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Evolution of the North African Mediterranean Margin: A perspective from Libya

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REPORT

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Summary:

The proposed model
The Paleozoic geohistory of Libya is considered from a point of view of rift development leading to the subsequent formation a passive margin of Permian age along the present East-Mediterranean coast. Sea floor spreading started in Late Permian and followed two stages of rifting in Carboniferous and Early Permian. The East-Mediterranean passive margin shows an overall upper plate geometry associated with failed rifts and rim basins in the back-country.

The evolution of the East-Mediterranean as the NW extension of NeoTethys will be discussed in a consistent plate-tectonic model for the Paleozoic. It can be demonstrated that the major phases of deformation are related to important changes in plate motion.

The critical data
The relevant data to assess and demonstrate the validity of the proposed model are: the depositional and subsidence history in the context of rifting followed by subsequent evolution of a passive margin; the nature and age of the East-Mediterranean sea-floor; and the information to create a plate-tectonic model connecting NeoTethys to the East-Mediterranean.

Alternative models not considered here
A number of alternative models have been proposed, most of which revolve around the idea of a Cretaceous opening of the East-Mediterranean Sea.

Keywords: Geophysics	Plate Motion	NeoTethys
Tectonics	Libya	East-Mediterranean
N-Africa	Passive margin	Scientific report

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Pontides; LI, Ligerian; MD, Moldanubian; MS, Moravo-Silesicum; HE Helvetic; SA, south-Alpine; PE, Penninic; AA, Austroalpine; CR, Carpathian; TN, north-Tarim.
Hun Gondwana terranes: blocks forming the northern margin of PaleoTethys, from west to east: IB, Iberic; AR, Armorica; MO, Moesia; CT, Cantabria; AQ, Aquitaine; AL, Alboran; IA, intra-Alpine (Adria, Carnic, Austro-Carpathian); DH, Dinaric-Hellenic; KR, Karakum-Turan; PA, Pamirs; TS, south-Tarim; QA, Qantang.
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Eurasian margin: BÜ, Bükk; AA, Austroalpine; CA, Carpathians; RH, Rhodope; DO, Dobrogea; IS, Istanbul; SK, Sakarya; EP, east-Pontides.
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Star symbols represent volcanic arc.

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Cimmerian terrane from west to east: LA, Lagonegro; AP, Apulia *s.str.*; GR, Greek autochthonous; BD, Bey-Daglari; MN, Menderes-Taurus; SS, Sanandaj-Sirjan; AL, Alborz; LT, Lut-Tabas.
Marginal oceans: VA, Vardar; IZANSI, Izmir-Ankara-Sirjan ocean.
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1. ABSTRACT

The proposed model
The Paleozoic geohistory of Libya is considered from a point of view of rift development leading to the subsequent formation a passive margin of Permian age along the present East-Mediterranean coast. Sea floor spreading started in Late Permian and followed two stages of rifting in Carboniferous and Early Permian. The East-Mediterranean passive margin shows an overall upper plate geometry associated with failed rifts and rim basins in the back-country.

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A number of alternative models have been proposed, most of which revolve around the idea of a Cretaceous opening of the East-Mediterranean Sea (see f.ex. Ricou 1994; Dercourt et al. 1993).

2. INTRODUCTION

The spatial and temporal evolution of the East-Mediterranean Basin and its relationship with the NeoTethys domain, with a perspective from Libya are examined and reassessed.

There has been and still is considerable debate about the nature and evolution of the Tethys ocean and what part of Tethys existed at what time (Stöcklin 1974, Hsü 1977, Sengör 1985, Zonenshain et al. 1985, Tozer 1989, Stampfli et al. 1991, Ricou 1994, Robertson et al. 1996, Stampfli & Mosar 1999, Stampfli et al. in press). Tethys was named first by Eduard Suess in 1893 based on work by Melchior Neumayer (see Sengör 1985). In Greek mythology Tethys is one of the supreme gods, the Titanides. The daughter of Uranus and Gaea, she married her brother, the Titan Oceanus and their offspring are the Oceanides - the river gods (some 3000). Numerous different Tethyan oceans have been named, hence Sengörs question "How many wives did Oceanus have?" (Sengör 1985). Common in literature are: Paleo-, Proto- Permo-, Meso-, Neo-Tethys, or Alpine Tethys and Tethys *sensu lato* to name but a few. From a perspective of the Paleozoic to Present evolution of the geological history of Libya, only Paleo- and NeoTethys are relevant. Herein we will follow the formal definition of Stöcklin (1974): the PaleoTethys was the ocean that separated the Variscan domain from the Cimmerian blocks, and the NeoTethys was the ocean that separated the Cimmerian blocks from Gondwana during the Late Permian. A third ocean - the Alpine Tethys (Favre & Stampfli 1992, Stampfli et al. 1998b) transected the Pangea super-continent in an east-west direction since the Middle Jurassic. This ocean has no direct relation with Neo- or PaleoTethys and is an extension of the Central Atlantic. These three oceanic realms form the

Tethyan domain *s.l.* that extended from Morocco to the Far-East (Sengör & Hsu 1984). To this one should add the numerous marginal basins which developed mainly along the Eurasian margin during Permian to Cenozoic times due mostly to backarc opening above continued north-directed subduction. From an "African-Gondwana perspective" the Tethyan ocean s.l., including the ProtoTethys, can also be defined as oceans developing successively along the northern margin of Gondwana. PaleoTethys developed between Gondwana and the Hun terranes which subsequently converge with Laurasia to form the Hercynian mountain belts of Europe and Morocco and extends eastwards into the Black Sea area and possibly further East. The younger NeoTethys separates Gondwana from the Cimmerian terranes that will collide with Eurasia to form the Cimmerides/Alpine mountain chain (Sengör 1987, Sengör et al. 1988).

