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Crustal flow driving twin domes exhumation and low angle normal faulting in the Menderes Massif of western Anatolia

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Abstract

Lower crustal flow in regions of post-orogenic extension has been inferred to explain the exhumation of metamorphic core complexes and associated low-angle normal (detachment) fault systems. However, the origin of detachment faults, whether initially formed as high-angle or low-angle shear zones, and the extension is symmetric or asymmetric remains enigmatic. Here, we use numerical modeling constrained by geophysical and geological data to show that symmetric extension in the central Menderes Massif of western Anatolia is accommodated by the crustal flow. Our geodynamic model explains how opposite dipping Gediz and Büyük Menderes detachment faults are formed by $\sim 40^\circ$ footwall rotation. Model predictions agree with seismic tomography data that suggests updoming of lower crust beneath the exhumed massifs, represented as "twin domes" and a flat Moho. Our work helps to account for the genetic relation between the exhumation of metamorphic core complexes and low-angle normal faulting in both Cordillera and Aegean orogenic regions and has important implications on crustal dynamics in extensional provinces.

Keywords: Geodynamic Modeling, Lower Crustal Flow, Western Anatolia

1. Introduction

Extensional tectonics following mountain building- orogenic- processes lead to special geological characteristics of the crust, such as rapid pulses of magmatism, exhumation of metamorphic core complexes, and associated low-angle normal (detachment) faults with tens of kilometers of displacements (Lister et al., 1984; Dewey, 1988; Malavieille, 1993; Jolivet, 2001; Rey et al., 2001). However, the extent and magnitude to which post-orogenic extension is involved in strain localization in the brittle upper crust and in ductile flow in the lower crust are not well understood. In this work, we test the hypothesis that the origin of central Menderes metamorphic core complex and symmetric detachment faulting in western Anatolia, a post-Alpine extended region, are formed by ductile flow of lower crust and footwall rotation of high-angle normal faults.

The exhumation of metamorphic rocks in the central Menderes Massif has been accommodated by two symmetrically developed, outward-facing ductile-brittle high strain (low-angle) detachment faults (<20°). The northern detachment fault, the Gediz/Alaşehir detachment, is associated with the top-to-the N-NNE shear sense, whereas the southern one is associated with the top-to-the S shear sense (Hetzel et al., 1995; Emre and Sözbilir, 1997; Gessner et al., 2001; Bozkurt, 2001; Lips et al., 2001; Işık et al., 2003; Göğüş, 2004; Çemen et al., 2006; Nilius et al., 2019; Heineke et al., 2019b) (Fig. 1a, b). Proposed ages for the initiation of normal faulting for each detachment system vary spatially and temporally. For instance, based on isotopic dating along fault rocks cataclasites, Hetzel et al. (2013), suggest that the onset of Büyük Menderes detachment faulting is ∼22-20 Ma, whereas the shearing across the Gediz detachment began at ∼16 Ma inferred from dating of syn-extensional intrusions (Catlos and Çemen, 2005; Glodny and Hetzel, 2007; Rossetti et al 2017). Further, magnetostratigraphic analysis and isotopic ages over the supra-detachment basins suggest that these two detachments have operated simultaneously since 16 Ma (Şen and Seyitoğlu, 2009). A detachment-controlled symmetric cooling pattern related to unroofing of central Menderes has also been suggested through thermochronological studies (Gessner et al 2001; Ring et al., 2003). Such symmetric configuration of core complex exhumation has been interpreted with respect to the Küçük Menderes graben (rift), located above a central axis of a large-scale syncline where both detachments and the folding evolved contemporaneously (Gessner et al., 2001; Seyitoğlu et al., 2004).

While strain localization across exhumed ranges in the central Menderes Massif region is characterized by detachment faults, high-angle normal faults are well documented along the graben boundaries - akin to rifted margins (Fig. 1a,b). Based on field observations and seismic reflection data, a number of studies interpret that detachment faults in central Menderes were initially formed at higher dip angles ($> 30^{\circ}$) and rotated to shallower orientations (Cohen et al., 1995; Bozkurt, 2000; Gessner et al., 2001; Seyitoğlu et al., 2002; Bozkurt and Sözbilir, 2004, Çiftci and Bozkurt, 2010; Demircioğlu et al., 2010). Gessner et al (2001) propose a rolling-hinge model to address the synchronous evolution of bivergent detachment fault systems. Namely, flexural isostatic footwall uplift is associated with fault rotation as well as the formation of younger high-angle normal faults on the hanging-wall of the main breakaway which successively rotates to low-angle shear zones (Buck, 1988, Wernicke and Axen, 1988). Accordingly, a rolling hinge type tectonic evolution for discrete shear zones in the footwall of the Gediz (Alaşehir) detachment (Seyitoğlu et al., 2002) and the Büyük Menderes detachment (Sümer et al 2020; Türesin and Seyitoğlu, 2021) has been invoked through geologic mapping, stratigraphic analyses, and data from seismic reflection studies. On the other hand, Öner and Dilek (2011) suggests that the detachment fault was initially formed as a low angle shear zone (without rotation) in which the early Miocene-late Pleistocene sediments of Alaşehir (Gediz) graben were deposited in a supradetachment basin. While the studies above provide a conceptual geological framework, however, the origin of core complex exhumation and evolution of ductile-to-brittle localized strain have not been addressed in the context of whole-crust extensional dynamics.

