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12

Early-to-mid Pleistocene Tectonic Transition Across the Eastern Mediterranean Influences the Course of Human History

Uri Schattner
Dr. Moses Straus Dept. of Marine Geosciences, Charney School of Marine Sciences, University of Haifa, Israel

1. Introduction

1.1 Out of Africa through the Levantine corridor

The widely accepted 'Out-of-Africa' hypothesis considers northeastern Africa as the cradle of humankind, based on early archaeological evidence, (e.g., Bar-Yosef and Belfer-Cohen, 2001; Templeton, 2002). The earliest evidence for hominin activity was found in Kenya, Ethiopia and Tanzania. Amongst these findings were remnants of Oldowan and Acheulian stone tools, remnants of animals and remains of hominins in sites dated to the Pliocene (>1.8 Ma)(e.g., Semaw, 2000).

Hominin remains outside Africa are dated to the Pleistocene and Holocene periods, from ~1.8 Ma to present. Remains were found in sites spanning from northeast Africa to the Far East (Carto et al., 2009; Stringer, 2000). The Levantine corridor (Fig. 1), is a narrow land bridge connecting Africa with eastern Europe, central Asia, India and the Far East. It extends along the land area of the Sinai plate, between the Dead Sea fault and the Mediterranean continental margin. This corridor is considered to be one of the main pathways of hominin dispersal due to the discovery of some of the oldest prehistoric remains outside of Africa (Bar-Yosef and Belfer-Cohen, 2001).

Erk-el-Ahmar (1.96-1.78 Ma) is located in the central part of the Dead Sea Fault and reflects the first of three recognized dispersal pulses (Bar-Yosef and Belfer-Cohen, 2001). At this site, the earliest hominin related flint artifacts were found outside of Africa (Braun et al., 1991; Horowitz, 2001) (Fig. 2). The next pulse is indicated in Ubediya, ~1.4 Ma (Tchernov, 1987; Klein, 1989) where a rich complex of hominin and fauna remains was found. This complex represents numerous returns to the same location close to a lake. Acheulian artifacts of Ubediya are very similar to those found in contemporary assemblages of Upper Bed II of Olduvai Gorge (Bar-Yosef and Belfer-Cohen, 2001). The third pulse is represented by the mid-Acheulian site of Gesher Benot-Ya'aqov (Goren-Inbar et al., 2000) (0.78 Ma) and Ruhama (Laukhin et al., 2001) (0.99-0.85 Ma).

Dispersal routes of hominins to the rest of the world have been hypothesized based on archaeological evidence (Bar-Yosef and Belfer-Cohen, 2001)(Fig. 1). Reproduction of a species in a closed area limits the population growth and hence, given the right environment, this species will tend to quickly expand into the newly accessible area. Therefore, both physiographic modifications and their timing are crucial for the onset of dispersal.

