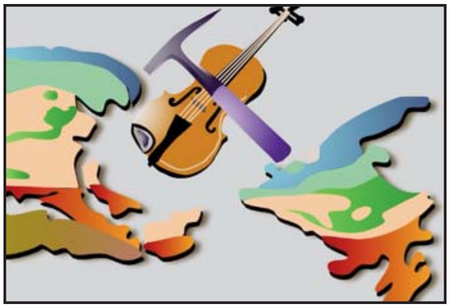
HAROLD WILLIAMS SERIES



The Ayyubid Orogen: An Ophiolite Obduction-Driven Orogen in the Late Cretaceous of the Neo-Tethyan South Margin

A.M. Celâl Şengör[1] and Joann Stock[2]

[1]*İstanbul Teknik Üniversitesi, Maden Fakültesi, Jeoloji Bölümü ve Avrasya Yerbilimleri Enstitüsü Ayazağa 34469 İstanbul, Turkey E-mail: sengor@itu.edu.tr*

[2]*California Institute of Technology Division of Geological and Planetary Sciences, MC 170-25 1200 E. California Blvd. Pasadena, CA 91125 USA*

"*The coastal lands of the southeastern Mediterranean are little-known from the viewpoint of their regional tectonics*"
Erich Krenkel 1924

SUMMARY
A minimum 5000-km long obduction-driven orogeny of medial to late Cretaceous age is located between Cyrenaica in eastern Libya and Oman. It is herein called the *Ayyubid Orogen* after the Ayyubid Empire that covered much of its territory. The Ayyubid orogen is distinct from other Alpide orogens and has two main parts: a western, mainly germanotype belt and an eastern mainly alpinotype belt. The germanotype belt formed largely as a result of an aborted obduction, whereas the alpinotype part formed as a result of successful and large-scale obduction events that choked a nascent subduction zone. The mainly germanotype part coincides with Erich Krenkel's Syrian Arc (*Syrischer Bogen*) and the alpinotype part with Ricou's Peri-Arabian Ophiolitic Crescent (*Croissant Ophiolitique péri-Arabe*). These belts formed as a consequence of the interaction of one of the now-vanished Tethyan plates and Afro-Arabia. The Africa-Eurasia relative motion has influenced the orogen's evolution, but was not the main causative agent. Similar large and complex obduction-driven orogens similar to the Ayyubids may exist along the Ordovician Newfoundland/Scotland margin of the Caledonides and along the Ordovician European margin of the Uralides.

SOMMAIRE
Entre la Cyrénaïque dans l'est de la Libye et Oman, se trouve un ceinture orogénique d'au moins 5 000 km de longueur créé par obduction au Crétacé moyen et tardif. Nous le nommons ici l'orogène ayyoubide d'après l'empire ayyoubide qui couvrait une grande partie de son territoire. L'orogène ayyoubide qui est distincte des autres orogènes alpides, comporte deux parties principales : une bande occidentale, principalement germanotype, et une bande orientale principalement alpinotype. La bande germanotype s'est formée en grande partie à la suite d'une obduction avortée, tandis que la partie alpinotype s'est formée par des épisodes d'obduction à grande échelle qui ont étranglé une zone de subduction naissante. La partie principalement germanotype coïncide avec l'arc syrien d'Erich Krenkel (*Syrischer Bogen*), alors que la partie alpinotype correspond au croissant ophiolitique péri-Arabe de Ricou (Croissant ophiolitique péri-Arabe). Ces bandes se sont formées par l'interaction de l'une des plaques de la Téthys, maintenant disparues, avec l'Afro-Arabie. Le mouvement relatif Afrique-Eurasie a influencé l'évolution de l'orogène, mais ça n'a pas été le principal facteur. Des orogènes grandes et complexes résultant de mécanismes d'obduction similaires à l'orogène Ayyoubide peuvent exister le long de la marge des Calédonides de l'Ordovicien de Terre-Neuve/Écosse et le long de la marge européenne des Uralides de l'Ordovicien.

INTRODUCTION
That the entire Eastern Mediterranean Sea, east of the Gulf of Sirte (Gulf of Sidris; ancient 'Great Sirte': *Σύρτις μεγάλη* or *Syrtis major*), is framed by late Cretaceous and Cainozoic orogenic structures has been known ever since the great German geologist Erich Krenkel coined the term *'der syrische Bogen'* (= the Syrian Arc); 'An examination of all the observations leaves no doubt that the Levantine basin is surrounded by a *unified, nowhere interrupted fold bundle*. It may have the name "the *Syrian Arc*". In Middle Syria, a new

Geoscience Canada, v. 41, http://dx.doi.org/10.12789/geocanj.2014.41.042

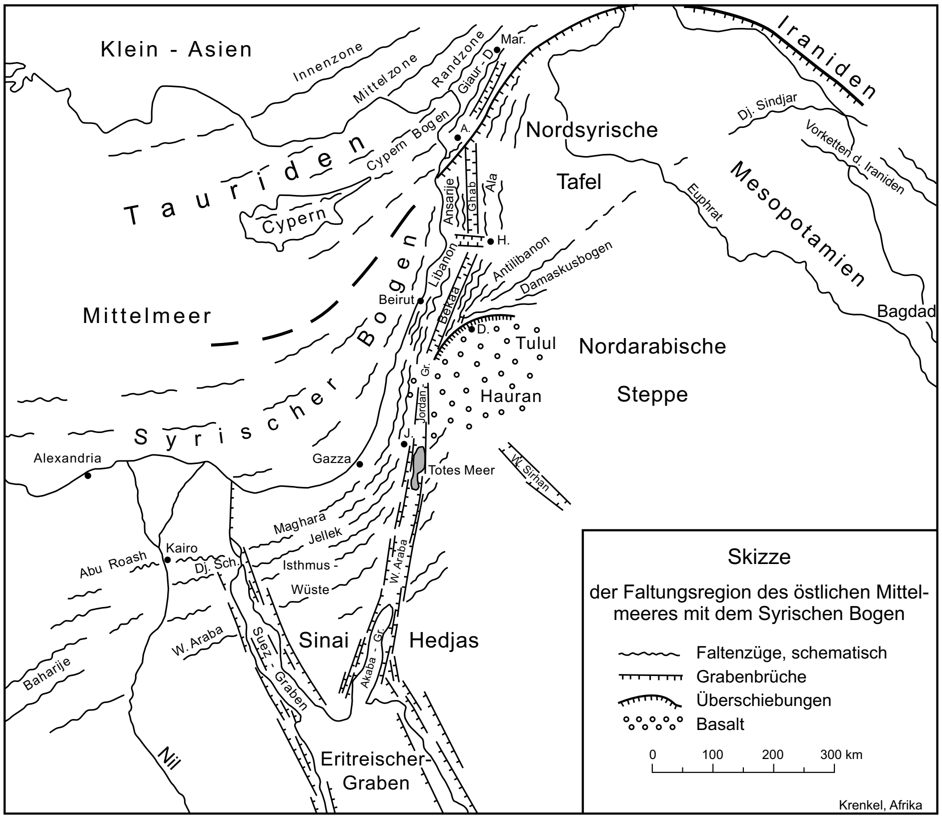


Figure 1. Krenkel's map of the Syrian Arc from his *Geologie Afrikas* (Krenkel 1925, p. 101). The Libyan and western Egyptian parts are not shown here. Translation of the legend: ' Sketch of the folded region of the Eastern Mediterranean with the Syrian Arc. Fold trains, schematic; Graben faults; Overthrusts; Basalt.' On the figure itself, *Syrischer Bogen* is Syrian Arc, *Damaskusbogen* means Damascus or Damascene Arc. *Nordsyrische Tafel* is north Syrian table or plate; *Cypernbogen* is Cyprus Arc, *Tauriden* is Taurides, separated into *Innenzone* = inner zone, *Mittelzone* = middle zone and *Randzone* = marginal zone. The Red Sea is termed *Eritreischer Graben* on this figure. The *Isthmuswüste* is the Isthmian Desert, *Totes Meer* is the Dead Sea and the *Vorketten der Iraniden* is fore-ranges of the Iranides. The abbreviations, A., D., H., J. and Mar., represent the cities of Antakya (Turkey; ancient *Antioch*), Damascus, Hama (Syria; Biblical *Hamāth*), Jerusalem (Israel and Palestine; *al Quds* in Arabic) and Kahramanmaraş (Turkey; formerly just Maraş; ancient *Germanicea Caesarea*), respectively.

branch shoots off from it: "the *Damascus Arc*" (Krenkel 1924, p. 275, emphases his; see Fig. 1) [or the *Damascene Arc*: (Krenkel 1925, p. 100)]. Krenkel considered the Syrian and the Damascene Arcs as frontal arcs (=*Vorbögen*) of the Dinaric-Tauric stem of the Mediterranean Orogen[1]. In the first volume of his great classic, *Geologie Afrikas* (Krenkel 1925, 1928, 1934, 1938), he pointed out that the 'Syrian arc extends, following the margin of the Eastern Mediterranean, in the coastal regions of Syria as part of the Mediterranean folded region. It moved towards Syrabia [Syria+Arabia in Krenkel's terminology]. Similarly, the Oman arc pushed against Syrabia as a fore-range of the Iranian arc; its continuation towards the Indus is hinted at in the submarine features of the Gulf of Oman' (Krenkel 1925, p. 38). On the next page, Krenkel pointed out that the Syrian arc also continued into Egypt and Libya: 'As it does in Syria, the Syrian Arc also goes through Egypt. Numerous fold waves can be recognised that begin as free branches[2] in the Libyan desert. Through the Suez Graben they are separated from their continuation in southern Syria' (Krenkel 1925, p. 39). Already in his 1924 paper, he continued the Syrian Arc into the Lower Cretaceous-cored east-west striking folds of the region of the salt lakes of Chott el Rharsa, Chott el Djerid and Chott el Fedjedj in central Tunisia, the 'Gafsa Ranges' of Krenkel (1943, see his fig. 7), and considered the Cyrenaican structures (which he called the 'Barka folds': Krenkel 1924, pp. 276, 278–279) as belonging to an 'inner' zone of the Syrian Arc. Krenkel held on to the interpretation of the 'Chott-folds' (i.e. his 'Gafsa Ranges') as the westernmost representatives of the Syrian Arc throughout his professional life (see Krenkel 1938, p. 1558; 1943, p. 56; 1957, p. 32), despite the fact that the onset of deformation in the Chott region is much younger, namely Miocene (see his map in Krenkel 1938, plate 44[3]). Elsewhere, he indicated that the inner zones in Cyrenaica had sunk below the sea. Krenkel correctly pointed out that the deformation in the Syrian Arc had started in the Senonian (Krenkel 1925, p. 102) and lasted into the Miocene–Pliocene.

As to the geological structure of the Syrian Arc, Krenkel (1924) pointed out that it was a folded mountain range resembling in general the Jura Mountains in Switzerland and France. He nevertheless emphasized, however, that the overall shortening seemed less both in the individual folds and in the concentration of individual fold trains in any given cross-section. The fold axes, he observed, plunged and re-emerged in short distances creating short, rounded plan views of folds. He noted that large rotund uparchings appeared between long axes of narrower elevations. Many faults broke the outer flanks of the folds and Krenkel wrote that these gave the entire belt the aspect of a *Bruchfaltengebirge* (fault-fold[4] mountain). This is significant, because the same term had long been in use to characterize the Mesozoic field of deformation in central Europe that had formed in the Alpine foreland and had been interpreted by Suess already in 1875 as the bursting of the foreland, similar to drifting and jostling pack-ice, as a consequence of the Alpine orogeny to the south (Suess 1875, p. 156). Although Stille (1925a, p. 206) denied the interpretation of the central European Mesozoic deformation as being entirely due to the Alpine orogeny to the south, he nevertheless admitted that it was affected by it. To emphasize the great difference between the deformation in the Alps, characterized by penetrative deformation creating large nappes, closely spaced fold bundles with pervasive foliations, and that in the Mesozoic and Cainozoic central Europe represented by non-penetrative blocky deformation characterized by brachyanticlines and brachysynclines which only rarely displayed foliation, Stille had created the terms 'alpinotype' for the first and 'germanotype' for the second style (Stille 1920; for a precise definition, see Stille 1940, p. 654)[5]. He later noted that the alpinotype mountain ranges always grow out of orthogeosynclines consisting of mio- and eugeosynclinal couples (we now know the eugeosynclines to have been oceans plus their Pacific-type margins and miogeosynclines to have been Atlantic-type continental margins including the shelves) and the germanotype mountains form out of parageosynclines, fault-bounded troughs of limited extent and subsidence (commonly rifts of diverse types). Stille further observed that the commonly abundant magmatism associated with the alpinotype mountain-building was always 'Pacific' type (i.e. calc-alkalic), whereas germanotype mountains displayed limited 'Atlantic' type (i.e. alkalic) magmatism.

These observations of his great countryman were not lost on Krenkel, who, in 1957, declared that the Syrian Arc was a germanotype mountain range (Krenkel 1957, p. 144). However, most germanotype mountains have 'behind' them large alpinotype mountains providing the stresses to create the germanotype block structure (e.g. the germanotype US Rockies have behind them the alpinotype US Cordillera, or the germanotype Sierras Pampeanas of Argentina have behind them the alpinotype Andes, or the germanotype Mesozoic–Cainozoic structures of central Europe have behind them the alpinotype Alps). Krenkel's Syrian Arc does not seem to have its own associated alpinotype mountain companion. Krenkel assumed that it once existed, but now lies sunken below the Eastern Mediterranean.

Blanckenhorn (1925), somewhat injured by Krenkel's blunt statement forming the motto of this paper, because he believed that Krenkel's further accusation of misinterpretation of the tectonics of the area was aimed at his work of the 'past 37 years', questioned the existence of Krenkel's *Syrischer Bogen* (for a summary of Blanckenhorn's work, see Avnimelech (1963) with an autobiography of Max Blanckenhorn and a bibliography of his writings on the Middle East). Nearly all his objections; that there was an east-west directed shortening in the Sinai area that also formed the gulfs of Suez and the Aqaba as large synclines, that the Egyptian and the Sinaitic folds cannot be grouped into a single orogenic system, that Krenkel got the 'push direction' wrong, were all shown by later research to be incorrect, however, and Krenkel's interpretations eventually prevailed. De Vaumas (1950) later followed Dubertret (1932) in correctly interpreting Krenkel's fault-folds of the *Syrischer Bogen* as *plis de fond* (i.e. 'basement folds'[6]) in Argand's terminology (1924), but he was in turn contradicted by Dubertret himself (1951). Dubertret's objections were similar to, and in part based on, Blanckenhorn's, some of which had been disposed of already by Dubertret's own earlier mapping, but he continued to misinterpret the E-W to ENE-WSW-trending fold structures as normal fault-related and NNE-SSW-trending ones as shortening-related, refusing to follow Kober's (1915) and Krenkel's (1924, 1925) interpretations associating them with the orogeny to the north (Dubertret 1930, 1932, 1934). Dubertret continued to deny the existence of thrusts in Syria and Lebanon (Dubertret 1948), but the engineer-geologist Henri de Cizancourt (1948), in the memoirs Dubertret edited, had already presented a picture of the Palmyran mountains as a thrust-bounded basement fold and thrust belt, similar to the US Rockies and little differing, in essence, from our present interpretations (see below).

Krenkel's observations and interpretations and the debates that followed have recently given rise to a number of interpretations regarding the Syrian Arc as 'far-field' effects of compressive stresses with unclear causes or of non-existent collision events taking part in the Alpides to the north (Bosworth et al. 1999, 2008; Abd El-Motaal and Kusky 2003). The purpose of this paper is to show that the Syrian Arc has indeed its own alpinotype mountain range behind it, that this mountain range has long been known and that the reason why the two had never been put together lies in the fragmented knowledge of the Syrian Arc. Krenkel's (1924) '*Syrischer Bogen*' and Ricou's (1971) '*croissant ophiolitique péri-arabe'* are two parts of one great orogenic system extending from Oman to Cyrenaica for 5000 km. It is independent from the other Tethyan orogens to its north and has both alpinotype and germanotype parts. We propose to call this orogen the Ayyubid orogen, because its best-known parts, from Cyrenaica and Egypt through Israel, Lebanon, Syria, Iraq to southeastern Turkey (Fig. 2) formed, in the twelfth century CE, the Empire of the Ayyubids (see Kinder and Hilgemann 1982, p. 136), founded by Ṣalāḥ al-Dīn Yūsuf ibn Ayyūb (1137–1193), better known to the western world under the name of Saladin, the most gentlemanly commander of the Crusader wars.

THE AYYUBID OROGEN: SPATIAL EXTENT

Figure 2 shows the spatial extent and the main subdivisions of the Ayyubid Orogen (for localities named below and not shown on this map, refer to Figs. 8, 9 and 11). It begins in Cyrenaica in the Al-Jabal al-Akhdar (the Green Mountain) and continues through the northern parts of the Libyan desert south of the Bay of Bamba (*Khalij Bamba*; also spelled Bambo) and the Western Desert of Egypt. In the Sinai Peninsula, it turns northeastward. In offshore Sinai and Israel, its existence is shown by the folds and thrusts under the uppermost Cretaceous beds and in Israel it turns entirely into a north-south direction. In Mt. Lebanon and Antilebanon it bifurcates: one branch goes off to form the Palmyra Arc (Krenkel's 'Damascene Arc' = Picard's (1958) 'Palmyraides'). The other branch continues through coastal Syria (Ruske 1981) and enters southeastern Turkey, forming the Mesozoic folds and thrusts of the Turkish border folds (Ketin 1966; Şengör and Yılmaz 1981) or 'Assyrides' (Şengör et al. 1982). From Turkey it goes though the inner Zagros chains [see the papers in Jassim and Goff

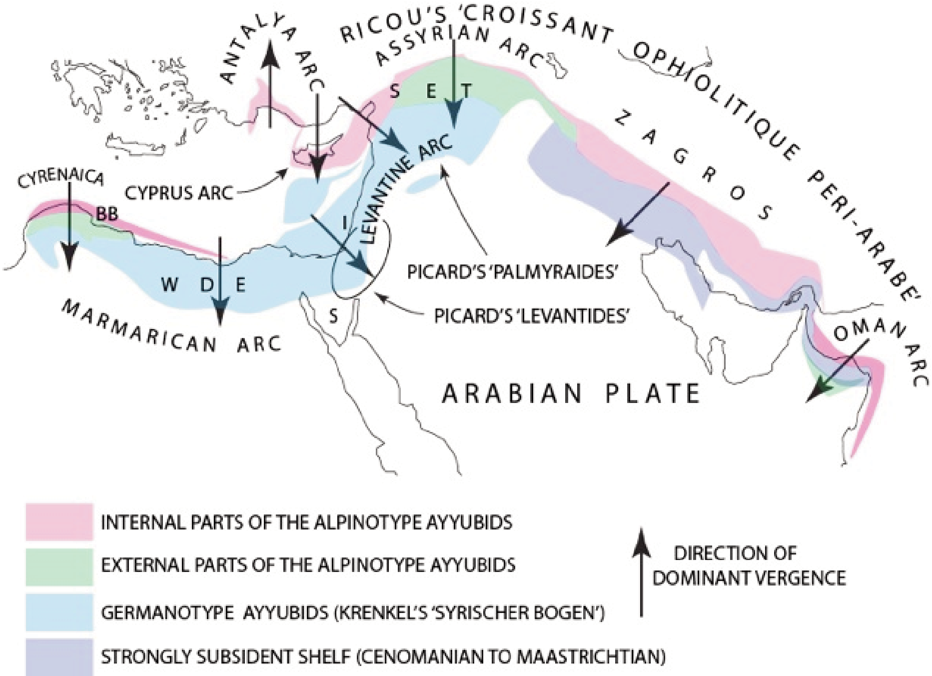


Figure 2. The extent and major tectonic subdivisions of the Ayyubid Orogen (compare with Krenkel's 1924, 'Syrischer Bogen' and Picard's 1958, 'Levantides'). The parts that continue into Pakistan and India are not shown mainly because the time of major deformation there was later (Paleocene in Pakistan; Eocene in northern Pakistan and India) and because a paper concerning the early tectonics of those parts of the Ayyubids is currently in preparation by Oliver Jagoutz, Leigh H. Royden and Şengör. Note that the external parts of the Ayyubids are discontinuous as shown in this map. This is largely because the Cretaceous structures in southeastern Turkey (in the 'Assyrian Arc') and in the Zagros Mountains are strongly overprinted and thus masked by the later Cainozoic deformational events. Wherever detailed work is available, the Cretaceous folds are recognized (e.g. Saura et al. 2011). We have little doubt that the green region, standing for the external parts in southeastern Turkey (SET), must extend all the way down to Oman maintaining very much the same width. The Antalya Arc has no external parts, because whatever external parts had formed in the Cretaceous were later overridden and structurally superimposed by the Cainozoic nappes. From offshore eastern Libya, Egypt to Israel and Lebanon, the publicly available seismic data are insufficient to distinguish with any confidence the external parts from the germanotype Ayyubids. In the sector of the Ayyubids depicted in this map, the indicated vergence direction is the same as the orogenic polarity. Key to abbreviations: BB Bay of Bamba, WDE Western Desert of Egypt, S Sinai Peninsula, I Israel, SET south-eastern Turkey.

(2006) and Leturmy and Robin (2010)] and reaches Oman (Robertson et al. 1990, and references therein). Thus a grandiose virgation exists in the northern part of the Arabian plate[7]. It is a free virgation, not constrained by any resistant mass in the lithosphere, except for its northernmost branch, which is constrained by the ophiolitic crescent of which we shall speak below.

Throughout the length of the Ayyubid orogen, the timing of the main deformation is very tightly constrained between the Turonian and Middle Maastrichtian, in many localities even between the early Santonian and late Campanian. In the western Ayyubids, the most external parts in the germanotype structure are better developed, whereas in the eastern half of the orogen, it is the alpinotype parts that are the most conspicuous. Along the entire belt the orogenic facing direction and the main sense of vergence is towards Afro-Arabia, with the singular exception of the Antalya Nappes in southern Turkey (Şengör and Yılmaz 1981). In the following paragraphs we describe individual sectors of the Ayyubids from west to east, whereby the main emphasis will be largely on the lesser known western part between Libya and Syria, because the eastern part is comparatively well-known.

Cyrenaica and the Libyan Desert

Cyrenaica (Fig. 2) forms a mountainous promontory east of the Gulf of Sirte. Gregory (1911, 1916) gives delightful and very detailed accounts, with abundant references even including the classical authors[8], of the earlier geographical and geological expeditions leading to the discovery of Cainozoic and Mesozoic rocks there and in his 1911 paper there is much, still useful, geological information. For the geomorphology and Quaternary geology of Cyrenaica, in places reflecting the skeleton of its structure established during its paleotectonic evolution, see Mühlhofer (1923), McBurney and Hey (1955), Hey (1968a, b) and Völger (1968). For the seismotectonics in Cyrenaica, see Campbell (1968), Goodchild (1968) and Al-Heety (2013).

Cyrenaica is crowned by the mainly east-west to ENE-WSW-striking mountain range Al-Jabal al-Akhdar (the Green Mountain) covered with wild olive trees and in places thick Mediterranean maquis shrubland. It rises from beneath the Eocene Apollonia Formation (Fig. 3A, B, C, D) consisting mainly of light coloured, massive, fine-grained siliceous limestone with flint debris near its base (Fig. 3B, C). It is in places chalky, but only rarely marly. It shows faint indications of grading probably because of deposition by turbidity currents. In many places it displays spectacular slump features with chaotic bedding, penecontemporaneous folding and thrusting plus flat channels produced by submarine erosion (Fig. 3D). The Apollonia Formation interfingers with the younger Dernah Formation, which is generally assigned a Priabonian age (Röhlich 1974, pp. 32–34; Klen 1974, pp. 25–26; Barr and Berggren 1980; Banerjee 1980).

