have been computed with the Rayinvr code of Zelt and Smith (1992). Clear arrivals from the lower crust and Moho interfaces are unambiguously identified on 2 OBS only from among near surface multiples that are generally dominant, and will be discussed later.

3.1.2. Evidence for a crystalline basement constrained by OBS refraction seismics beneath the pre-kinematic basement imaged by MCS data

Modeling the OBS first arrivals recorded from 5 km to about 80 km establish that there is a crystalline basement of velocity close to 5.7 km/s at some depth beneath the basement of the sedimentary basin as this is defined and imaged from MCS (Bécel, 2006). This refractor is for instance at 8 km depth right below the OBS 10, whereas the deepest reflector distinguished by MCS as forming the base of the finely-stratified basin infill sediments and that is thus considered as their pre-kinematic basement is at 6 km depth. The crystalline basement top seen by OBS refraction appears unreflective on the MCS profiles, probably due to a rather small impedance contrast between the two basements. OBS first arrivals allow to follow its strong topographical variation that follows the bottom to the sedimentary infill imaged on MCS profiles.

3.1.3. Depth variations constrained from Pn and PmP travel-times compensated from shallow structure by MCS and OBS-refraction

Even to offsets greater than 80 km where onsets are clear, there is no significant change of waveform or velocity that would indicate first arrivals of refracted waves from a layer deeper than the crystalline basement interface. At all offsets, the OBS in the basins record a long lasting wavefield that, as illustrated on the MCS sections, contains multiples, peg legs and lateral echoes in addition to possible primary reflections from interfaces deeper than the top of the crystalline basement. However, such primary reflections on deeper interfaces have been recorded on OBS 02 and 11 (Fig. 4) and are discussed hereafter.

In a further step, observational data at larger offsets, from land stations to the west and east of the Sea of Marmara, have been integrated to the OBS dataset. They give additional information on the deep structure and extend the model towards greater depth (Fig. 3). For the 2D modeling of the deep crustal structure, we considered land stations, which are as much as possible aligned with the OBS profile SM-1/23: Stations ELM and PAL to the west, and KOR to the east.

On the record-sections of these land stations (Fig. 5), we identify as first arrivals Pn waves that are the refracted waves in the upper mantle, and in the two opposite directions of propagation. For the land station ELM, we observe Pn waves for offsets ranging from 138 to 185 km, for PAL, from 120 to 165 km and land station KOR to the east from 130 to 170 km. The sampling of the Moho discontinuity with direct and reverse profiles gives critical constraints on the depth and velocity of the top of the lithospheric mantle. The modeling of the times of reversed Pn waves establishes that the Moho interface is at about 26 km depth under the NMT (Fig. 3). From the different velocities tested for the upper mantle, the value of 8 km/s seems the most appropriate, but since the segment of Moho that is sampled by reversed and overlapping observations on both sides is short because of the large basin size, this true refractor velocity cannot be estimated with an accuracy better than 0.2 km/s.

In the reflected part of the wave field, the PmP, Moho-reflection is strong but does not have the clearest onsets because it is in the coda of an earlier reflection from a shallower intracrustal interface. This

**Fig. 6.** a) Top: seismic section from land station Pala (same as Fig. 5b), bottom: seismic section with calculated travel-times through the model superimposed. b) Top: seismic section from land station Elma (same as Fig. 5b), bottom: seismic section with calculated travel-times through the model superimposed. Pg: Refracted waves through the basement, PI P: the waves reflected on the top of the lower crust, PmP: waves reflected on the Moho, Pn: and the refracted through the upper mantle.
intraprofile phase, PnP has the clearest arrivals (Fig. 6). PnP waves are observed clearly on the land station ELM ranging from offsets 70 to 110 km and on the land station PAL from offsets 90 to 120 km. This has the usual character of a reflection at the top of a reflective lower crust. Modeling of the two reflected waves (Pip and PnP) defines the top and bottom of the lower crust. These wide-angle data establish hence that the reflective lower crustal unit is present under the NMT itself, whereas the MCS vertical reflection does not clearly image it in the complex side-echo wavefield generated by the basin rims. If it is clearly identified, however, its thickness cannot be well constrained due to the trade-off with velocity.

Among the late arriving waves on the OBS record-sections, only OBS 02 and OBS 11 clearly exhibit identifiable waveforms reflected at great depth among well-identified sea-bottom multiples in the late response of the structure. The reflector of the last one that is observed can be identified as the Moho consistently with the depth derived from Pn waves recorded by land stations and they can then be modeled (Fig. 4). Moho reflections PnP and reflections at the top of lower crust reflections, Pip, have been observed and considered in the modeling. However, the PnP do generally not provide clear onsets on land-stations records, where the energy is difficult to separate from the response of the lower crust convolved with the ringing in the water layer and sediment under shot points. The top of the lower crust is thus constrained mainly by the OBS data, being less clear on land station data.