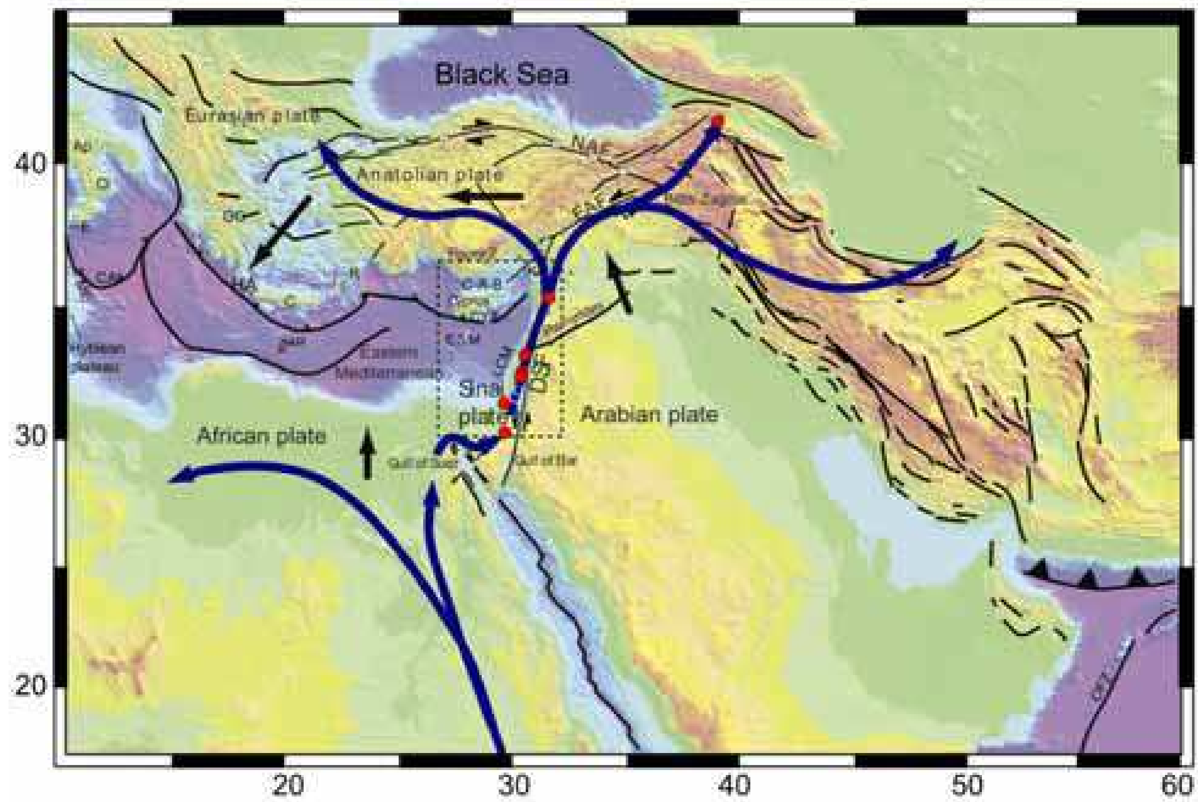

Fig. 1. Geodynamic settings of the eastern Mediterranean basin (after Reilinger et al., 2006) showing the differential convergence (subduction/collision) between Africa–Sinai–Arabia and Eurasia–Anatolia. Large black arrows show the direction of the general plate motion; small black arrows indicate relative plate motion; thin black lines show the major faults and plate boundaries. Black triangles symbol active subduction zones (overriding plate). Blue arrows indicate major pathways of hominin dispersal out of Africa; red dots denote the location of major Hominin sites along the Levantine Corridor (Bar-Yosef and Belfer-Cohen, 2001) (1.81–1.7 Ma). Ap — Apennine mountains; C — Crete; CA — Calabrian arc; C–A-B — Cilicia–Adana basins; Cl — Calabria; CYA — Cyprus Arc; DSF – Dead Sea fault; E — Etna volcano; EAF — East Anatolian fault; ESM — Eratosthenes Seamount; GC — Gulf of Corinth; HA — Hellenic arc; IK — Iskanderun bay; LC – Levantine Corridor; LCM – Levant continental margin; ME — Malta escarpment; N&M — Napoli and Milano mud volcanoes; NAF — North Anatolian fault; OFZ – Owen Fracture Zone; R — Rhode.Numbers along the NAF and DSF represent the following basins that developed as pull-aparts and were crossed by a diagonal through-going fault: (1) Erzincan, (2) Suşehri–Gölova, (3) Erbaa–Niksar, (4) Bolu–Yeniçağa, (5) Izmit–Sapanca, (6) Marmara Sea and (7) Hula. Inset: regional tectonic settings. Black rectangle shows the location of Fig. 2

This chapter highlights the temporal coincidence of the first out of Africa pulse with the initiation of major physiographic modifications of the mild topography of the Levantine corridor and the entire eastern Mediterranean basin. The chapter summarizes the works of Ben-Avraham et al. (2005), Schattner and Weinberger (2008), Schattner and Lazar (2009), Lazar and Schattner (2010), and Schattner (2010). It pieces together a wealth of accumulated knowledge published in the literature into a coherent tectonic reconstruction of the eastern Mediterranean convergence system during the beginning of the Pleistocene. The importance of prehistoric dates in this context is whether they are dated to pre, syn or post tectonic transition.

1.2 Eastern Mediterranean geodynamics

The eastern Mediterranean basin is a fossil remnant of the Meso and Neo-Tethys oceans (Fig. 1). The basin started to develop during the late Paleozoic when Eratosthenes block, amongst other platelets, drifted away from the Gondwana super-continent (Garfunkel, 1998; Schattner and Ben-Avraham, 2007). During the Paleogene–Neogene, fore-arc extension and subduction hinge rollback of the northern Afro-Arabian plate dominated the development of the eastern Mediterranean basin (Robertson, 1998). Towards the terminal stages of the Neo-Tethys, during the Oligocene to the present, the entire northern flank of the Afro-Arabian plate progressively subducted northward beneath Eurasia.

Until early–middle Eocene a single active subduction zone occupied the northern margin of the eastern Mediterranean basin, in southeastern Turkey, where arc volcanism continued into the early Neogene (Robertson, 1998). From the Neogene, the northwards subduction and hinge rollback continued with varying rates across the Alpine-Himalayan orogen (Ben-Avraham and Nur, 1976; Faccenna et al., 2006). While much of the subducted slab was consumed, some seafloor features were accreted onto the overriding plate (Şengör et al., 2003). During the Miocene the subduction front jumped to its present location south of Cyprus (Kempler and Ben-Avraham, 1987; Robertson, 1998).

At the forefront of subduction, slab break-off developed along the northern underthrusting edge of the Arabian plate (Faccenna et al., 2006). The break-off was suggested to propagate from the Owen fracture zone westward towards the northern edge of the Dead Sea fault, concurrent with increasing northwards indentation of Eurasia by Arabia (Fig. 1). The remaining of Arabian oceanic crust gradually collided and accreted to form the Bitlis–Zagros suture by the late Miocene to early Pliocene (Robertson, 1998; Faccenna et al., 2006).

In contrast, west of the Dead Sea fault, subduction of the eastern Mediterranean basin under the Cyprus Arc persisted without prominent interruptions until the early Pleistocene (Kempler, 1998; Robertson, 1998). Increased slab pull towards the north resulted in accelerated slab retreat and arc propagation southwards (Faccenna et al., 2006). Further to the south the continental margins of the Sinai and African plates remained passive throughout the Neogene.

2. Early-to-mid Pleistocene tectonic transition across the eastern Mediterranean

A major kinematic transition occurred across the eastern Mediterranean region during the early Pleistocene - the early-to-mid Pleistocene tectonic transition (Schattner, 2010). Approximately synchronous structural modifications were recorded along plate boundaries as well as within the plates are integrated below and Fig. 3.