Below the Apollonia Formation, across a low-angle unconformity, the Al-Athrun Formation characterizes the coastal areas of Cyrenaica (Fig. 3A; Röhlich 1974, pp. 27–28; Banerjee 1980, pp. 12–13; see also fig. 8c in

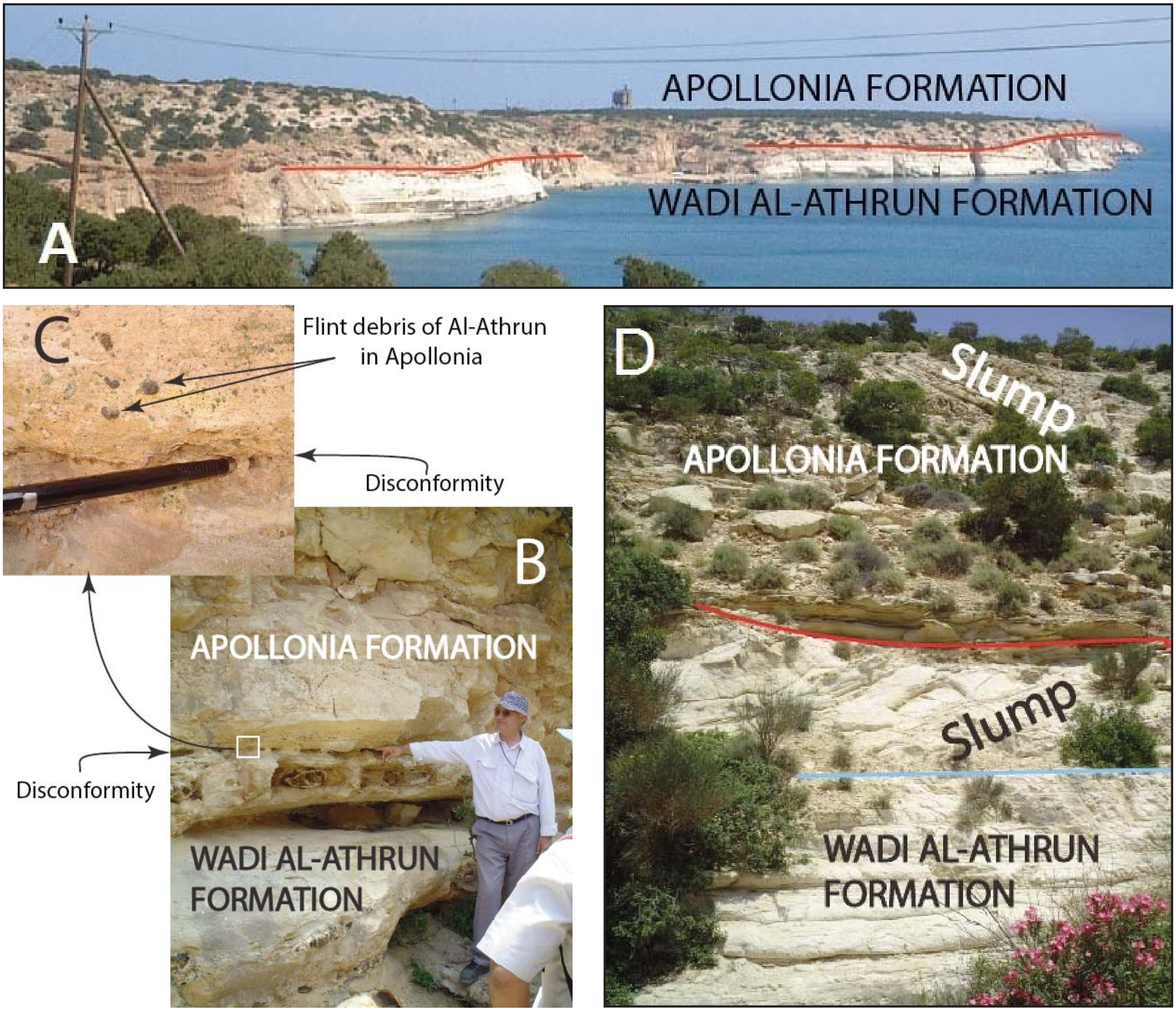


Figure 3. Various outcrop aspects of the Apollonia Formation: A. Apollonia Formation sitting with slight angular unconformity on the Wadi al-Athrun Formation in the Marsa al-Hilal (near the northern apex of Cyrenaica). The red line delineates the unconformity. B. Apollonia Formation sitting with very slight unconformity on the Wadi al-Athrun Formation. Professor Ali El-Arnauti points at the unconformity. C. Chert debris at the base of Apollonia Formation plucked from the cherts of the Wadi al-Athrun Formation. D. Apollonia Formation sitting across an unconformable contact on Wadi al-Athrun Formation. Both formations here exhibit slump structures. See text and figure 9 in Barr (1968) showing part of the same outcrop. All photos are by Michael A. Martin.

Duronio et al. 1991). It is a white to buff, thinly bedded, richly fossiliferous uppermost Campanian to Maastrichtian limestone deposited on the deeper parts of the Cyrenaica shelf. It exhibits impressive slump structures resembling those in the Apollonia Formation (see Fig. 3D). In the western extremity of Cyrenaica (e.g. in the well A1-NC 120; see Figs. 8 and 9 for location), the uppermost Albian or Cenomanian to Coniacian Al-Hilal Formation (Röhlich 1974, pp. 25–27; Banerjee 1980, pp. 24–25), made up of brownish to greenish grey, thinly-bedded, commonly glauconitic shales bearing pyrite clusters and passing upwards into more calcareous layers, conformably underlies the Al-Athrun Formation, but as one goes landward and eastward, it pinches out and gives place to a stratigraphic gap. In the Benghazi Basin (south of the town of Benghazi) it was deposited in a shallow, neritic environment on a restricted platform top (Duronio et al. 1991).

Landward and eastward, the Al-Athrun Formation is replaced by the dolomitic Wadi Dukhan (or Ducchan) Formation of Maastrichtian age (Pietersz 1968; Kleinsmiede and van den Berg 1968; Klen 1974; Röhlich 1974), which underlies, in places conformably, but in other places across a slight unconformity, the Al-Uwayliyah Formation of Upper Paleocene age consisting of whitish chalk and greenish marl, and then, again across a slight unconformity, the Middle to Upper Eocene Dernah Formation (Duronio et al. 1991). As pointed out above, the Dernah Formation is in general younger than the Apollonia Formation, but because of their interfingering relationship, it may in places be older than Priabonian. In general, to the south, i.e. landward, the Dernah Formation replaces the Apollonia Formation entirely and where the Al-Uwayliah Formation also falls out, the Wadi Dukhan Formation comes to underlie only the Dernah Formation. The stratigraphic and structural relationships thus formed are exactly the same as those between Apollonia and Al-Athrun formations: gently dipping Paleocene/Eocene units unconformably overlying only slightly steeper Maastrichtian units. Elsewhere, the Campanian Al-Majahir Formation, consisting dominantly of cream-coloured, in part chalky, neritic limestone and lesser dolomitic limestone, dolomite and marl (Röhlich 1974), partly equivalent to the coastal Al-Athrun Formation, unconformably overlies the Al-Baniyah Formation along the southern slope of the al-Jabal al-Akhdar (Röhlich 1974, 1978, 1980).

The structural picture changes considerably when one enters the area of the two large anticlines, one south of Marawa and the other extending southwestward from Jardas al Abid (for locations see Fig. 8[9]). Figure 4 shows two outcrops in the southern part of the Marawa anticline. In Figure 4A one sees an open but very near being a closed, mesoscale fold in the Al-Baniyah Formation of Cenomanian to Coniacian age consisting of pinkish, medium-bedded microcrystalline limestone, in places marly, in other places dolomitic. This formation is much more tightly folded, around mostly E-W (Fig. 4B) and SW-NE axes, than the overlying Al Athrun and the Wadi Dukhan formations. The folds are without exception open to closed and of flexural slip type. Figure 5A shows a group of Al-Baniyah beds in the Jardas al-Abid anticline displaying the structures shown in Figure 5B. The beds dip 50° to the NNE. The fossils in these formations (e.g. rudists) are essentially undeformed (Fig. 5C). Calcite-filled cracks probably formed as extrados extensional features. Perpendicular to these are stylolitic surfaces that localized slip along bedding-parallel planes to allow flexural slip.

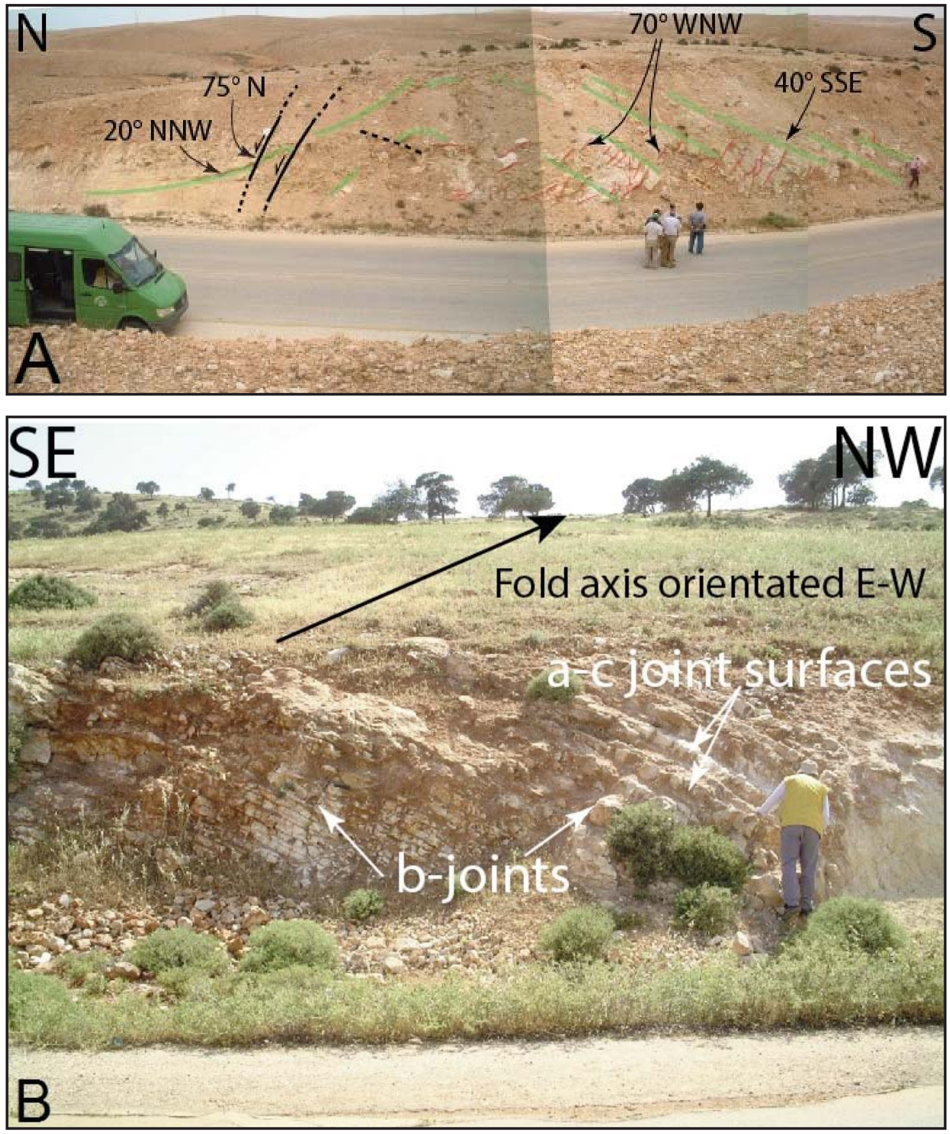


Figure 4. A. Flexural slip fold in the al-Baniyah Formation within the large anticlinal core south of Marawa. Green lines indicate bed-form surfaces. Red lines are joints. Black are faults with sense of movement shown by half-arrow. B. Another flexural slip fold within the al-Baniyah Formation. The axis of the fold trends E-W. Photos by Michael A. Martin.

On the basis of similar observations, Röhlich (1974, 1978, 1980) proposed a phase of shortening in the al-Jabal al-Akhdar area that

> "*had stronger deformational effects than any subsequent tectonic phase. Folding stress produced an ENE-WSW-striking anticlinorium composed of several wide folds. Some anticlines have the character of elongated domes. The beds usually dip 10 to 20 degrees in the limbs, but locally as much as 50 degrees. High angle faults striking NE-SW, E-W and NW-SE [appeared] too; some of them were not reactivated later. The anticlinorium emerged as a ridge, probably an elongated island, and its axial part was relatively deeply eroded (to some hundreds of metres) during the intra-Senonian interval*" (Röhlich 1974, p. 57).

Röhlich (1974) also pointed out that in the northeastern part of the

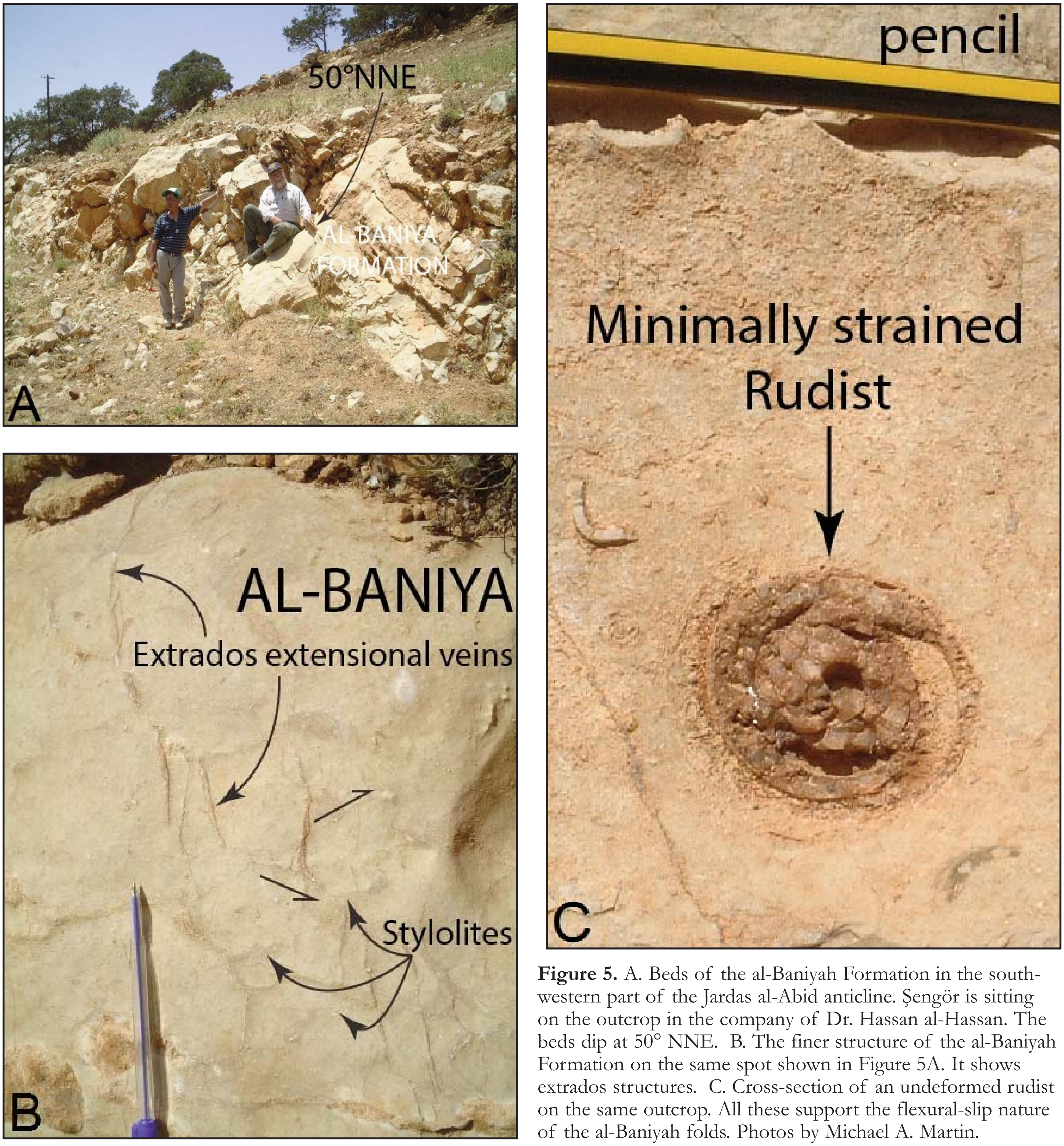


Figure 5. A. Beds of the al-Baniyah Formation in the southwestern part of the Jardas al-Abid anticline. Şengör is sitting on the outcrop in the company of Dr. Hassan al-Hassan. The beds dip at 50° NNE. B. The finer structure of the al-Baniyah Formation on the same spot shown in Figure 5A. It shows extrados structures. C. Cross-section of an undeformed rudist on the same outcrop. All these support the flexural-slip nature of the al-Baniyah folds. Photos by Michael A. Martin.

al-Jabal al-Akhdar the effects of the intra-Senonian folding could not be seen and that the medial to late Cretaceous sedimentation there was uninterrupted.

Röhlich (1980, p. 929) wrote that the folding of the pre-Campanian sedimentary rocks of the al-Jabal al-Akhdar range was of 'mediotype', i.e. between those of orogenic belts and platforms[10]. This corresponds to Stille's germanotype deformation, but of the kind that produced more elongate structures, closer in shape to those in alpinotype orogenic belts. Such type of folding Stille (1940, pp. 4–5, and 656) had indeed called 'mediotype'.

We entirely agree with Röhlich's excellent observations. We would only add that the Al-Jabal al-Akhdar structures commonly have steeper southeast sides than northwest ones. There is thus a gentle south to southeasterly vergence.

Following a period of marine transgression in the latest Cretaceous, Al-Jabal al-Akhdar re-emerged, but this time in the form of a broad uplift, the boundaries of which were outside the Al-Jabal al-Akhdar range (see fig. 1 in Völger 1968). This up-doming produced the slight unconformity above the Wadi Dukhan and al-Athrun formations. We ascribe this to a falcogenic deformation, i.e. essentially faultless bending of the entire lithosphere below al-Jabal al-Akhdar. Röhlich (1980) pointed out that the axis of the Eocene upheaval was parallel with the axes of folding during the Senonian. What this upheaval was caused by we shall discuss at the end of this section.

Röhlich (1974, 1978, 1980) made no attempt to find out what actually caused the Senonian folding in Cyrenaica. Bosworth et al. (1999, 2008) ascribed it to 'far field stresses' on incorrect and partly misinterpreted information. An alleged Africa-wide shortening, which Bosworth et al. (1999) saw as the cause of the Senonian folding, left wide areas west of Cyrenaica untouched, although at the same time the Benue aulacogen in Nigeria was also shortened, as indicated by the angular unconformity between the folded Albian to Santonian section (ending with the Awgu Formation) and the Maastrichtian Lafia Sandstone in the Middle Benue Trough (Obaje 2009, pp. 62–63). This folding shifted the focus of shortening westward in the Lower Benue Trough forming the Anambra Basin, where there occurred another folding event at some time between a certain Maastrichtian and a certain Paleocene (Obaje 2009, pp. 60–62). Argand (1924, p. 206), already suspected the presence of these late Cretaceous to earliest Cainozoic folding events; the pre-Awgu folding was firmly established by 1952 (de Beauregard et al. 1952) and was used in the earliest plate tectonic interpretations of the region (Burke et al. 1971, 1972; Burke and Dewey 1974; Freeth 1978). The cause of the shortening in various parts of northern Africa was therefore most likely more localized. Using Bosworth's and his colleagues' earlier publications, one might say that the far field stresses were able to deform only earlier rifted areas. Cyrenaica, however, was located on a high and not in a rift basin before the Senonian folding. Bosworth et al. (2008) saw the cause of the localization of folding in Cyrenaica as its prominent position and tried to underpin this interpretation by arguing that the Marmarican basins in Egypt were not at the same time folded, an assertion which is incorrect (Moustafa 2008, especially fig. 11, showing the sharp intra-Khoman, i.e. 'Santonian', angular unconformity above folds in the Abu Gharadig Basin). Even if it were correct, the proposed strain shadow next to the so-called Cyrenaican 'shock absorber' (Bosworth et al. 2008) cannot account for the intense folding in the Sinai (which would have been in the deepest recess of the proposed 'strain shadow') and Israel. Therefore the 'far field' model cannot explain the peculiarities of the Syrian arc deformation. The observations that the deformation was weaker in the northernmost part of the al-Jabal al-Akhdar and was stronger in the east, as shown by the time gaps represented by unconformities (Fig. 6), and the 'Santonian' unconformity in the subsurface of Egypt, disprove this model. The proposal by Abd El-Motaal and Kusky (2003) depends on the 'closure of the Tethys' to the north. They imply a northern closure, but they do not say where (Tethys at the time had several branches). But, as there was *no* Tethyan closure during the Senonian (cf. Şengör and Natal'in 1996; Şengör 2009), their proposal does nothing to explain the origin of the Syrian arc.

A more promising approach to the origin of the deformation in Cyrenaica than seeking the origin of the entire Syrian Arc deformation in ill-defined concepts or non-existent events is perhaps to look at the available data on the structure of the area in detail.

First, the surface observations: we have earlier established that the folding during the 'Senonian' orogeny was one of flexural slip. Röhlich (1974) commented that many faults break mostly the southern flanks of the folds, which Şengör (unpublished observation) was able to corroborate in the field. Flexural-slip folding, to be maintained across large areas, implies either a décollement or a thick, incompetent layer at depth. Since there was no basin under the al-Jabal-al Akhtar, there could be no thick incompetent layer below the 'Senonian' folds to absorb the shortening. Thus, what is here needed is a décollement. Figure 7A shows schematically what such a décollement might have looked like. The 'Senonian' orogeny would deform the rocks above the décollement as shown in Figure 7B. These would then be eroded – Röhlich (1974) noted erosion of anticlinal cores for hundreds of metres as quoted above) and the overlying formations would be laid down on the planed surfaces (Fig. 7C). Renewed deformation, but this time only in the form of a broad falcogenic rise would create the pre-Apollonia and Dernah formations surface, the planation of which would prepare the ground for the deposition of these formations (Fig. 7D).