In the MCS time-domain stack-section, the top of the lower crust from the WARR model would be more or less at the same time as the lateral echoes. Even on a dip-line, we have very little chances to remove the dipping event without altering the signal that may be reflected from the lower crust.

3.2. The deep crust and Moho under the southern flank of the NMT (line SM-3a–19–15)

Reflection seismic evidence of the reflective lower crust, with the Moho at its base as commonly considered is only imaged on the profiles located on the margin where the sediment cover is thin. Such an image has been shown locally on line SM-36, on the southern flank of the Cinarcik basin (Laigle et al., 2008). Unequivocal identification of this base of the reflectivity as the top of the mantle can only be achieved by mantle velocity measurement by refracted waves at distant stations, as it has been achieved for the NMT. There is no deep reflectivity image that comes out clearly through the side-echoes and multiples on the coincident MCS profiles.

However, both the reflection and refraction data are clear in the case of the East–West profile along the southern margin. This line is made of several pieces of MCS profiles, SM-15–19–3a, that were tie-lines between transects and comprises 5 OBS (Fig. 7a). The MCS data show conspicuous late energy that we interpret as lower crustal reflectivity, an interpretation that can here be checked by the coincident wide-angle reflection–refraction from OBS and land stations. The same approach as for the main regional line discussed above has been used to build the velocity–depth model.

In a first step, an initial supracrustal velocity model has been derived from the layered sediments and their basement as imaged by the coincident MCS profiles displayed in the subsequent Section 4. Then, arrival-times at OBS with mainly basement refracted waves have been modeled. Third, wide-angle reflection–refraction data, from land stations ELM, PAL, MAH to the west and land station FEV to the east (Fig. 7a) have been added to reveal the deep structure on the southern margin of the NMT. Fig. 7 displayed the seismic profile from land station ELM, where three distinct arrivals are observed. The first arrivals, up to the offset 125 km, are attributed to Pg waves whereas the first arrivals at greater offsets ranging from 138 to 162 km are attributed to refracted waves in the upper mantle. This land station exhibits also late arrivals corresponding to waves reflected on the top of the layered lower crust (Pip) with possibly refractions as first-arrivals between 125 and 138 km, and on the top of the upper-mantle (PnP).

On the final velocity depth model displayed in Fig. 7b, the Moho depth ranges from 35 km depth beneath the land station ELM in the west to about 26 km beneath the OBS 31, north of the Imrali Basin. The Moho remains at that shallow depth between OBS 32 and OBS 29, i.e. towards the deep basins of the NMT. The Moho shallowing takes place where the basement is becoming deeper, thus documenting an even more substantial thinning of crust under the NMT. This confirms quantitatively our earlier interpretation of the MCS data on the E–W profile along the southern rim (Laigle et al., 2008). We indeed propose that the crustal thinning observed as being eastwards on an E–W profile was in fact basinwards, since the NMT broadens there to include the North Imrali basin.

We thus suggested that a south–north basinward crustal thinning might exist. The Moho reflection observed on this southern profile beneath the south-western edge of the Cinarcik Basin, between OBS 32 and OBS 29, is at the same, shallow, depth as the Moho found beneath the parallel profile SM-1/23 between the OBS 12 and OBS 04 which are located 10 km further north. There is no significant N–S Moho depth variation between these two profiles in the Cinarcik–Imrali Basins. On both strike-lines, the Moho gets deeper where the top of the basement gets shallower, but this does not occur at the same longitude. For the profile SM-1/23, it is occurring beneath the Tekirdag basin, between OBS 12 and 13, at about 25 km from the shoreline, and for the profile SM-3a–19–15, it is occurring south to the Central High, between OBS 33 and 32, at about 40 km from Marmara Island (Fig. 7a).

3.3. Evidence for the lateral variation of Moho depth

Due to the large dimension of the NMT, it is difficult from any station to have a good coverage of the Moho sampling throughout it because Pn or PnP are only clear in a restricted range of offsets. In addition the clarity of the onsets are perturbed by the considerable upper crustal lateral heterogeneity. The dataset provides hence very limited cross-sampling in three-dimensional (3D). It does not fulfill requirements such as could be met for an approach to crustal thickness variation by a 3D tomography from PnP as achieved by Zelt et al. (2005) in the Gulf of Corinth which is of smaller size and with subdued upper crustal complexity.

3.3.1. E–W variations in the Moho depth

A decrease of Moho depth beneath the eastern and western rim of the Marmara Trough is underlined by the refraction–wide angle reflection modeling of OBS and land station of the E–W profile SM-1/23. An E–W spatial variation in the absolute depth of the Moho of about 5 km is observed as well as a significant increase in basement depth. This is in agreement with the commonly accepted idea of crustal thinning under extensional basins. It is also confirmed with the modeling of the WARR data of the E–W southern profile, from which a crustal thinning beneath the eastern part of the NMT is also inferred.