2.1 Arabian plate and the Dead Sea fault

Eurasia indentation by the underthrusting northern Arabia resulted in the uplift of the Zagros belt (McQuarrie et al., 2003; Faccenna et al., 2006; Reilinger et al., 2006) and activation of the East Anatolian fault since the Pliocene (Fig. 1; Westaway and Arger, 2001). Along the River Euphrates in southeast Turkey, northern Syria and western Iraq uplift of terrace deposits increased in the late early Pleistocene (Demir et al., 2007). Further south, branches of the Dead Sea fault ruptured the Palmyrides and the stable part of the Arabian plate from the early Pleistocene (Rukieh et al., 2005). The regional stress field shifted from N to NW striking compression as evident by structural variations along the northern border of Arabia during the early Pleistocene (Zanchi et al., 2002).

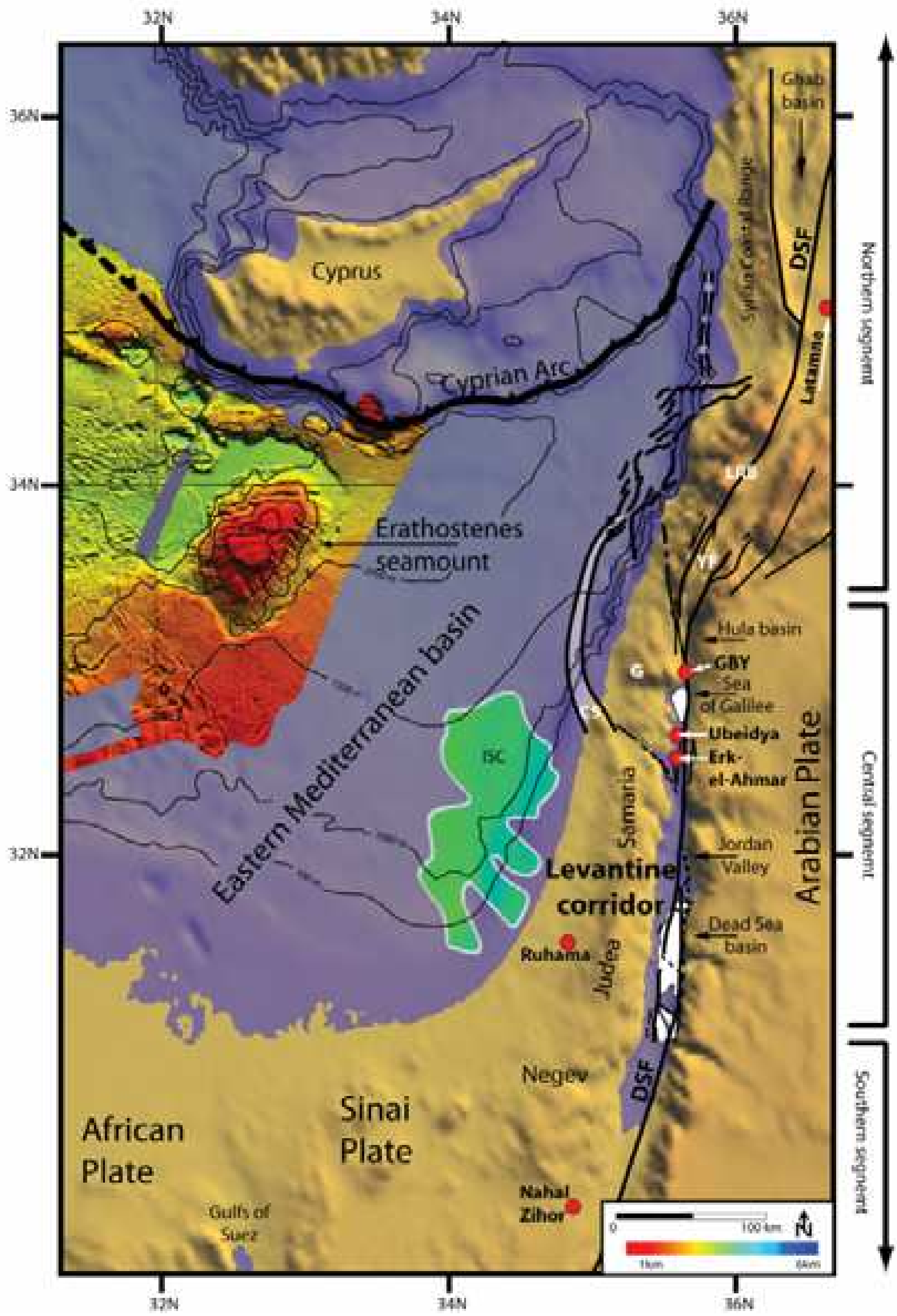

Fig. 2. Relief map of the eastern Mediterranean and its bathymetry (thin black lines) highlighting the bathymetry of Eratosthenes Seamount (Medimap Group et al., 2005; J.K. Hall, personal communication); main Plio-Pleistocene faulting along the Levant margins (Schattner et al., 2006 and references therein); and location of the Israel slump complex — ISC (Frey Martinez et al., 2005). Red dots denote the location of major Hominin sites along the Levantine Corridor mentioned in the text. DSF — Dead Sea fault; LRB — Lebanese restraining bend; YF — Yammunneh fault; KB — Korazim block; G — Galilee; C — Carmel fault; BTT — Beirut Tripoli thrust; Tr — Trodos massif; Kr — Kyrenia range