A multi-scale full seismic waveform inversion for crustal and upper-mantle structures by Fichtner et al. (2013) demonstrates that the upper crust of the Menderes Massif is associated with anomalously high velocities with respect to neighboring regions. This has led to the interpretation of upward displacement of the lower crust typically associated with higher velocities than the upper crust (Figure 1c). Seismic analyses of Karabulut et al. (2013) also imply a flow of lower crust beneath the central Menderes Massif, inferred from the locally flat Moho at ~25 km depth to accommodate isostatic compensation through lower crustal dynamics (Block and Royden, 1990). Further, a lower velocity upper crustal anomaly has been imaged beneath the Küçük Menderes graben (Figure 1c) in addition to the updoming lower crustal material shown in the relatively long-wavelength (> 150 km) variations of the isotropic S velocities (Fichtner et al., 2013).

In this work, by using thermo-mechanical models we investigate the dynamics of the post-orogenic extension where lower crustal flow plays a key role in strain distribution. Specifically, the evolution of an array of high and low-angle normal faults is explored and a genetic relationship between these faults and the exhumation of lower crustal rocks is demonstrated. Further modeling is used to test the role of varying extension rates and reduced upper crustal strength. Results are reconciled with the last ~ 15 Ma evolution of the central Menderes Massif of western Anatolia, where symmetrically arranged detachment systems are prominent and seismological data suggests lower crustal flow. Overall, our work will provide new insight into how crustal dynamics control the tectonics of rifted margins and the evolution metamorphic core complexes around the globe.

2. Data and Methods

2.1. Numerical technique, rheological characteristics, and model setup

To reach the objectives of this work, we use the finite element code ASPECT (Heister et al., 2017; Kronbichler et al., 2012), which has been extensively used for conducting geodynamical experiments ranging in scale from the crust to the deeper mantle. ASPECT is employed to solve the extended Boussinesq equations of momentum, mass, and energy as well as advection equations for each compositional field. We compute the 2-D visco-plastic deformation within the lithosphere and sub-lithospheric mantle of a model domain that is 500 km wide and 165 km deep (Figure 2). Adaptive mesh refinement is used to adapt the precision and optimization of the computational calculations. We employ coarse, intermediate, and maximum resolutions beneath 50 km depth, between 50 and 20 km depth, and above 20 km depth, each represented by a resolution of 2500, 1250, and 625 m respectively.

In terms of rheological setup, the visco-plastic effective viscosity η_{eff} is computed from either a composite of dislocation and diffusion creep $\eta_{eff}^{diff|disl} = \frac{1}{2}(\frac{1}{A})^{1/n}\dot{\varepsilon}_e^{(1-n)/n}\exp(\frac{Q+PV}{nRT})$, or Drucker-Prager plasticity $\eta_{eff}^{pl} = \frac{C\cos(\phi)+P\sin(\phi)}{2\,\dot{\varepsilon}_e}$, depending on whether viscous stresses remain smaller than the yield stress or not (Glerum et al., 2018); with pressure P, temperature T, stress exponent $n = 1$ for diffusion creep and $n > 1$ for dislocation creep, pre-exponential factor A, the effective deviatoric strain rate $\dot{\varepsilon}_e = \sqrt{\frac{1}{2}\dot{\varepsilon}'_{ij}\dot{\varepsilon}'_{ij}}$, activation energy Q, activation volume V, gas-constant R, cohesion C, and friction angle ϕ (see Glerum et al., 2018 for numerical implementation and our supplement for employed parameter values).

The initial conditions of our model design aim to approximate the first-order lithospheric structure at the onset of extension in the central Menderes Massif region of western Anatolia, approximately 15 million years ago. The model accounts for four layers of materials: upper crust, lower crust, mantle lithosphere, and sub-lithospheric mantle (Table 1 and Fig. 2). The crustal domain consists of an upper crust (25 km thick) with wet quartzite rheology (Gleason & Tullis, 1995), and a lower crust (25 km thick) with wet anorthite rheology (Figure 2) (Rybacki & Dresen, 2000). Our assumption of an initial 50 km thick crust is based on geological inferences (e.g., paleoelevation and metamorphic grades of exhumed rocks) (Şengör et al., 1985) in which extension in this region started after plate shortening by the Alpine orogeny. A relatively thin (30 km thick) mantle lithosphere is included in the model setup with dry olivine rheology (Hirth & Kohlstedt, 2003), since a number of geological studies suggest that portions of mantle lithosphere have been removed from beneath the region through lithospheric delamination or convective removal (Aldanmaz et al., 2000; Ersoy et al., 2010; Van Hinsbergen et al., 2010; Gessner et al., 2013; Göğüş, 2015). The overall initial thickness of the lithosphere is 80 km.

2.2. Mechanical boundary conditions, thermal field and weakening effects

Kinematic boundary conditions are implemented by prescribing half of the extension velocity at each lateral model boundary. The top boundary is a free surface, while the bottom boundary features a free tangential motion in conjunction with a constant vertical inflow of material that balances the outflow through the lateral model sides. A constant temperature is prescribed at the top (0°C) and bottom of the model domain (1300°C), as well as isolating boundaries at the sides. The initial temperature follows a steady-state geotherm in the lithosphere and an adiabat below. Radiogenic, shear heating, and adiabatic heating are included in the energy equation.

We include strain weakening/softening, where a strain-dependent friction angle and cohesion decrease linearly over a given strain interval (Huismans & Beaumont, 2003; Le Pourhiet et al., 2017), which is also predicted in progressive microscale numerical simulations (Dinç Göğüş et al., 2023). Here, the frictional weakening factor for the upper crust is set to 0.1 (for EXP-1, 2, 3) or 0.05 (EXP-4) over the accumulated plastic strain interval of 0 to 1. If strain exceeds 1, the friction angle and cohesion remain constant at their most weakened value.

We show two model suites where we investigate the impact of key parameters within a plausible range: (1) we vary the extension velocity between V_{ext} = 1-4 cm/year full rate and (2) we vary the friction angle strain weakening factor of the upper crust (0.1 - 0.05).