In the following discussion we will focus on the history of the NeoTethys and what is regarded here as its western extension - the East-Mediterranean oceanic basin. Although the evolution of the NeoTethys is now fairly well constrained, its relation with the East-Mediterranean Basin is more controversial. The NeoTethys opened between the Cimmerian micro-continent(s) and Gondwana during the latest Paleozoic and earliest Tertiary. Depending on the authors, the oceanic East-Mediterranean-Ionian Sea basin is thought to have opened as early as during the Late Paleozoic (Vai 1991) or as late as during the Cretaceous. (*e.g.* Dercourt et al. 1993, Dercourt et al. 1985, Dercourt et al. 1986), with most of them suggesting that it began to open during Late Triassic or Early Jurassic (*e.g.* Catalano et al. 1996, Catalano et al. 2001, Finetti 1985, Garfunkel & Derin 1984, Robertson & Dixon 1984, Sengör et al. 1984), and, therefore, was possibly related to the opening of the Alpine Tethys. The interpretation favored here shows that *the East-Mediterranean domain corresponded to an oceanic basin since the Late Paleozoic and new plate reconstructions show that the East-Mediterranean Basin and the NeoTethys are directly connected and distinct from the Alpine Tethys.*

NeoTethys rifting occurred in two phases, namely during the late Early Carboniferous and during the Early Permian. The latter was followed by sea-floor spreading starting in the Late Early Permian. The East-Mediterranean area was also affected by these two rifting phases, the Carboniferous one being related to Variscan suturing of Euramerica and Gondwana. The Permian rifting phase is concomitant with the rifting phase that culminated in the opening of the NeoTethys. *The oceanic floor of the East-Mediterranean and Ionian sea is therefore thought to be of Late Permian to Early Triassic age.* In the present model Apulia and the autochthonous sequences from the Hellenides and Taurides (we call these three domains Greater Apulia) are detached from Gondwana during the Permian through the opening of the east Mediterranean ocean.

The Carboniferous to Triassic evolution of the western Tethys and East-Mediterranean Basin is characterized by the closing of PaleoTethys and opening of NeoTethys and the concomitant opening of back-arc marginal basins along the Eurasian margin. The Mesozoic evolution of these regions is marked by the northward subduction of NeoTethys and the opening of new marginal basins along the Eurasian margin. The Tethysides are mainly issued from the closing and these marginal basins and, therefore, the suture of the main oceans - PaleoTethys and NeoTethys, are in many cases occulted and cryptic.

3. GENERAL COMMENTS

A careful use must be made with parts of the terminology classically used in North Africa. Thus a number of "events" are designated with terms imported from other parts in the world and where they are generally used to describe the development of orogenic mountain belts. Whereas in their original meaning they refer to the formation of a specific mountain belt in a determined part of the world at a given time, they have been exported to North Africa - often on false assumptions and erroneous interpretation of sedimentary unconformities. There, they are presently used as indicators of a given period in time (f.ex. Hercynian, Caledonian, Taconic).

- The Hercynian "event": reality or fiction? in Europe and Morocco the Hercynian (or Variscan) is associated with the development of the hercynian mountain belt. In North Africa it refers to an erosional unconformity - the Carboniferous-Permian erosion surface; thus designating merely a stratigraphic/sedimentological record at a given time period.
- Similarly, the Caledonian "event" relates to the development of the Caledonian mountain belt sensu lato in North Europe. In North Africa it relates to an erosional unconformity in Silurian/Devonian and/or in Ordovician.
- The Taconic "event" related to the formation of an orogenesis in North America and is associated with the emplacement of an important unmetamorphosed allochton onto N-America around 452Ma (Late Ordovician).

This ambiguous use of terminology creates confusion and misunderstandings and should be avoided when describing "events" in North Africa!

4. THE AGE AND NATURE OF THE NEOTETHYS - EAST-MEDITERRANEAN BASIN

- An overview of the West- and East-Mediterranean sea-floor structure and its geodynamic evolution can be found in: (Boote et al. 1998, Catalano et al. 1996, Del Ben & Finetti 1991, Doglioni et al. 1999, Finetti 1985, Garfunkel 1998, Gueguen et al. 1998, Jolivet & Faccenna 2000, Mantovani et al. 1997, Morelli 1985a, Rehault et al. 1984a, Rehault et al. 1984b, Vanney & Gennesseaux 1985).

The Mediterranean can be divided into two major provinces (Jolivet & Faccenna 2000): the West-Mediterranean and the East-Mediterranean. The overall structure is related to the northward convergence of Africa towards Europe. Continued north-directed subduction of the African plate under Europe and the collage of terranes at the southern margin of Europe led to all the important tectonic features such as: the Atlas mountains, the Rif-Betic Chain, the Alps, the Carpathians, the Appenines, the Liguro-Provencal and Tyrrhenean Basins, and the Aegean Basins. In the West-Mediterranean the Liguro-Provencal and Tyrrhenean Basins developed as backarc basins of the W-NW subducting oceanic lithosphere of the Alpine Tethys and the East-Mediterranean. The East-Mediterranean forms the remains of the NW NeoTethyan ocean that subducts northward under Greece and Turkey, and causes the development of backarc basins and core-complexes in the Aegean Sea.