3.3.2. Can we sense a N–S variation in the Moho depth?

North–South variations on the MCS profiles are difficult to sense under the basins themselves on account of the structure-generated noise level as discussed earlier. All the wide angle reflection refraction data from land stations have been screened in order to find examples where a North–South variation in the Moho depth could be established. We searched for them on in-line profiles and fan-profiles. We considered only land stations sufficiently distant to have recorded Pn waves without ambiguity as first arrivals or clear PnP waves. Moreover, we selected seismic sections for which we have a good control of the supra-crustal structure from refraction modeling (Bécel, 2006) or MCS profiles, which have been post-stack depth-migrated. We proceeded to several simple tests, some of them turned out to be conclusive.
3.3.3. Small Moho N–S dip at 40 km south to the NMT

Fig. 8 displays as an example the seismic sections of the land station Oren, located 80 south of the NMT which recorded the shots of the N–S profile SM-5 through the Central High and Kumburgaz basin and the NNW-SSE profile SM-3 through the Central Basin. Both of these seismic sections exhibit clear Moho reflections, PmP waves.
Fig. 8. North–South variation in the absolute depth of the Moho Discontinuity sensed on land station Oren which recorded the shot of the North–South profile 3 across the Central Basin. a) Location map — black square in the south indicates the location of the land station Oren, the black line indicates the line of shots profile SM-3 and the black square offshore the OBS locations. The red arrow indicates the Moho portion sampled with the WARR data. b) Ray paths of the reflected to the Moho waves (PmP) through the model with a flat Moho at 30 km depth, vertical exaggeration of 1.5.  LC for lower crust. c) Corresponding calculated travel-times superimposed to the seismic section of land station OREN. d) Ray paths of the PmP waves through the model with a Moho topography. e) Corresponding calculated travel-times superimposed to the seismic section. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
These two lines have already been modeled with the OBS, giving us access to the supra-crustal structure (Bécel, 2006).

To this starting model with the supra-crustal structure, we then added two flat interfaces at depth, which correspond to the top of the LLC and the Moho discontinuity at 20 and 30 km depth respectively. The computed travel times for the PmP are superimposed to those observed on the land station seismic section (Fig. 8b,c). The results show a time delay of about 0.2 s between the observed and computed PmP waves, the calculated PmP waves arriving too early. This time-delay is larger than the uncertainty that we have on the PmP phase picking and can be compensated if we consider a shallower Moho towards the NMT (Fig. 8d,e). Modeling of these PmP waves leads us to conclude that a N–S Moho depth variation exists 10 km south to the positions where the Moho depth is constrained by the two crossing E–W profiles. This variation is not dramatic but progressive. For the case of the profile SM-5, a depth difference of about 2 km (from 28 km to 30 km depth) is measured which corresponds to a dip of 4.3° over 30 km (Fig. 8d).

The land station data considered here constrain the depth variation, not just beneath the deep basins but at midway between the station and the profile. For this profile, Moho is sampled between 62 and 77 km south to the OBS 32 (Fig. 8a,b). Here we establish that the Moho sampled at midpoint between receiver and shots, which is indeed at the southern margin not of the NMT but of the Sea of Marmara itself is still deeper than that at the southern rim of the NMT itself, in part because of less sedimentary multiples at later times, and is thus able to image down to the lower crust as it has been shown for the southern E–W line SM-3a-19-15.

4.1. Detachment: a shallow northwards dipping reflector in the upper crust of the southern NMT

Fig. 9 allows to follow a bright seismic event, labeled “D”, over the whole length of lines SM-2–3a-3 which is 43 km long. This event is seen at later times than the top of the basement under the stratified sediment reflectivity. This figure displays a stack time section as a view taken from the south of the Central Basin (Fig. 9a). These lines have been obtained with the largest source of 8100 cu. in., with a shot interval of 150 m and a 15 fold-coverage.

From the course-change point cc1 (middle of Fig. 9b), profile SM-2 crosses the NMT through the so-called Western High towards NNW. Between course-change points cc1 and cc2 (Fig. 9b), profile SM-3a is striking E–W along the southern rim of the Central Basin. To the right of course-change point cc2, profile SM-3 which strikes N–S crosses the Central Basin.

On profile SM-2 of Fig. 9b, the base of the sediments (“B”), a seismic event (“D”) and lateral echoes (“LE”) discussed and modeled from Fig. 2 in Section 2.3, are clearly imaged. The event called “D” can be modeled as corresponding to a reflector (Fig. 2). The latter has a dip of about 20° on this line towards the Western High, which is a lower bound value, if the line is not exactly a dip-line. Such a dip of 20° is larger than expected from any lithological layer boundary in the lithosphere, except in particular cases for the top of the basement.

The other option is thus to consider this event “D” as a tectonic boundary like a fault. This reflector appears with an even smaller apparent dip on the time-section SM-3a of Fig. 9 which strikes east. The apparent dip remains also smaller on profile SM-3 into the Central Basin. In the time-sections, this apparent dip has been taken with caution because of the varying thickness of water and sediments above the event “D”, and it will be hereafter checked. However, in the interpretation of the reflector as a fault, its dip is then rather small, suggesting a rather low-angle crustal normal-fault or detachment.