3. Results

We show predictions of four experiments selected from a series of numerical experiments (>100) in which the role of major controlling parameters on continental extension is examined. Details on the model design, including the initial temperature field and boundary conditions applied, are given in the methods section. Figure 2 and Table 1 include further information on various parameters of the numerical experiments. The description of the section below begins with an explanation of the reference experiment (EXP-1) that best approximates the late Cenozoic geological evolution of the central Menderes Massif and follows with presentations of EXP-2, 3, and 4. Extension in the brittle crust begins through a random initial strain as a representation of deformation distribution which further evolves into strain localization. Results are only shown for the central part of the model domain, where the respective approximate distances of the central Menderes to the plate boundaries in the north and south are nearly close.

3.1.Evolution of core complex and the low-angle (detachment) faults

Figure 3 shows the evolution of our reference experiment EXP-1, where V_{ext} = 2 cm/yr extension rate has been applied, which approximates present-day GPS-derived N-S extension in western Anatolia (Aktuğ et al., 2009). By 6 Myr (Fig. 3a), strain is localized along opposing conjugate shear bands/fault systems, dipping 50º-55º in the center of the model domain. Later by 12 Myr with more stretching (Fig. 3b), flow of the lower crust occurs where there is a higher magnitude of extension of the upper crust, beneath two major faults, owing to the lateral lithostatic pressure gradients. The net crustal thickness is now 30 km beneath the central region, expressed as approximately flat Moho, while there is lower crustal thickening and convergence beneath the zone of maximum upper crustal extension. The influx of crustal material accommodates differential isostatic compensation, and upper portions of the shear zone rotate to shallower dips around two domal/antiformal cores. By 15 Myr (Fig. 3c), more rocks from the deeper crust are dragged upward near the surface along two major symmetrically (bivergent) arranged detachment systems, for instance, lower crustal rocks are exhumed as shallow as <10 km depth below the surface. The lowest dip angle of these faults near the surface is now ~11°-14° and new high-angle normal faults develop on the hanging-wall of these main breakaway fault systems. Note that the Moho variation is uniformly sub-horizontal (25 km) beneath the central domain although thinning of the upper crust is heterogeneous. For instance, there is a higher magnitude of stretching along upper (brittle) crustal extension above detachment faults, and below that there is a culmination of a twin-domes type core complex.

We test how variable rates of lithospheric extension can modify model predictions on the timescale of 15 Myr. These results may provide insight into the evolution of extensional tectonics both on the eastern and the western margin of the central Menderes, where the former margin accounts for lower and the latter for higher rates of extension due to its proximity to the central Aegean basin since subduction retreat has caused back-arc spreading (Jolivet and Brun, 2010; Van Hinsbergen et al., 2010a).

EXP-2 shows the model development in which the extension rate is halved with respect to EXP-1 (so V_{ext} = 1 cm/yr) (Figure 4a). The vertical flow and thickening of the lower crust and thinning of the upper crust are less pronounced, consequently, there is no significant core complex exhumation. By 15 Myr, the widespread distribution of shallow dipping normal faults across the entire crust does not develop, unlike in EXP-1, such fault plane rotation is very limited. The Moho variation remains subdued throughout the entire model domain and the strain localization mechanism may be characterized by wide rift type, in the sense defined by Buck (1991) and shown in numerical experiments of Brune et al., (2017).

On the other hand, with an increased extension velocity (V_{ext} = 4 cm/yr) in EXP-3, strain localization, ductile flow, and the associated exhumation of metamorphic cores and dome-like structures are accelerated (Figure 4b). For instance, by 5 Myr there is up to ~ 20 km lower crustal exhumation between closely spaced faults associated with shallow dip angles near the surface. By 10 Myr, the lower crust reaches the surface at the left limb of the symmetrically arranged detachment faults. Results show similarities with that of EXP-1, for example, oppositely dipping detachment systems warping around domal cores, however, the entire evolution develops much more rapidly. Note that the crustal thickness varies significantly across the model domain, for instance, it is undulated and thinned around the margins, nevertheless, the crustal thickness is almost constant beneath bivergent detachments where the upper crustal thinning is more amplified and the crustal flow process controls isostatic compensation.

In Figure 4c we show the evolution of EXP-4 in which the strength of the upper crust is reduced by amplifying the frictional plastic strain softening. The motivation for such parameter change comes from previous studies where Huismans and Beaumont (2002) suggest that extensional tectonics may be symmetric or asymmetric depending on the implementation of this variable. By 6 Myr, instead of conjugate shear zone development, strain localizes along a specific shear zone, such as the one on the right side of the model where asymmetry prevails in earlier stages of the extension (Huismans and Beaumont, 2002). After 15 Myr, the unroofing of the metamorphic dome reaches ∼5 km depth below the surface as the deformation pattern becomes distinctly asymmetric by the development of a detachment zone across the brittle crust on the domed limb. This is similar to the simple shear extension model of Wernicke (1985), but a major detachment in our model soles in the lower crust rather than cutting through the lithosphere. Depending on the investigated width of the model domain, such extension type may also be symmetrical in terms of detachment fault development (Grasemann et al., 2012). Namely, another low-angle high strain shear zone develops (not shown in this frame), dipping in the opposite direction to the main shear zone, but it is in the distal end of the frame, 300 km further away from it. Although footwall rotation and the differential uplift of the lower crust are well pronounced in this experiment, we note that the results of EXP-4 do not account for the observed tectonics of the central Menderes region. For example, EXP-4 does not predict <100 km distance between Alaşehir/Gediz and Büyük Menderes detachment faults, symmetrical exhumation of metamorphic cores, and the flat Moho between these faults, inferred from seismological studies (e.g., Zhu et al., 2006; Fichtner et al., 2013; Karabulut et al., 2013).