A first attempt to sort out the overall architecture of Cyrenaica was made by El-Arnauti et al. (2008) by combining surface observations with seismic reflection data. We here build on their work by also considering the gravity data provided by Elakkari (2005) and Suleiman and Saleem (2008). Figure 8 is a Bouguer gravity anomaly map of Cyrenaica and surrounding areas and the large positive anomaly that extends to the two anticlines of Jardas al-Abid and Marawa immediately arrests the attention (see also fig. 6 in Suleiman and Saleem 2008). Considering the fact that the northern parts of Cyrenaica are low coastal areas almost at sea-level and that the al-Jabal al-Akhdar itself hardly rises to a height of 500 metres this anomaly may be interpreted as indicating the presence at no great depth of a body a few km thick, denser than upper crustal rocks. We interpret it as part of the Eastern Mediterranean oceanic crust torn from its original place and shoved under the continental rise of Cyrenaica. In a way it may represent an abortive attempt at ophiolite obduction as shown in Figure 17B of this paper. Such an interpretation is entirely consistent with the stratigraphic, sedimentological and structural information we have from Cyrenaica: 1) the Senonian deformation did not affect a former basin, but a platform area; 2) while deformation was going on in the al-Jabal-al Akhdar area, sedi-

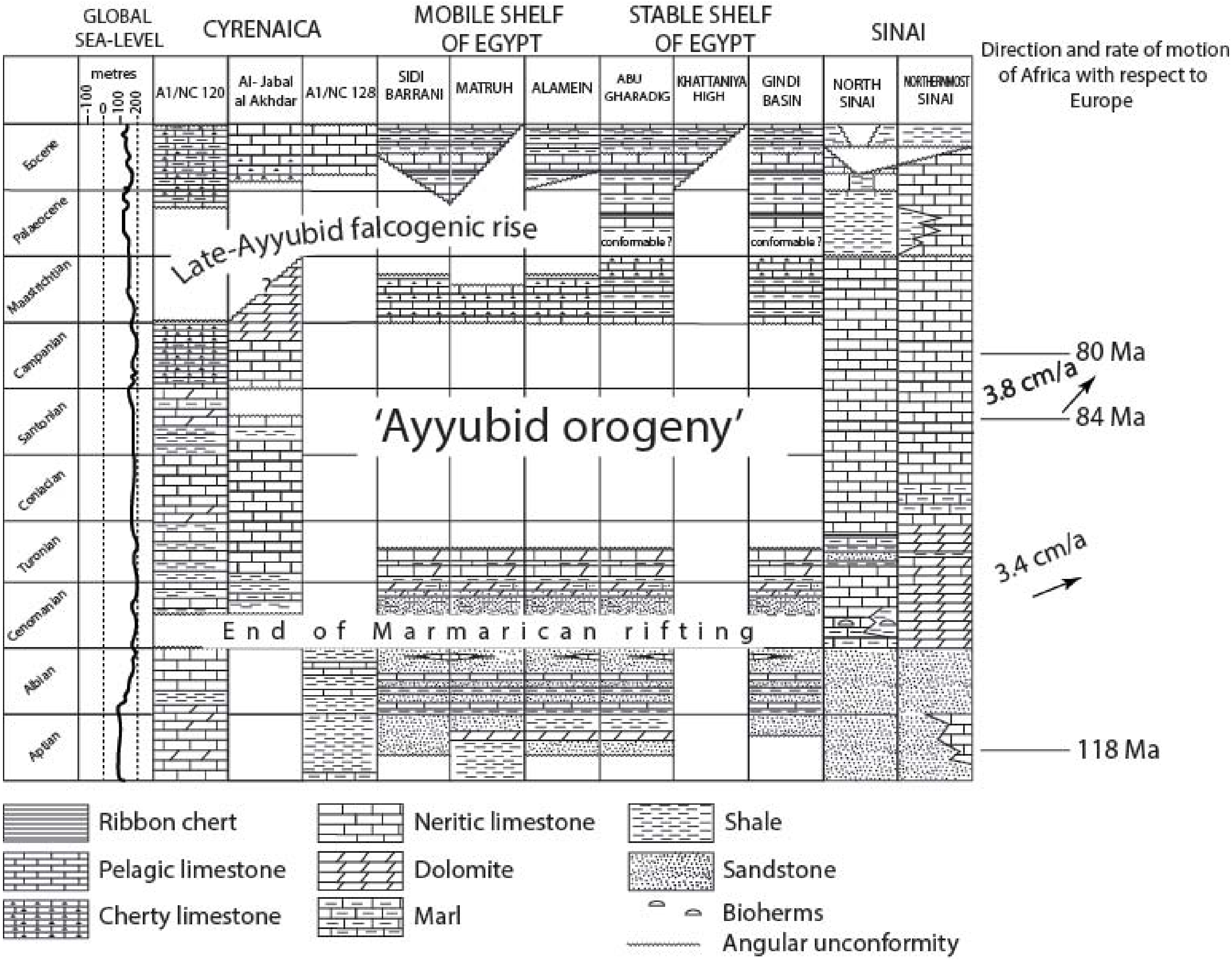


Figure 6. Stratigraphic outline of the Ayyubid orogen from Libya to the Sinai Peninsula. The sources are as follows: the two wells A1/NC 120 and A1/NC 128 and al-Jabal al-Akhdar (i.e. the 'Green Mountain') are from Duronio et al. (1991); Sidi Barrani, Matruh, Alamein, Abu Gharadig, Khattaniya and Gindi are from Kerdany and Cherif (1990) and Said (1990a); north and northernmost Sinai are from Jenkins (1990). The global sea-level curve is from the large coloured fold-out 'Geologic Time Scale 2012' in Gradstein et al. (2012).

mentation in places to the north remained uninterrupted; 3) strong deformation was a short- to very short-lived event: it took place between the late Santonian and the earliest Campanian (Fig. 6, column labeled al-Jabal al-Akhdar), i.e. within a time span of 1.5 to 7 Ma, depending on when Campanian sedimentation commenced on the deformed edifice. Suleiman and Saleem (2008, fig. 6) ignored the deformation in Cyrenaica in their interpretation of the gravity observations and were compelled to assume that no thickness difference affected the continental crust all the way to the Eastern Mediterranean. They accounted for the positive anomaly by thinning the upper crust and thickening the lower crust, but do not give a reason for the origin of this unusual geometry. Their interpretation contradicts what we know of the geological history of the margin (first rifting in the Permian and Triassic, then orogeny in the Santonian). Although the interpretation of gravity observations is not unequivocal, our hypothesis is based on the gravity data *together* with the topography and the geological history of the area and seems now to be the best constrained of the available suggestions.

Long-sustained stresses deforming entire continents are unlikely to act on such short timespans, for example, as is now seen in Asia: its widespread internal deformation has been going on at least for the last 55 million years (e.g. Şengör 1997). In Europe, foreland deformation has been even longer lived: since at least the late Cretaceous; and finally 4) the notable southerly vergence betrays an asymmetry in the deformation. The detached masses need a backstop, indeed a piston, which was most likely somewhere in the north.

But before we reach Figure

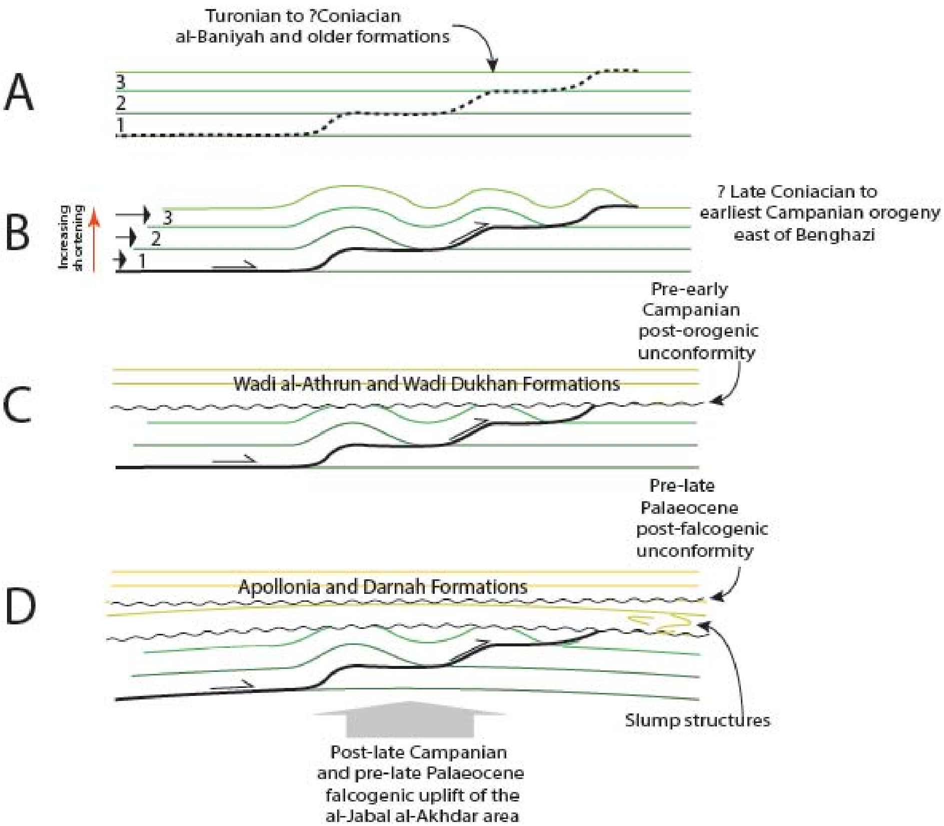


Figure 7. A schematic outline of the sequence of events in the progress of the 'Senonian' orogeny in Cyrenaica. Notice the completely different natures of the events during the 'Senonian' orogeny and during the later falcogenic phase before the deposition of the Darnah and Apollonia formations.

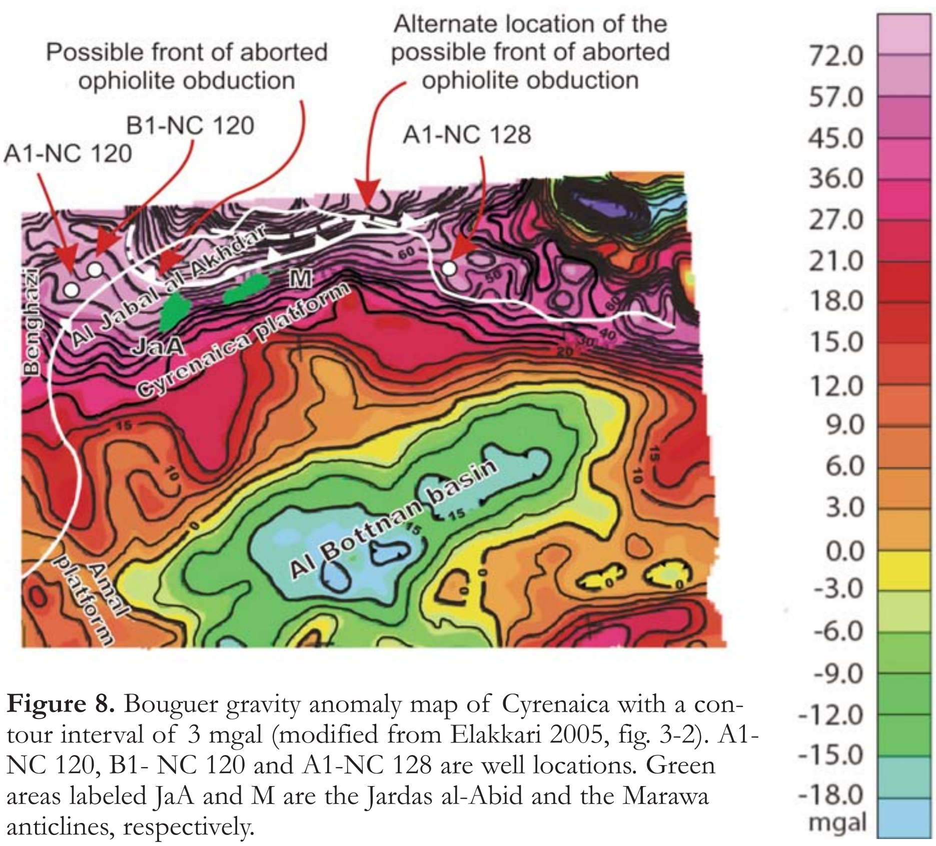


Figure 8. Bouguer gravity anomaly map of Cyrenaica with a contour interval of 3 mgal (modified from Elakkari 2005, fig. 3-2). A1-NC 120, B1- NC 120 and A1-NC 128 are well locations. Green areas labeled JaA and M are the Jardas al-Abid and the Marawa anticlines, respectively.

17B, we continue our tour of the Ayyubid orogen to see whether its other parts provide any support for this hypothesis. We note that between Cyrenaica and the Egyptian frontier, the entire Libyan desert is underlain by Chattian to Pliocene sedimentary rocks (Geological Map of Libya, 1: 1,000,000, 1985, sheet NE), so no surface data on the 'Senonian' orogeny are available there. Figure 9 shows the subsurface structures in Libya and Egypt drawn on the basis of isopachs of the Senonian sedimentary rocks (Yanılmaz et al. 1989; Hantar 1990). It is clearly seen on this map that Senonian structures do not continue for any appreciable distance to the south in Libya, but they do so as far south as the Bahariya-Diyur High in Egypt and that the Libyan and the Egyptian structures constitute a single, united field of deformation. That the depicted structures in Egypt (Yanılmaz et al. 1989; Hantar 1990) are real folds of large dimensions is seen in numerous industry seismic profiles, one example of which has been published by Bosworth et al. (2008, fig. 2); for other examples, see Moustafa (2008). However, depicting such structures using the entire Senonian interval is far too coarse to give any idea on their temporal evolution. Data exist in Egypt to follow their development stage by stage through the late Cretaceous.

Egypt Including the Sinai Peninsula

Figure 10 displays a set of non-palinspastic palaeogeographic maps showing the displacement of the shoreline during the Cenomanian to Campanian interval. Note that during the Cenomanian (Fig. 10A) a line of three islands marked the southern master fault block of the Abu Gharadig rift (cf. Bosworth et al. 2008, fig. 2). The end of the Abu Gharadig Basin in the Qattara Basin was marked by a large land area that has a peninsula jutting out in the direction of the three islands. That the three islands corresponded to the top of a northerly-tilted normal fault-bounded block is shown in figure 2 of Bosworth et al. (2008). Already during the Turonian (Fig. 10B), two of the three islands moved south and the easternmost one was enlarged in an easterly direction. At the same time, the western land

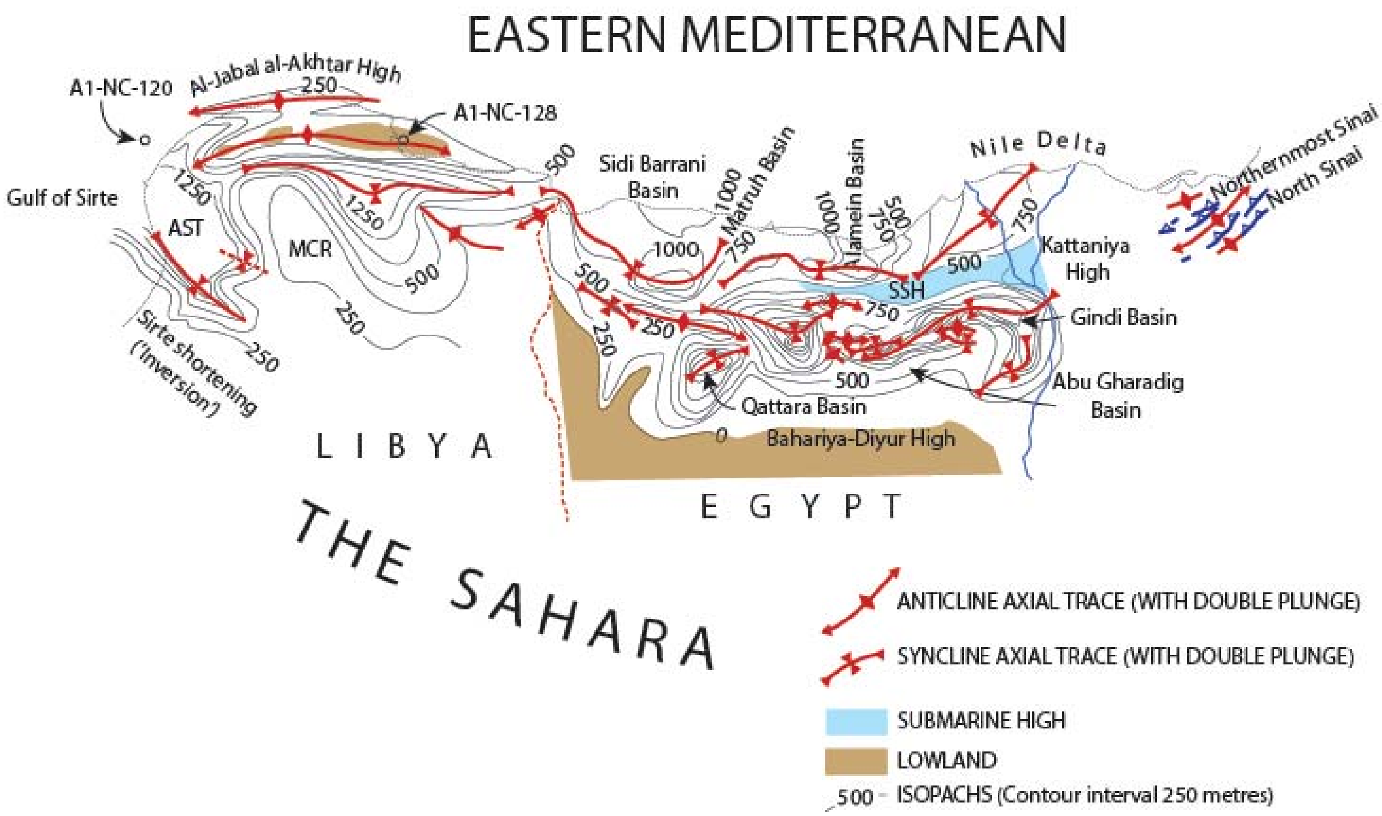


Figure 9. Senonian structures in northern Libya and Egypt (Libya is from Yanılmaz et al. 1989; the two well locations, those of A1-NC,120 and A1-NC-128, are from Duronio et al. 1991; Egypt is from Hantar 1990; the structures in the Sinai Peninsula are from Jenkins 1990; Moustafa and Khalil 1990; Flexer et al. 2005). Key to abbreviations: AST = Ash Shulaydimah (=Sciuduma=Suluq) Basin, MCR = Mid-Cyrenaican Ridge, SSH = Sharib-Sheiba High.

area became wider as shorelines everywhere migrated seawards from their Cenomanian positions. In addition, an elongated island appeared to its east almost connecting the peninsula with the three islands. This cannot be because of extension, for in an extensional regime the major south-dipping normal fault of the Abu Gharadig Basin would have tilted the footwall and caused a migration of its edge northward, not southward. Also any area in extension, provided it does not sit on a plume-generated uplift, would subside, and not become uplifted. The enlargement of the westerly land area is thus anomalous, especially at a time of very high world-wide sea level (see Fig. 6).

During the Turonian to Coniacian interval (Fig. 10C), the three islands north of the Abu Gharadig Basin became united into an ENE-WSW trending 'cordillera' (*sensu* Argand 1916) while the shore to the south of it retreated. This is most likely due to thrust loading by the southerly-marching cordillera (for the geometry of thrusts using former normal faults in this area, see Moustafa 2008). The new island that had appeared in the Turonian was also displaced southward. It too was probably a part of the large thrust mass underlying the cordillera.

In the Coniacian to Santonian (Fig. 10D) interval, the number of cordilleras increased as new island chains appeared in the north along axes parallel with the earlier cordilleran axis. During the Santonian to Campanian interval, the land area reached its maximum size, as many thrust masses become uplifted as island chains on their bounding thrust systems. Thus we see a very similar picture in Egypt to what is seen farther west in Cyrenaica and the agreement in timing, structural style and orientation leave no doubt that here we simply see a westerly prolongation of the same structures as in Libya. The claim by Bosworth et al. (2008) that Egypt had been protected by an alleged Cyrenaican 'shock absorber' is certainly not true (also cf. Moustafa 2008). The only difference in the Egyptian Western Desert is that we have only subsurface seismic reflection data to study the structures. How misleading seismic reflection profiles can be in terms of yielding structural detail, we shall see below in the case of the Damascene Arc, but they certainly have greater resolution than paleogeographic data, as a comparison of our Figure 10 with Moustafa's figures shows. Even in Egypt, though, as soon as the structures come to the surface in the Sinai Peninsula, we again see mesoscale folds and thrusts, exactly as in Cyrenaica (Fig. 11), suggesting that the same style also dominates the subsurface in the Western Desert of Egypt, where seismic profiling allows us to see the crustal architecture only through a haze.

A glance at Figure 6 shows that the entire Upper Turonian to Maastrichtian interval is missing. We suggest, on the basis of what we see in Figure 10, that the deformation here

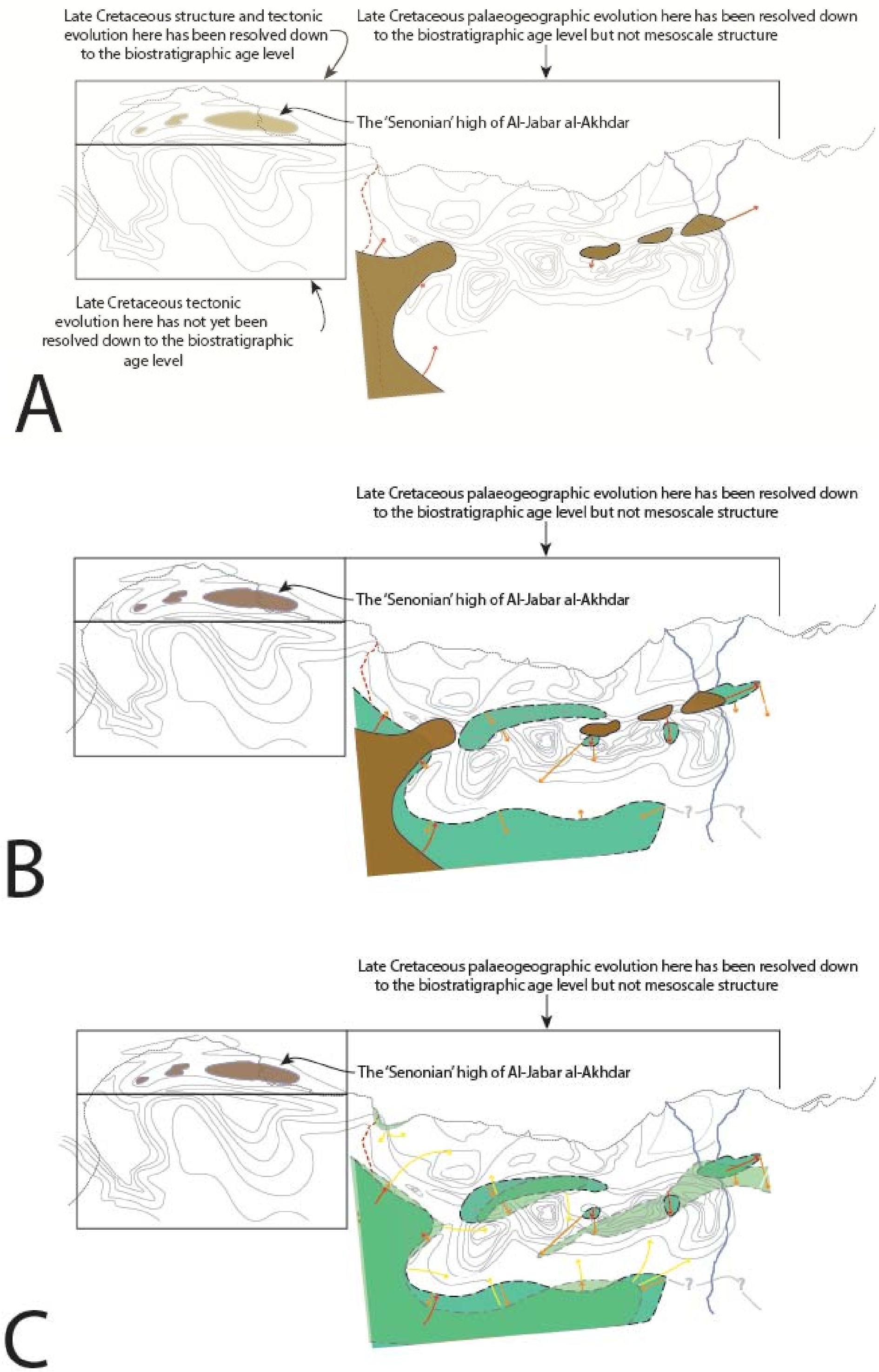


Figure 10. (*on this and following page*). The paleogeographic evolution of Egypt during the late Cretaceous. The Senonian isopachs are also shown to emphasize that they are far too crude to reveal the true structure. In Libya, south of al-Jabal al-Akhdar one has to make do with them, because any finer data are not publicly available. But in Egypt one can follow step by step the evolution of the large-scale folds throughout the late Cretaceous in the form of 'cordilleras' in Argand's (1916) sense (A through E). Despite the rising sea-levels at the time, the land area continuously increased until the end of the Santonian and then again decreased somewhat in the Campanian, when better-defined anticlinal axes in the form of long and narrow cordilleras appeared. The cordilleras marking anticlinal crests generally migrated southward throughout the late Cretaceous. This is consistent with the mainly southerly vergences of the associated structures. The paleogeographic data in Egypt are from Said (1990b). A. Cenomanian paleogeography (legend for all time frames is the same as the one shown in Fig. 10E; in all frames the little arrows show the displacement of shorelines.). B. Turonian palaeogeography. C. Coniacian palaeogeography. D. Santonian palaeogeography. E. Campanian palaeogeography.

had probably commenced during the later Turonian, but it reached its maximum intensity during the late Senonian as indicated by the largest extent of the land surface during the Senonian to Cenomanian interval and by the sharp unconformity in the Khoman Formation (Santonian to Maastrichtian: fig. 11 in Moustafa 2008); that unconformity was also folded later, as documented in other seismic profiles by Moustafa (2008). That this interval was also the time of maximum intensity of deformation in southern Israel, just to the northeast of the Sinai Peninsula we shall see in the minutely studied Hatira anticline there.