An additional set of reflectors on profile SM-3, across the Central Basin, is also observed. At the southern edge of the Central Basin, there are segments of reflectors with a southward dip, contrary to the basinward dip of “D”. With respect to the latter being a detachment normal fault, such reflectors would be expected as marking the top of tilted blocks, which are therefore labeled “TB” in Fig. 9. In this image, “D” is overlain by a basement wedge with an opposite, south-westwards dipping top. This basement wedge is interrupted to the north by a steep fault that separates the basement material from the sedimentary deposits down to large depths in the subsided Central Basin (Laigle et al., 2008). The shape of the wedge of basement between “D” and “TB” reflectors is indeed the geometry expected for a tilted block above a detachment. Moreover, another north-dipping event can be followed from the outcrop at sea-bottom down to the detachment “D”. This event has a dip decreasing with depth, since it appears splaying from the detachment “D”, and is thus labeled “SF” for splay-fault in Fig. 9.

Fig. 10 displays a profile close to the one discussed just above, but with different orientations. This profile SM-40–41 makes also a turn at the southern shelf. From the course change point cc3 to the left, profile SM-41 strikes N 50°W across the southern part of the Western High whereas to the right of cc3, profile SM-40 strikes N 40°E towards the southern edge of the Central Basin.

On profile SM-41 of Fig. 10, the same main structures as the ones imaged on the profile SM-2 are found: a bright seismic event corresponding to the detachment “D” on top of which the basement reflector “B” is imaged under the layered sediments. Around 4.5 s, events with strong amplitude, which can be attributed to lateral echoes, are also observed. The intra-crustal detachment can be followed over the whole
30 km length of profiles SM-40-41 whereas it is more difficult to identify tilted block. The same fault "SP" as the one observed on Fig. 9 is imaged as splaying from the detachment around CDP 6400 with their outcrops being close to each other. Profiles SM-2-3a-3 (Fig. 9) and SM-40-41 (Fig. 10) hence reveal the continuity of the detachment, in the form of a detachment fault penetrating the basement, under tilted blocks, and from which the steep normal-faults that outcrop appear to splay. These tectonic features that may be considered as the NMT southern basin bounding faults are here imaged in depth for the first time. They will be discussed as such later in Section 4.4.

Obtaining such an image of faults and structures is unexpected with respect to the commonly transparent images of the continental upper crust in reflection seismic sections, except for such examples of tectonically active regions as in the North Aegean Sea (Laigle et al., 2000; Vignier, 2002) and the Gulf of Corinth (Sachpazi et al., 2003) or in the Basin and Range region of the western US (Klemperer et al., 1986).

4.2. Polarity of the intra-basement detachment reflector

Recorded with a higher frequency source and a tighter 27 m shot interval, profile SM-46 (Fig. 11a) samples with a 60-fold coverage the same tilted block in the right hand side of CDP 6400 with their outcrops being close to each other. Profiles SM-2-3a-3 (Fig. 9) and SM-40-41 (Fig. 10) hence reveal the continuity of the detachment, in the form of a detachment fault penetrating the basement, under tilted blocks, and from which the steep normal-faults that outcrop appear to splay. These tectonic features that may be considered as the NMT southern basin bounding faults are here imaged in depth for the first time. They will be discussed as such later in Section 4.4.

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The reverse polarity of this reflection with respect to the sea bottom reflection could also be related to its nature of being a fault zone and hence associated to fluids on the detachment plane itself, in a damage or gouge zone. If it were a thin layer inclusion, its particularly bright response on line SM-46 could suggest a quarter wavelength maximum response hence a thickness on the order of 50 m.

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On the profile SM-45-46 (Fig. 11b,c), the high amplitude and reverse polarity with respect to the sea bottom reflection is most clearly seen where we interpreted a tilted block as overlying the detachment. Furthermore, the geometry of this block, which appears back-tilted, confirms the idea of an intracrustal detachment. The polarity of the reflection becomes negative only when back tilted blocks surmount the detachment, it stays positive elsewhere. According to the refraction modeling, tilted blocks have high velocity of about 4.5 km/s and are interpreted as being cretaceous limestone basement, as it can be supported from the well data on the southern margin (Ergun and Ozel, 1995).

We may hence attribute the difference of polarity to a gouge zone with low velocity (<4.5 km/s) where the hanging wall is not soft sedimentary rock.

Fig. 9. a) Location map of part profiles SM-2-3a-3 (thick solid black line) on the south western part of the NMT and part of the SM 40-41 (thin dashed line). b) Stack time section of profile SM-2-3a-3 — cc1 indicates the first course change of the R/V le Nadir and cc2 the second change. B is for basement, D for detachment, TB for tilted blocks, LE for lateral echoes and SP for splay fault — vertical exaggeration of 3 with respect to the sea bottom.
4.3. Geometry in space of the detachment, description of the dip and strike

The northwards dipping reflection interpreted as an intra-crustal detachment can be followed on almost all the MCS interconnected profiles south to the EW profiles SM-1/23 and between the Western High and the Central High (Fig. 1b). However, the reflections on the detachment are lost when the sedimentary sequence becomes too thick, limiting hence the penetration but possibly also when it gets steeper in the time-section from the pull-down effect of the deepest basins. The two causes appear indeed to occur simultaneously.