The arguments developed in this paper are supported by a long list of:

- geophysical data concerning the East Mediterranean area: refraction, reflection seismic and seismic velocity distribution, isostatic equilibrium of the oceanic basement after back stripping, large elastic thickness of the subducting slab and tomographic data, and

- geological arguments: seismostratigraphy of the Ionian Sea, subsidence pattern of surrounding margins and stratigraphic and sedimentologic records from these margins.

4.1 The East-Mediterranean passive margin and related basins:

Associated with the NeoTethys passive margin development and already prior to the rift-drift transition (that is prior to oceanic lithosphere production) a series of basins developed along the East-Mediterranean coast between Morocco and the Levant:

- Jeffara Basin - rift in continental crust

The Tunisian-Libyan Jeffara Basin represents an intracontinental aborted rift that is located at the western termination of the East-Mediterranean Basin. During the Carboniferous, this rift zone graded westwards into the active transpressional system of the Gondwana-Laurussia plate boundary. Between these two realms, isolated continental basins are found along the Moroccan Atlas shear zones (*e.g.* Nedjari 1994). The synthetic cross-section for the Jeffara basin (modified after Burollet et al. 1978, Busson 1970, Busson & Burollet 1973), suggests that it was located near the margin of a major rift. In the rift shoulder zone (ST1 to KR1), Late Carboniferous (Namurian) to Late Permian sediments rest unconformably on Silurian to Cambro-Ordovician series, indicating about 1.7 km of uplift and erosion of Lower Paleozoic strata. According to subsidence analysis (Takhrist 1991), this rift zone can be followed westward up to the Algerian border where one branch turns southward into the Ghadames Basin, containing 400 to 500 m of Triassic series. Another branch extends further West in the Algerian Saharian Atlas (Vially et al. 1994), possibly located along or close to the Variscan suture zone.

In the rim basin located to the South of the cross-section given in figure 13, the occurrence of an unconformity between Lower and Upper Carboniferous series indicates that the area was uplifted during the late Early Carboniferous. The up to 5 km thick Upper Permian sequence occurring within the rift zone (well TB1) represents a syn- to early post-rift series. A rapid lateral thickness variation of this series implies the presence of major normal faults between Kerchaoui and Medenine. However, for want of access to relevant reflection seismic data, we are unable to document these faults and thus have to rely on published data (Burollet et al. 1978).

During the Triassic, the rift shoulder and flanking rim basin were broadly overstepped by the post-rift series. The presence of a Late Triassic (Carnian) radiolaritic bed indicates episodic connections with a pelagic domain, whereas Late Triassic-Early Jurassic salts give evidence for a restricted basin environment.

The geometry of the Late Paleozoic (Permian) erosional surface can therefore be reconciled with a model of elastic flexure of an extensionally unloaded footwall block. The Late Paleozoic syn-rift sediments contain only shallow marine fauna, suggesting that the graben was completely filled with 8 km of sediments. The theoretical elastic deflection near the rift shoulder amounts to about 0.25 km; this is clearly less than the observed maximum uplift (about 1,7 km) even if we take into account erosional unloading of the rift shoulder (about 0,7 km of uplift). However, the shape of the subsidence curve and the geometry of the pre-rift series implies that this flexure is the combined effect of thermal uplift and tectonic and erosional unloading of the footwall block.

- Sicily

In the Sosio area of western Sicily, Late Permian Hallstatt-type pelagic limestone, similar to those found in Oman where they sometimes rest directly on MORB (Niko et al. 1996), have been reported from the Sosio complex (Kozur 1995). This Late Permian pelagic macro-fauna presents affinities with both Oman and Timor, this implies a Late Permian direct deep water connection of the East-Mediterranean Basin with the NeoTethys. Similarly the Triassic Lagonegro sequence of the southern Apennines (Ciarapica et al. 1990a) contains a reworked Late Permian marine fauna. The Late Permian Sicani Basin (Catalano et al. 1996) is considered as the north-western end of the East-Mediterranean rift system that terminated northwards against the transform plate boundary between Gondwana and Laurussia along which older Permian sediments involved in the PaleoTethys suture zone were recycled. A north-westward aborted continuation of this rift towards the central Iberian rift system (Salas & Casas 1993) can be envisaged.

The Sicani Basin was connected with the Lagonegro basin of southern continental Italy. The latter is characterized by deep water sedimentation from Middle Triassic to Paleogene times. Subsidence of the Lagonegro Basin commenced during the Late Permian, and persisted during the Triassic, as indicated by the transition from Lower Triassic outer platform facies to a Middle Triassic basinal sequences (Miconet 1988). These sequences are separated by Middle Anisian slope deposits containing olistostromes and olistolithes, some of them derived from an Upper Permian carbonate platform (Ciarapica et al. 1990a, Ciarapica et al. 1990b). Pelagic sedimentation dominated the Lagonegro Basin from Middle Triassic till Oligocene times, and precludes any important phase of deformation either extensional or compressional. This is also compatible with subsidence curves for wells penetrating the autochthonous sequences of southern Sicily, such as the Gela 1 well, that is located close to the Malta escarpment. This well reflects a continuous and undisturbed thermal subsidence from the Middle Triassic to the Early Tertiary.

- Hyblean Plateau - passive margin edge - continental crust

Recent investigation on the Hyblean plateau on the southeastern corner of Sicily has demonstrated its long-lived sedimentary history. The plateau is located on the African foreland at the edge of the Malta escarpment and forms the transition to the Ionian Basin with its oceanic crust. A carbonated platform developed since the earliest Triassic with Triassic dolomites and evaporites in excess of 3000m (Catalano et al. 1996, Stampfli & Mosar 1999, Yellin-Dror et al. 1997). Subsidence curves indicate a thermal subsidence since the earliest Triassic and possibly the Late Permian as in the Gela area further to the NW. It is reasonable to admit that rifting started in the Permian and that the prolonged thermal subsidence since the Early Triassic is linked to postrift thermal subsidence.