Increasing convergence between the Sinai and Arabian plates along the central and northern segments of the Dead Sea fault (Gomez et al., 2007) during the early Pleistocene, resulted in synchronous structural modifications along the Dead Sea fault axis. Along the northern Dead Sea fault, the Ghab basin subsided continuously during the Plio-Pleistocene (Brew, 2001). During the late early Pleistocene the Jisr ash Shugshur basalts (1.3–1.1 Ma; Sharkov et al., 1994) flowed to the northern part of the basin, covering a possible northern transverse faulting (Kopp et al., 1999). A notable phase of transpressive motion, folding and uplift deformed the Syrian coastal ranges (Gomez et al., 2006) and uplifted of the Nahr El-Kabir at the Mediterranean shore of NW Syria during the Pleistocene (Hardenberg and Robertson, 2007); uplifted the Lebanese restraining bend (Dubertret, 1955; Butler et al., 1998; Walley, 1998; Griffiths et al., 2000; Tapponnier et al., 2004; Elias, 2006); and branching developed along the Dead Sea fault Yammunneh main fault (Gomez et al., 2007). However, timing of initiation of the transpressive phase in Lebanon and Syria is often reported to the entire Pliocene – Pleistocene period.

Further south, basaltic units dated from the early Pleistocene onwards are found exclusively east of the Dead Sea fault axis (Weinstein et al., 2006). One of these units, the 1.5-0.5 Ma old Hazbani basalt (Sneh and Weinberger, 2003) flowed southward from its source in southern Lebanon along the Dead Sea fault axis towards the subsiding Hula basin (Heimann, 1990). Similar to the 1.3–1.1 Ma Jisr ash Shugshur basalts, the Hazbani basalt also covered the northern part of the basin, precluding a possible transverse fault from being identified at the surface (Sneh and Weinberger, 2006). At that time the Hula basin ceased to develop as a pull apart, when a diagonal through-going strike-slip fault propagated between its SE and NW corners. The trajectory of this fault was found to be parallel to the present-day motion along the Dead Sea fault (Schattner and Weinberger, 2008). In temporal and structural coincidence with the development of the Hula diagonal fault, block rotation initiated along the Korazim block (Heimann and Ron, 1993); rapid subsidence of the Sea of Galilee basin was accompanied by depocenter migration to the northeast (Hurwitz et al., 2002); two NE-striking anticlines developed immediately south of the Sea of Galilee (Rotstein et al., 1992; Zurieli, 2002); subsidence of two basins in the southern Jordan valley was accompanied by bending of overlaying monoclines (Lazar et al., 2006). At the same time (early Pleistocene) accelerated subsidence and extension prevailed across the southern and central Dead Sea fault segments (ten Brink and Ben-Avraham, 1989; Ben-Avraham et al., 2005). The Dead Sea basin subsided rapidly accumulating ~4 km of sediments that facilitated salt removal into prominent diapirs (Lisan and Sedom diapirs; Al-zoubi and ten Brink, 2001; Horowitz, 2001; Larsen et al., 2002; Ben-Avraham and Lazar, 2006). In addition, the depocenter of the southern Dead Sea basin migrated southeastwards while the largest vertical displacement occurred across its southern boundary. Further south a general tendency of fault localization towards the axis of the Dead Sea fault is recorded along the margins of the Gulf of Elat (Aqaba; Marco, 2007). This differential lateral motion along the entire Dead Sea fault, extensional in the south as opposed to contractional in the north, result from a change in relative plate motion which developed during the early Pleistocene (Schattner and Weinberger, 2008; Weinberger et al., 2009).

2.2 Sinai plate and the easternmost Mediterranean margins

The northward increase in contractional strike-slip motion across the Dead Sea fault (Gomez et al., 2007; Schattner and Weinberger, 2008) induced extensive deformation across the Sinai plate, between the Dead Sea fault and the eastern Mediterranean continental margin during the early Pleistocene. Folding, faulting, uplift and tilting of the Syrian coastal ranges (Gomez et al., 2006), the Lebanese restraining bend (Walley, 1998), and the northern Galilee (Matmon et al., 2003), extended to the edge of the continental margin where shelf marginal wedges were tilted basinward (Ben-Avraham et al., 2006; Schattner et al., 2006). A 40-60 km wide N-trending arch developed during the early Pleistocene across the Sinai plate in the northern Galilee (Matmon et al., 1999, 2003), Samaria and Judea regions (Wdowinski and Zilberman, 1996, 1997) and the Negev, which as a result was uplifted and tilted (Ginat, 1997; Ginat et al. 1998; Avni, 1998; Avni et al., 2000).

As a consequence of the topographic uplift the catchment area of rivers from the Arabian plate to the Mediterranean Sea was severely shortened by a new water divide that developed close to the shore. At the same time the anomalously high subsidence rate of the Levant continental shelf which prevailed during the Pliocene diminished in the Pleistocene when the dominant landward aggradation shifted to basinward progradation (Ben-Gai et al., 2005). At the continental margin the Oligo-Miocene Jaffa basin became completely buried under a thick sedimentary cover (Gvirtzman et al., 2008). Marine channels crossing the continental shelf were gradually starved and buried by Nile derived sediments rather from onland. Seismic data reveals a succession of incised channels that drained land area during its tectonic uplift however today has no trace on the shelf seabed.

Offshore, several fault systems re-activated the northern Levant continental margin during the Pleistocene (Schattner and Ben-Avraham, 2007) from the marine extension of the Carmel fault (Schattner et al., 2006), along the Lebanese continental slope (Daëron et al., 2001, 2004; Carton et al., 2007, 2009; Elias et al., 2007), and were traced to a possible juvenile triple junction with the easternmost part of the Cyprus Arc (Butler et al., 1998; Schattner et al., 2006). These along-margin fault systems originate as westward branches of the Dead Sea fault, which dissect the uplifted folds of the Sinai plate. Over 15 slump bodies, 400 km^3 in volume, dated to the early Pleistocene, have been identified off the southern Levant margin (Frey Martinez et al., 2005).