The spatial sampling by our numerous profiles in this region allows us to show that the detachment has an apparent dip that differs with respect to the azimuth of the profiles, away from their turn at the southern margin. The velocity pull-up/pull-down effects on the time-section due to the strong lateral variations of the sea-bottom and the basement topography and thus of the sedimentary infill skews the view that we have on the dips. Therefore, to further discuss the geometry of "D", its arrival time has been transformed to depth and then migrated.

The migrated 2D image in depth of the detachment surface is displayed in Fig. 12b. In a first step (Fig. 12a), points regularly spaced on the interconnected lines where the detachment is followed have been converted to depth using realistic velocities.

The resulting image shows that the detachment surface is dipping northeastwards. At shallow depths (less than 4.5 km) the surface has a local steeper dip under profile 40 striking N40°E with a maximum true dip of 28°. It indicates an overall strike of N 130°E for the detachment. However the surface is not strictly a plane and has a slightly smaller dip in N direction with a strike more E-W as observed on profiles SM-2 and SM-44. The true dip of the intracrustal detachment fault reflector for this south-western part at shallow depths is overall N30°E through the grid of profiles. At depths greater than 5 km, the steepest apparent dip is observed along profile SM-3a-19-15, where the detachment reaches 13.5 km depth at the eastern tip of profile SM-19. There, a true dip direction would be N50°E although it is less well constrained. Both its trace at outcrop and its termination at depth, close to the lower crust, strike ESE to SE. This is obliquely away towards ESE, from the western end of the alignment of deep-sea basins in the Marmara Trough.

In the turn of profile SM-2-3a-3 (Fig. 9) on the southern shelf NE of Marmara Island, detachment-reflections almost come up to the seafloor. At this point, around CDP 39000, the detachment gets as shallow as about 1 km depth below the sea level. A simple interpretation would be then to consider that this detachment outcrops at Marmara Island, if the dip remains the same or even at a more northern position, if the dip becomes steeper at the outcrop. Marmara Island would be thus a geologic window of the unit below the detachment. On the Marmara Island Mesozoic rocks outcrop and more particularly metamorphic exhumed rocks, (Okay et al., 1996) which are consistent with the proposed interpretation.

4.4. North Marmara Trough southern basin-bounding fault: splay-fault from the detachment?

On profile SM-3a (Fig. 9), and profile SM-40-41 (Fig. 10) a steeper fault, which splays upwards from the detachment, is underlined. This
fault is not among the ones documented at sea bottom and has not been previously mapped. This normal fault, which outcrops around CDP 39650 appears to bound the southern shelf and to merge with the detachment. Its true dip is steeper than the 25° apparent dip on profile SM-3a since this is not a dip-line.

This fault imaged at depth could be the rim of a structural element, which was not documented before. On the profile SM-3a, this fault is located to the south of a tilted block lying on the detachment. It is likely that it is the same splay-fault “SF” which is imaged on the two profiles SM-3a (Fig. 9) and SM-40 (Fig. 10), since the outcrops are close to one
another. This however then prevents from having a good accuracy in measuring its strike, which is here of N 80°E (Fig. 15). This fault which outcrops, is still active. This implies that the detachment itself is active.

In summary, the low-angle reflector we called detachment fault is bounding at their base the basement blocks perched at the southern margin of the NMT. This new image and view would suggest that the southern rim of the NMT as seen in the break of the sea-bottom topography is not a steep normal master fault breaking deep down into the crystalline crust with the top of the basement significantly and sharply downthrown across it (Armijo et al., 2002; Parke et al., 2002). The southern rim appears hence to be made of tilted blocks lying on a northeastwards dipping detachment.

This low angle fault, with about 20° dip on the line 2 (Fig. 9) may be an inherited structure with its normal sense of motion taken only in the recent tectonic evolution. It could have reactivated a feature originated as a thrust fault in the regional context of convergence–subduction at the Intra-Pontide suture zone prior to the Cenozoic. Indeed, the detachment has a direction that may be that of the Paleogene Thrace Basin, which is interpreted as a forearc basin related to the Intra-Pontide Suture Zone, which separates the Pontides from the Sakarya Zone (Wong et al., 1995, Yilmaz et al., 1997), as confirmed by the ophiolite belt found on land at the two extremities of the Marmara Sea.: through the Ganos and the Armutlu Peninsula.

4.5. Detachment observed down to the top of the reflective lower crust?

On the E–W profile along the southern rim of the NMT (Figs. 1b and 13), the detachment is followed dipping east, towards the top of the lower crust. On this stack time section, the detachment reflector seems going close to the top of LLC on profile 15, which is here at about 7 s. The profile 3a-19-15 is an EW profile, hence the detachment cannot be migrated properly, the dip line being close to the profile SM-40 direction (see Section 4.3).