Israel Including Offshore Sinai

As seen in Figures 2 and 11, the Ayyubid orogen swings to the northeast in the Sinai Peninsula and then turns almost completely northward in Israel. The part that includes the Sinai Peninsula and Israel was called the 'Levantides' in a figure that Leo Picard published in a 1958 paper. Because both this term and the paper in which it first appeared are little known, we reproduce the figure here in Figure 12 (in a

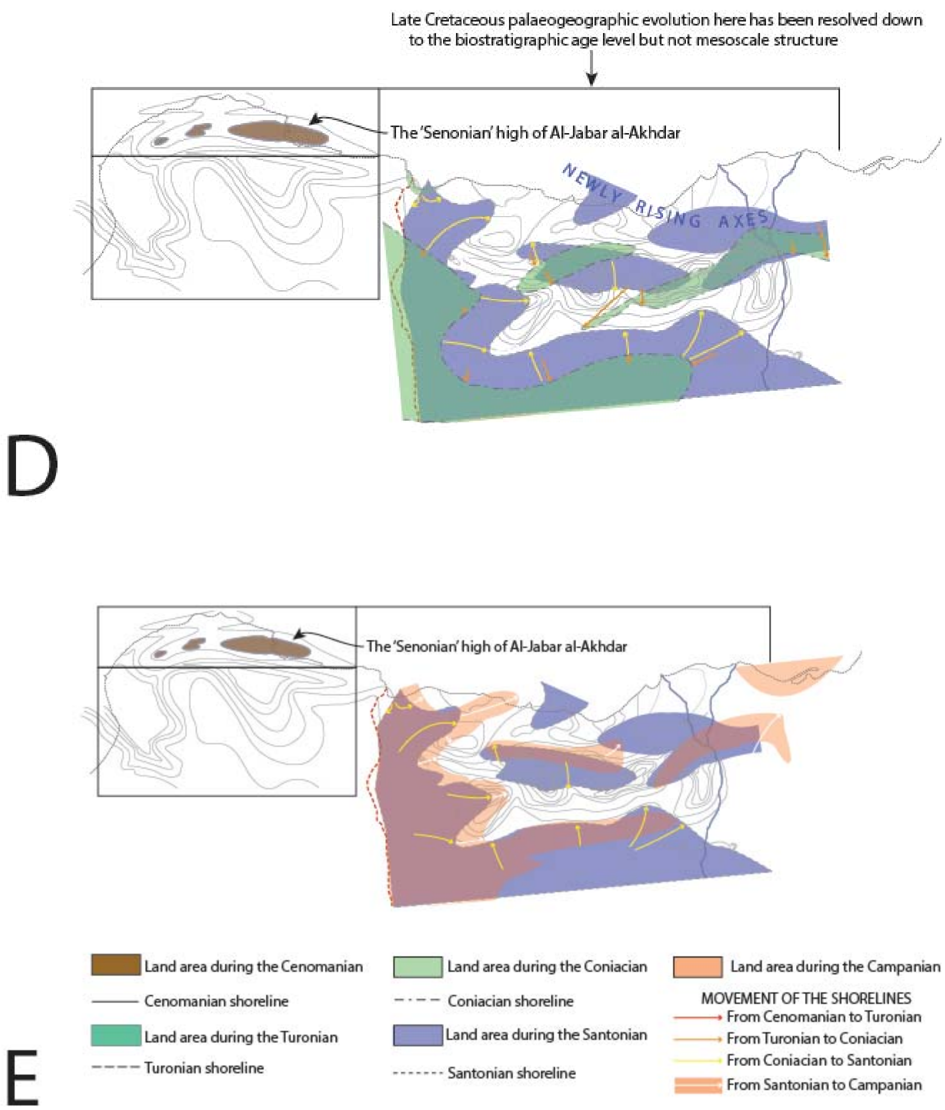


Figure 10. (*continued*).

redraughted version, because some parts of the original are barely legible). Seismic data from the offshore in this region show that intra-Upper Cretaceous deformation is also known in the offshore (grey area in Fig. 11), where Neev et al. (1985) and Gardosh and Druckman (2006) documented the presence of strong pre-end Cretaceous folding and thrusting. Tapponnier et al. (2004) argued that the continental margin of the Levant is along the dotted line marked cm in Figure 11 (see Carton 2005). Inboard of that line they reported Eocene shortening structures. Because the Eocene structures are nucleated on, and continue the shortening of, the 'Senonian' structures everywhere else in the Ayyubid orogen, we assume, on the basis also of the unconformities here, mapped by Neev et al. (1985) and Gardosh and Druckman (2006), that most of these structures have a Senonian ancestry.

In fact, just to the southeast of these structures, on-land in southern Israel, is the Hatira (Makhtesh-Hagadol or Kurnub) Anticline (Fig. 13A, B). This structure was noticed as early as 1886 (Hull 1886 – frontispiece geological map: structure labeled as 'strata disturbed'). Eyal (2011) recently documented that the most rapid folding of what at a first glance appears to be a simple structure that formed all at once after the Maastrichtian, in fact had formed mainly during the late Campanian and early Maastrichtian (3.2°/m.y.). Only the lowest rate of folding was during the Paleocene (0.4°/m.y.) (Fig. 13B). Without Eyal's detailed studies, no one would have suspected from this field sketch that the most rapid phase of folding was during the late Campanian-early Maastrichtian interval. Eyal's work very nicely shows what sort of detailed studies must be undertaken to sort out the timing and geometry of even the apparently simplest structures, known for nearly a century and a half. In seismic reflection profiles, even structures such as the Hatira Anticline itself are difficult to delineate in the sort of detail that is possible in the case of surface exposures; imagine how very difficult, nay impossible, it must be to peel off the sort of detail Eyal (2011) has managed to obtain in the case of the Hatira Anticline. Without such detail, however, our understanding of mountain belts would be very incomplete.

The Damascene Arc

North of Israel, in Lebanon, a grand virgation characterizes the Ayyubid structure (Fig. 11; Suess 1909, p. 314). The entirely germanotype Damascene Arc splits off towards the northeast and builds a low-altitude desert mountain range, which, in itself, is a smaller virgation. The structures of this arc, which Picard (1958) called the 'Palmyraides' (Fig. 12) have been studied intensively during the last two decades, and our description of its structure and evolution is based on that recent work (Chaimov et al. 1990; McBride et al. 1990; Al-Saad et al. 1992; Searle 1994; Brew et al. 2001; Sawaf et al. 2001) plus Şengör's own earlier study (O'Keefe and Şengör 1988).

Figures 2 and 11 show the extent of the Damascene Arc and Figure 14 illustrates three cross-sections across its northeastern (Fig. 14A), middle (Fig. 14B) and southwestern (Fig. 14C) sectors on the basis of seismic reflection profiling. Figures 14A' and A'' are two cross-sections drawn by Searle (1994) across a tiny part of the northeastern traverse on the basis of surface geology. All of these cross-sections leave little doubt that the Dama-

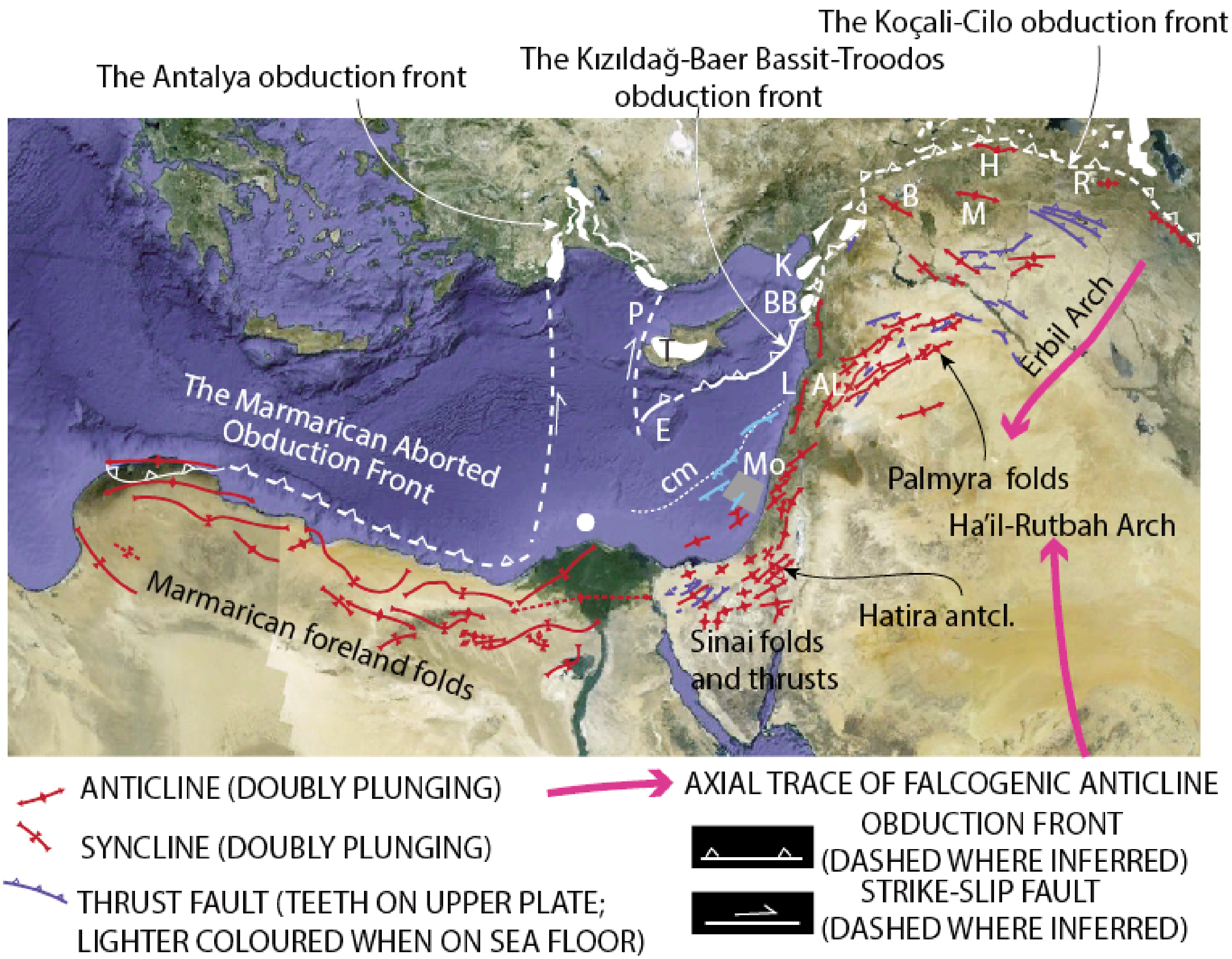


Figure 11. The major structures of the Ayyubids west of Iran. The more easterly parts, including the Zagros Mountains, Oman, Pakistan and the Himalaya are not shown, because, with the exception of the central and eastern Himalaya and Myanmar, they are relatively well-known. The sector shown is where the Syrian Arc ('Syrischer Bogen') concept was first defined in 1924 by Erich Krenkel. It is therefore the 'type' area of the Ayyubid orogen. Sources of data for this map: Libya is from Yanılmaz et al. (1989), El-Arnauti et al. (2008); Egypt is from Hantar (1990); the structures in the Sinai Peninsula are from Jenkins (1990), Moustafa and Khalil (1990), Flexer et al. (2005); Israel is from Schulman et al. (1959), de Sitter (1962), Flexer et al. (2005); Syria is from Chaimov et al. (1990), McBride et al. (1990), Al-Saad et al. (1992), Searle (1994), Brew et al. (2001), Sawaf et al. (2001); Antalya Nappes are from Dumont et al. (1972); Monod (1976); Robertson and Woodcock (1981a, b), Woodcock and Robertson (1982), Reuber et al. (1984), Théveniaut et al. (1993), Bağcı and Parlak (2009), Varol et al. (2007); Cyprus and the Baer Bassit ophiolites are from Whitechurch and Parrot (1974), Al-Riyami et al. (2000, 2002), Morris et al. (2006), Chan et al. (2007); Southeastern Turkey is from Türkünal (1953), Sungurlu (1974), Yılmaz (1985, 1993), Robertson (1986), Bağcı et al. (2005), Karaoğlan et al. (2013); Zagros is from Saura et al. (2011); offshore Egypt, Israel, Lebanon: Neev et al. (1985), Carton (2005). Key to abbreviations: AL= Anti-Lebanon thrust-bound anticline, B = Bozova High, BB = Baer Bassit ophiolite nappe, cm = continental margin according to Tapponnier et al. (2004) taken from Carton (2005), E= Eratosthenes High, H = Hazro High, K = Kızıldağ ophiolite nappe, L = Mt. Lebanon Anticline, M = Mardin High, Mo = 'Mediterranean offshore structures' of Flexer et al. (2005), fig. 18K.6; seismic reflection profiles and wells in the area indicated in grey clearly show Early Tertiary is unconformable on 'Late Cretaceous' and that, in turn, is unconformable on 'Middle Cretaceous'. The unconformities are clearly caused by thrusting with a dominant east vergence (see fig. 12.9 in Neev et al. 1985), P= Paleo-Paphos fault, R = Ricgar (Gare) Anticline.

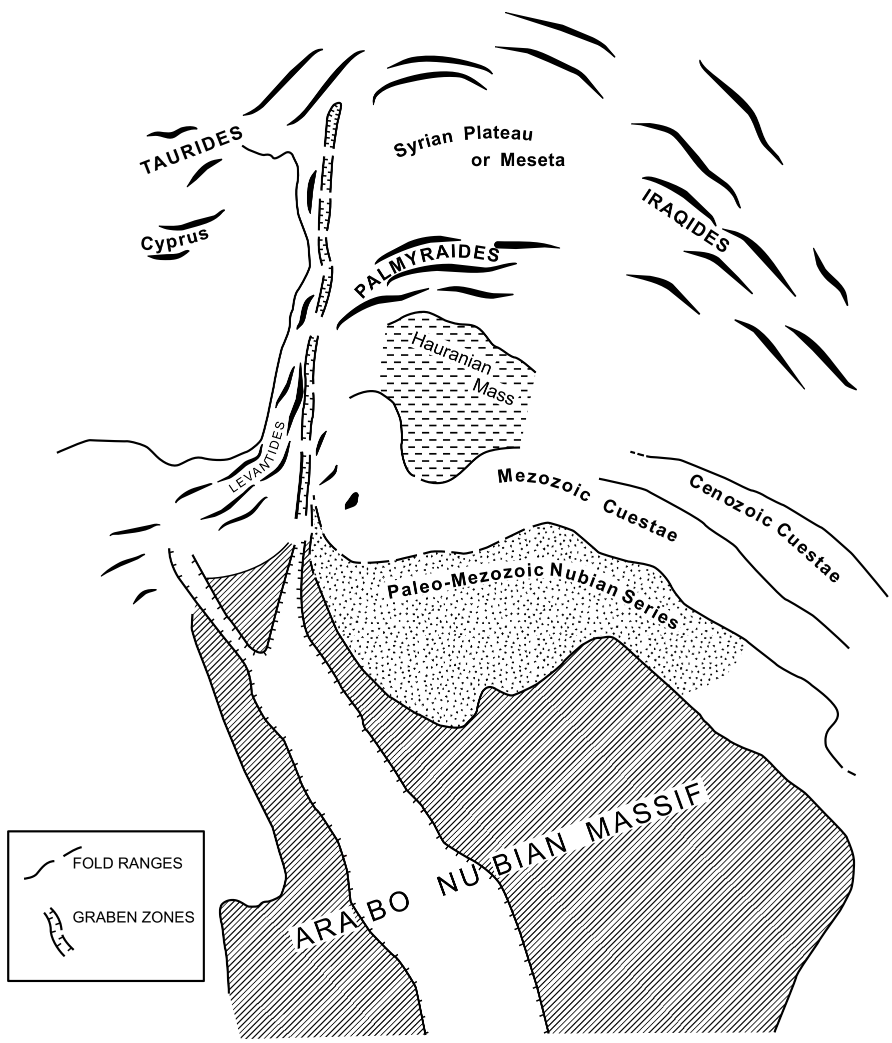


Figure 12. A retracing of Leo Picard's map of the 'morphotectonic elements of the Middle East'. The term 'Levantides' first appeared in this map (Picard 1958, p. 21); Picard did not use it in his 1958 text, but indicated in 1959 that 'Israel, as part of the Levantides fold belt (Picard 1958) is crossed by a series of frequently asymmetric anticlines and synclines which strike mostly NE-SW' (Picard 1959, p. 312). Compare this map with that in Figure 1 of this paper.

scene arc is a germanotype mountain belt dominated by open to close folding and associated thrusting in a previous rift basin that probably was an aulacogen (de Cizancourt 1948; O'Keefe and Şengör 1988; Searle 1994). Chaimov et al. (1990) estimated that the southwesternmost sector probably accommodated some 20 to 25 km shortening, which supposedly decreases to some 1–2 km in the northern sector. These estimates are made entirely on seismic reflection profiles and they betray a severe problem when compared with Searle's (1994) estimate in the tiny Jabal Mazar area near the middle part of the cross-section A in Figure 14. Along the fold shown in Figure 14A', Searle estimated a total shortening of some 0.9 km. However along that cross-section the seismic profiles show nothing like what Searle mapped. On Searle's map there are at least four such folds that would require a shortening of some 4 km, i.e. twice that estimated by consulting the seismic reflection profiles alone. But even that would be an underestimate, for such limestone sequences, as seen to be folded in the Damascene Arc, tend to absorb much ductile (up to 10%: Engelder and Engelder 1977) and elastic strain (up to 2%: Engelder 1979) before the actual buckling begins to build the folds and they retain even the elastic strain for very long time intervals (since the late Paleozoic in the Appalachians, for example) (Engelder and Engelder 1977; Engelder 1979). This means that along the Palmyran traverse A one should perhaps add another 4 to 6% shortening to that computed from Searle's folds. This would raise the shortening to 8 km at least. Therefore where seismic reflection profiling allows an estimate of 1–2 km shortening, a minimum of 8 km shortening may in fact have taken place on the basis of surface structural mapping! In cross-section C, we would not be surprised if the actual shortening would exceed 100 km in the southwestern end of the Damascene arc, dwindling to perhaps some 10 km or less in the northeastern end. De Cizancourt's (1948) superb model, based on gravity observations and field mapping more than sixty years ago, was already pointing in that direction.

Although nearly all workers agree that the main folding of the Damascene arc took place during the Miocene, there was also significant folding during the Cretaceous. Searle (1994), for example, noted that there is minor on-lap in the Upper Cretaceous sequences and the shortening observed in the Triassic to Cretaceous (inclusive) shows a greater extent than that in the Eocene rocks in Jebel Abiad. When we have studies from this area of the kind that Eyal (2011) undertook in Israel, we shall have a much clearer picture of the distribution in time of the Palmyran folding and thrusting.

The Cornell workers (Chaimov et al. 1990; McBride et al. 1990; Al-Saad et al. 1992; Brew et al. 2001; Sawaf et al. 2001) have ascribed folding in the Damascene arc to the events around the Arabian plate and Searle (1994) has shown that a push from the west (not from the north) must have been active. As the Cretaceous structures have not been separately mapped, it is impossible to tell what the strain picture during the Cretaceous Ayyubid orogeny was. However, the very geometry of the Damascene Arc must have been established during the late Paleozoic and the Triassic rifting phases creating the Palmyra aulacogen (O'Keefe and Şengör 1988), so that the Cretaceous folding could

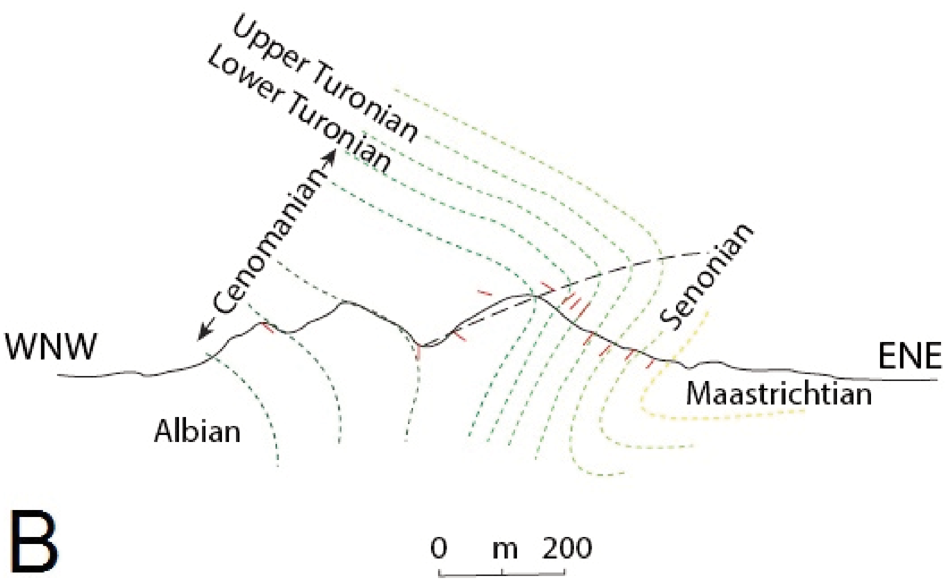


Figure 13. A. A general view of a part of the Hatira Anticline (from https://picasaweb.google.com/the.jookies.2006/HatiraAnticline#5276974114307103538). Note the plunging nose. B. A cross-section across the overturned southeastern limb of the Hatira (Makhtesh-Hagadol or Kurnub) Anticline (redrawn from Schulman et al. 1959, fig. 1). Note that the anticline is so drawn to imply that all the folding was post-Maastrichtian, i.e. entirely Cainozoic. However, Eyal (2011) recently showed that the most rapid folding occurred here during the late Campanian and early Maastrichtian (3.2°/m.y.) and the lowest rate was during the Paleocene (0.4°/m.y.).

not have had a very different geometry from the Miocene one. This is supported by the fact that no serious and regionally significant interference between the Cretaceous structures and the Miocene structures has been reported.

Under these circumstances, we propose that it was the well-known massive ophiolite obduction from the west in the Baer Bassit and Hatay regions in Syria and Turkey that must have provided the necessary push. This brings us to the discussion of the mainly alpinotype parts of the Ayyubid orogen.

The 'Croissant Ophiolitique Peri-Arabe'

The late Luc-Emmanuel Ricou pointed out as early as 1971 that a complete belt of large ophiolite nappes of late Cretaceous obduction age embraces the Arabian plate from the east, north and the northwest (Ricou 1971). He called this structure the circum-Arabian ophiolitic crescent (=*croissant ophiolitique péri-arabe*: see Fig. 2). Since then a huge amount of work has been carried out on the different members of this ophiolitic crescent, all members of which have been shown to be supra-subduction zone ophiolites (e.g. general: Şengör and Natal'in 1996; Oman: Lippard et al. 1986; Kermanshah: Whitechurch et al. 2013, Neyriz: Babaei et al. 2005; Cilo: Yılmaz 1985; Kızıldağ: Tekeli and Erendil 1986; Baer-Bassit: Al-Riyami et al, 2000, 2002; Chan et al. 2007; Troodos: Miyashiro 1973; Pearce and Robinson 2010; Antalya Okay and Özgül, 1984). What is of primary interest from the viewpoint of this paper is the timing of obduction of each of these large ophiolite nappes. Figure 15 is a summary of the data that reveal an amazing synchroneity of both the time of spreading and the time of obduction of all the ophiolites of the circum-Arabian ophiolitic crescent and that this time is the same as that of the so-called 'Senonian' orogeny along Krenkel's Syrian Arc (compare Figs. 6 and 15).