4. Discussion

The results of the reference numerical experiment (EXP-1), where lower crustal flow is associated with footwall exhumation of "twin-domes" and rotation of steep normal faults in the upper crust into two major shallow-dipping high strain zones, comprise many of the first-order features that are invoked to explain the tectonic evolution of the central Menderes metamorphic core complex in western Anatolia. Figure 5a illustrates the spatial and structural agreement between the predicted bivergent tectonics in the form of an observed present-day axisymmetric array of low-angle shear zones projected across an N-S cross section of the region (see Fig 1a X-X' for the location of the cross section). Further, the results satisfy the approximate distance between the Gediz (Alaşehir) (north) and Büyük Menderes (south) detachment faults and associated flanking and exhumation of "twin-domes shaped" footwall metamorphic rocks (Rey et al., 2011; Whitney et al., 2015). Moreover, the simulated geometry approximates the syncline/downward arc beneath Küçük Menderes graben in the center, with both detachments lying on the margin of this syncline. As the model predicts, there is no known detachment fault associated with the formation of Küçük Menderes (axial) graben where high-angle normal faults control its evolution since the Middle Miocene (Emre and Sözbilir, 2007; Rojay et al., 2005; Seyitoğlu and Işık, 2009; Bozkurt et al., 2008), similar to the formation of a localized and narrow rift in the distributed extensional domain (Brun and Choukroune, 1983).

Our model findings show that high-angle normal faults (50°-55°) have rotated progressively ~40° to shallower dips about a horizontal axis accommodated by the crustal flow. This is consistent with several geological interpretations in western Anatolia. Namely, by interpreting the thermochronology data, cooling ages, and reconstruction of Eocene (flat) foliation planes to present day tilted structures, Gessner et al. (2001) suggest that the dip angle of both Gediz (Alaşehir) and Büyük Menderes detachment faults were initially 40°-60° during the earlier stages of post-Alpine extension. The higher dip angle origin of both Gediz (Alaşehir) and Büyük Menderes low-angle detachment faults has also been suggested through current (non-horizontal/tilted) dip angles of the bedding planes of syn-tectonic sedimentary strata, dipping towards the fault plane (Bozkurt, 2000; Sözbilir, 2001; Seyitoğlu et al 2002; Sümer et al., 2000; Çiftçi and Bozkurt, 2010; Türesin et al., 2021). According to fault plane restoration estimates, based on vertical distributed shear (Westaway and Kusznir (1993)) that develops contemporaneously with crustal extension and sedimentary deposition, the initial dip angle of the Büyük Menderes detachment ranges from 44° to 54° (Bozkurt, 2000), and an original dip of 56°-58° was estimated for the Gediz/Alaşehir detachment (Cohen et al., 1995). We note that during the model evolution a younger higher-angle normal fault forms on the hanging-wall of the low-angle detachment akin to rider blocks (Reston and Ranero, 2011; Choi and Buck, 2012). This new steeply dipping splay fault merges with the major detachment along the brittle-ductile transition while its dip angle reduces through time. In EXP-1 such a fault system starts to develop 12 Myr after model initiation, (e.g., comparable to 3-4 Ma before present in central Menderes geological evolution). The timing is in accord with the onset of higher-angle normal faults and the deposition of Late Miocene-Pliocene sediments proximal to the margin of the Gediz and Büyük Menderes grabens (Sarıca, 2000; Bozkurt and Sözbilir, 2004; Kent et al., 2016).

Short wavelength (≤150 km) variations of the crustal structures across the central Menderes Massif region are shown in Figure 5b based on a regional high-resolution seismic tomography model (Fichtner et al., 2013). Notably, slow seismic velocities (Vs < 3.4 km/s, < 20 km depths, light blue region) in the upper part of the crust are observed in the center, beneath the Küçük Menderes graben to which lower crustal exhumation and detachment faulting have not been ascribed. However, on both margins of the Küçük Menderes graben, there are higher seismic velocities (3.4 ≤ Vs ≤ 3.8 km/s black and yellow regions) at shallower depths associated with the uprising of the lower crust (Fichtner et al., 2013; Çubuk-Sabuncu et al., 2017). This is consistent with findings of the EXP-1, where a twin-domes feature develops by crustal flow (Fig. 5a).

In Figure 5a, we show the predicted Moho depth variation and it is sub-horizontal beneath the central Menderes Massif region where the Moho depth ranges from 25 to 28 km. Results of EXP-1 are mainly consistent with seismologically derived Moho variations for this part of western Anatolia's extended terrane (Zhu et al., 2006; Karabulut et al., 2013; Fichtner et al., 2013). It is worth noting that Moho beneath western Anatolia shows undulated pattern along the N-S transect, however, it is locally flat underneath the Menderes Massif. (Karabulut et al., 2013). Again, the flat Moho is controlled by the ductile flow of the mid-lower crust to isostatically compensate for the differential thinning of the upper crust and variations in crustal thickness, and this process has been suggested to account for the evolution of cordilleran core complexes in the western US and the Aegean region (Klemperer et al, 1986; Gans, 1987; Block and Royden, 1990; McKenzie et al., 2000; Tirel et al., 2004).

In Figure 5c, we reconcile the exhumation history of the central Menderes Massif derived from low temperature thermochronological constraints against the predictions of EXP-1. For EXP-1, the exhumation rates are calculated (for every 1 Myr) through the trajectory of the lower crust at minimum depths with respect to the zero elevation (see yellow dots shown in the figure inset). These points on both domes correspond to the exhumed crust beneath the two detachment faults and follow approximately similar patterns and magnitudes since the beginning of the model (with uncertainties of ±0.2 km/My). Namely, there is an increase in the exhumation rate between 15- 10 Ma, from 1 km/My to ~1.7 km/My and then it starts to decrease to ~0.6 km/My by 5 Ma. The change after 5-0 is relatively minor and it is > 1.0 km/My. Two-stage evolution predicted by the EXP-1 is in accord with the findings of Wölfler et al. (2017) for the exhumation of the footwall rocks of the Büyük Menderes detachment. According to the authors, the exhumation rate is ~0.9 km/My during middle Miocene and it decreases to ~0.43 km/My during the Miocene to Pliocene. For the footwall rocks of the Gediz detachment, Buscher et al. (2013) estimates that the rate of exhumation is 0.6-2 km/My and such variation is in the range of our model calculations.