The critical part which could show us that the detachment reflector stops before reaching the top of the lower crust, or alternatively is extending deeper and then whether it either merges with, or cuts into the lower crust is not clear in the time-section. Indeed it would be an interference domain of the response of the two reflections with different dips forming a slant bow-tie diffraction shape in the time-section. To really resolve such a feature by seismic processing, the signal to noise ratio and resolution would need to be as good as in shallow sedimentary reflection seismics of the oil industry where such processing can succeed. Obtaining such data quality at 15 km depth within the crust would need focused acquisition efforts not commonly within reach at least with a one-pass reconnaissance profile.

Further to the east, the detachment is not observed on the transect profiles SM-5 and SM-6 (Bécel, 2006). If it were indeed absent, this could be consistent with the fact that it is already too deep and/or has reached the top of the LLC further WSW as described above. This is also the case of profile SM-7 further east that reaches as far south as Imrali Island.

4.6. Backtilted supra-crustal blocks on the margin of the continental shelf

On profile SM-3a-19-15, a succession of conspicuous opposite dipping and down-stepping reflection bands can be consistently interpreted as the top of tilted blocks on top of the detachment (Fig. 13).

4.6.1. Block “a”

On the profile SM-2-2a-3 (Fig. 9), a tilted block is imaged in the hanging-wall of the low-angle fault-plane between CDP 40700 and CDP 43000. The crustal culmination observed around CDP 42500 with
an anticlinal shape is hence not a compressional feature but indeed a roll-over anticline formed by extension. When reaching under the deep Central Basin, profiles do not show reflections of a detachment or back tilted block.

This same tilted block which lies on the detachment fault is identified on some other profiles, such as on profile SM-40-41 (Fig. 10), where we distinguish, although less clearly, the top of this tilted block, on profile SM-46 (Fig. 11) and profile SM-19, not shown. With these interconnected lines, we have hence access to the tilted block’s strike by mapping its crest (Fig. 14). The strike is here of about N 120°E, the same strike as the detachment.

4.6.2. Block “b”

On the profile SM-3a-19-15 (Fig. 13) a succession of two tilted blocks is underlined. The first one to the west corresponds to that block “a” just described before. Down-stepped by a fault that interrupts the top of this tilted block to form a crest, a huge tilted block “b” appears in the hanging-wall of “a”. Sediments deposited on top are tilted and prograding towards the east in the plane of section, which has a basinward component in this direction. This indicates continuing differential subsidence with a sediment deposition rate that kept pace, since there is no marine basin that is sea-bottom depression, along the section.

The huge tilted block “b” is observed more to the east, on the profile SM-19 and profile SM-15. These profiles show under the stratified reflective layer on top of it what may have been two smaller tilted blocks before, “b1” and “b2”. The step between the latter appears sealed by a stratified reflective layer. This layer above the top of this pair of tilted blocks could correspond to the top of the Upper Cretaceous Limestone as found in the Marmara I borehole (Ergun and Ozel, 1995) (Fig. 13).

The huge tilted block “b” is also imaged on profile SM-5 and on profile SM-6 (Bécel, 2006). Its crest is mapped in Fig. 14 and has approximately the same direction as the tilted block “a”. The crest of the smaller tilted block “b1” seen on profile SM-15 (Fig. 13) is also observed on the profile SM-18 (not shown) and its crest is sketched in Fig. 14.

4.6.3. Another tilted block further north

A tilted block is imaged on profile SM-5, which cuts across the Central High, and also on profile SM-38 and 39. We have no seismic evidence of the detachment on this profile. It may have died out as a reflector, or be too deep for imaging or have merged with the deep crustal reflectivity. This block (Fig. 14) would hence not correspond to one among several tilted block (as a roll over anticline) anymore but rather to a rider block (Gibbs, 1984), which would have been separated from its former neighbors on top of the detachment.

The detachment is overlain by back-tiled supra-crustal blocks in a zone that widens eastwards. These new features are located beneath the southern shelf of the NMT, south to the Central Basin and Central High, which obliquely broadens towards SE, with respect to the E–W direction of the NMT. In this region, the southern boundary is indeed striking SE along the North Imrali fault system. In this shelf, there are basement deeps that appear subdued at sea-bottom because of strong sedimentation, and highs that are indeed revealed to be crests of tilted or rider blocks covered by sediments. Tilted blocks are overlain unconformably by their sediments. Inherited tilted blocks are

![Stack time section of part the profiles SM-3a-19-15 which forms an EW line along the southern rim of the NMT. D is for detachment, LLC is for layered lower crust, “a” and “b” are for the 2 huge tilted blocks, “b1” and “b2"are for smaller tilted blocks sealed by a stratified reflective layer, SP is for splay fault, black star indicates the location of the Marmara I borehole (Parke et al., 2002).](image)
delineated by the outcropping normal faults that may connect onto the detachment.