The isotopic age data, almost all from the mafic plutonic foundation of the obducted ophiolites using U–Pb ages on zircon grains, indicate that all the ophiolites formed during the Cenomanian with Kızıldağ and Troodos having formed possibly a little bit later during the early Turonian. There might be a very slight younging towards the present northwest from Oman to Antalya, but the obduction times seem to be best bracketed between the Turonian (or perhaps even the Coniacian) and the Upper Campanian. This corresponds to a time interval of some 3 to 5 million years and is precisely the same as the time of deformation along the Syrian Arc.

All along the ophiolitic crescent, the large ophiolite nappes moved onto an Atlantic-type continental margin of normal crustal thickness for such margins, as revealed by the pre-

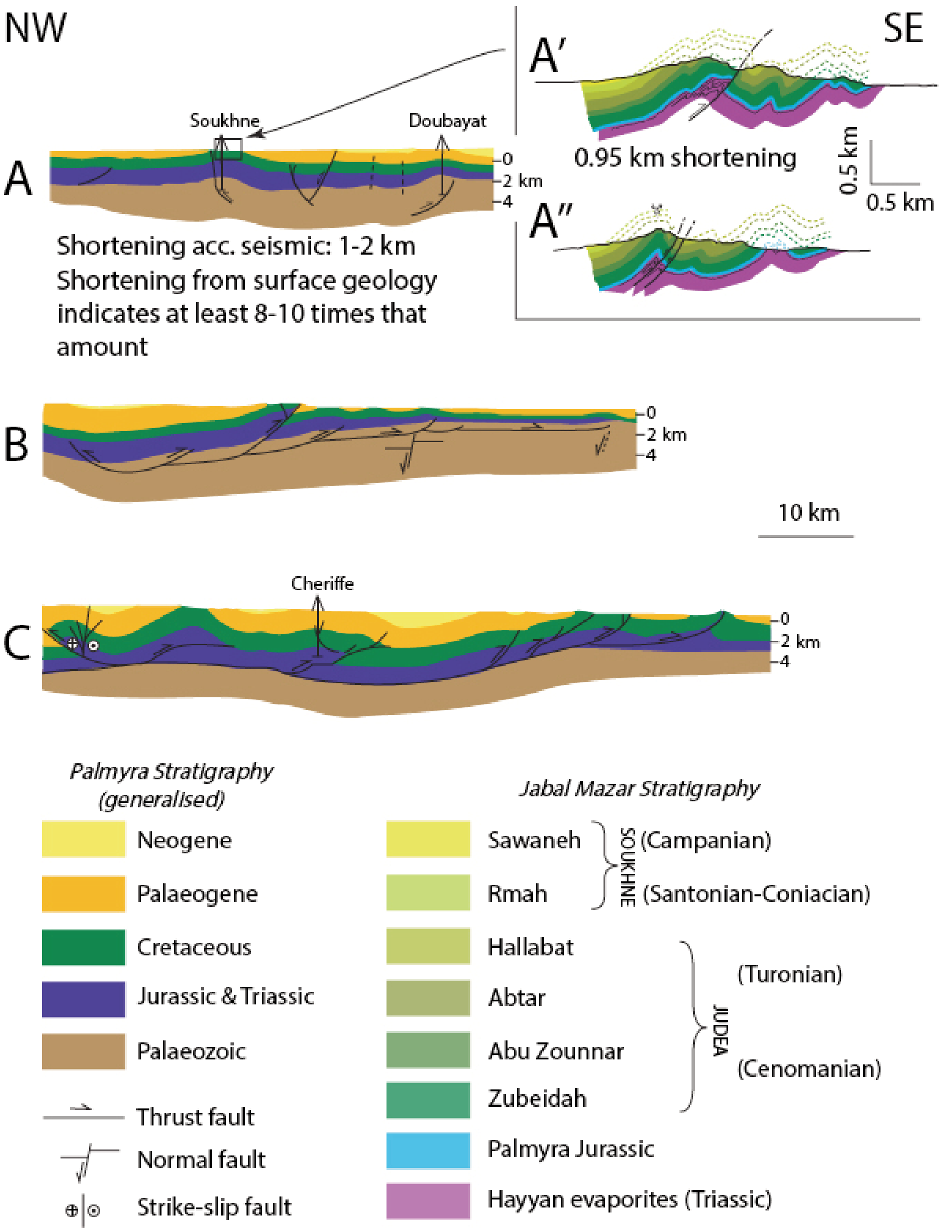


Figure 14. Three cross-sections across the Palmyran fold and thrust belt (A, B, C) redrawn after Chaimov et al. 1990). According to these authors the total amount of shortening across the belt is some 20 km along the cross-section C and 1–2 km along the cross-section A. A' and A" are cross-sections based on field mapping by Searle (1994). Along these cross-sections across the Jabal Mazar (Grave Mountain) structure located very near cross-section A the total shortening is about 0.95 km. This shows immediately that the figures cited by Chaimov et al. (1990) for the total shortening across the Palmyran fold and thrust belt cannot possibly be correct. They are an order of magnitude off. This results from the inability of seismic reflection data to resolve detailed structure. Also the basal décollement drawn in the Paleozoic rocks is probably not there, because there is no orogen north of the Palmyran fold and thrust belt to absorb the shortening. It is more likely that the soft rocks filling the Palmyran aulacogen took up the shortening by a variety of means (homogeneous bulk shortening and thickening, kinking, etc.) that cannot be recognized on seismic reflection profiles, as already implied by de Cizancourt (1948) and also pointed out by Searle (1994).

dominance of neritic shelf deposits on all of them (see Şengör and Natal'in 1996). In some, the continental margins began subsiding as the ophiolitic armada (consisting of the Semail, Neyriz, Kermanshah, Cilo, Kızıldağ/Baer-Bassit and the Troodos massifs), was approaching, most likely by being pulled down by the associated nascent subduction zones. Only in front of the Troodos ophiolite was the continental margin considerably thinner than elsewhere, and some even think entirely oceanic (e.g. most recently Tapponnier et al. 2004). However, it has long been clear that the Eratosthenes seamount is a continental structure most likely torn from the Afro-Arabian margin as the eastern Mediterranean was opening (see Kempler 1998; Rybakov and Segev 2004 and the references therein). Here the Troodos was thrust for a very considerable distance and underwent an anticlockwise rotation for some 20° between the Turonian and the Campanian around a pole somewhere east of the present day Hatay, because it has recently turned out that the Hatay ophiolites have rotated in unison with the Troodos nappe (Morris et al. 2006; Fig. 16). During the thrusting, it may be that the Antalya segment acquired a different vergence from the nappe front to its east for the reasons explained in Figure 16D and E. This would explain why the Antalya and the Troodos nappes were separated and moved into opposite directions.

This concludes our tour of the Ayyubids. In the next section we outline how we think this grand orogenic belt may have formed.

THE AYYUBID OROGEN: MECHANISM OF FORMATION

The review of the tectonics of the Ayyubids in the preceding section demonstrates five very significant characteristics of this remarkable structure of the face of our earth: 1) as a unified orogen it formed in a surprisingly short time period: at most between the Turonian and the Campanian, although parts of it have later become reactivated during the Cainozoic. 2) Its structures are remarkably continuous from eastern Libya to Oman; they only change character somewhere between Syria and Antalya. Along the entire

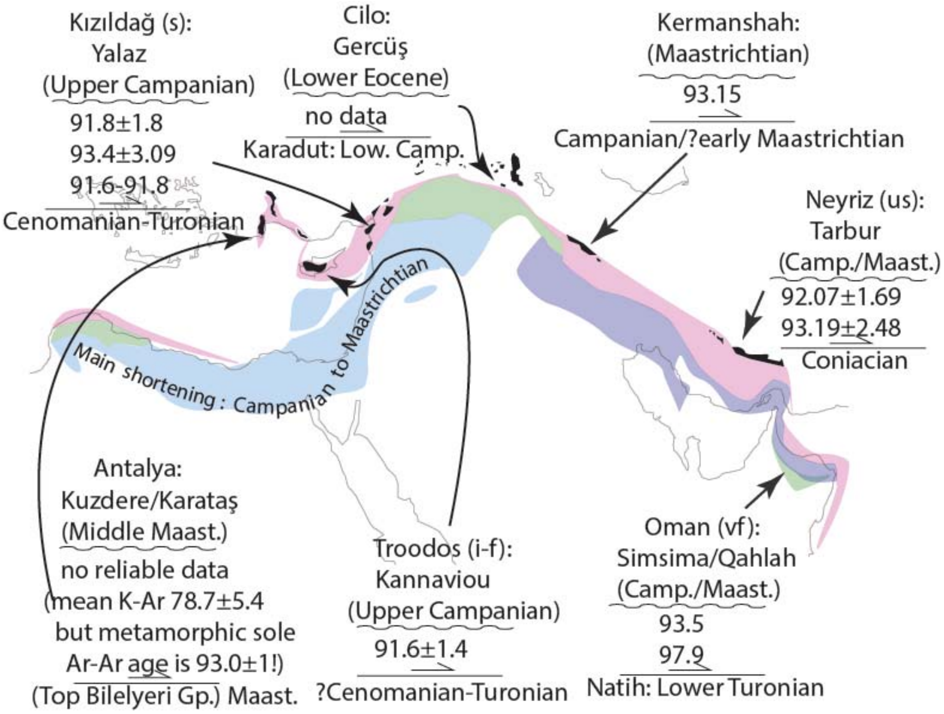


Figure 15. Timing of events along the Ayyubid orogen. In each locality the top name is that of the ophiolite nappe as commonly known in the literature. The letters in parentheses next to them indicate the inferred speed of spreading (us = ultra-slow, s = slow, i = intermediate, f = fast, vf= very fast). Beneath it is the name of the formation that seals the nappe contacts and below that the age of that formation. Below the wavy line signifying an unconformity are zircon U–Pb ages. When other ages are used this is indicated. Below that, across the thrust symbol, is the age of the youngest underlying rocks. Where this is a single formation, its name is given; where not, only the youngest age of a sequence is indicated. Ophiolite genesis and obduction from Oman to Cyprus all happened synchronously within the resolution of the isotopic and biostratigraphic data we now have. The obduction events were also synchronous with the shortening all along the Syrian Arc from eastern Libya to southeastern Turkey. The sources are as follows: Antalya: Robertson and Woodcock (1981b), Montigny et al. (1983), Lagabrielle et al. (1986), Karaoğlan et al. (2013); Troodos: Mukasza and Ludden (1987), Allerton and Vine (1991), Premoli-Silva et al. (1998), Peybernès et al. (2005), Karaoğlan et al. (2013); Baer Bassit: Kızıldağ: Tekeli and Erendil (1986), Dilek and Delaloye (1992), Steuber et al. (2009), Karaoğlan et al. (2013); Cilo: Fontaine (1981), Yılmas (sic [Yılmaz]) (1985), Yılmaz and Duran (1997); Kermanshah: Braud (1987), Whitechurch et al. (2013); Neyriz: Ricou (1976), Janessary and Whitechurch (2008); Oman: Tilton et al. (1981), Hanna (1990), Abdelghany (2003).

orogen the structures verge towards Afro-Arabia, with the singular exception of the Antalya nappes which moved north towards the Menderes-Taurus Block (Şengör and Natal'in 1996). 3) Nowhere along the entire belt does one see highly metamorphic core regions characterized by HT/LP rocks affecting the entire continental crust as seen in collisional orogens and in core regions of magmatic arcs. Instead, wherever metamorphism is seen, it is invariably of HP/LT type and is related to subduction under ophiolite nappes. Slices of HT/LP rocks are also related to obduction. 4) The marginal fold and thrust belts commonly seen spectacularly developed in front of and behind collisional orogens or behind Andean arc orogens (compressive arcs: Dewey 1980; Jarrard 1986; Şengör 1990) have only feebly developed in front of the Ayyubids and those that have developed have been superimposed and largely masked by similar structures related to later collisional orogenies (see Şengör and Natal'in 1996). Some structures in Al-Jabal al-Akhdar in Libya, in Sinai and in Israel resemble mini-marginal fold and thrust belts. No hinterland thrusting has so far been reported. 5) Germanotype foreland structures are well-developed in the western part of the orogen (but not in front of the Antalya Nappes), but are almost non-existent in the eastern sector. By contrast the alpinotype structures are best developed in the east (here Cyprus and Antalya together are exceptions).

Figure 17 shows our attempt to account for the origin of the Ayyubid orogenic belt, while also explaining all of its peculiarities listed above. In all segments of the Ayyubid orogen, orogeny was preceded by the establishment of an Atlantic-type continental margin. This margin was formed by rifting events during the Permian and in Oman oceanic conditions were also established offshore already during the Permian. In Iran and southeastern Turkey, intracontinental stretching probably continued well into the Triassic. By the Lias, a shelf edge had been established everywhere. Sometime during the middle Cretaceous, a subduction zone formed all the way from the Antalya area in Turkey to Oman and most likely beyond (Şengör and Natal'in 1996). This subduction zone consumed the small width (as judged by the short time interval between spreading and obduction) of oceanic lithosphere between itself and the Atlantic-type continental margin of Afro-Arabia (in the case of Antalya, between itself and the Anatolide/Tauride margin) and that margin descended to depths of 20 to 30 km producing blueschist- and eclogite-facies rocks in the Antalya Nappes (in Alanya, Turkey: Okay and Özgül 1984) and in Oman (e.g. el Shazly and Coleman 1990), finally choking the nascent subduction zone. In front of the ophiolitic sheets of the Ayyubids, extensive flysch and molasse basins developed as far west as Turkey, where large slabs of oceanic crust and upper mantle were stranded on the continental crust as obducted ophiolite nappes; farther west we do not see such basins, nor the obducted ophiolites. Because shallow water shelf and platform sedimentary rocks are preserved where there is no record of large flysch or molasse basins, the continental crust must not have been

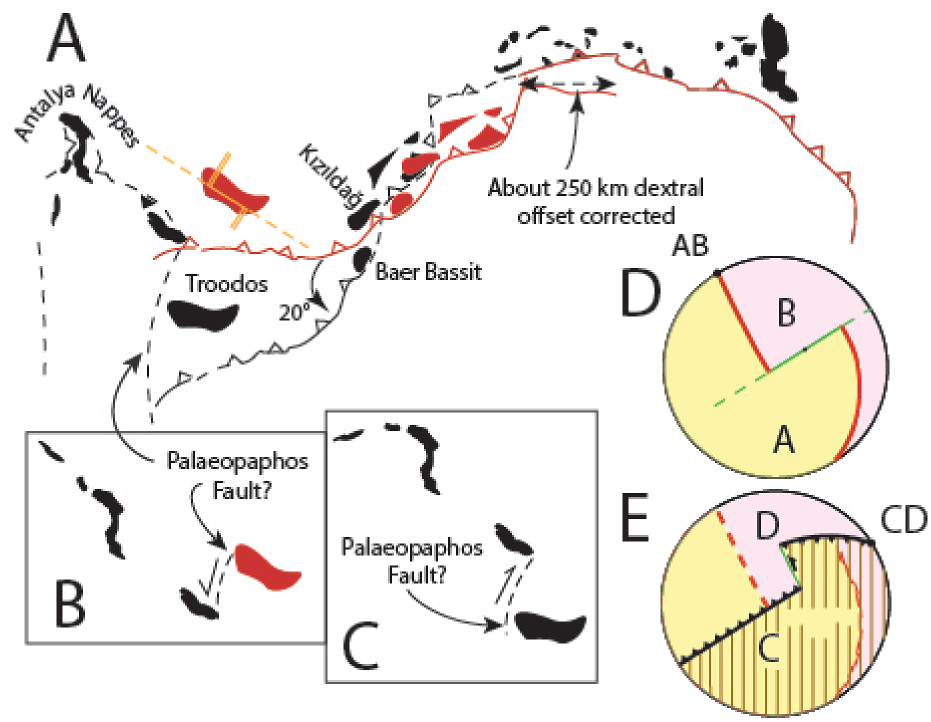


Figure 16. Schematic maps showing the idea that the Troodos and the Kızıldağ and the Baer Bassit ophiolite bodies rotated 20° anticlockwise during the emplacement of the circum-Arabian ophiolitic crescent as a result of the pinning of the nappe front of the crescent somewhere in southeastern Turkey. This model elegantly explains the cause of the rotation and how the Cyprus and the Antalya nappes may have been torn from each other as a result of initial nucleation of opposite-verging thrust faults along an oceanic transform fault. In Figure 16A, the ridge and oceanic transform fault orientations are shown in the pre-rotational geometry (after Morris et al. 2006, fig. 6b). This suggests that the giant nappe front may have been localized on an oceanic transform fault parallel with the one preserved in Cyprus in the form of the Limassol Forest Complex. This is supported by the observation by Morris and Anderson (2002) that sub-vertical dykes strike parallel with the thrust emplacement direction of the ophiolite nappe. If a former fracture zone indeed turned into a subduction zone, the two pieces, namely the Antalya and the Troodos nappes, may have become separated along a Paleopaphos fault in the manner shown in Figure 16B and C. It is also suggested, following Şengör and Yılmaz (1981), that the Antalya nappes may have acquired their extreme curvature later than the main Maastrichtian obduction, during the Cainozoic rethrusting into the Isparta Angle. Figure 16D and E show how a significant change of location of a rotation pole (from AB to CD) between two plates may turn a former fracture zone into a subduction zone with segments displaying opposing facings (for more details on this issue, see Şengör in press). The former plates A and B have been replaced by new plates C and D by the deactivation of the spreading centre between plates A and B, because the pole shift had made the old spreading centres impractical.

pulled down a subduction zone. The only exception to this statement is the Antalya Nappes. There the HP/LT metamorphism occurred in an entirely oceanic environment and the metamorphosed rocks were emplaced onto the continental margin later during the Paleocene as an already assembled package of nappes (Okay and Özgül 1984).

Where flysch or molasse basins end in southeastern Turkey, large germanotype foreland structures first make their appearance. Fontaine (1981) showed that it was during the thrusting of the ophiolitic nappes that the Hazro uplift (Fig. 11) made its first appearance. It is now in the form of a package of thrusts (Fig. 17C) and occupies a position not dissimilar to that of the external massifs in the Alps. The Bozova and the Mardin highs (Fig. 11) were similar but more subdued structures.

As soon as the ophiolites turn the corner in Hatay (the Kızıldağ and the Baer-Bassit ophiolitic massifs), the Damascene germanotype mountain belt appears to its south and that style remains dominant all the way to Libya: we have now entered the classical ground of the Syrian Arc. No large ophiolite nappes are seen to burden the continental margin. Neither are there any HP/LT rocks along the continental margin. In fact, there is no metamorphism at all.

Figure 17 shows the difference between the western Ayyubids and the eastern Ayyubids (the Antalya Nappes being the exception; but they do not affect the Afro-Arabian margin). In the eastern Ayyubids, major ophiolite obduction overwhelmed and totally destroyed the continental margin and choked the infant subduction zone. This is probably a consequence of a long oceanic appendage to the continental crust that pulled it down under the ophiolite, which almost effortlessly rode across the continental margin during the obduction (see Şengör 1990, fig. 17 showing the steps of such an obduction that happened during the late Eocene in New Caledonia: Aubouin et al. 1977; Paris and Lille 1977). Westwards, the Neo-Tethys became narrower and the subduction zone seems to have formed closer to the continental margin. To the north of the Syrian Arc, the length of the oceanic appendage to the continental crust was no longer sufficient to pull the margin down to allow a quiet passage of the ophiolites. Here instead, the nose of the upper plate in the newly-formed juvenile subduction zone hit the base of the continental margin as shown in Figure 17B, because the continental margin was not pulled down to allow the oceanic lithosphere to pass over it, i.e. there was not even a 'nascent' subduction zone here. As long as the rest of the continent was being dragged under the overriding nappes to the east, the African continental margin in the west continued to be pulled along and pushed against the risen nose of the ophiolite nappe, but because it would not

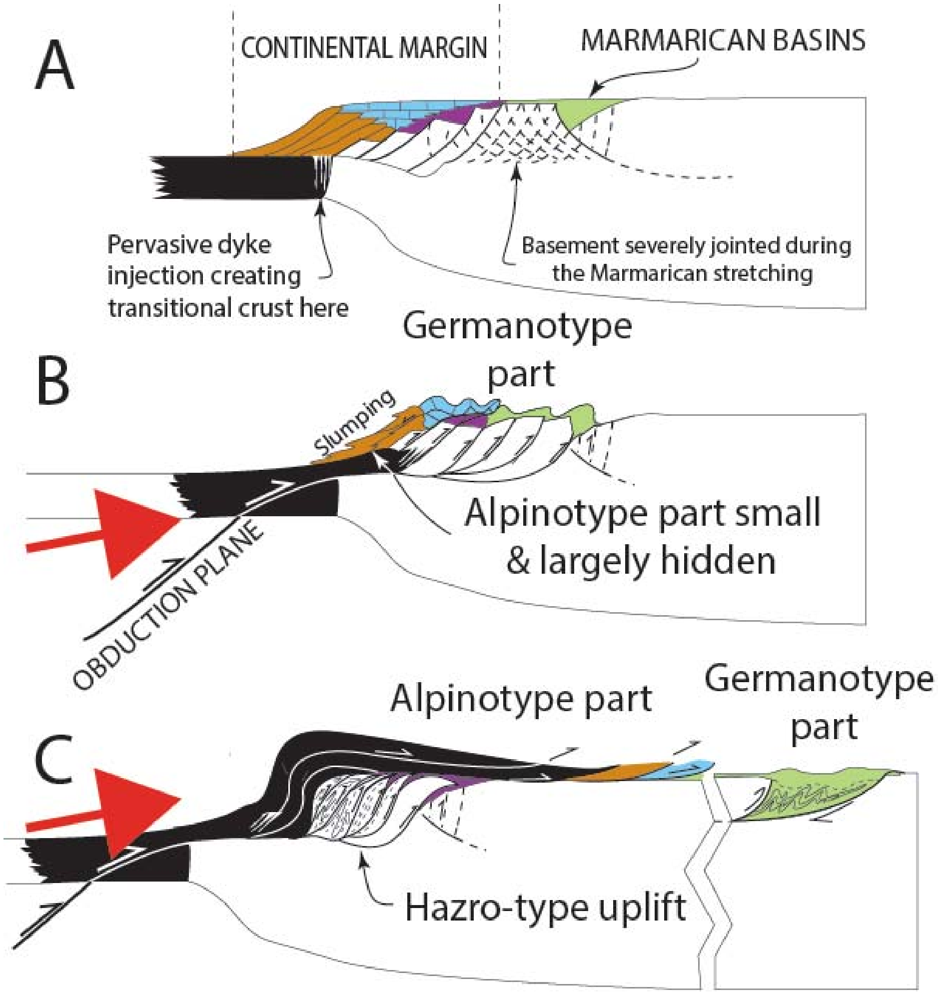


Figure 17. Three schematic cross-sections illustrating the nature of orogeny in the Ayyubid orogen. A is a generalized N-S cross section across the Libyan-Egyptian sector of the orogen before the onset of orogeny. The Marmarican rifting had just ended creating the normal fault-bounded basins of the Marmarican taphrogen. The rifting must have also severely faulted and jointed the basement creating numerous conjugate sets. Orogeny in Libya and northern Egypt is presented as a result of an aborted ophiolite obduction in B. The ophiolite (moving in the direction of the red arrow) just compressed the continental margin and created the structures seen in al-Jabal al-Akhdar, created the gravity anomaly shown in Figure 8, and compressed the former rift basins of the Marmarican taphrogen turning their region into the germanotype parts of the Ayyubids in this area. Starting with Cyprus, the ophiolite was obducted for long distances onto the continental crust of Afro-Arabia, which is illustrated in C. This obduction event created a rich assortment of foreland structures: in Syria, the obduction created the Cretaceous parts of the Palmyran germanotype dextral transpressional system. In Turkey, Iran and Oman, Hazro-type basement thrust packages formed external massifs under the obducting ophiolite, whereas in front of the ophiolite wide flexural flysch and eventually molasse basins formed. Germanotype reactivation of the foreland in these places was minimal, possibly because the early to medial Cretaceous rifting had done little damage to the lithosphere here.

descend, the heavy oceanic lithosphere had no means of overthrusting the continent. Instead it bored its nose into the base of the continental margin, deformed whatever was in front of it and stopped, as soon as the obduction events, i.e. the pulling of the continental crust under the ophiolites farther east, was over.