We note that our models do not account for all geological processes involved in the late Cenozoic evolution of western Anatolia, such as surface erosion, partial melting of the lower crust, and plate rotation. For instance, based on thermochronological data and available erosion rates, Buscher et al. (2013) suggest that surface erosion also controls the exhumation and the landscape evolution of the Bozdağ block, an exhumed range displaced by the Gediz detachment fault. Recent cosmogenic nuclides study by Heineke et al. (2019a) suggests that the erosion plays relatively minor role for rock exhumation along Gediz and Büyük Menderes (~10% for the Gediz), however, it is more effective along both sides of the Küçük Menderes graben (~50% rock exhumation). In our models, the flow of the lower crust, hence, the double dome formation (footwall exhumation) is mainly controlled by isostasy rather than lower crustal buoyancy. Rey et al. (2011) presents numerical experiments where partially molten lower crust becomes buoyant and double dome formation occurs with a small magnitude of extensional strain rate. Buoyancy driven dome formation may apply some of the cordilleran type metamorphic core complexes in North America where regional extension is decoupled from lower crustal flow under pure shear extension (Levy et al., 2023; Zuza and Cao, 2022). Although it may exert control to some degree, the purely buoyancy driven crustal upwelling model does not apply to the Aegean-west Anatolia, because the lithospheric extension in the region has been active since ~25 Ma (Seyitoğlu et al 1996; Bozkurt and Sözbilir, 2004; Çemen et al., 2006; Gessner et al., 2013; Roche et al., 2018; Heineke et al 2019a,b). In brief, the isostatic footwall uplift by stretching is the primary mechanism driving twin dome formation in the area as predicted by our experiments in this work. Further, van Hinsbergen et al. (2010a) provides paleomagnetic evidence for a vertical axis rotation of ~25°-30° in the central Menderes region with respect to the northern and southern part of the Menderes Massif to account for crustal exhumation and detachment faulting since ~16 Ma. This would certainly be the focus of future 3-D modeling studies. Overall, the results of our reference experiment (EXP-1) are consistent with the present-day flat Moho inferred from seismological studies as well as with the twin-dome type exhumed massifs along bivergent detachment faults as documented by geological studies.

The exhumation of metamorphic core complexes and associated low-angle normal faults are common features of large-magnitude extensional provinces across Cordilleran and Mediterranean regions that develop in the terminal phase or in the aftermath of orogenesis. The results of this work in which progressive rotation of the high-angle normal faults to low-angle orientations as well as the exhumation of symmetrical (twin) domes are accommodated by the distributed flow of the lower crust may explain how post-orogenic lithospheric extension operates around the globe.

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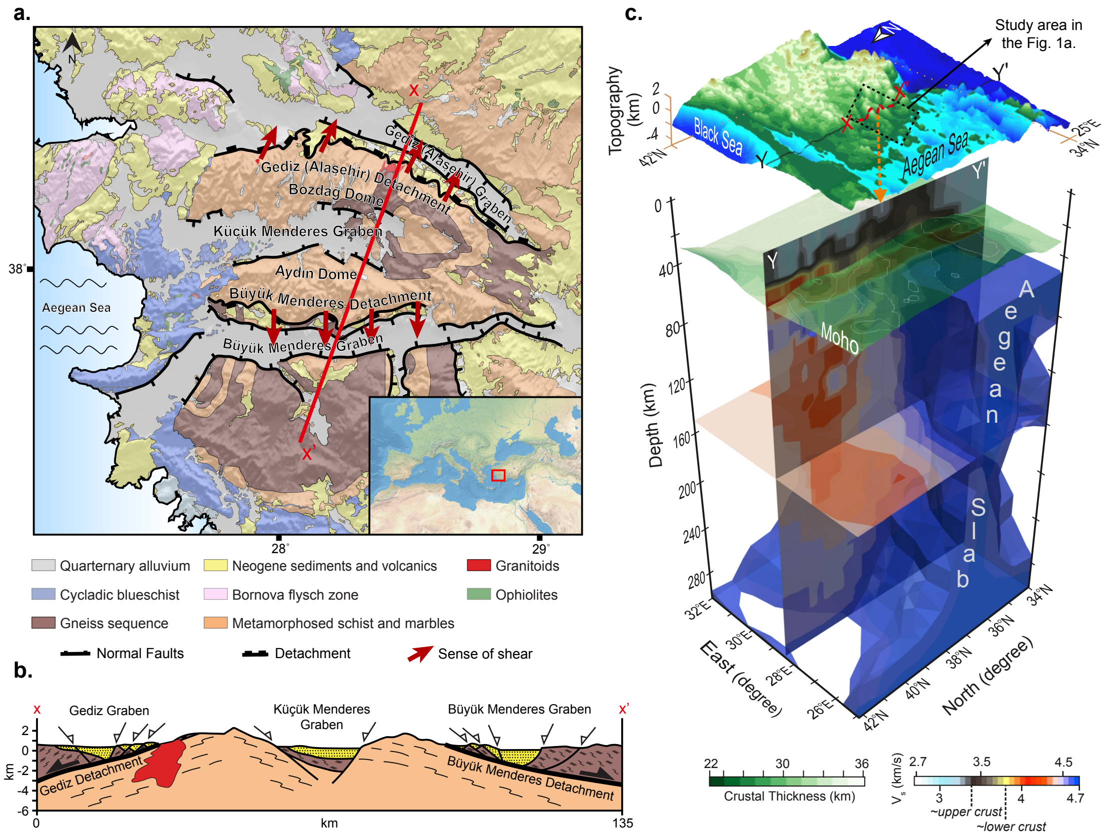