5. Discussion–conclusion

5.1. Narrow basins: a challenge for deep seismic investigations

The design of the SEISMARMARA-Leg 1 survey as a multi-method approach allowed to reveal the crustal scale architecture of the narrow and deep bathymetric NMT with complex shallower structures, even if these are rather adverse situations for seismic data acquisition.

Investigating the deep crust and Moho under tectonically active sedimentary basins is a challenge to seismic methods. The reflection seismic response can be obscured by the response of the multiples from the sea-bottom and from the sedimentary part and/or by shallow structure diffractions.

In the case of a narrow basin, propagation paths in the vertical plane of the profile may not be the first arrivals since side propagation may be faster. Here strong side-swipes from its rims arrive at such echo times that they overprint the response of the deep structure, as experienced in the Gulf of Corinth where only oblique angle under-shooting in wide-angle geometry could identify deep crustal structure (Clément et al., 2004; Zelt et al., 2005). In the Marmara Sea, in order to cope with the identification of seismic events recorded in MCS profiles, SEISMARMARA Leg1 comprised a large number of profiles with diverse orientations, which allowed discriminating a seismic event on a strike-line with respect to the NMT as being a side-echo rather than real.

5.2. Topography of Moho and lower crustal layer constrained under the deep basins

The imaging of the deep structure in reflection seismics straight below the narrow trough is thus quite impossible since the deeper reflections are usually hidden by lateral echoes. In order to constrain the deep structure under the NMT, MCS profiles had been therefore complemented by coincident wide-angle and refraction profiles, in order to provide diverse views from waves interacting differently with the surface and deep structure.

The major result obtained from wide-angle reflection–refraction on reversed profiles is the identification of the top of the upper mantle beneath the NMT. An E–W Moho depth variation and a more subtle and smoother N–S one have been highlighted by modeling of Pn and PmP waves arrival times. In the East–West direction, the Moho depth decreases quite sharply from onshore to offshore beneath the eastern and western ends of the NMT. A spatial variation of about 5 km is measured in the absolute depth of the Moho, which is stepping up at the eastern and western edges of the NMT to reach 26 km depth. The depth under the NMT cannot be resolved by the data to be significantly different from grossly constant. The southern limit of the area of thinnest crust strikes obliquely away from the alignment of marine depressions, in a south-eastward direction from the Central Basin towards the Imrali Island.
The variation in the Moho depth from the south towards the NMT has been sensed and shown quantitatively as well as qualitatively. This variation is progressive rather than sharp and a new result is that the reduction in Moho depth already begins far to the south from the NMT, under the southern shelf composed of several basins. Crustal thinning may be thus involving the whole Marmara Sea and southern margin, with a maximum finite deformation located under the NMT with respect to Moho depth measured on its western and eastern rims. Recording and modeling of reflections from the top of the lower crust on land stations and OBS establishes that a reflective lower crustal unit is also present under the NMT itself. The topographical variation of the top of the lower crustal layer cannot be distinguished as being different from that of the Moho boundary given the resolution of the arrival time pickings at wide-angle, and also the uncertainties on velocity. With the basement depth deepening within the NMT, this supports the common expectation of crustal thinning under extensional basins. More specifically here, since the Moho and the top of the lower crust appear to upwarp together, the thinning of the crust is documented to result principally from the thinning of the upper crust.

The most important structural result of the present investigation is the thinning of the crust, and principally the upper part, under the NMT, by as much as 11 km locally, with a 5 km of mantle upwarp adding to the 6 km of basement foundering.

5.3. Upper crustal thinning by a detachment fault and tilted blocks

A set of MCS profiles with diverse azimuths in the region along the southern rim of NMT allows us to image a lower crustal reflectivity. They also reveal structural elements through the upper crust, and allow to resolve the sedimentary record that may help to link these structures to the present active evolution.

An intra-basement reflector imaged by the set of MCS profiles in the western half, appears as a major feature, a probable intracrustal detachment fault, which was completely unknown before this survey. The mapping of the detachment gives us a true dip of about 28° in the N40°E direction. The dipping event has been interpreted this way from its reflective character and from its rather low-angle dip and upper-crustal scale. There is also the normal throw shown by the overlying back-tilted basement blocks separated by steep normal-faults and splay faults, with the shallowest one outcropping. The normal sense of motion on the detachment is thus suggested to be active at present. However, the low-angle geometry might have been inherited from previous episodes of crustal evolution in a convergence context.

On the E–W profile along the southern margin, the apparent dip of the detachment is moderate consistently with its northeastward true dip direction, and so the reflector is seen on a long profile segment. This allows to see the detachment dipping towards the top of the reflective lower crustal layer, while the latter is upwarping eastwards under the detachment hanging-wall. Deep seismic imaging thus reveals that the thinning of the upper-middle crustal layer is controlled by detachment faulting and block-tilting. The geometrical relationship between the features in the upper crust such as the detachment and tilted blocks and the lower crust suggests that they are linked together in the recent evolution. However, the detailed geometry of the possible link at depth of the detachment with the top of the lower crust and hence whether reflectivity is forming there at present could only be possibly resolved by a targeted experiment with increased acquisition capacities.