Because the ophiolite nappe along the Syrian Arc acted as a piston, rather than as a burden on the margin, it compressed the rift basins that had formed during the evolution of the Marmarican taphrogen (Şengör and Natal'in 2001). The same thing happened in the Damascene Arc: a previous aulacogen collapsed under the push of the Troodos nappe which failed to climb onto normal thickness continental crust, but instead compressed the entire Levantine continental margin.

Ophiolite obduction in the east and attempted obduction in the west along the Afro-Arabian margin thus seem to be the cause of the Ayyubid orogeny. The Ayyubids are in fact what Şengör (1990, p. 94) called an 'obduction-type orogen'. Obduction-type orogens are very different from subduction-related orogens (Şengör 1990). In the latter, the main orogenic activity is on the hinterland (or overriding) plate forming arc-related structures including forearc accretionary prisms, fore-arc basins, arc massifs and hinterland fold and thrust belts (in compressional arcs) or marginal basins (in extensional arcs). By contrast, in obduction-type orogens, the main action is in the down-going, i.e. the foreland, plate and it is very short-lived (the duration of the obduction). Almost nothing happens in the hinterland plate, except to ride over the foreland plate. Extensive alpinotype and/or germanotype deformation takes place in the foreland plate with accompanying HP/LT metamorphism and flysch sedimentation. Where obduction fails, as it seems to have done in the western Ayyubids, almost no metamorphism occurs and no large flysch or molasse basins come into being. In such cases, the foreland deformation is almost exclusively germanotype. Obduction-type orogens may pass laterally into collision type and/or subduction type orogens, or even into transpressional ones (Şengör 1990).

As far as we know, the Ayyubids are the largest obduction-related orogen in the world, extending for at least 5000 km from eastern Libya to Oman (and most likely beyond into Pakistan and the Himalaya). That it was obduction-related explains why it was so short-lived. It also explains the very remarkable synchroneity of the events all along its trend. It further explains why it has no associated large meta-

Table 1. Rotation Parameters Describing Past Relative Positions of Plates

Rotated Plate	Fixed Plate	Chron	Age, Ma	Latitude	Longitude	Angle	Source of rotation
Africa	North America	M11	136	66.02	-19.07	-57.81	Sibuet et al. 2012
Africa	North America	M0	125	65.95	-20.46	-54.56	Sibuet et al. 2012
Africa	North America	34o*	118	66.3	-19.9	-54.3	Sibuet and Collette 1991
Africa	North America	34y	83.64	76.81	-20.59	-29.506	Müller et al. 1999
Africa	North America	33o	79.9	78.64	-18.16	-26.981	Müller et al. 1999
North America	Europe	M11	136	69.67	154.26	23.17	Sibuet et al. 2012
North America	Europe	M0	125	69.67	154.26	23.17	Sibuet et al. 2012
North America	Europe	34o*	118	74.1	159.1	24.7	Sibuet and Collette 1991
North America	Europe	34y	83.64	66.54	148.91	19.7	Srivastava et al. 1988
North America	Europe	33o	79.9	65.8	149.9	18.8	Sibuet and Collette 1991
Africa	Europe	M11	136	-45.2676	174.193	44.0133	This paper
Africa	Europe	M0	125	-42.9282	174.4605	38.752	This paper
Africa	Europe	34o	118	-43.181	174.9167	38.3823	This paper
Africa	Europe	34y	83.64	-36.1131	165.6693	17.9221	This paper
Africa	Europe	33o	79.9	-35.4638	166.1001	15.8798	This paper

Ages of chrons are from Gradstein et al. (2012). Positive rotations are counterclockwise.

*These rotations were labeled M0 by Sibuet and Collette (1991) but since they assigned an age of 118 Ma to these rotations, and were following the time scale of Kent and Gradstein (1986), this corresponds to the old edge of chron 34 (34o).

morphic core complexes and marginal fold and thrust belts of any significant size. Finally, it explains the remarkable asymmetry in the distribution of its alpinotype and germanotype parts. It was not only the presence of older rifts that caused the germanotype deformation, but also the strong sideways push, like a piston, of the ophiolite that failed to climb up the margin (or, more correctly, the margin did not come down to receive it) that led to the folding and thrusting of the rift contents (as in the Egyptian Western Desert and in Syria) *and* the platform sedimentary rocks (as in Cyrenaica, Sinai and Israel). Contrary to what Bosworth et al. (2008) wrote, the Egyptian structures are not less deformed than those in eastern Libya; only their style of deformation was different because of the presence of the deep rift basins. This difference is not dissimilar to the difference between the deformation styles in Israel and along the Damascene Arc.

After having seen that a major ophiolite obduction was the cause of the Ayyubid orogeny, we may now ask what triggered the onset of the obduction. The popular position has always been to relate it to the sudden northerly swing of Africa with respect to Europe. To test that idea, we have replotted, using the data in Srivastava et al. (1988), Sibuet and Collette (1991) and Sibuet et al. (2012) the North America to Europe rotations and M0 and M11 with respect to Eurasia and Müller et al. (1999) for the 34o, 34y and 33o for Africa-North America. The result is shown in Table I and Figure 18. The rotations in Müller et al. (1999) have uncertainties included, but the other publications we used do not and because of that we cannot show the 95% confidence in our motion paths. Actually for the North Atlantic (Europe to North America) there are no marine magnetic anomalies of Mesozoic age so the M0 and M11 rotations are based on various assumptions by different authors in dealing with the pre-breakup continental extension. Although for Iberia there are M-series anomalies, regrettably, Iberia was moving by itself at the time and not with the rest of Europe. Therefore they are of no use to us in drawing the Africa–Europe relative motion vectors.

What Figure 18 clearly shows is that the abrupt northerly turn of Africa with respect to Europe was after 84 million years ago, i.e. during the most intense phases of the Ayyubid orogeny. Moreover, during the time interval from 34y to 33o the violet-coloured point was moving with a speed of some 45.4 km/m.y. and the pink point was moving with a speed of 52.2 km/m.y. This was the fastest that Africa ever moved with respect to Europe during the late Cretaceous. But the motion of Africa with respect to Europe was northeasterly rather than north-south and the Ayyubid orogeny had started much earlier than the 34y time, as Figures 6 and 10 clearly show. In Oman, obduction probably started during the Coniacian, i.e. some 2 to 3 Ma *earlier* than the sharp northerly swing of Africa with respect to Europe. However a possible swivel of Africa between M11 and M0 times and another change of course from M0 to 34o times may well have been responsible for initiating the Omani subduction. Figure 10 shows that shortening in Egypt most likely started already during the Turonian. Even in Cyrenaica, where the timing of the main Ayyubid deformations can be most narrowly bracketed, they were mostly *before* the main swing of Africa with respect to Europe that began at 84 Ma. So, how are we to account for this discrepancy?

The first step is to recognize that between Africa and Eurasia there were other, now entirely vanished plates[11]. We can only find out about their motions through the thick haze of structural interpretation in the field, assisted even more feebly by seismic tomography. John F. Dewey's distress-

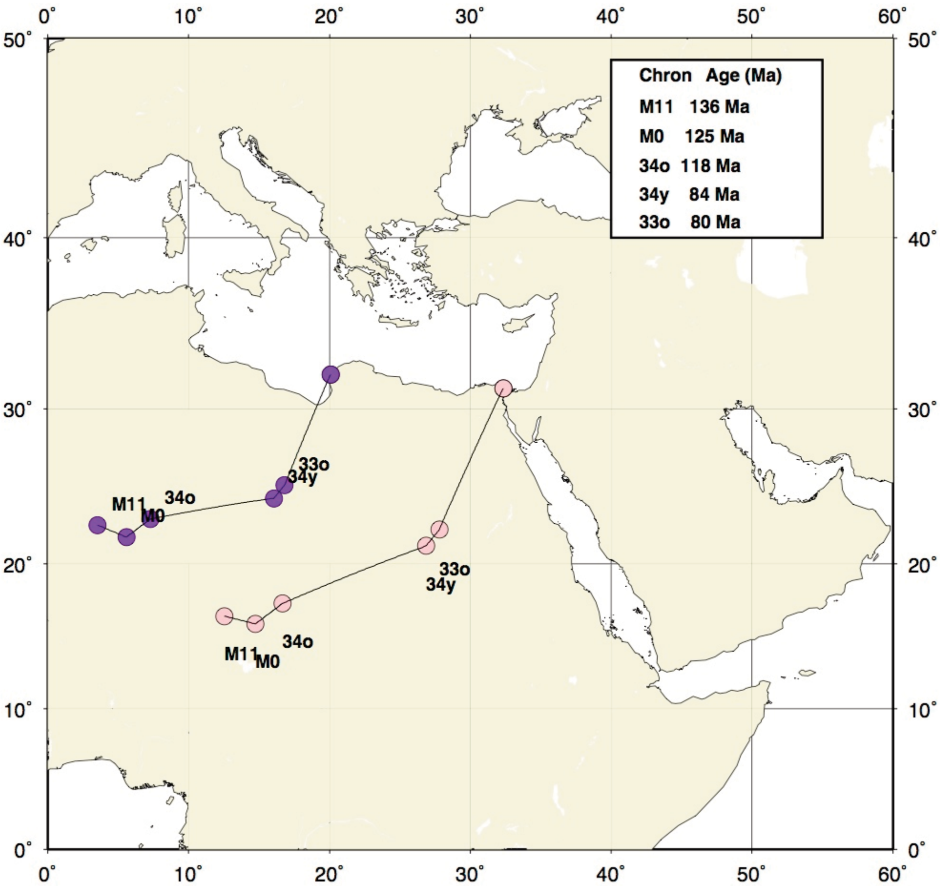


Figure 18. The movement of two points fixed to the North African shore between 136 (Valanginian) to 80 (Campanian) Ma. The chronological scale used for the magnetic anomalies is from Ogg (2012). Note that between M0 and 34y times, Africa moves more easterly than northerly, yet this is the time when the Ayyubid deformations begin in Egypt and in Oman. The major ophiolite obduction event in Oman began before 34y time, i.e. when Africa was moving eastward. That the other major obduction events along the Ayyubids fall between the 34y and 33o times might have been thought to explain nicely why the Ayyubid subduction zone tore westward, but the high pressure metamorphism in SE Turkey and in Antalya and the compressional events in Egypt had started well before this time. Therefore even the propagation westward of the Omani subduction zone occurred well before Africa's motion with respect to Europe became more northerly.

ing message, so masterfully argued in his 1975 paper (Dewey 1975), that plate tectonics destroys evidence, stares us in the face in its full force here. And yet we may not be so hapless. We noted above that the Ayyubid subduction probably began in the east, in Oman, at some time in the medial Cretaceous and the subduction zone tore westwards with great alacrity, rapidly turning into an obduction front. In Oman, the ages of HP/LT metamorphic rocks give a broader age span from 100 Ma to 80 Ma than the HP/LT rocks farther west, but subduction metamorphism was going on in the entire stretch from Alanya to Oman during the Santonian. There is hardly a difference in the ages of the younger HP/LT metamorphic rocks in Oman and those in southeastern Turkey and the same is probably true for the Antalya Nappes within the Alanya Window (cf. Okay and Özgül 1984). This means that the nascent subduction zone was pulling the whole of Afro-Arabia down under some Tethyan plate and causing orogeny above it even before Africa turned northward with respect to Eurasia. Only in north Africa there was not enough slab to pull the continental margin down to allow a full-blown ophiolite obduction as in other segments of the Ayyubids, but the general relative motion of Africa with respect to the unknown Tethyan plate was sufficient to maintain relative motion across the Syrian Arc in the Levant and north Africa. In the Levant margin, the subduction zone was there, but far away in Antalya and in Cyprus. In these places the continental margin was simply shortened horizontally without being pulled down into a giant shear zone to cause alpinotype orogeny.

The Ayyubid orogen was thus an entirely ophiolite obduction-driven orogen related to a plate that today has entirely vanished (except for bits still present in the eastern Mediterranean). Are there similar orogens of similar size elsewhere? The only ones we can think of are the Ordovician ophiolite-driven orogeny in Scotland and Newfoundland (cf. Dewey 2005) and the Ordovician obduction in the Urals (Puchkov 2002). Whether a similar medial Jurassic one existed in Tibet is as yet open to question. The early Paleozoic germanotype deformations of the Boothia uplift in central northern Canada (one during the Arenigian, roughly about 485 Ma, and the other during the Caradocian, about 450 to 454 Ma: Okulitch et al. 1986), before the major deformation set in during the Devonian synchronously with the major collision events along the Appalachian/Caledonian Orogen (Kevin Burke, pers. comm., 1980) may very well represent the germanotype foreland features of the alpinotype ophiolite obduction events along the Laurentian margin during the so-called 'Taconic Orogeny' in medial Ordovician time, circa 470 Ma (Williams 1975; Dewey and Casey 2013). Another candidate is the vast Ordovician ophiolite obduction area of the Urals (see Puchkov 2002, especially fig. 6), where the Orenburg Rift (see Nikishin et al. 1996) is the only coeval deformational structure at high angles to the obduction front. There are many basin and ridge structures of dimensions similar to the Damascene Arc in front of the Urals, but their detailed structural evolution still awaits analysis.

CONCLUSIONS

The Turonian to Maastrichtian shortening structures from Cyrenaica, northern Egypt, coastal Levant and the

Palmyra Mountains forming Krenkel's (1924) Syrian and Damascene arcs and Ricou's (1971) peri-Arabian ophiolitic crescent constitute a major, obduction-driven orogenic belt along the north-eastern and northern margin of Afro-Arabia as well as in the Antalya region of southwestern Turkey, herein called the Ayyubids. The north African and the Levantine parts of this major orogenic belt resulted from an aborted ophiolite obduction (the only other example of an 'aborted' obduction we know of is in the southern Chilean Sarmiento ophiolite complex [de Wit and Stern 1981, especially fig. 3, section YY']), which, elsewhere along the chain, was successful. Although the most intense phases of the orogeny during the Santonian were coeval with a change in the motion of Africa with respect to Europe to a more northerly orientation than before, the Ayyubid orogeny had started earlier than this time by the onset of subduction in Oman during the medial Cretaceous and the extremely rapid propagation of this subduction zone westward also during the medial Cretaceous. It seems that, with the exception of the Antalya Nappes, ophiolite obduction was already over everywhere by late Campanian time, when Africa turned onto a more northerly track. The explanation of this mismatch between the motion of Africa with respect to Europe and the timing of the deformational events in the Ayyubids is that a third plate (or more) existed between Europe and Africa during the Cretaceous.

Obduction-driven orogenies can be spatially as long as any subduction or collision-driven mountain range, although they do not create fully developed magmatic arcs or collisional magmatic cores as in collisional orogens. If the ophiolite obduction is aborted while convergence is going on, major germanotype foreland structures may form, especially if the foreland had been disrupted by rifting just before the orogeny. Obduction-driven orogens, which habitually take a very short time to form (cf. Dewey 2005), may become totally obliterated by subsequent subduction- or collision-driven orogenies and their record may be misinterpreted as an earlier phase in the evolution of these later mountain ranges. The Ayyubids have thus long been thought of part of the Alpides. We now recognize that it is a separate orogen on its own right.

ACKNOWLEDGEMENTS

We are extremely pleased and honoured to contribute this paper in memory of a great geologist, Professor Harold ('Hank') Williams, a master of ophiolite geology, among other subjects. For Şengör, the pleasure and honour are related to a long friendship with Professor Williams. Williams was among his earliest field instructors. When Şengör was only a second-year student, his teacher John F. Dewey sent him to work under Williams and W.S.F. Kidd for a month in Newfoundland. That was his first encounter with ophiolites in the field. That initial acquaintance turned into a life-long friendship, during which he continued to learn from Williams. We thank Jim Hibbard, who was the other student with Şengör during that memorable field work in Newfoundland, for inviting us to contribute to the Williams' memorial volume. Şengör's friends in Libya, especially Emin Yanılmaz and Ali El-Arnauti, who took him around in Cyrenaica, showed him the critical outcrops and provided him with a massive amount of literature, probably would have been co-authors in this paper, had the unfortunate events that befell both Libya and Syria not led to his losing their tracks. We heard that Emin Yanılmaz safely left Syria to return to Turkey, but we have not been able to get any news of Ali El-Arnauti. We can only hope that he managed to survive safely the destruction of his country. Professor Hubert Whitechurch kindly supplied us with a pre-print of his important new paper on the Kermanshah ophiolites in Iran and Professor Leigh H. Royden discussed with us her work with Oliver Jagoutz in the Himalayan Neo-Tethys that much encouraged us in our interpretations. She was also helpful in evaluating the Libyan gravity anomalies. Professor Ali Polat helped with literature research in Canada. We thank Professors Brian P. Wernicke , B. Clark Burchfiel and an anonymous reviewer for a critical reading of an earlier draft of this paper. Jim Hibbard, the guest editor of Hank's volume and Brendan Murphy, the editor-in-chief of *Geoscience Canada* also made us work hard to make our presentation clearer. We thank Cindy Murphy for her great assistance during the final production of the paper.

REFERENCES

Abdelghany, O., 2003, Late Campanian–Maastrichtian foraminifera from the Simsima Formation on the western side of the Northern Oman Mountains: Cretaceous Research, v. 24, p. 391–405, http://dx.doi.org/10.1016/S0195-6671(03)00051-X.

Abd El-Motaal, E., and Kusky, T.M., 2003, Tectonic evolution of the intraplate S-shaped Syrian Arc fold-thrust belt of the Middle East region in the context of plate tectonics: The Third International Conference on the Geology of Africa, Assiut, Egypt, v. 2, p. 139–157.

Al-Heety, E.A., 2013, Seismicity and seismotectonics of Libya: as an example of intraplate environment: Arabian Journal of Geosciences, v. 6, p. 193–204, http://dx.doi.org/10.1007/s12517-011-0347-y.

Allerton, S., and Vine, F.J., 1991, Spreading evolution of the Troodos ophiolite, Cyprus: Geology, v. 19, p. 637–640, http://dx.doi.org/10.1130/0091-7613(1991)019<0637:SEOTTO>2.3.CO;2.

Al-Riyami, K., Robertson, A.H.F., Xenophontos, C., Danelian, T., and Dixon, J., 2000, Tectonic evolution of the Mesozoic Arabian passive continental margin and related ophiolite in Baer-Bassit region (NW Syria), *in* Parnayides, I., Xenophontos, C., and Malpas, J., *eds*., Proceedings of the Third International Conference on the Geology of the Eastern Mediterranean, Nicosia, Cyprus, 23-26 September 1998: Geological Survey Department, Ministry of Agriculture, Natural Resources and Environment, p. 61–81.

Al-Riyami, K., Robertson, A., Dixon, J., and Xenophontos, C., 2002, Origin and emplacement of the Late Cretaceous Baer-Bassit ophiolite and its metamorphic sole in NW Syria: Lithos, v. 65, p. 225–260, http://dx.doi.org/10.1016/S0024-4937(02)00167-6.

Al-Saad, D., Sawaf, T., Gebran, A., Barazangi, M., Best, J.A., and Chaimov, T.A., 1992, Crustal structure of central Syria: The intracontinental Palmyride mountain belt: Tectonophysics, v. 207, p. 345–358, http://dx.doi.org/10.1016/0040-1951(92)90395-M.

Argand, E., 1916, Sur l'arc des Alpes Occi-

dentales: Eclogae Geologicae Helvetiae, v. 14, p. 145–191.
Argand, E., 1924, La tectonique de l'Asie: Comptes-rendus du XIIIe Congrès Géologique International 1922, Vaillant-Carmanne, Liège, p. 171–372.
Ashgirei, G.D., 1956, Strukturnaya Geologiya: Izdatelstvo Moskovskogo Universiteta, Moskva, 339 p.+ 1 errata slip
Aubouin, J., Mattauer, M., and Allègre, C., 1977, La couronne ophiolitique périaustralienne: un charriage océanique représentatif des stades précoces de l'évolution alpine: Comptes Rendus hébdomadaires de l'Académie des Sciences (Paris), v. 285, p. 953–956.
Avnimelech, M., 1963, Max Blanckenhorn (16.4.1861-13.1.1947): Israeli Journal of Earth-Sciences, v. 12, p. 1–11.
Babaei, A., Babaie, H.A., and Arvin, M., 2005, Tectonic evolution of the Neyriz ophiolite, Iran: an accretionary prism model: Ofioliti, v. 30, no. 2, p. 65–74, http://dx.doi.org/10.4454/ofioliti.v30i2.241.
Bağcı, U., and Parlak, O., 2009, Petrology of the Tekirova (Antalya) ophiolite (southern Turkey): evidence for diverse magma generations and their tectonic implications during Neotethyan-subduction: International Journal of the Earth Sciences, v. 98, p. 387–405, http://dx.doi.org/10.1007/s00531-007-0242-7.
Bağcı, U., Parlak, O., and Höck, V., 2005, Whole-rock mineral chemistry of cumulates from the Kızıldağ (Hatay) ophiolite (Turkey): clues for multiple magma generation during crustal accretion in the southern Neotethyan ocean: Mineralogical Magazine, v. 69, p. 53–76, http://dx.doi.org/10.1180/0026461056910234.
Banerjee, S., 1980, Stratigraphic Lexicon of Libya: Department of Geological Researches and Mining Bulletin no. 13, Socialist People's Libyan Arab Jamahiriyah Industrial Research Centre, Tripoli, xviii+300+2 p.
Barr, F.T., 1968, Upper Cretaceous stratigraphy of Jabal al Akhdar, northern Cyrenaica, *in* Barr, F.T., *ed*., Geology and Archaeology of Northern Cyrenaica, Libya: Petroleum Exploration Society of Libya, Tenth Annual Field Conference, Holland-Breumelhof N.V., Amsterdam, p. 131–147.
Barr, F.T., and Berggren, W.A., 1980, Lower Tertiary biostratigraphy and tectonics of northeastern Libya, *in* Salem, M.J., and Busrewil, M.T., *eds*., The Geology of Libya, v. III Second Symposium on the Geology of Libya: Academic Press, London, p. 163–192.
Ben Ferjani, A., Burollet, P.F., and Mejri, F., 1990, Petroleum Geology of Tunisia: Entreprise Tunisienne d'Activités Pétrolières, Tunis, 194 p., 5 foldout plates.
Blanckenhorn, M., 1893, Die Strukturlinien Syriens und des Rothen Meeres, *in* Festschrift. Ferdinand Freiherrn von Richthofen zum Sechzigsten Geburtstag am 5. Mai 1893. Dargebracht von seinen Schülern: Dietrich Reimer, Berlin, p. 115–180, plate of cross-sections and a geological map at 1:2,400,000 scale.
Blanckenhorn, M., 1912a, Naturwissenschaftliche Studien am Toten Meer und im Jordantal. Bericht über eine im Jahre 1908 (im Auftrage S.M. des Sultans der Türkei Abdul Hamid II. und mit Unterstützung der Berliner Jagor-Stiftung) unternommene Forschungsreise in Palästina: R. Friedländer & Sohn, Berlin, VII+478 p.+2 foldout plates in back pocket.
Blanckenhorn, M., 1912b, Kurzer Abriss der Geologie Palästinas—Begleitworte zu einer neuen geologischen Karte von Palästina: J.C. Hinrichs'sche Buchhandlung, Leipzig, 27 p., 2 foldout plates[12].
Blanckenhorn, M., 1915, Syrien, Arabien und Mesopotamien, *in* Steinmann, G., and Wilckens, O., *eds*., Handbuch der Regionalen Geologie, v. 4, 159 p., 4 foldout plates.
Blanckenhorn, M., 1925, Der sogenannte "Syrische Bogen" und die erythräische Geosynklinale: Zeitschrift der Deutschen Geologischen Gesellschaft, v. 77, p. 206–226.
Bosworth, W., Guiraud, R., and Kessler, L.G., 1999, Late Cretaceous (ca. 84 Ma) compressive deformation of the stable platform of northeast Africa (Egypt): Far-field stress effects of the "Santonian event" and origin of the Syrian arc deformation belt: Geology, v. 27, p. 633–636, http://dx.doi.org/10.1130/0091-7613(1999)027<0633:LCCMCD>2.3.CO;2.
Bosworth, W., El-Hawat, A.S., Helgeson, D.E., and Burke, K., 2008, Cyrenaican "shock absorber" and associated inversion strain shadow in the collision zone of northeast Africa: Geology, v. 36, p. 695–698, http://dx.doi.org/10.1130/G24909A.1.
Braud, J., 1987, La Suture du Zagros au Niveau de Kermanshah (Kurdistan iranien): Reconstitution paléogéographique, évolution géodynamique, magmatique et structurale: Thèse présentée pour obtenir le grade de Docteur ès Sciences naturelles, Université de Paris-Sud, Centre d'Orsay, 489 p.+[1] p.
Brew, G., Barazangi, M., Al-Maleh, A.K., and Sawaf, T., 2001, Tectonic and geologic evolution of Syria: GeoArabia, v. 6, p. 573–616+ 1 foldout plate.
Burke, K., and Dewey, J.F., 1974, Two plates in Africa during the Cretaceous?: Nature, v. 249, p. 313–316, http://dx.doi.org/10.1038/249313a0.
Burke, K., Dessauvagie, T.F.J., and Whiteman, A.J., 1971, Opening of the Gulf of Guinea and geological history of the Benue Depression and Niger delta: Nature Physical Science, v. 233, p. 51–55, http://dx.doi.org/10.1038/physci233051a0.
Burke, K., Dessauvagie, T.F.J., and Whiteman, A.J., 1972, Geological history of the Benue Valley and adjacent areas, *in* Dessauvagie, T.F.J., and Whiteman, A.J., *eds*., African Geology: University of Ibadan Press, Ibadan, p. 187–205.
Campbell, A.S., 1968, The Barce (al Marj) earthquake of 1963, *in* Barr, F.T., *ed*., Geology and Archaeology of Northern Cyrenaica, Libya: Petroleum Exploration Society of Libya, Tenth Annual Field Conference, Holland-Breumelhof N.V., Amsterdam, p. 183–195.
Carton, H., 2005, Études Tectoniques en Mediterranée Orientale par Analyse de Données de Sismique Réflexion: Mer de Marmara (Bassin de Çınarcık) et Marge du Liban: École Doctorale des Sciences de la Terre, Institut de Physique du Globe de Paris, Laboratoire de Géosciences Marines, 297 p.
Chaimov, T.A., Barazangi, M., Al-Saad, D., Sawaf, T., and Gebran, A., 1990, Crustal shortening in the Palmyride Fold Belt, Syria, and implications for movement along the Dead Sea Fault System: Tectonics, v. 9, p. 1369–1386, http://dx.doi.org/10.1029/TC009i006p01369.
Chan, Gaven H.-N., Malpas, J., Xenophontos, C., and Lo, Ching-Hua, 2007, Timing of subduction zone metamorphism during the formation and emplacement of Troodos and Baer-Bassit ophiolites: insights from ^{40}Ar–^{39}Ar geochronology: Geological Magazine, v. 144, p. 797–810, http://dx.doi.org/10.1017/S0016756807003792.
de Beauregard, J., Bouchardeau, A., Roch, E., and Schwoerer, P., 1952, La chaîne crétacée de la Bénoué: Comptes Rendus hébdomedaires de l'Académie des Sciences (Paris), v. 234, p. 1697–1699.
de Cizancourt, H., 1948, La tectonique profonde de la Syrie et du Liban—essai d'interprétation géologique de