Along the north African margin a well-documented NW to N shortening event occurred during the early Pleistocene (Guiraud, 1990). E-trending folds and reverse faults developed, associated with NW–SE dextral and NNE–SSW sinistral strike-slip faults. Major intra-plate fault zones were rejuvenated and regional uplifts intensified. However, with time, the intensity of shortening strongly decreased (Guiraud et al., 2005). During the Pleistocene the extensional axis direction across the Gulf of Suez rotated counterclockwise from NE to N (Bosworth and Taviani, 1996).

2.3 Cyprus and Eratosthenes Seamount

The progressive subduction of the Sinai plate beneath the Cyprus Arc was significantly disrupted during the late Pliocene–early Pleistocene with the initiation of the Eratosthenes Seamount-Cyprus Arc collision (Kempler, 1998). On the verge of subduction, tectonic subsidence of Eratosthenes accelerated and the Seamount reached its current vertical position (Robertson, 1998). Directly overlaying the top Messinian unconformity, early Pliocene succession reveals bathyal pelagic microfossils (Robertson, 1998). These deep-marine sediments uncovered on the northern flank of the Seamount (sites 965-967) were interpreted as indicators for a rapid subsidence of the Seamount (Ibid.). This rapid submergence at this stage could have also been greatly affected by the accelerated post Messinian sea level rise (e.g., Meijer and Krijgsman, 2005). However, the late Pliocene benthic foraminifera covered by Nile derived sediments indicate further tectonic deepening of the seamount (site 967, Robertson, 1998), this time without accompanying sea level rise.

Deformation pattern of the Seamount changes considerably from north southwards. While Pleistocene compressional folding deform the base of the northern lower slope (site 967), the central part of the Seamount's plateau experiences flexural loading and faulting (Limonov et al., 1994; Robertson et al., 1995). The Seamount and its basement were faulted by a low angle approximately E-W trending detachment zone which tilted its flanks to the north and south. Deformation of the Seamount resulted from crustal flexure, induced by southward overthrusting of the Cyprus active margin. Large scale catastrophic mass movements developed on the northern and southern slopes of the Seamount. Since identification of these slides is limited to the lines covered by seismic data it is assumed that mass movement did not occur exclusively to the north and south. The present-day bathymetry of the Seamount is faulted and scarred in the north relative to the much smoother southern slopes.

In a tomography study of the eastern Mediterranean Faccenna et al. (2006) image three different states of convergence. Beneath the Arabia-Eurasia collision zone tomography does not show any clear continuation of the Arabian slab northwards. Further to the west across the Cyprus Arc where incipient collision takes place, seismic velocities indicate a ~120 km gap in the descending Sinai slab. In their westernmost profile Faccenna et al. (2006) show that a continuous African slab is being consumed beneath the Hellenic Arc, where subduction still prevails. The relatively narrow gap in the northern Sinai slab suggests a younger initiation of the Seamount-Arc collision than the slab tear-off north of Arabia. This westwards tear propagation is in agreement with the nature of propagation described along other convergence margins of the Mediterranean region (e.g., Wortel and Spakman, 2000). While Tomography cannot infer directly to the state of convergence it strongly suggests a linkage between the continuity of the descending slab and the degree of collision at the upper lithosphere.

Convergence across the central Cyprus Arc shifted from subduction to collision (Salamon et al., 2003; Hall et al., 2010) while its eastern part exhibits strike-slip motion (Ben-Avraham et al., 1995). Uplift of the island of Cyprus which persisted since the late Miocene accelerated during the early Pleistocene when the island completely emerged from the sea (Kempler, 1998; Harrison et al., 2004). The Troodos Massif and the Kyrenia Range were uplifted in accelerated rates (Robertson and Woodcock, 1986; Robertson et al., 1995; Poole and Robertson, 1991, 1998) while NW to NE trending strike-slip faulting became dominant in northern Cyprus (Harrison et al., 2004). After the mid Pleistocene, uplift rates decreased (Poole and Robertson, 1992, 1998). While the deformations of the Cyprus region are straightforwardly associated with the collision, they consist a part of the synchronous structural deformations across the entire eastern Mediterranean region mentioned above – together comprising the early-to-mid Pleistocene tectonic transition (Fig. 3).

Deformations were also recorded in the marine surroundings of Cyprus, from Iskenderun through Adana-Cilicia to Antalya basins, and along the Larnaka and Latakia Ridges. In these regions, which were studied in detail in the last decade, the Pliocene-Quaternary is interpreted as a single seismic stratigraphic unit (Unit 1 in Aksu et al., 2005a; Aksu et al., 2005b; Burton-Ferguson et al., 2005; Calon et al., 2005a; Calon et al., 2005b; Hall et al., 2005a; Hall et al., 2005b; Isler et al., 2005; Aksu et al., 2009; Hall et al., 2009). In light of the findings brought here the young fill of these basins should be reexamined.