- Tunesian Plateau/Shelf and Malta Platform: Pelagian Sea - pre-Jurassic rifting - continental crust

Structure and stratigraphy of the area between the African (Tunesia-Lybia) shelf and the Sicily-Malta-Hyblean shelf are described in Del Ben (Del Ben & Finetti 1991), Catalano *et al.* (Catalano et al. 1996) and Yellin-Dror *et al.* (Yellin-Dror et al. 1997). A sedimentary pile up to 9 km thick starting in in the Triassic has been described. Permo-Triassic rocks covering the Sirte basin as well as the Pelagian sea has been described by Finetti (Finetti 1985). These observations are consistent with rifting since the Late Permian. The most recent history, since the earliest Miocene, of this rift zone is reflected by the development of the Malta-Pantelleria rift zone (Dart et al. 1993).

- Gulf of Sirte/Sirte Basin – Lybia

The evolution of the Libyan Sirte Basin in the context of the NeoTethys passive margin development is specifically addressed in chapter 3.3.

- Egyptian Margin - passive margin with flexural basin in continental crust

The East-Mediterranean rift can be studied only by the evolution of the rift shoulder and rim basin that developed along its southern margin in Egypt. This margin was affected by Late Cretaceous and Paleogene intraplate compressional tectonics causing inversion of half-grabens. The continental slope of this margin remained, however, relatively undeformed. Geophysical data (depth of the Moho and magnetic anomalies; Ben-Avraham & Ginzburg 1990, Garfunkel 1998, Ginzburg & Gvirtzman 1979) indicate the presence of a rheologically strong oceanic crust or denuded continental mantle along the Egyptian coast and in the south-eastern part of the Mediterranean area.

The southern margin of the East-Mediterranean rift, in the area of the Sinai, Gulf of Suez, Nile Delta and Western Desert is characterized by a Mesozoic wedge reaching up to 10 km in thickness. In the proximal part of the margin, late Lower Carboniferous continental sediments, unconformably overlying a Precambrian basement, occur in the Wadi Araba-Ayun Musa rim basin.

- Levant Margin - continent-ocean transition - transform margin

The palinspastic profiles of the Levant transform margin are based mainly on the data from Glikson (1964), Weissbrod (1976), Bartov et al. (1980), Garfunkel & Derin (1984), Hirsch (1986), Bein & Sofer (1987), Cohen et al. (1988), Abdalla Hegab (1991). The Late Paleozoic Heletz high, of the Negev region, is interpreted as an uplifted shoulder of an East-Mediterranean rift. Rifting activity in this area began during the Carboniferous, as indicated by the occurrence of late Lower to Upper Carboniferous sediments that unconformably overlie Precambrian to Cambrian rocks in a rim basin that developed to the SE of the Heletz shoulder (Zarqa, Makhtesh Qatan; Eshet 1992, Glikson 1964, Habib et al. 1994). The occurrence of volcanics immediately below the Middle Triassic sediments in the Heletz well, suggests a Late Permian to Middle Triassic rift related magmatic event.

The very thick, Early Jurassic, volcano-detritic Asher formation that was penetrated by the Atlit- 1 well (Garfunkel & Derin 1984) near Haifa, implies an important reactivation of the Levant transform zone. On the Heletz high, located about 10 km East of the Mediterranean coast, a paleo-canyon cuts almost 1 km into the Jurassic limestones and is filled by Early Cretaceous black shales. It developed either in response to a Late Jurassic-Early Cretaceous tectonic event or a major eustatic lowstand in sea-level.

The multiphase tectonic history of the Levant transform is also reflected by the subsidence curves:
-The Maktesh Qatan curve displays syn- to pre-Late Carboniferous rift shoulder erosion. From the thickness of the eroded Paleozoic pre-rift series outcropping in Jordan, we estimate that the Heletz high was uplifted by some 1.7 km. This relatively early shoulder uplift could be either the consequence of elastic flexure of the lithosphere in response to extensional unloading, thermal expansion of the lithosphere during the initial phase of rifting, or a Late Carboniferous transpressive event.

-This event was followed by weak subsidence during the Permian and diachronous sedimentary overstepping of the shoulder. The uncertainties on paleontological dating of the Permian strata do not allow to clearly time this Late Paleozoic rifting phase.
-From the Late Permian (Maktesh Qatan) or the Early-Middle Triassic (Heletz, Maghara) onwards, the curves display the geometry of thermal subsidence and have been modelled accordingly. On the shoulder, crustal thinning factors are negligible.
-If the tectonic subsidence of the Maktesh Qatan well displays a continuous thermal contraction of the lithosphere during the Early Mesozoic, the curves of the wells and outcrops closer to the major fracture zones (Heletz, Haifa, Ramon) show a bending point at the transition from the Late Triassic to the Early Liassic (Norian-Sinemurian unconformity) that is followed by a accelerated Middle Liassic tectonic subsidence and local volcanism (Laws & Wilson 1997)). From the Middle to the Late Jurassic, the curves reflect a lithospheric thermal contraction. In the Heletz area, the modelling shows a final phase of accelerated subsidence after the Liassic event. This suggests that the crust of the Levant transform rift shoulder was principally thinned during the Early Jurassic.
-At the Late Jurassic-Early Cretaceous boundary, the subsidences curves for Haifa (Atlit 1) confirm reactivation of the Levant transform margin. However, the discrepancy between the Middle-Late Cretaceous tectonic subsidence and the trend of the extrapolated modelled thermal subsidence after the Liassic event is smaller than the bathymetric and stratigraphic uncertainties. Therefore it appears that the Early Cretaceous and younger tectonic reactivation of the Levant margin (Gvirtzman & Garfunkel 1998) had a negligible effect in lithospheric configuration.