Figure 1a. Geological map of the central part of western Anatolia extensional region (Menderes Massif) that shows the main geological features discussed in this work. Metamorphosed schist and marble are footwall rocks and represented by greenschist to amphibolite facies metamorphics and the hangingwall rocks are described by high-grade metamorphics (Gneissic core). The map is based on the geological map of Türkiye (MTA, 2002), including the differentiation of metamorphic units, according to Lips et al (2001) and Okay (2001) **b.** Simplified NNE-SSW cross section across the central Menderes Massif that shows bivergent detachment faults and symmetrically arranged exhumed massifs **c.** 3-D view of a seismic tomography model based on isotropic S velocities (Vs) along the across the crust as well as the upper mantle beneath western Anatolia (derived from Fichtner et al., 2013).

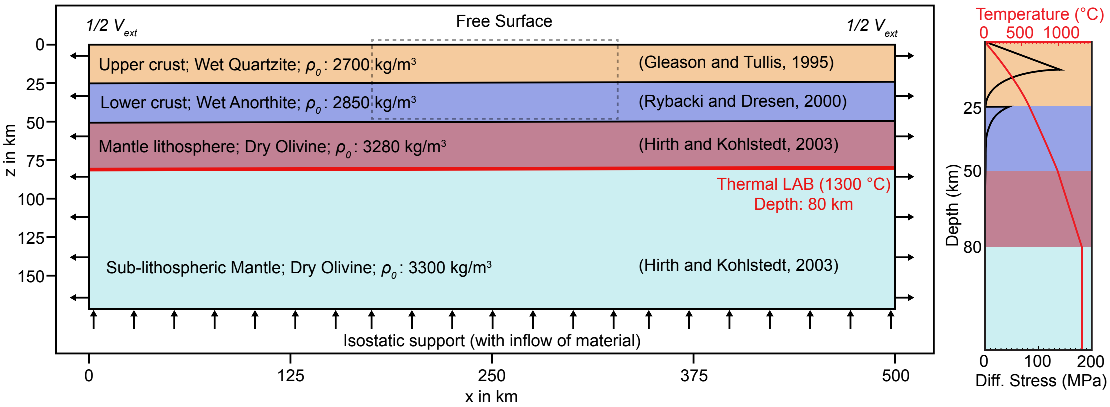

Figure 2. Illustration of the model geometry, set-up, material properties, and density fields for the experiments. The corresponding strength profile for the initial conditions and a strain rate of $1x10^{-15}$ is shown on the right. The dashed rectangle outlines the area shown in Figures 3 and 4. The density values given are the reference densities for each material.