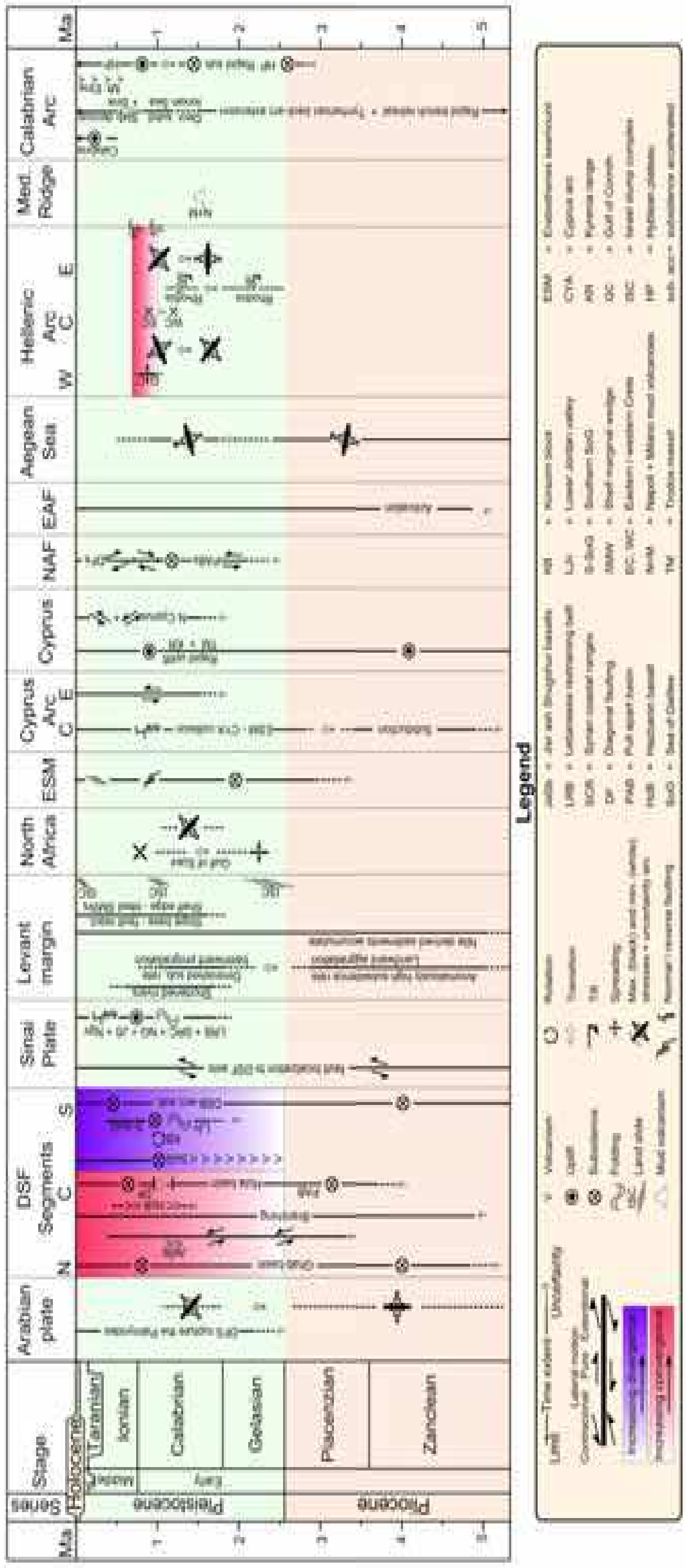

Fig. 3. Time–space diagram of the tectonic activity in the eastern Mediterranean region during the Pliocene to recent, modified after Schattner (2010). The time scale reflects the latest changes in the beginning of the Pleistocene (e.g., Mascarelli, 2009), however the data presented is based on works preceding this change

2.4 Anatolia and the Aegean sea

The early-to-mid Pleistocene tectonic transition was not limited to the easternmost part of the Mediterranean or the Levant but extended across the convergence front to the southern margin of Eurasia. Fault-kinematic analyses indicate a short-lived compressional stress regime across Anatolia and the North Anatolian fault around 0.7–1.0 Ma (mid Pleistocene; Over et al., 1997; Barka et al., 2000). As a result, a series of extensional basins located along the axis of the north Anatolian fault ceased to develop as pull-apart basins. The marginal faults of these basins (Erzincan, Suşehri–Gölova, Erbaa–Niksar, Bolu–Yeniçağa, Izmit–Sapanca and Marmara Sea basins, Fig. 1) became inactive when a through-going fault developed across each basin along the north Anatolian fault axis (Gürbüz and Gürer, 2009). This localization of motion to the axis of the main fault, in a contractional strike-slip motion, is similar to and synchronous, with the modifications along the Dead Sea fault (Marco, 2007). In particular, it is structurally similar to the through-going fault that developed at the same time diagonally across the Hula basin along the Dead Sea fault (Schattner and Weinberger, 2008). The short transpressional phase of the north Anatolian fault altered in the mid Pleistocene to transtension (Sorel et al., 1992; Bellier et al., 1997).