In summary, it appears that after an important Late Paleozoic rifting phase(s), the Levant Transform remained a transtensionally active zone at least up to Early Cretaceous times (an alternative point of view regarding the Levant area as a dip slip margin is presented by (Garfunkel 1998). Together with its southern continuation, the Pelusium line, this transform fault zone could be regarded as a major intra-continental fracture zone active at the scale of the whole NE Africa (Guiraud & Bosworth 1997).

- Palmyra rift - continental crust

The Palmyra rift is located in Syria and has subsequently to its formation developed into the Palmyride fold-and-thrust belt, which is an integrating part of the Syrian Arc system extending from the Euphrates through Syria, Israel, Jordan into Sinai and Egypt (Bosworth et al. 1999, Searle 1994). The onset of rifting in Palmyra is placed in the Early to Late Carboniferous, contemporaneous with the Early Carboniferous initial rifting phase observed in the NeoTethys margin of NW India and in the Tunisian Jeffara area. The subsidence patterns for the Dolaa borehole section (constructed after Al-Laboun 1988, Beydoun 1981, Lovelock 1984) suggest that the rifting phase ended during the Permian or the Early Triassic. The subsidence curve of Dolaa does not reflect the Jurassic fault reactivation observed laterally. The thermal subsidence ends during the Early Cretaceous, at the beginning of the Peri-Arabic obduction process, as in the Zagros and Oman.

- Iberian Basin

The Iberian Basin of central Spain (Arche & López-Gómez 1996, Salas & Casas 1993) can be considered a north-westward aborted continuation of the East-Mediterranean-NeoTethys rift. This basin was an intracontinental rift that developed since the Early Permian. Subsidence curves clearly indicate a thermal subsidence from Early Permian to Late Jurassic (Arche & López-Gómez 1996, Salas & Casas 1993). Subsequently several major events affected this basin: one Late Jurassic and Early Cretaceous (Kimmeridgian-middle Albian) and one Late Cretaceous (Late Albian-Maastrichtian) event.

4.2 Oceanic nature of the East-Mediterranean sea-floor; its structure and evolution

- Geophysical arguments

The oceanic nature of the East-Mediterranean basement is strongly supported by geophysical data such as refraction seismic (Ben-Avraham & Ginzburg 1986, Ginzburg & Gvirtzman 1979, Hsü 1977, Makris et al. 1983, Makris & Wang 1994, Morelli 1985b) and tomographic data (Spakman 1986a, b, Spakman et al. 1993). This basement is covered by an 8 to 10 km thick pre-Tertiary sedimentary sequence. Basement depths can reach 15 km or even 20 km if the layer at the base of the sedimentary sequence, sometimes considered as thinned continental crust, is interpreted as metamorphic (compacted) sediments. This oceanic nature of the basement is recognized from the Levant region up to the Ionian Sea (Avedik et al. 1995, Del Ben & Finetti 1991, Finetti 1985, Makris et al. 1986, Catalano et al. 2001).

Recent work on seismic reflection profiles in the Ionian Sea area (Catalano et al. 2001) conclude that the Ionian Sea was part of the "Mesozoic Tethys ocean" which was confined by the two passive margin edges of SE Sicily/Malta escarpment and the SW Puglia Platform of Italy. The basin between these two conjugate margins was about 330 km wide, suggesting a slow spreading rate. The Ionian sea is inferred to have a 8-9km thick oceanic crust overlain by a 6-8km thick sedimentary cover. Further data from seismicity of the Tyrrhenean subducting slab also suggest an oceanic nature for the subducting lithosphere.

The sediment thickness in the Levant Basin and the presence of Messinian salt preclude detailed seismostratigraphic analysis. In the Ionian Sea, the situation is more favorable and taking into account Finetti's proposed seismostratigraphy for the Mesozoic and Cenozoic sequence (line MS 26, Malta escarpment), and extrapolating it to the line of Avedik *et al.* (1995; Ionian Sea east of Malta) one can observe below the Late Jurassic markers a one to two seconds thick sedimentary section before reaching a series of relatively continuous reflectors resting on what we interpret as oceanic crust. Makris *et al.* (1986) give a refraction velocity of up to 6.3 km/s for this sedimentary section (similar to the 6.1 km/s basal sequence in the East-Mediterranean Basin: WA 3 profile; Morelli 1985a). This 6.3 km/s layer is characterized by relatively continuous reflectors on the line of Avedik *et al.* (1995). We regard this high velocity layer as compacted or metamorphosed basinal sediments, possibly interstratified with oceanic basalts. They are believed to consist of deep marine carbonates and siltstones.

Using the refraction WA3 profile of Morelli and Giese (1985) for the Levant Basin and the refraction profile of Makris *et al.* (1986) for the Ionian Sea, we estimated the minimum age of the oceanic crust by comparing its total subsidence (after backstripping and replacement of the sediments by sea water) with that of the Atlantic Ocean. An exponential decay of the porosity was assumed. Lithologies, as proposed by these authors, were accepted, except for the 6.3 km/s layer as discussed above. We find a total tectonic subsidence of 6.2 km for the Levant (WA3) and between 6.1 and 6.5 km for the Ionian Sea section. Such values are known only from isostatically equilibrated oldest oceanic crusts. The plate tectonic model proposed by Helwig (1976), which shows the best relation with the subsidence of old oceanic crusts ($z(t) = 6.4 - 3.2\exp[-t/62.8]$), would give a minimum Jurassic age for the East-Mediterranean oceanic crust. If thinned continental crust would be present, as proposed by some authors, and in order to obtain the observed burial depth, the original extensional β factor would have to be in the order of 5 to 10 ("pseudo-oceanic"; see Fig. 13 of Wilson et al. 1996). We consider this as unrealistic.