mesures gravimetriques, *in* Dubertret, L., *ed.*, Notes et Mémoires, v. 4: Études Géologiques et Géographiques sur le Liban, la Syrie et le Moyen-Orient, p. 156–190.

de Sitter, L.U., 1962, Structural development of the Arabian shield in Palestine: Geologie en Mijnbouw, v. 41, No. 3, p. 116–124.

de Vaumas, É., 1950, La structure du Proche-Orient. Essai de synthèse: Bulletin de la Société Royale de Géographie d'Égypte, v. 23, p. 265–320, plates 1–9.

Dewey, J.F., 1975, Finite plate evolution: some implications for the evolution of rock masses at plate margins: American Journal of Science, v. 275-A (John Rodgers volume), p. 260–284.

Dewey, J.F., 1980, Episodicity, sequence and style at convergent plate boundaries: Geological Association of Canada Special Publication, v. 20, p. 553–576.

Dewey, J.F., 2005, Orogeny can be very short: Proceedings of the National Academy of Sciences of the USA, v. 102, p. 15286–15293, http://dx.doi.org/10.1073/pnas.0505516102.

Dewey, J.F., and Casey, J.F., 2013, The sole of an ophiolite: the Ordovician Bay of Islands Complex, Newfoundland: Journal of the Geological Society, v. 170, p. 715–722, http://dx.doi.org/10.1144/jgs2013-017.

de Wit, M.J., and Stern, C.R., 1981, Variations in the degree of crustal extension during formation of a back arc basin: Tectonophysics, v. 72, p. 229–260, http://dx.doi.org/10.1016/0040-1951(81)90240-7.

Dilek, Y., and Delaloye, M., 1992, Structure of the Kızıldag ophiolite, a slow-spread Cretaceous ridge segment north of the Arabian promontory: Geology, v. 20, p. 19–22, http://dx.doi.org/10.1130/0091-7613(1992)020<0019:SOTKOA>2.3.CO;2.

Dubertret, L., 1930, Note préliminaire sur la structure géologique des états du Levant sous mandat français: Comptes Rendus Sommaire de la Société Géologique de France, no. 6, Séance du 17 mars 1930, p. 43–45.

Dubertret, L., 1932, Les formes structurales de la Syrie et de la Palestine: leur origine: Comptes Rendus hébdomadaires des Seances de l'Académie des Sciences (Paris), v. 195, p. 65–66.

Dubertert, L., 1934, Présentation d'une carte géologique au 1,000,000e de la Syrie et du Liban suivie d'une étude de l'hydrologie et de l'hydrographie: Revue de Géographie Physique et de Géologie Dynamique, v. 6, no. 4, p. III+160 p.+5 foldout plates.

Dubertret, L., 1948, Aperçu de géographie physique sur le Liban, l'Anti-Liban et la Damascene, *in* Dubertret, L., *ed.*, Notes et Mémoires, v. 4: Études Géologiques et Géographiques sur le Liban, la Syrie et le Moyen-Orient, p. 191–226, and 1 foldout coloured physical geographical map.

Dubertret, L., 1951, (Sur la géographie du Proche Orient) La structure du Moyen Orient d'après Étienne de Vaumas: Revue de Géographie de Lyon, v. 26, p. 367–374, http://dx.doi.org/10.3406/geoca.1951.2707.

Dumont, J.F., Gutnic, M., Marcoux, J., Monod, O., and Poisson, A., 1972, Le Trias des Taurides occidentales (Turquie) Définition du bassin pamphylien: Un nouveau domaine à ophiolithes à la marge externe de la chaîne taurique: Zeitschrift der Deutschen Geologischen Gesellschaft, v. 123, p. 385–409.

Duronio, P., Dakshe, A., and Bellini, E., 1991, Stratigraphy of the Offshore Cyrenaica (Libya), *in* Salem, M., Hammuda, O.S., and Eliagoubi, B.A., *eds.*, The Geology of Libya v. IV: Elsevier, Amsterdam, p. 1589–1620.

Elakkari, T.S., 2005, Structural Configuration of the Sirt Basin: Unpublished Master of Science Thesis, International Institute for Geo-Information Science and Earth Observation, Enschede, The Netherlands, Geological Resources Management and Environmental Geology (GRMEG), 58 p.

El-Arnauti, A., Lawrence, S.B., Mansouri, A.L., Şengör, A.M.C., Soulsby A., and Hassan, H., 2008, Structural Styles in NE Libya, *in* Salem, M.J., and El-Hawat, A.S., *eds.*, The Geology of East Libya: Gutenberg Press Ltd., Malta, v. 4, p. 153–178.

El-Shazly, A.K., and Coleman, R.G., 1990, Metamorphism in the Oman Mountains in relation to the Semail ophiolite emplacement, *in* Robertson, A.H.F., Searle, M.P., and Ries, A.C., *eds.,* The geology and tectonics of the Oman Region : Geological Society, London, Special Publications, v. 49, p. 473–493, http://dx.doi.org/10.1144/GSL.SP.1992.049.01.30.

Engelder, T., 1979, The nature of deformation within the outer limits of the Central Appalachian foreland fold and thrust belt in New York State: Tectonophysics, v. 55, p. 289–310, http://dx.doi.org/10.1016/0040-1951(79)90181-1.

Engelder, T., and Engelder, R., 1977, Fossil distortion and décollement tectonics of the Appalachian Plateau: Geology, v. 5, p. 457–460, http://dx.doi.org/10.1130/0091-7613(1977)5<457:FDADTO>2.0.CO;2.

Eyal, Y., 2011, The Syrian Arc Fold System: age and rate of folding (abstract): Geophysical Research Abstracts, v. 13, EGU2011-7401, EGU General Assembly 2011.

Flexer, A., Hirsch, F., and Hall, J.K., 2005, Tectonic evolution of Israel, *in* Hall, J.K., Krasheninnikov, V.A., Hirsch, F., Benjamini, C., and Flexer, A., *eds.*, Geological Framework of the Levant, v. II: The Levantine Basin and Israel: Historical Productions-Hall, Jerusalem, p. 523–537.

Fontaine, J.-M., 1981, La Plate-Forme Arabe et sa Marge Passive au Mesozoique: l'Exemple de Hazro (S.E. Turquie): Thèse presentée à l'Université de Paris-Sud, Centre d'Orsay pour obtenir le titre de Docteur 3e Cycle mention sedimentologie, xii+370 p.+1 foldout map.

Freeth, S.J., 1978, A model for tectonic activity in West Africa and the Gulf of Guinea during the last 90 m.y. based on membrane tectonics: Geologische Rundschau, v. 67, p. 675–687, http://dx.doi.org/10.1007/BF01802811.

Gardosh, M.A., and Druckman, Y., 2006, Seismic stratigraphy, structure and tectonic evolution of the Levantine Basin, offshore Israel, *in* Robertson, A.H.F., and Mountrakis, D., *eds.*, Tectonic Development of the Eastern Mediterranean Region: Geological Society, London, Special Publications, v. 260, p. 201–227, http://dx.doi.org/10.1144/GSL.SP.2006.260.01.09.

Geological Map of Libya, 1985, Geological Map of Libya: Socialist People's Libyan Arab Jamahiriya Industrial Research Centre, Geological Researches and Mining Department, map first edition, scale: 1:1,000,000, 4 sheets+ 1 booklet of legends.

Goodchild, R.G., 1968, Earthquakes in ancient Cyrenaica, *in* Barr, F.T., *ed.*, Geology and Archaeology of Northern Cyrenaica, Libya: Petroleum Exploration Society of Libya, Tenth Annual Field Conference, Holland-Breumelhof N.V., Amsterdam, p. 41–44.

Gradstein, F.M., Ogg, J.G., Schmitz, M.D., and Ogg, G.M., *eds.*, 2012, The Geologic Time Scale 2012: Elsevier, Amsterdam, v. 1–2, 1144 p.

Gregory, J.W., 1911, The geology of Cyrenaica: Quarterly Journal of the Geological Society, v. 67, p. 572–615,

http://dx.doi.org/10.1144/GSL.JGS.1911.067.01-04.22.
Gregory, J.W., 1916, Cyrenaica: The Geographical Journal for May 1916, p. 321–345+1 map *hors-texte*.
Hanna, S.S., 1990, The Alpine deformation of the Central Oman Mountains, *in* Robertson, A.H.F., Searle, M.P., and Ries, A.C., *eds*., The Geology and Tectonics of the Oman Region: Geological Society, London, Special Publications, v. 49, p. 341–359, http://dx.doi.org/10.1144/GSL.SP.1992.049.01.21.
Hantar, G., 1990, North Western Desert, *in* Said, R., *ed*., The Geology of Egypt: Egyptian General Petroleum Corporation, Conoco Hurghada Inc., and Repsol Exploration, S.A., A.A. Balkema, Rotterdam, p. 293–319.
Hey, R.W., 1968a, The Quaternary geology of the Jabal al Akhdar coast, *in* Barr, F.T., *ed*., Geology and Archaeology of Northern Cyrenaica, Libya: Petroleum Exploration Society of Libya, Tenth Annual Field Conference, Holland-Breumelhof N.V., Amsterdam, p. 159–165.
Hey, R.W., 1968b, The geomorphology of the Jabal al Akhdar and adjoining areas, *in* Barr, F.T., *ed*., Geology and Archaeology of Northern Cyrenaica, Libya: Petroleum Exploration Society of Libya, Tenth Annual Field Conference, Holland-Breumelhof N.V., Amsterdam, p. 167–171.
Holmes, W.H., 1876, Report on the geology of the northwestern portion of the Elk Range, *in* Hayden, F.V., *ed*., Annual Report of the United States Geological and Geographical Survey of the Territories, Embracing Colorado and Parts of Adjacent Territories; being a Report of Progress of the Exploration for the Years 1874: Government Printing Office, Washington, p. 59–71+ 8 unnumbered plates.
Hull, E., 1886, The Survey of Western Palestine—Manual of the Physical Geography and Geology of Arabia Petræa, Palestine, and Adjoining Districts with Special reference to the Mode of Formation of the Jordan-Arabah Depression and the Dead Sea: The Committee of the Palestine Exploration Fund, Adelphi, viii+[i]+145 p.+ 2 coloured foldout maps.
Janessary, M.R., and Whitechurch, H., 2008, The birth of an oceanic crust at the foot of Gondwana margin (Neyriz ophiolite, Iran): Geosciences, v. 17, p. 40–49.
Jarrard, R.D., 1986, Relations among subduction parameters: Reviews of Geophysics, v. 24, p. 217–284, http://dx.doi.org/10.1029/RG024i002p00217.
Jassim, S.Z., and Goff, J.C., *eds*., 2006, Geology of Iraq: Dolin, Prague and Moravian Museum, Brno, 341 p.+1 errata sheet.
Jenkins, D.A., 1990, North and Central Sinai, *in* Said, R., *ed*., The Geology of Egypt: Egyptian General Petroleum Corporation, Conoco Hurghada Inc. and Repsol Exploracion, S.A., A.A. Balkema, Rotterdam, p. 361–380.
Karaoğlan, F., Parlak, O., Klötzli, U., Thöni, M., and Koller, F., 2013, U–Pb and Sm–Nd geochronology of the Kızıldağ (Hatay, Turkey) ophiolite: implications for the timing and duration of suprasubduction zone type oceanic crust formation in the southern Neotethys: Geological Magazine, v. 150, p. 283–299, http://dx.doi.org/10.1017/S0016756812000477.
Kempler, D., 1998, Eratosthenes seamount: the possible spearhead of incipient continental collision in the Eastern Mediterranean, *in* Robertson, A.H.F., Emeis, K.-C., Richter, C., and Camerlenghi, A., *eds*., Proceedings of the Ocean Drilling Program, Scientific Results, v. 160, p. 709–721.
Kent, D.V., and Gradstein, F.M., 1986, A Jurassic to Recent chronology, *in*, Vogt, P.R., and Tucholke, B.E., *eds*., The western North Atlantic Region: Geological Society of America, Boulder, CO, The Geology of North America, v. M, p. 351–378.
Kerdany, M.T., and Cherif, O.H., 1990, Mesozoic, *in* Said, R., *ed*., The Geology of Egypt: Egyptian General Petroleum Corporation, Conoco Hurghada Inc. and Repsol Exploracion, S.A., A.A. Balkema, Rotterdam, p. 407–449.
Ketin, İ., 1966, Tectonic units of Anatolia (Asia Minor): Bulletin of the Mineral Research and Exploration Institute of Turkey, no. 66, p. 23–34+3 foldout plates.
Kinder, H., and Hilgemann, W., 1982, Atlas zur Weltgeschichte, 2. erweiterte Auflage: R. Piper & Co. München, v. 1, 287 p.
Kleinsmiede, W.J.F., and van den Berg, N.J., 1968, Surface geology of the Jabal al Akhdar, northern Cyrenaica, Libya, *in* Barr, F.T., *ed*., Geology and Archaeology of Northern Cyrenaica, Libya: Petroleum Exploration Society of Libya, Tenth Annual Field Conference, Holland-Breumelhof N.V., Amsterdam, p. 115–123.
Klen, L., 1974, Geological Map of Libya 1: 250,000 Explanatory Booklet Sheet: Benghází NI 34-14: Libyan Arab Republic, Industrial Research Centre, 56 p.+7 pp.+6 photographic plates.
Kober, L., 1914a, Die Bewegungsrichtung der alpinen Deckengebirge des Mittelmeeres: Dr. Petermanns Mitteilungen aus Justus Perthes' Geographischer Anstalt, Jahrgang 60, Mai-Heft, p. 230–256 + Plates 36 and 37.
Kober, L., 1914b, Alpen und Dinariden: Geologische Rundschau, v. 5, p. 175–204, http://dx.doi.org/10.1007/BF01768569.
Kober, L., 1915, Geologische Forschungen in Vorderasien I. Teil A. Das Taurusgebirge B. Zur Tektonik des Libanon: Denkschriften der mathematisch-naturwissenschaftlichen Klasse der kaiserlichen Akademie der Wissenschaften in Wien, v. 91, p. 379–427 + 3 plates.
Kober, L., 1921, Der Bau der Erde: Gebrüder Borntraeger, Berlin, [II]+324 p.+ 1 foldout map.
Krenkel, E., 1924, Der Syrische Bogen: Centralblatt für Mineralogie, Geologie und Paläontologie, Jahrgang 1924, p. 274–313.
Krenkel, E., 1925, Geologie Afrikas—Erster Teil: Geologie der Erde, Gebrüder Borntraeger, Berlin, X+461 p.+22 plates.
Krenkel, E., 1928, Geologie Afrikas—Zweiter Teil: Geologie der Erde, Gebrüder Borntraeger, Berlin, XII p.+p. 463–1000+15 plates.
Krenkel, E., 1934, Geologie Afrikas—Dritter Teil, erste Hälfte: Geologie der Erde, Gebrüder Borntraeger, Berlin, VIII p.+p. 1003–1304+2 plates.
Krenkel, E., 1938, Geologie Afrikas—Dritter Teil (Schluß des Werkes): Geologie der Erde, Gebrüder Borntraeger, Berlin, VIII p.+p. 1305–1918+10 plates.
Krenkel, E., 1943, Struktur und Relief Nordafrikas: Beiträge zur Kolonialforschung, v. 5, p. 47–105+Plates 11–13 and 2 coloured foldout maps.
Krenkel, E., 1957, Geologie und Bodenschätze Afrikas, 2. stark veränderte Auflage: Akademische Verlaggesellschaft Geest & Portig K.-G., Leipzig, XV+597 p.
Lagabrielle, Y., Whitechurch, H., Marcoux, J., Juteau, T., Reuber, I., Guillocheau, F., and Çapan, U., 1986, Obduction-related ophiolitic polymict breccias covering the ophiolites of Antalya (southwesternTurkey): Geology, v. 14, p. 734–737, http://dx.doi.org/10.1130/0091-7613(1986)14<734:OOPBCT>2.0.CO;2.
Leturmy, P., and Robin, C., *eds*., 2010, Tectonic and Stratigraphic Evolution of Zagros and Makran During the Meso-

zoic–Cenozoic: Geological Society, London, Special Publications, v. 330, viii+360 p., http://dx.doi.org/10.1144/SP330.0.

Lippard, S.J., Shelton, A.W., and Gass, I.G., 1986, The Ophiolite of Northern Oman: Geological Society, London, Memoirs, v. 11, 178 p.+2 coloured foldout maps.

McBride, J., Barazangi, M., Best, J., Al-Saad, Sawaf, T., Al-Otri, M., and Gebran, A., 1990, Seismic reflection structure of intracratonic Palmyride Fold-Thrust Belt and surrounding Arabian Platform, Syria: American Association of Petroleum Geologists Bulletin, v. 74, p. 238–259.

McBurney, C.B., and Hey, R.W., 1955, Prehistory and Pleistocene geology in Cyrenaican Libya—a record of two seasons' geological and archaeological fieldwork in the Gebel Akhdar Hills, with a summary of prehistoric finds from neighbouring territories: University Press, Cambridge, 315 p.

Miyashiro, A., 1973, The Troodos ophiolitic complex was probably formed in an island arc: Earth and Planetary Science Letters, v. 19, p. 218–224, http://dx.doi.org/10.1016/0012-821X(73)90118-0.

Monod, O., 1976, La «courbure d'Isparta»: une mosaïque de blocs autochthones surmontés de nappes composites à la jonction de l'arc helléniques et de l'arc taurique: Bulletin de la Société Géologique de France, série 7, v. 18, p. 521–531.

Montigny, R., Whitechurch, H., Reuber, I., Thuizat, R., and Juteau, T., 1983, K–Ar ages of the Antalya ophiolites (Turkey) geological implications: Bulletin de Sciences Géologiques, v. 36, p. 61–72.

Morris, A., and Anderson, M.W., 2002, Palaeomagnetic results from the Baër–Bassit ophiolite of northern Syria and their implication for fold tests in sheeted dyke terrains: Physics and Chemistry of the Earth, v. 27, p. 1215–1222, http://dx.doi.org/10.1016/S1474-7065(02)00123-7.

Morris, A., Anderson, M.W., Inwood, J., and Robertson, A.H.F., 2006, Palaeomagnetic insights into the evolution of Neotethyan oceanic crust in the eastern Mediterranean, *in* Robertson, A.H.F., and Mountrakis, D., *eds.*, Tectonic Development of the Eastern Mediterranean Region: Geological Society, London, Special Publications, v. 260, p. 351–372, http://dx.doi.org/10.1144/GSL.SP.2006.260.01.15.

Moustafa, A.R., 2008, Mesozoic–Cenozoic basin evolution in the northern Western Desert of Egypt, *in* Salem M.J., and El-Hawat, A.S. *eds.*, The Geology of East Libya: Gutenberg Press Ltd., Malta, v. 3, p. 29–46.

Moustafa, A.R., and Khalil, M.H., 1990, Structural characteristics and tectonic evolution of north Sinai fold belts, *in* Said, R., *ed.*, The Geology of Egypt: Egyptian General Petroleum Corporation, Conoco Hurghada Inc. and Repsol Exploration, S.A., A.A. Balkema, Rotterdam, p. 381–389.

Mühlhofer, F., 1923, Beiträge zur Kenntnis der Cyrenaika (Mit Besonderer Berücksichtigung des Höhlen- und Karstphänomens): Speläologische Monographien, v. 4, Verlag der Österreichischen Bundeshöhlenkommision, Wien, VIII+67 p.+1 foldout plate.

Mukasza, S.B., and Ludden, J.N., 1987, Uranium–lead isotopic ages of plagiogranites from the Troodos ophiolite, Cyprus, and their tectonic significance: Geology, v. 15, p. 825–828, http://dx.doi.org/10.1130/0091-7613(1987)15<825:UIAOPF>2.0.CO;2.

Müller, R.D., Royer, J.-Y., Cande, S.C., Roest, W.R., and Maschenkov, S., 1999, New constraints on the Late Cretaceous/Tertiary plate tectonic evolution of the Caribbean, *in* Mann, P., *ed.*, Caribbean Basins: Sedimentary Basins of the World, v. 4, p. 33–59, http://dx.doi.org/10.1016/S1874-5997(99)80036-7.