5.4. Asymmetric extension and partitioning of displacement

An unsuspected view revealed in this study is that the extensional sidewalls of the Central Basin are more complex than if they were just pull-apart basin sidewalls separating along a gash at an extensional jog or side-step of a single strike-slip fault. The SW sidewall of the Central Basin appears indeed as surmounting, and possibly being part of a system of branching faults probably forming a negative flower structure at the scale of the whole crust (Fig. 15). This system includes a dipping detachment that is imaged almost from outcrop at the sea-bottom down to the top of the lower crust. Fault-tilted blocks are imaged in its hanging-wall, above which the bottom of the Central Basin has subsided.

These features which were unknown before this survey have been partly suggested in Laigle et al. (2008), but their detailed seismic imaging documented here can give an indication to the mechanism of thinning of the upper crust as well as to the general partitioning mechanism of the displacement. With respect to stable Eurasia to the north, there is displacement and deformation in the NMT and southern margin, with asymmetric extension and transport towards the southwest. The crustal thinning may occur in part by the detachment mapped south of the Central Basin, since this detachment allows the escape of its footwall, towards the SW.

The crustal material transport may occur from above the lower crust and from under the detachment fault, which partly exhumes its footwall south-westwards. This mechanism may provide the space for subsidence of the upper crustal material in the hanging wall of the detachment and for an upwarp of the lower crustal layer beneath, at the tip of the detachment at depth.

The detailed features of basin and basement heterogeneity imaged south of the Central Basin confirm that this large-scale extension and corresponding right-lateral displacement are not restricted to the deep basins only but is accommodated by localization of displacement on several faults in the broad region of the NMT and southern rim or even margin.

5.5. E–W variation of size of element and similar process

The mechanism suggested here for the Central Basin and margins may be active in a transposed way, in the eastern third of the Marmara Sea comprising the Cinarcik Basin and North Imrali basin where it would have a more important width as described in details in Laigle et al. (2008). The southern rim of the North Imrali basin could be
viewed as a detachment normal-fault. In its hanging-wall, the Cinarcik basin may subside between conjugate normal faults, as in a rotational glide-block system (e.g. Groshong, 1990). This has also a pull-apart character in 3D, with the changing strike of the bounding faults. Extensional sidewalls of the Central and Cinarcik Basin could thus be considered as intra-basins faults of the NMT, with tilted blocks of different sizes on their southwestern rim.

South of the Central Basin, the detachment is too shallow for having in its hanging-wall a tilted basement block as it appears south of the Cinarcik basin where the North Imrali basin is established.

In the whole North Marmara Sea, the deformation seems partitioned asymmetrically and occurring with similar processes with, along the NMT, a change of size of its elements that relay one another and change strike along transfer zones. The geometry of the fault system can be regarded as a broad flower structure at crustal or even lithospheric scale.

5.6. Elements of a principal displacement zone in horizontal plane shear strain

At the scale of the two hundred kilometers length of the Sea of Marmara and of several tens of kilometers of relative strike-slip displacement across it, it is not only the basement but also the whole lithosphere that is revealed to be activated. This is shown by other markers of deformation than the subsidence of basins, that can be accessed with the deep seismic penetration and imaging, such as the change in topography of deeper levels, like the top of the lower crust and the Moho.

Previous interpretations of the Sea of Marmara could only use depth-limited observations recalled in the introduction such as the sea bottom swath-bathymetry and reflection seismic sections limited to depth by the sea bottom multiple. They have emphasized model elements as diverse as the importance of extension (Aksu et al., 2000; Parke et al., 2002; Valtirak, 2002), block or plate-like junctions or geometry (Okay et al., 2000), pull-apart or transtensional jog activity at diverse scales (Armiyo et al., 2002), evolution to strike-slip localization (Le Pichon et al., 2001). The new seismic observations from the SEISMARMARA Leg-1 survey lead us to embed them in a broader and deeper context. The original insights through the whole crust brought by the data presented here allow to suggest a consistent model of thinning, extension and transtension at the scale of the lithosphere. The Sea of Marmara can indeed be regarded as a large negative flower structure at the scale of the whole crust where the principal amount of finite deformation is localized at depth under the North Marmara Trough.

The diverse nature and orientation of the different faults and structure can be viewed as the evolution of a principal displacement zone resulting from horizontal plane simple-shear applied at depth at lithospheric scale as suggested by the large-scale dimension and propagation of the North Anatolian Fault (Laigle et al., 2008).

The crustal thinning newly evidenced here would involve principally the North Marmara Trough but also affect its southern margin, the whole Marmara Sea. This is suggested to occur in part by the detachment mapped south of the Central Basin, allowing the escape of upper crustal material in its footwall. With respect to stable Eurasia to the north, there is displacement and deformation in the NMT and southern margin, with an asymmetric extension, a partitioning of displacement and with a transport of material towards the southwest.

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