Tomographic images of the East-Mediterranean subduction zone, presented by Wortel & Spakman (Wortel & Spakman 1992) and Spakman et al. (Spakman et al. 1993), clearly show a 1000 km subducted slab dipping northward under Crete. The elastic thickness of this subducting slab, calculated from flexural modelling, is in the order of 70 km for the Hellenic trench (Sinclair 1996). Such a value is usually found in old continental lithosphere. However the oceanic nature of the subducting East-Mediterranean slab has been demonstrated above, therefore if this value is plotted on the diagram of Burov & Diament (Burov & Diament 1995) for oceanic lithosphere elastic thickness, it confirms the very old age of the East Mediterranean Basin oceanic lithosphere; the minimum age of which could be Late Triassic.

- Sedimentological arguments

Finetti (1985) and Catalano *et al.* (2001), based on seismic profiles, propose a Triassic to Early Jurassic sea-floor spreading for the Ionian Sea. Enay *et al.* (1982) proposed a similar age based on samples collected from the Malta escarpment, where important submarine volcanism was recognized in Jurassic and Triassic sequences. However, cross-sections through the Sirte Rise and the Pelagian Sea (*e.g.* Del Ben & Finetti 1991, Torelli et al. 1995) do not show any unconformities within the Mesozoic sequence. In the Pelagian Sea, the Triassic sequence is locally very thick (3.5 km), shows important lateral variations and can be regarded either as a syn-rift or early post-rift sequence. A continuous Upper Triassic to Cretaceous sequence is reported from the Malta escarpment (Enay et al. 1982) and a complete Mesozoic sequence, ending in Middle Triassic dolomites, has been drilled in the Gela-1 well in the autochthonous series of southern Sicily (Yellin-Dror et al. 1997). The Gela-1 subsidence curve does not reflect any important post Middle Triassic rifting event.

Considering the thickness of sediments that accumulated in such an old oceanic basin, it is interesting to note that such basinal sequences do not necessarily have to be very thick, as evidenced by known sections from the NeoTethys and the Mediterranean area:
- The Lagonegro hemipelagic Scythian to the Middle Miocene pelagic sequence (Amodeo 1998) is estimated to be 1 to 2 km thick: Scythian to Rhetian 750 m (18m/Ma); Jurassic 50m (0.7m/Ma); Cretaceous 400m (6m/Ma); Tertiary 350m (7m/Ma). However, this condensed sequence may have been deposited on a seamount-type high or on a distal tilted block of the Ionian ocean margin.
- In Oman, the Late Permian to Early Cretaceous slope facies sequence (Sumeini region; (Watts 1990)) is 2 km thick (15m/Ma) and the pelagic Hawasina Basin deposits covering the same time span, are only a few hundred meters thick (3 to 4 m/Ma for the Triassic and 1 m/Ma for the Jurassic and Cretaceous radiolaritic sediments).
In the East-Mediterranean Basin, a great sediment thickness is reached mainly through important Tertiary to Quaternary clastic input. Increased subsidence rates during the Cretaceous along the East-Mediterranean southern margin (e.g. the Pleshet basin; Cohen et al. 1988) are related to the Early Cretaceous low-stand in sea-level and the development of deep water clastic wedges.

4.3 Libya and the Sirte Basin : a possible rift branch of NeoTethys ?

As part of the northern Gondwana margin Libya has had a longstanding history of basin development in the Early Paleozoic (Boote et al. 1998, Keeley & Massoud 1998; for information on Libya see: www.nationalgeographic.com/ngm/0011/feature1/index.html). Many of these basins most likely formed in response to the geodynamic processes (subduction and passive margin development) along the northern Gondwana margin (Stampfli 2000, Stampfli & Mosar 1999, von Raumer et al. 2001). In northern Libya, control on Late Paleozoic sequences is restricted to subsurface data (Mansouri et al. 1993). A Permo-Carboniferous sequence is known from wells drilled on the Cyrenaica platform and the Jabal Akhdar, close to the Mediterranean coast (Keeley & Massoud 1998). Major inversion and uplift affected both areas since the Late Cretaceous. However, a synthetic sequence can be reconstructed, consisting of 2 km of Permo-Carboniferous sediments, resting unconformably on the Devonian, and overlain by Triassic strata with a minimum thickness of a few hundred meters, which, in turn, are conformably overlain, at least in one well, by Lower Jurassic strata. Therefore, it is likely that, at least locally, a relatively continuous Late Carboniferous to Jurassic marine sequence was deposited in this area. Yet it is not possible to decide whether we are dealing with a rim basin or a syn-rift environment. In four wells on the Cyrenaica platform, the occurrence of Devonian below Carboniferous strata could be interpreted as suggesting that this platform formed a potential rim basin that suffered little uplift. This basin is characterized by increased subsidence, commencing in the Carboniferous (Ameed & Ghori 1991). The important Late Cretaceous and Paleogene inversion of the Jabal Akhdar and the coastal region suggest that this area was a possible former rift zone, the normal fault of which was compressionally reactivated, as seen also in the northern Sinai and Western Desert.