A marked change is observed during the same time frame in the direction of extension across the Aegean Sea (Figs. 1). NNE-SSW to N-S trending extensions which prevailed during the Pliocene shifted to NNW-SSE to N-S in the early-to-mid Pleistocene (Angelier et al., 1981; Mercier et al., 1989; Jolivet, 2001). Meanwhile the tectonic regime of the Hellenic arc changed drastically around 1 Ma (e.g., Duermeijer et al., 2000; Mantovani et al., 2002). In the western part of the arc, central and northern Greece, NE-SW tensional stresses shifted to NNW-SSE while the Gulf of Corinth underwent a rapid phase of rifting. An intense short-lived compressional phase separated the extensional periods (Sorel et al., 1992). Concurrently, at the eastern end of the Hellenic arc, N-S tension of western Turkey shifted toward NE-SW. The entire Hellenic arc was uplifted while several discontinuities developed in Crete-Rhodes region (Buttner and Kowalczyk, 1978; Armijo et al., 1992). Crustal stretching ceased in western Crete basin and initiated in the eastern Crete basin, with a roughly NW-SE extensional trend (Duermeijer et al., 2000). Rhodos, which was tilted southeastwards and submerged at depths of 500–600 m between 2.5 and 1.8 Ma, tilted to the northwest between 1.5 and 1.1 Ma. As a result some of its submerged relief re-emerged from the sea (van Hinsbergen et al., 2007).

2.5 Helenic and Calabrian arcs

At the outer circumference of the Hellenic arc, convergence rates increased from 1 to 3 cm/year (Rabaute and Chamot-Rooke, 2007). This shift probably occurred during the short but intense compressional phase of 1 to 0.7 Ma that prevailed between two extensional regimes (Sorel et al., 1992). The initiation of mud volcanism in Milano and Napoli volcanoes is dated to ~1.5 Ma (Kopf et al., 1998). The activity of both centers along the western and eastern branches of the Mediterranean Ridge was probably triggered by these changes in stress regime rather than from the on-going process of subduction. Rabaute and Chamot-Rooke (2007) suggest that most of the deformations took place along the Mediterranean ridge and the Hellenic backstop rather than along the north African margins.

Further west rapid trench retreat of the Calabrian arc (Figs. 1) and consequent Tyrrhenian back-arc extension were dominant since the Tortonian (Goes et al., 2004). However, towards the end of the early Pleistocene (~0.8 Ma), subduction rate of the Ionian oceanic lithosphere decreased considerably (Mattei et al., 2007). Back-arc opening in the southern Tyrrhenian

Sea and arc migration was suggested to die out when the arc arms ended their outward rotation (Sicily and Calabria - CW; Southern Apennines – CCW). The subducting slab began to decouple and sink into the mantle about 0.7-0.5 Ma, triggering the upward rebound of Calabria (Ibid.). This decoupling and associated uplift allowed an astanospheric sideflow which resulted in the initiation of volcanic activity of Mt. Etna at ~0.5 Ma (Gvirtzman and Nur, 1999). At the western end of the eastern Mediterranean, the late Pliocene-early Pleistocene rapid subsidence of the Hyblean Plateau (Branca et al., 2008) shifted to uplift at about 1 Ma while all its volcanic activity migrated northwards to the Catania Plain (Yellin-Dror et al., 1997).

3. Subduction to collision triggered the early-to-mid Pleistocene tectonic transition

The disruption in subduction of the Sinai plate is a relatively long-term process and not a singular moment in time. Unlike prehistoric developments, geodynamic effects on surface development often take several millions of years. However, in this case the effect was rapid (by geological terms) and somewhat catastrophic. The tectonic integration clearly shows that during the short period of the Early Pleistocene, the entire eastern Mediterranean was massively deformed and its mild regional topography was accentuated. Motion of tectonic plates are primarily driven by subducting slabs and their downwelling descent beneath overriding plates (Conrad and Lithgow-Bertelloni, 2002; Kincaid and Griffiths, 2003; Billen, 2008). When the Eratosthenes Seamount hindered the subduction of the Sinai plate a transition to collision commenced. The large dimensions and buoyant continental of the seamount (Ben-Avraham et al., 2002) (~25 km) induced a major regional disruption and uplift of the adjacent forearc. The incipient collision may have altered the lower crustal flow pattern (e.g., Westawey et al., 2009) which resulted in the regional change in topography. This major tectonic event could be considered as the immediate trigger for the synchronous structural modifications that occurred across the entire eastern Mediterranean region, constituting an Early Pleistocene tectonic transition.

Motion of tectonic plates is primarily driven by subducting slabs and their downwelling descent beneath overriding plates (Steiner and Conrad, 2007). The major driving force of this process, slab-pull, may vary depending on the stage of development of the overall subduction. Along the northern flanks of the African-Sinai-Arabian plates the stage of convergence plays a central role in the tectonic reorganization since the Neogene (Faccenna et al., 2006): Eurasia indentation by northward motion of Arabia, break-off of the northern Arabian slab beneath the Bitlis-Zagros suture, and consequent hinge rollback acceleration across the Hellenic trench. Tomography emphasizes the difference between a broken slab beneath Bitlis-Zagros (northern flank of Arabia) and a continuous Hellenic slab (northern flank of Africa; Ibid.). The transition between the two slabs was suggested to lie in the region between Rhodes and Cyprus where the slab extension to the north is not clear (Faccenna et al., 2006). These unique settings provoked the uplift of the Turkish–Iranian plateau, allowed the westward motion of the Anatolian plate and induced the formation of the north Anatolian fault. However, the occurrence of the sharp early-to-mid tectonic transition across the eastern Mediterranean still needs to be explained.