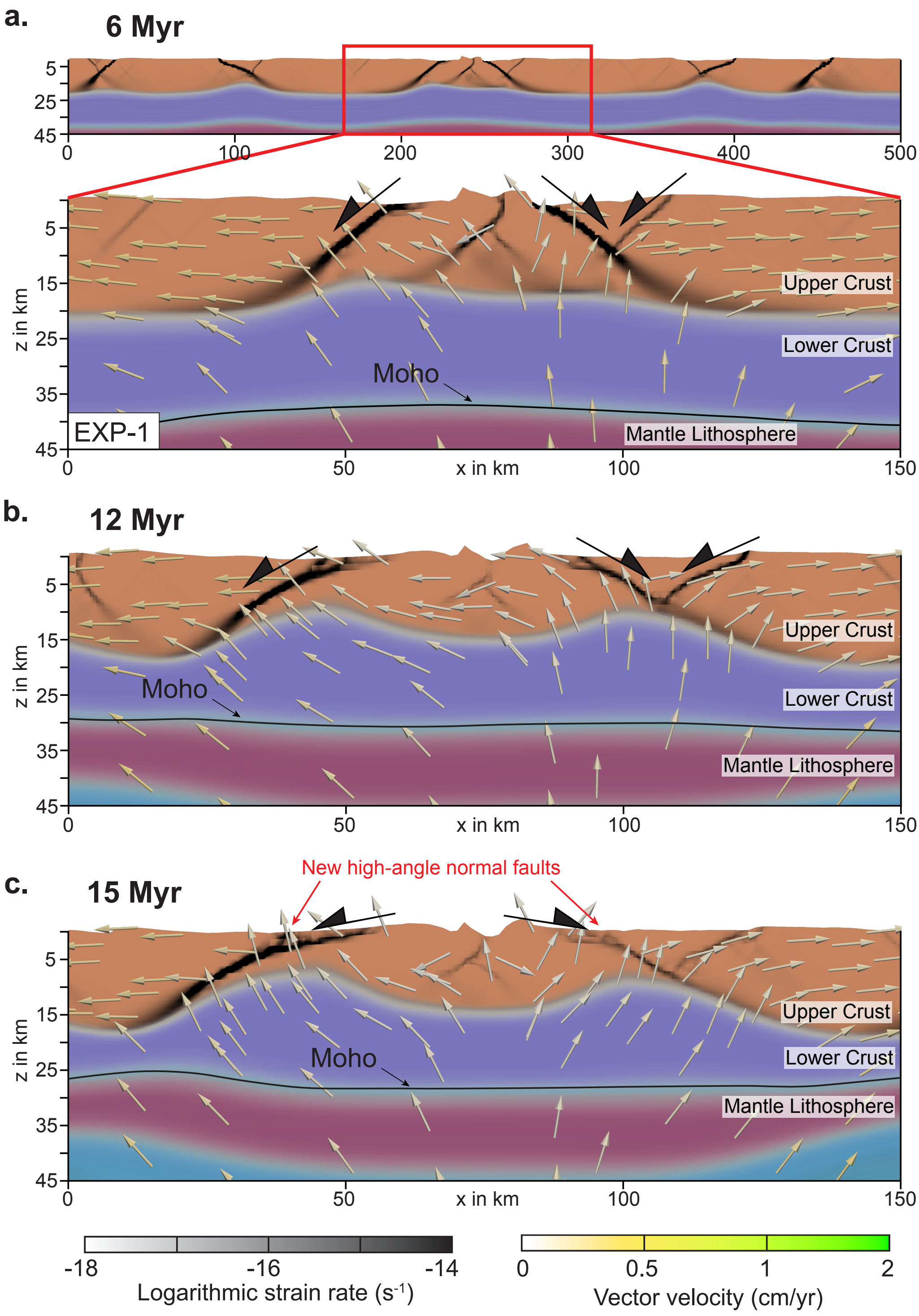

Figure 3. Geodynamic evolution of our reference experiment (EXP-1) that approximates the last 15 Myr geological evolution of the central Menderes Massif of western Anatolia. Note that central part of the model domain is shown (150 km wide). Strain localization is interpreted in accordance with shear zone development where footwall rotation is associated with detachment fault evolution. The vectors show flow field over the experiment.

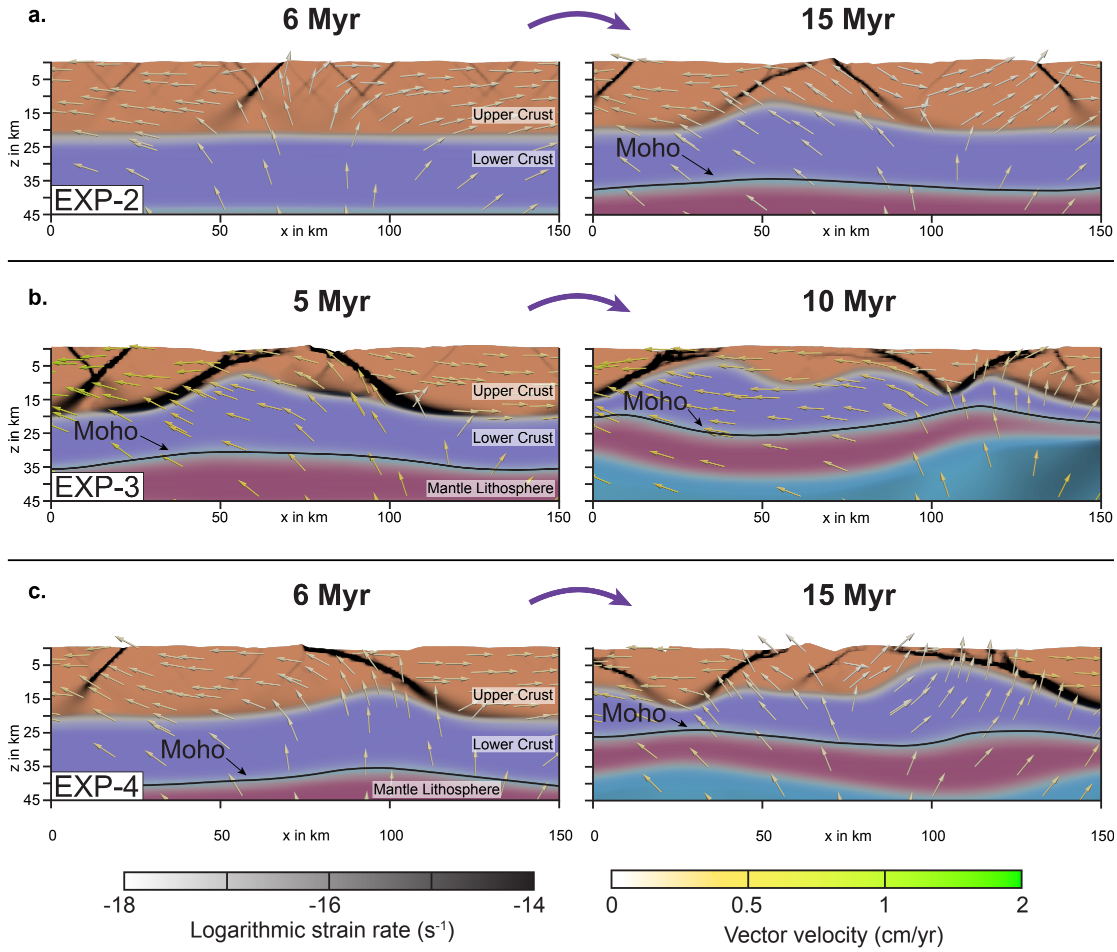

Figure 4. Geodynamic evolution of EXP- 2, 3, and 4. All model parameters are kept constant with respect to the reference experiment (EXP-1) except, **a.** in EXP-2 the extension rate (1 cm/yr), **b.** in EXP-3 the extension rate (4 cm/yr), and **c.** in EXP-4 the upper crustal strength is reduced by decreasing the frictional weakening to 0.05.