The recognition of Triassic in the eastern Sirte Basin possibly indicates early subsidence along a basin reaching form the passive margin into the craton (Keeley & Massoud 1998).

4.4 Summary of the East-Mediterranean - NeoTethys – Libya passive margin evolution

The key arguments supporting the presented model are:

- the sedimentological/stratigraphical records for a Permian age of rifting. The subsidence associated with thermal cooling postdating rifting is seen in all places to begin in the Triassic – Early Jurassic
- the oceanic nature of the sea floor: geophysical investigation demonstrated that the East-Mediterranean sea-floor between the Ionian Sea and the Levant is of oceanic nature and modeling suggests a minimum age of Early Jurassic probably even as old as Later Permian.

The data from the East-Mediterranean can be interpreted in terms of passive margin evolution of the NW end of the NeoTethys. We are dealing here again with two rifting phases, i] during the Late Carboniferous, that is well expressed along the present Mediterranean margin from Palmyra to Tunisia and possibly more to the West in Algeria, ii] during the Late Permian, that preceded the onset of sea-floor spreading in the NeoTethys. The Permian phase is quite difficult to recognize along the African margin, although the 5 km thick Upper Permian series points rather to a syn-rift sequence than to a post-rift Carboniferous rift topography infill. The

Permian rifting event, however, is well marked along the Levant transform margin, as well as by the sedimentary record of the Sicanian basin seen in the nappes of western Sicily.

Observation and modeling of subsidence in Egypt and Jeffara are compatible with a lower plate setting with the development of a rim basin further inland as in the Wadi Araba-Ayun Musa in Egypt. The thermal subsidence of these areas began at the end of the Carboniferous. The development of major arches and basins within Africa (*e.g.* structures of the Algerian Sahara, van der Weerd & Ware 1994) is related to compressional intraplate deformation resulting from the Variscan collision of Gondwana and Laurussia.

During the Late Jurassic, the Levant transform extended southward through Egypt along a proto-Pelusium fault, possibly as an intracontinental fracture. During the Late Cretaceous-Early Tertiary it was compressionally reactivated giving rise to the development of Syrian arc structures (Ben-Gai & Ben-Avraham 1995). The early Late Cretaceous subsidence acceleration of in the Abu-Gharadig basin (Egypt; Janssen et al. 1995), preceded the large scale basin inversion affecting the whole margin since the Senonian. Its origin may be flexural in nature and related to the stress reorganisation preceding ophiolite obduction onto the N and NE margin of the Arabic peninsula. This event is marked by the reactivation of pre-existing normal faults, suggesting a rifting phase, however coeval subsidence acceleration is also evident in other African basins that are located far away from the Mediterranean region (Janssen et al. 1995).

On the Lampedusa shelf west of the Hyblean plateau, Torelli *et al.*, (1995) recognised an Early Cretaceous rifting phase that is also evident in the Syrte Basin and Tripolitania Basin of Tunisia and that correlates with a major rifting event in Central Africa (Guiraud & Maurin 1992).

5. ALTERNATIVE MODEL - CRETACEOUS EAST-MEDITERRANEAN

Two different models for the evolution of the East Mediterranean can be found in the scientific literature. The fundamental differences are the age of the rifting and opening of the East-Mediterranean and hence their roles in a global plate model.
Thus, depending on the authors the East-Mediterranean-Ionian Sea ocean opened: **1]** as early as the Late Paleozoic (Stampfli 2000, Stampfli & Mosar 1999, Stampfli et al. 1998a, Stampfli et al. in press, Vai 1991) or **2]** as late as during the Cretaceous. (Dercourt et al. 1993, Dercourt et al. 1985, Dercourt et al. 1986, Gealey 1988, Golonka & Ford 2000, Ricou 1994).
Included into the second group are a number of large projects or plate "consortiums" such as:

- the "french School" around the TETHYS program (Dercourt et al. 1993, Dercourt et al. 1985, Dercourt et al. 1986, Ricou 1994)
- the PALEOMAP Project by M: Ross and Ch. Scotese (http://www.scotese.com/) (Golonka 2000, Scotese et al. 1999, Scotese & Langford 1995)
- the Exxon Plate Project (Yilmaz et al. 1996)
- the PLATES Project of the University of Texas at Austin Institute of Geophysics (http://www.ig.utexas.edu/research/projects/plates/plates.html)
- the ODSN (Ocean drilling stratigraphic network) at GEOMAR-Kiel reconstruction package (http://www.odsn.de/odsn/services/paleomap/adv_map.html).

Cretaceous sea-floor spreading of the Mediterranean Sea was once envisaged in an attempt to explain the presence of numerous Cretaceous ophiolites in Greece and Turkey that are related to the Mesogean domain. However, most of these ophiolites occur in relatively high structural positions within the Hellenides and Taurides. Therefore, they were usually assigned to more northerly (internal) oceanic realms, such as the Pindos or the Vardar oceans, or they are referred more generally as of NeoTethyan origin. We stress here that such an attribution is erroneous, as the term NeoTethys *sensu* Stöcklin (1974) should be restricted to the main oceanic domain located to the South of the peri-Arabian ophiolitic nappes (Pillevuit et al. 1997). In view, especially of the age data and the related subsidence patterns discussed here, it appears that a Cretaceous age for the opening of the East-Mediterranean-Ionian Sea must be considered erroneous. Hence the plate tectonic models based on such an assumption must be used with caution and should be reassessed.

Many authors adopt a rather conservative attitude by suggesting that it began to open during Late Triassic or Early Jurassic (Catalano et al. 1996, Catalano et al. 2001, Finetti 1985, Garfunkel & Derin 1984, Robertson & Dixon 1984, Sengör et al. 1984).