The Hellenic and Calabrian arcs propagated southwards mainly by hinge rollback of the African semi stationary plate which remained passive along its eastern Mediterranean margins (e.g., Rosenbaum and Lister, 2004; Doglioni et al., 2007). Similar relative motion could have prevailed also across the Cyprus Arc as part of the general closure of the eastern Mediterranean. However, since the late Pliocene–early Pleistocene, subduction across the Cyprus Arc was choked by the ~25 km thick and buoyant crust of the Eratosthenes Seamount. South of the Seamount, ~4 km of Nile derived sediments accumulated on the Mediterranean seafloor of the Sinai plate since the Pliocene (e.g., Segev et al., 2006). This thick load is exclusively linked with the vertical development of the Levant margin (Fig. 3 in Gvirtzman et al., 2008). Several works (e.g., Tibor et al., 1992) attributed both subsidence of the Levant margin and the uplift of Sinai plate (mainly in the Samaria and Judea) to the Nile sedimentary load. Others argue that the margin subsidence commenced prior to the Pliocene and therefore only part of it is attributed to the Nile sedimentary load (Gvirtzman et al., 2008). Either ways, Schattner (2010) assumes that this flexural load was not uni-directional but also affected regions north and west of the Nile cone. Instead, subsidence of the Seamount and its surroundings was recorded from the late Pliocene into the Pleistocene (e.g., Robertson, 1998) and no subduction-related flexural uplift was recorded in south of the Seamount. In this context the Nile load is likely to have inhibited the development of a flexural crustal fore-bulge. It is further suggested that the combined effect of Seamount buoyancy and fore-bulge depression resulted in lack of hinge roll-back and lower angle of subduction. These ultimately resulted in increased subduction friction, slab break-off and transition from subduction to collision.

The initial Eratosthenes Seamount collision with the Cyprus Arc fits into the center of the semi counterclockwise plate motion described above (black arrows on Fig. 1): Eurasia indentation by Arabia, westwards Anatolia escape and southwards Aegean propagation. However, in contrast to these long lasting motions the incipient Seamount-arc collision is constrained in time. With the transition to collision the relative convergent motion between Sinai and Anatolia increased abruptly, altering the stress regime. The immediate effect was recorded as the exhumation of Cyprus, however regional shortening was induced throughout the eastern Mediterranean, mainly radial to the collision. This convergence phase is exhibited by the folding, faulting and tilting that developed along the Levant margin (Sinai plate), and the diagonal faulting which ruptured basins along both the North Anatolian and the Dead Sea faults. On the other hand, extensional stresses were recorded in regions which are tangent to the collision, e.g., transtentional motion across the southern Dead Sea fault and the E to NE extensional event of northern Africa. In addition, cessation of subduction and slab break off also reflects reorganization of lower crustal flow (Westaway et al., 2009) and possibly the mantle lithosphere, which affect the regional isostasy distribution in the in the vicinity of the Cyprus Arc. Coupling between lower and upper crustal processes complexes the response pattern of local areas in the eastern Mediterranean, however, the identical timing points to a common trigger of the early-to-mid Pleistocene tectonic transition.

4. Conclusions

Subduction is a central geodynamic mechanism, which drives the entire plate-tectonic motion. Throughout Plio-Pleistocene time the progressive closure of the eastern Mediterranean basin continued across the Bitlis–Zagros collision zone and the Cyprus, Hellenic and Calabria arcs at variable rates. This convergence was severely interrupted during a relatively short period, late Pliocene–early Pleistocene, when subduction at the Cyprus Arc was transformed into collision due to the arrival of Eratosthenes Seamount from the south. Consequently, the entire eastern Mediterranean was massively deformed during the early-to-mid Pleistocene and its mild regional topography was accentuated. The Levantine corridor land-bridge became a more convenient gateway for hominin dispersal out of Africatowards Eurasia.

Co-occurrence of the structural modifications recorded across the entire region was synchronous rather than a cascade of events, with the exception of the somewhat younger deformations at the Calabrian arc. Timing of the deformations points to a temporal modification in the differential convergence across the boundary between Africa–Sinai–Arabia and Eurasia–Anatolia, which has been modified in the easternmost part of the basin by the Seamount–Arc collision and migrated to the western end. Timing of the transition is concurrent with the first pulse of hominin dispersal out of Africa. While the exact causes for human dispersal out of Africa are still highly debated, geodynamic forces set the stage for the first wave with the rest to follow.

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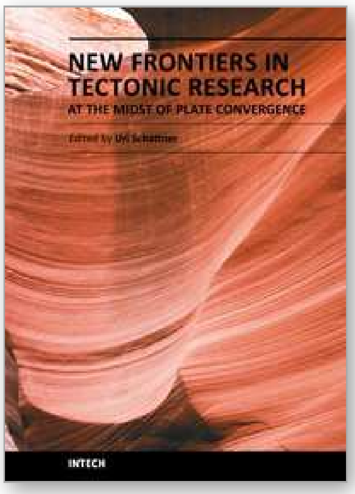

New Frontiers in Tectonic Research - At the Midst of Plate Convergence
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Ocean closure involves a variety of converging tectonic processes that reshape shrinking basins, their adjacent margins and the entire earth underneath. Following continental breakup, margin formation and sediment accumulation, tectonics normally relaxes and the margins become passive for millions of years. However, when final convergence is at the gate, the passive days of any ocean and its margins are over or soon will be. The fate of the Mediterranean and Persian Gulf is seemingly known beforehand, as they are nestled in the midst of Africa-Arabia plate convergence with Eurasia. Over millions of years through the Cenozoic era they progressively shriveled, leaving only a glimpse of the Tethys Ocean. Eventually, the basins will adhere to the Alpine-Himalaya orogen and dissipate. This book focuses on a unique stage in the ocean closure process, when significant convergence already induced major deformations, yet the inter-plate basins and margins still record the geological history.

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Phone: +86-21-62489820
Fax: +86-21-62489821

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