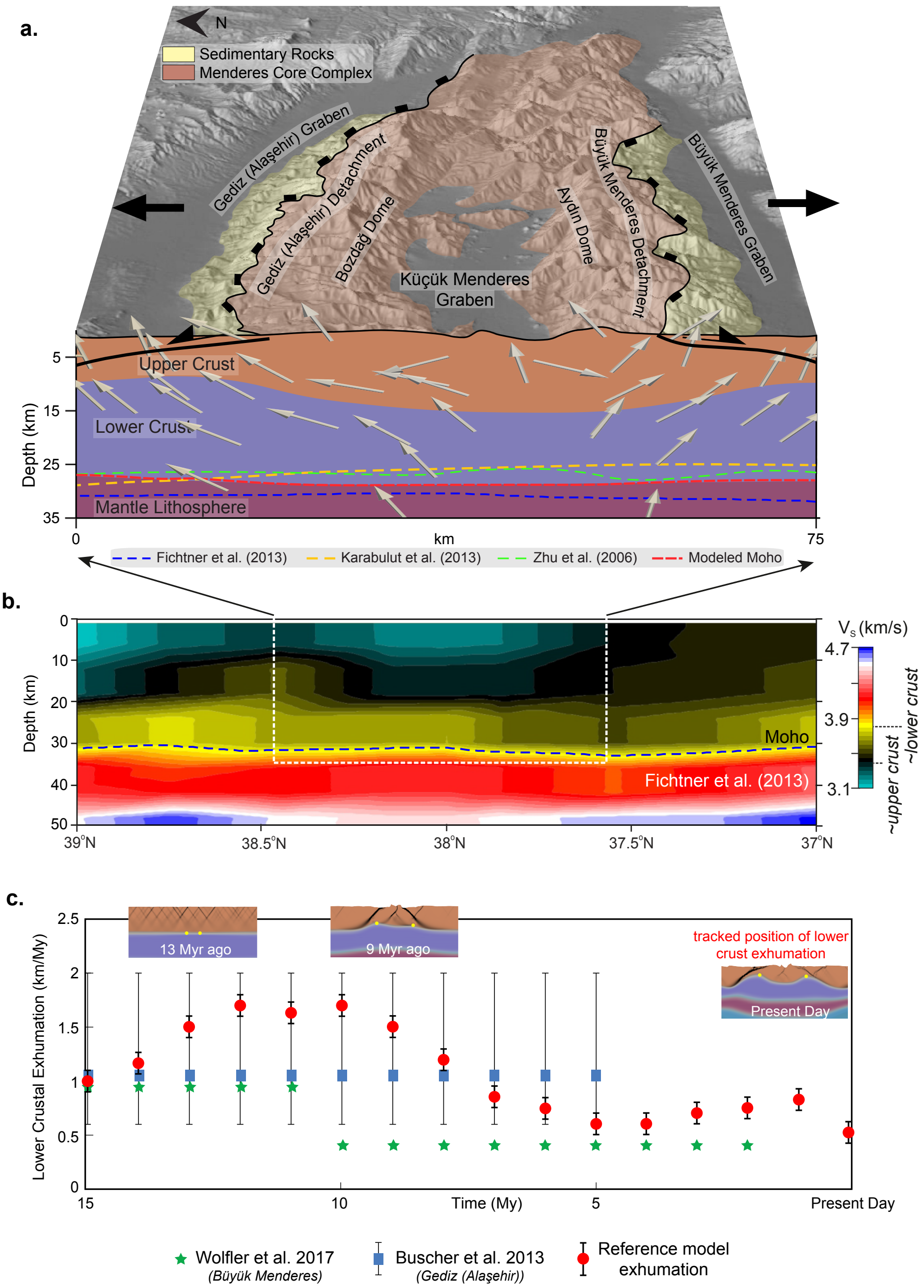

Figure 5. Reconciling predictions of EXP-1 with geological, geophysical and thermochronological studies in the central Menderes massif of western Anatolia. **a.** NNE-SSW profile that shows the map projection of bivergent Gediz and Büyük detachment faults and their approximation to the model result. Note the agreement between modeled and observed Moho (see text for references). **b.** 2-D seismic (V_s) variations across the same profile of western Anatolia region to a depth of 50 km. Higher velocities in the shallow part of the crust (dark grey-black) are interpreted as the ascending of the normally associated lower crustal material. **c.** A diagram that shows the exhumation rate variation of the model and based on studies from low temperature thermochronology (Buscher et al., 2013 and Wölfler et al., 2017).

Parameter	Units	Upper Crust	Lower Crust	Lithospheric Mantle	Sub-Lithospheric Mantle
Reference Density	kgm^{-3}	2700	2850	3280	3300
Thermal expansivity	$10^{-5}K^{-1}$	2.7	2.7	3.0	3.0
Thermal diffusivity	$10^{-7}m^2s^{-1}$	7.7160	7.3099	8.3841	8.3333
Radiogenic heat production	μWm^3	1.5	0.2	-	-
Friction Weakening		0.1 and 0.05*	0.05	0.05	0.05
Cohesion strain weakening		0.05	0.05	0.05	0.05
Rheology		*Wet Quartzite*	*Wet Anorthite*	*Dry Olivine*	*Wet Olivine*
Pre-exponential constant for diffusion creep	$Pa^{-1}s^{-1}$	5.97x10^{-19}	2.99x10^{-25}	2.25x10^{-9}	2.25x10^{-9}
Grain size exponent		2.0	3.0	0	0
Activation energy for diffusion creep	$kj\ mol^{-1}$	223	159	375	375
Activation volume for diffusion creep	$cm^3\ mol^{-1}$	0	38.0	6.0	6.0
Pre-exponential constant for dislocation creep	$Pa^{-n}s^{-1}$	8.57x10^{-28}	7.13x10^{-18}	6.52x10^{-16}	6.52x10^{-16}
Power law exponent for dislocation creep		4.0	3.0	3.5	3.5
Activation energy for dislocation creep	$kj\ mol^{-1}$	223	345	530	530
Activation volume for dislocation creep	$cm^3\ mol^{-1}$	18.0	38.0	18.0	18.0

Table 1: Model parameters for reference experiment. *0.05 is used in experiment 4. Initial parameters of temperature=293 K, adiabatic surface temperature=1557 K, heat capacity=1200 Jkg−1K−1, internal friction angle=20◦, cohesion=20 MPa, and grain size=1 mm are defined in all layers.