Neev, D., Greenfield, L., and Hall, J.K., 1985, Slice tectonics in the Eastern Mediterranean Basin, *in* Stanley, D.J., and Wezel, F.-C., *eds.*, Geological Evolution of the Mediterranean Basin: Springer-Verlag, New York, p. 249–269, http://dx.doi.org/10.1007/978-1-4613-8572-1_12.

Nikishin, A.M., Ziegler, P.A., Stephenson, R.A., Cloetingh, S.A.P.L., Furne, A.V., Fokin, P.A., Ershov, A.V., Bolotov, S.N., Korotaev, M.V., Alekseev, A.S., Gorbachev, V.I., Shipilov, E.V., Lankreijer, A., Bembinova, E.Y., and Shalimov, I.V., 1996, Late Precambrian to Triassic history of the East European Craton: dynamics of sedimentary basin evolution: Tectonophysics, v. 268, p. 23–63, http://dx.doi.org/10.1016/S0040-1951(96)00228-4.

Obaje, N.G., 2009, Geology and Mineral Resources of Nigeria: Springer, Berlin, 221 p.

Ogg, J.G., 2012, Geomagnetic polarity time scale, *in* Gradstein, F.M., Ogg, J.G., Schmitz, M.D., and Ogg, G.M., *eds.*, The Geologic Time Scale 2012: Elsevier, Amsterdam, v. 1, p. 85–113, http://dx.doi.org/10.1016/B978-0-444-59425-9.00005-6.

Okay, A.I., and Özgül, N., 1984, HP/LT metamorphism and the structure of the Alanya Massif, southern Turkey: an allochthonous composite tectonic sheet, *in* Dixon, J.E. and Robertson, A.H.F., *eds.*, The Geological Evolution of the Eastern Mediterranean, Geological Society, London, Special Publications, v. 17, p. 429–439, http://dx.doi.org/10.1144/GSL.SP.

O'Keefe, F.X., and Şengör, A.M.C., 1988, Tectonic evolution of the Palmyra zone, Syria: The American Association of Petroleum Geologists Bulletin, v. 72, p. 1017.

Okulitch, A.V., Packard, J.J., and Zolnai, A.I., 1986, Evolution of the Boothia Uplift, arctic Canada: Canadian Journal of Earth Sciences, v. 23, p. 350–358, http://dx.doi.org/10.1139/e86-037.

Paris, J.P., and Lille, R., 1977, New Caledonia: evolution from Permian to Miocene. Mapping data and hypotheses about geotectonics: Geodynamics in Southwest Pacific, Editions Technip, Paris, p. 195–208.

Pearce, J.A., and Robinson, P.T., 2010, The Troodos ophiolitic complex probably formed in a subduction initiation, slab edge setting: Gondwana Research, v. 18, p. 60–81, http://dx.doi.org/10.1016/j.gr.2009.12.003.

Peybernès, B., Fondecave-Wallez, M.-J., and Cugny, P., 2005, Diachronism in the sedimentary cover around the Troodos ophiolitic massif (Cyprus): Bulletin de la Société Géologique de France, v. 176, p. 161–169, http://dx.doi.org/10.2113/176.2.161.

Picard, L., 1958, Geological background on petroleum drilling in Zikhron Yaakov (S. Carmel): The Bulletin of the Research Council of Israel, section G: Geosciences, v. 7G, p. 20–26.

Picard, L., 1959, Geology and oil exploration in Israel: Fifth World Petroleum Congress—Proceedings Geology and Geophysics, Section I, paper 16, Fifth Petroleum Congress, Inc., New York, p. 311–336.

Pietersz, C.R., 1968, Proposed nomenclature for rock units in northern Cyrenaica, *in* Barr, F.T., *ed.*, Geology and Archaeology of Northern Cyrenaica, Libya: Petroleum Exploration Society of Libya, Tenth Annual Field Conference, Holland-Breumelhof N.V., Amsterdam, p. 125–130.

Premoli-Silva, I., Spezaferri, S., and D'Angelantonio, A., 1998, Cretaceous foraminiferal bio-isotope stratigraphy of hole 967E and Paleogene planktonic foraminiferal biostratigraphy of hole 966F, Eastern Mediterranean, *in* Robertson, A.H.F., Emeis, K.-C., Richter, C., and Camerlenghi, A., *eds.*, Proceedings of the Ocean Drilling Program, Scientific Results, v. 160, p. 377–394.

Puchkov, V., 2002, Paleozoic Evolution of the east European continental margin involved in the Uralide orogeny, *in* Brown, D., Juhlin, C., and Puchkov, V., *eds.*, Mountain Building in the Uralides: Pangea to the Present: Geophysical Monograph Series 132, American Geophysical Union, Washington, D.C., p. 9–31, http://dx.doi.org/10.1029/132GM02.

Reuber, I., Juteau, T., and Whitechurch, H., 1984, Genèse en contexte transformant des ophiolites d'Antalya (Turquie): Bulletin de la Société Géologique de France, série 7, v. 26, p. 945–954.

Ricou, L.-E., 1971, Le croissant ophiolitique péri-arabe, une ceinture de nappes mises en place au Cretacé supérieur: Revue de Géographie Physique et Géologie Dynamique, v. 18, p. 327–349.

Ricou, L.-E., 1976, Évolution Structurale des Zagrides la Région Clef de Neyriz (Zagros Iranien): Mémoires de la Société Géologique de France, nouvelle série, v. 55, no. 125, 140 p.

Robertson, A.H.F., 1986, The Hatay ophiolite (southern Turkey) in its eastern Mediterranean tectonic context: a report on some aspects of the field excursion: Ofioliti, v. 11, p. 105–119.

Robertson, A.H.F., Searle, M.P., and Ries, A.C., *eds.*, 1990, The Geology and Tectonics of the Oman Region: Geological Society, London, Special Publications, v. 49, viii+xviii+845 p.

Robertson, A.H.F., and Woodcock, N.H., 1981a, Gödene Zone, Antalya Complex: volcanism and sedimentation along a Mesozoic continental margin, S.W. Turkey: Geologische Rundschau, v. 70, p. 1177–1214, http://dx.doi.org/10.1007/BF01820188.

Robertson, A.H.F., and Woodcock, N.H., 1981b, Bilelyeri Group, Antalya Complex: deposition on a Mesozoic passive continental margin, south-west Turkey: Sedimentology, v. 28, p. 381–399, http://dx.doi.org/10.1111/j.1365-3091.1981.tb01687.x.

Röhlich, P., 1974, Geological Map of Libya 1: 250,000 Explanatory Booklet Sheet: Al Bay á' NI 34-15: Libyan Arab Republic, Industrial Research Centre, 70 p.+10 photographic plates.

Röhlich, P., 1978, Geological development of Jabal al Akhdar, Libya: Geologische Rundschau, v. 67, p. 401–412, http://dx.doi.org/10.1007/BF01802797.

Röhlich, P., 1980, Tectonic development of al Jabal al Akhdar, *in* Salem, M.J., and Busrewil, M.T., *eds.*, The Geology of Libya, v. III: Second Symposium on the Geology of Libya, Academic Press, London, p. 923–931.

Ruske, R., 1981, Geologie des Syrischen Küstengebirges: Freiberger Forschungshefte C 362 Geowissenschaften Geologie, VEB Deutscher Verlag für Grundstoffindustrie, Leipzig, 90 p.+3 foldout tables.

Rybakov, M., and Segev, A., 2004, Top of the crystalline basement in the Levant: Geochemistry, Geophysics, Geosystems (G^3), v. 5, Q09001, http://dx.doi.org/10.1029/2004GC000690.

Said, R., 1990a, Cenozoic, *in* Said, R., *ed.*, The Geology of Egypt: Egyptian General Petroleum Corporation, Conoco Hurghada Inc. and Repsol Exploracion, S.A., A.A. Balkema, Rotterdam, p. 451–486.

Said, R., 1990b, Cretaceous paleogeographic maps, *in* Said, R., *ed.* The Geology of Egypt: Egyptian General Petroleum Corporation, Conoco Hurghada Inc. and Repsol Exploracion, S.A., A.A. Balkema, Rotterdam, p. 439–449.

Saura, E., Vergés, J., Homke, S., Blanc, E., Serra-Kiel, J., Bernaola, G., Casciello, E., Fernández, N., Romaire, I., Casini, G., Embry, J.C., Sharp, I.R., and Hunt, D.W., 2011, Basin architecture and growth folding of the NW Zagros early foreland basin during the Late Cretaceous and early Tertiary: Journal of the Geological Society, v. 168, p. 235–250, http://dx.doi.org/10.1144/0016-76492010-092.

Sawaf, T., Brew, G., Litak, R., and Barazangi, M., 2001, Geologic evolution of the intraplate Palmyride Basin and Euphrates Fault System, Syria, *in* Ziegler, P.A., Cavazza, W., Robertson, A.H.F., and Crasquin-Solelau, S., *eds.*, Peri-Tethys Memoire 6: Peri-Tethyan Rift/Wrench Basins and Passive Margins, Mémoirs du Muséum National d'Histoire Naturelle, v. 186, p. 441–467.

Schulman, N., Karcz, Y., and Freund, R., 1959, An overturned fold on the SE flank of the Makhtesh-Hagadol anticline: The Bulletin of the Research Council of Israel, section G: Geosciences, v. 8G, p. 148–150.

Searle, M.P., 1994, Structure of the intraplate eastern Palmyride Fold Belt, Syria: Geological Society of America Bulletin, v. 106, p. 1332–1350, http://dx.doi.org/10.1130/0016-7606(1994)106<1332:SOTIEP>2.3.CO;2.

Şengör, A.M.C., 1990, Plate tectonics and orogenic research after 25 years: A Tethyan perspective: Earth Science Reviews, v. 27, p. 1–201, http://dx.doi.org/10.1016/0012-8252(90)90002-D.

Şengör, A.M.C., 1997, Asia, *in* Moores, E., and Fairbridge, E., *eds.*, Encyclopedia of European and Asian Regional Geology: Chapman and Hall, London, p. 34–51, http://dx.doi.org/10.1007/1-4020-4495-X_6.

Şengör, A.M.C., 2009, Tectonic evolution of the Mediterranean: a dame with four husbands: Trabajos de Geologia, v. 29, p. 45–50.

Şengör, A.M.C., in press, Transform faults, *in* Harff, J., Meschede, M., Petersen, S., and Thiede, J., *eds.*, Encyclopedia of Marine Geosciences: Encyclopedia of Earth Sciences Series, Springer Verlag, Berlin.

Şengör, A.M.C., and Natal'in, B.A., 1996, Palaeotectonics of Asia: Fragments of A Synthesis: *in*, Yin, A., and Harrison, M., *eds.*, The Tectonic Evolution of Asia: Rubey Colloquium, Cambridge University Press, Cambridge, p. 486–640.

Şengör, A.M.C., and Natal'in, B.A., 2001, Rifts of the world, *in* Ernst, R.E., and Buchan, K.L., *eds.*, Mantle Plumes: their identification through time: Geological Society of America Special Papers, v. 352, p. 389–482, http://dx.doi.org/10.1130/0-8137-2352-3.389.

Şengör, A.M.C., and Yılmaz, Y., 1981, Tethyan evolution of Turkey: a plate tectonic approach: Tectonophysics, v. 75, p. 181–241, http://dx.doi.org/10.1016/0040-1951(81)90275-4.

Şengör, A.M.C., Yılmaz, Y., and Ketin, İ., 1982, Remnants of a pre-Late Jurassic ocean in northern Turkey: Fragments of Permian–Triassic Paleo-Tethys?: Discussion and reply: Geological Society of America Bulletin, v. 93, p. 932–936, http://dx.doi.org/10.1130/0016-7606(1982)93<932:ROAPJO>2.0.CO;2.

Sibuet J.-C., and Collette, B.J., 1991, Triple junctions of Bay of Biscay and North Atlantic: New constraints on the kinematic evolution: Geology, v. 19, p. 522–525, http://dx.doi.org/10.1130/0091-7613(1991)019<0522:TJOBOB>2.3.CO;2.

Sibuet, J.-C., Rouzo, S., and Srivastava, S., 2012, Plate tectonic reconstructions and paleogeographic maps of the central and North Atlantic oceans: Cana-

dian Journal of Earth Sciences, v. 49, p. 1395–1415, http://dx.doi.org/10.1139/e2012-071.
Srivastava, S.P., Verhoef, J., and Macnab, R., 1988, Results from a detailed aeromagnetic survey across the northeast Newfoundland margin, Part II: Early opening of the North Atlantic between the British Isles and Newfoundland: Marine and Petroleum Geology, v. 5, p. 324–337, http://dx.doi.org/10.1016/0264-8172(88)90026-8.
Steuber, T., Özer, S., Schlüter, M., and Sarı, B., 2009, Description of *Paracaprinula syriaca* Piveteau (Hippuritoidea, Plagioptychidae) and a revised age of ophiolite obduction on the African-Arabian Plate in southeastern Turkey: Cretaceous Research, v. 30, p. 41–48, http://dx.doi.org/10.1016/j.cretres.2008.04.001.
Stille, H., 1920, Über Alter und Art variscischer Gesirgsbildung: Nachrichten der königlichen Gesellschaft der Wissenschaften zu Göttingen, mathematisch-physikalische Klasse, Jg. 1920, p. 218–224.
Stille, H., 1925a, Die saxonischen Brüche (Schlußwort zu den Göttinger Beiträgen zur saxonischen Tektonik 1923-1925): Abhandlungen der Preußischen Geologischen Landesanstalt, Neue Folge, no. 95, p. 149–207 + 1 plate.
Stille, H., 1925b, The upthrust of salt masses of Germany [translated from the German by Donald C. Barton]: Bulletin of the American Association of Petroleum Geologists, v. 9, p. 417–441.
Stille, H., 1940, Einführung in den Bau Amerikas: Gebrüder Borntraeger, Berlin, XX+717 p.
Suess, E., 1875, Die Entstehung der Alpen: W. Braumüller, Wien, IV+168 p.
Suess, E., 1883, Das Antlitz der Erde, v. Ia (Erste Abtheilung): F. Tempsky, Prag and G. Freytag, Leipzig, 310 p.
Suess, E., 1901, Das Antlitz der Erde, v. III1 (Dritter Band. Erste Hälfte): F. Tempsky, Prag and Wien, and G. Freytag, Leipzig, IV + 508 p.
Suess, E., 1909, Das Antlitz der Erde, v. III2 (Dritter Band. Zweite Hälfte. Schluss des Gesamtwerkes): F. Tempsky, Wien and G. Freytag, Leipzig, IV + 789 p.
Suleiman, A., and Saleem, M., 2008, The tectonic setting of the Cyrenaica Platform and adjacent areas, NE Libya: implication of gravity investigations, *in* Salem, M.J., and El-Hawat, A.S., *eds.*, The Geology of East Libya, Gutenberg Press Ltd., Malta, v. 4, p. 285–294.
Sungurlu, O.,1974, VI. Bölge kuzey sahalarının jeolojisi (Geology of the northern part of Petroleum District - VI), *in* Okay, H., and Dileköz, E., *eds.*, Türkiye İkinci Petrol Kongresi (Second Petroleum Congress of Turkey) Tebliğler (Proceedings), Türkiye Petrol Jeologları Derneği, Ankara, p. 85–107.
Tapponnier, P.E., Daëron, M., Sursock, A., Jomaa, R., Briais, A., Carton, H., Singh, S., Elias, A., King, G.C., and Jacques, E., 2004, Passive-active margin inversion along the Levant plate-boundary: Subduction birth and growth of Mt Lebanon (abstract): AGU Fall Meeting Abstract T52B-05, San Francisco.
Tekeli, O., and Erendil, M., 1986, Geology and petrology of the Kızıldağ ophiolite (Hatay): Bulletin of the Mineral Research and Exploration Institute of Turkey, v. 21, p. 21–37.
Théveniaut, H., Gallet, Y., Besse, J., Krystyn, L., and Marcoux, J., 1993, Origin and evolution of Turkish blocks during the late Triassic from magnetostratigraphic data: Exploration Geophysics, v. 24, p. 295–300, http://dx.doi.org/10.1071/EG993295.
Tilton, G.R., Hopson, C.A., and Wright, J.E., 1981, Uranium–lead isotopic ages of the Samail Ophiolite, Oman, with applications to Tethyan ocean ridge tectonics: Journal of Geophysical Research, v. 86, p. 2763–2775, http://dx.doi.org/10.1029/JB086iB04p02763.
Türkünal, S., 1953, Géologie de la Région de Hakkâri et de Başkale (Turquie)—Thèse présentée à la Faculté des Sciences de l'Université de Genève pour obtenir le grade de Docteur ès Sciences Géologiques et Minéralogiques, Université de Genève, Faculté des Sciences, Laboratoire de Géologie et de Paléontologie, Thèse no. 1199: M.T.A. Enstitüsü Matbaası, Ankara, 43+34 pp.+XIV plates.
Varol, E., Tekin, U.K., and Temel, A., 2007, Age and geochemistry of middle to late Carnian basalts from the Alakırçay Nappe (Antalya Nappes, SW Turkey): implications for the evolution of the southern branch of Neotethys: Ofioliti, v. 32, p. 163–176.
Völger, K., 1968, The geographic subregions of the Jabal al Akhdar in aerial photographs, *in* Barr, F.T., *ed.*, Geology and Archaeology of Northern Cyrenaica, Libya: Petroleum Exploration Society of Libya, Tenth Annual Field Conference, Holland-Breumelhof N.V., Amsterdam, p. 173–181.
Whitechurch, H., and Parrot, J.-F., 1974, Les écailles métamorphiques infrapéridotitiques du Baër-Bassit (Nord-Ouest de la Syrie): Cahiers ORSTOM, série Géologie, v. 4, no. 2, p. 173–183.
Whitechurch, H., Omrani, J., Agard, P., Humbert, F., Montigny, R., and Jolivet, L., 2013, Evidence for a Paleocene–Eocene evolution of the foot of the Eurasian margin (Kermanshah ophiolite, SW Iran) from back-arc to arc: Implications for regional geodynamics and obduction: Lithos, v. 182–183, p. 11–32, http://dx.doi.org/10.1016/j.lithos.2013.07.017.
Williams, H., 1975, Structural succession, nomenclature, and interpretation of transported rocks in western Newfoundland: Canadian Journal of Earth Sciences, v. 12, p. 1874–1894, http://dx.doi.org/10.1139/e75-166.
Woodcock, N.H., and Robertson, A.H.F., 1982, Wrench and thrust tectonics along a Mesozoic–Cenozoic continental margin: Antalya Complex, SW Turkey: Journal of the Geological Society, v. 139, p. 147–163, http://dx.doi.org/10.1144/gsjgs.139.2.0147.
Yanılmaz, E., Ahmed, S., and Emsaad, I., 1989, Regional geology of northeast Libya: Unpublished Arabian Gulf Oil Company Report, Benghazi, unpaginated.
Yılmaz, E., and Duran, O., 1997, Güneydoğu Anadolu Bölgesi Otokton ve Allokton Birimler Stratigrafi Adlama Sözlüğü "Lexicon": Türkiye petrolleri A. O. Araştırma Merkezi Grubu Başkanlığı Eğitim Yayınları No. 31, Ankara, 460 p.
Yılmas [*sic*] (Yılmaz), Y., 1985, Geology of the Cilo ophiolite: an ancient ensimatic island arc fragment on the Arabian Platform, SE Turkey: Ofioliti, v. 10, p. 457–484.
Yılmaz, Y., 1993, New evidence and model on the evolution of the southeast Anatolian orogen: Geological Society of America Bulletin, v. 105, p. 251–271, http://dx.doi.org/10.1130/0016-7606(1993)105<0251:NEAMOT>2.3.CO;2.
Zumoffen, G., 1926, Géologie du Liban: Henry Barrere, Paris, 165 p.+ 1 foldout coloured geological map.

Received July 2013
Accepted as revised January 2014

ENDNOTES:

1 Here is a mixture of terminologies: The 'Dinaric-Tauric' is a term that is

taken from Suess (1901) and combining it with the terms 'stem' and 'orogen' betrays Kober's (1914a, b, 1921) influence. Kober considered every orogen to have two oppositely verging 'flanks' or 'stems' separated by a scar (=*Narbe*) or 'betwixt mountains' or 'median massifs' (=*Zwischengebirge*). Stille used Kober's terminology and Krenkel may have followed Stille as he did later again as we shall see below.

2 This is Eduard Suess' terminology: by 'free branches', often as 'free ends', Suess meant fold trains that diverge and dwindle away without encountering any obstacle, such as a resistant massif. He used these designations when describing virgations.

3 The present knowledge indicates that a gentle folding along east-west axes here may have begun in the Ypresian, but these were very open folds that could hardly have been noticed in the field. See Ben Ferjani et al. (1990, p. 78).

4 The German term *Bruchfalten* was translated into English as 'fault-folds' and *Bruchfaltung* as 'fault-folding' by Donald C. Barton in his translation of Stille's article on the upthrust of the salt masses of Germany (see Stille 1925b, p. 420). The term fault-fold is, however, originally an American invention and it was introduced to explain the structure of the Elk Range in the United States Rockies by Holmes (1876, pp. 68 and 71), accompanied by his magnificent and famous cross-sections and relief model (unnumbered plates in Holmes 1876, between pp. 70 and 71; for reproductions of these, see Suess 1883, pp. 214–215, figs. 22 and 23). Suess translated Holmes' term into German as *Bruchfalte* in the first volume of his monumental *Das Antlitz der Erde* (Suess 1883, p. 215). Holmes' detailed cross-sections, displayed on his last foldout plate, will give a good idea of what a 'germanotype' structure looks like, identified as such by Stille (1940, p. 242, fig. 59).

5 In the translation of Stille's article cited in the previous endnote, Donald C. Barton translated Stille's *alpinotype Gebirgsbildung* as 'Alpine type of mountain building' and his *germanotype Gebirgsbildung* as 'Germanic type of mountain building' (see Stille 1925b, p. 420). We prefer alpinotype and germanotype as the term 'germanic' has also other connotations in English.

6 It should be remembered that Argand's *plis de fond* often appeared not only as folds of large dimensions involving the basement, but also as a stack of basement nappes or imbrications (see Argand 1924, pp. 334–335, see his fig. 5).

7 This virgation was noticed by Suess (1909, p. 314), on the basis of Blanckenhorn's (1893) summary, but he thought it was the branching off the *normal faults* ('*Sprünge*') into the Palmyran direction that was responsible for its appearance. Blanckenhorn did not correct his interpretation in his later publications until 1925 (Blanckenhorn 1912a, b, 1915, 1925). Leopold Kober seems to have been the first to notice the fold-and-thrust belt character of the Palmyran ranges (Kober 1915) and Krenkel (1924) followed him. Zumoffen (1926) also recognised similar folds affecting the Cretaceous rocks in Lebanon. By the 1930s, there was little doubt that Krenkel's recognition of folding was right.

8 Cyrene, the prosperous Greek colony from which Cyrenaica derives its name, was the birth place of the first earth scientist we know of, namely Eratosthenes (284 or 274 to 202 or 194 BC), who introduced the term geography for a remarkable book he wrote. Eratosthenes was the third director of the Museion in Alexandria, the institution that also housed the great Library of Alexandria.

9 The most detailed publicly available geological maps of these two anticlines are the map sheets Benghazi (N1 34-14) and Al-Bayda (N1 34-15; Röhlich 1974) of the 1:250,000 Geological Map of Libya. The NE sheet of the four-sheet 1:1,000,000 Geological Map of Libya (1985) shows them in much less, but still useful detail. Şengör was able to consult all three of these maps in the field.

10 The distinction of platform type folds versus orogenic belt (or 'geosynclinal') type folds has long been a Russian practice adopted by other iron curtain countries after World War II. See Ashgirei (1956, pp. 187–225).

11 Recent unpublished work by Oliver Jagoutz and Leigh H. Royden of the Massachusetts Institute of Technology promises much precision concerning the now vanished plates of the Neo-Tethys east of Oman.

12 A coloured geological map is one of the two plates and it is identical to the one published in Blanckenhorn (1912a).