The model retained here proposes a Late Permian rifting of the East-Mediterranean and considers it as a NW extension of the NeoTethys ocean (http://www-sst.unil.ch/research/plate_tecto/index.htm; and more specifically about NeoTethys: http://www-sst.unil.ch/research/plate_tecto/neotet.htm).

6. TECTONIC EVENTS IN AFRICA

A detailed review of the tectonic structure and the geohistory of the African plate are beyond the scope of this report. An overview and references for the Late Paleozoic and Mesozoic history can be found in the following publications: (Binks & Fairhead 1992, Burke 1996, Guiraud & Bellion 1995, Guiraud & Bosworth 1997, 1999, Guiraud & Maurin 1992, Janssen et al. 1995, Maurin & Guiraud 1993).

Africa, because of its position within Gondwana has recorded much of the complex history leading to the break-up of the Gondwana/Pangea supercontinent since Late Permian and throughout the Mesozoic. The African continent is a collage of several continents and arcs that assembled during the Proterozoic (Unrug 1996). Three distinct provinces are relevant here: the West African Craton, the Congo Craton (which impinges also on South America prior to the Atlantic Ocean opening) and the Arabian-Nubian block. The latter can be subdivided into the Nubian shield separated by a collage of arc terranes from the Arabian Shield to the East. The main rifting events that led to the break-up of Gondwana and hence Africa, followed old suture zones and tend to form epi-sutural basins in the Jurassic and Cretaceous. The Mesozoic structures are thus strongly basement-controlled features. To the north it is the evolution of Paleo- and NeoTethys that governs the structural evolution. The opening of the Atlantic, mainly the South Atlantic, and the separation of Madagascar and India are at the origin of the Cretaceous basins developing in Central Africa and that failed to break Africa into several pieces. Finally the opening of the Red Sea and the East African Rift system since Late Oligocene and Middle Miocene, respectively (Guiraud & Bosworth 1999), led to the break-away of the Arabian Shield and East Africa (Binks & Fairhead 1992).
During the Mesozoic several relevant tectonic events characterize the North African plate and can be summarized:

- Permian-Triassic: this rifting (Stampfli & Mosar 1999, Stampfli et al. in press) event is mainly located along the Northern margin of Africa and is related to the opening of

NeoTethys as discussed above. Tectonic instability affected the margin underscored by block tilting and uplift in the Cyranaica while at the same time continental crustal thinning was going on in the East-Mediterranean (Guiraud & Bosworth 1999).

- Late Jurassic: wide-spread rifting throughout the Jurassic period with different events in different areas as discussed in Guiraud & Bosworth (Guiraud & Bosworth 1999).
- Early Cretaceous intracontinental rift basins: two periods of rifting are observed in Western and Central Africa (Guiraud & Maurin 1992): i) from Neocomian to early Albian, roughly E-W trending basin form as a consequence of submeridional extension in Central Africa; the N-S trending Transahara fault zone acted as a sinistral wrench zone; ii) from Middle Aptian to Late Albian NW trending rift systems opened in response to a NE extensional regime while along the Central African fault zone pull-apart basins are generated by dextral strike-slip.
- Syrian Arc: Late Cretaceous (Santonian ca. 83Ma) compressive deformation: A brief period of shortening/compression affects the entire African plate during this period. Along North Africa this deformational event creates the Syrian Arc and leads to the formation a fold-and-thrust belt as well as whole basin inversions. The deformation belt stretches from Syria across Israel, Jordan into Egypt (Bosworth et al. 1999). In Egypt it can bee seen in the northern Arabian peninsula, but crosses the Gulf of Suez to extend into the Western Desert up to Libyan border. This deformational event is contemporaneous and linked to the peri-arabic ophiolite obduction.
- Red Sea rifting and East African Rift development: with the separation of the Arabian plate developed the Red Sea rift and the Gulf of Suez Rift. Rifting in the Red Sea area began in Oligocene (Voggenreiter et al. 1988) and the Gulf of Suez in Late Oligocene. In Oligocene and Early Miocene, rifts in Kenya and Ethiopia and the Red Sea linked and expanded to form the East African Rift System (Bosworth 1994).
- Recent events related to opening of oceanic realms and suprasubduction related extension in the Mediterranean started roughly around 32Ma are related to a complex interaction of Africa moving at a slower rate towards the North and the retreating subducting slab (northward dipping slab of Africa under Eurasia) opening new space by creating back-arc rifting (Jolivet & Faccenna 2000). These events are contemporaneous with the final closure of the Alpine Tethys and the slab break-off of the European plate (subducting southward under the Adriatic "promontory) (Stampfli & Marchant 1997, Stampfli et al. 1998b).

7. PLATE MODELS, CHANGES IN PLATE MOTION AND DRIFT VELOCITIES

Earlier quantitative and kinematic reconstructions of the evolution of the Mediterranean region began in the early 1970, by determining the relative motions of Iberia, Africa and Eurasia (f.ex. Dewey et al. 1989, Le Pichon et al. 1988, Savostin et al. 1986). Improved models were based on better understanding of the regional geology and integrating the opening of the Atlantic Ocean (f.ex. Klitgord & Schouten 1986, Rowley & Lottes 1988, Stampfli & Mosar 1999).

Plate reconstructions discussed herein are based on a new compilation and re-examination of global paleomagnetic data sets, including Euler poles derived from best-fit magnetic anomaly fits, as well as virtual geomagnetic poles (see Torsvik et al. 2001 and references therein). The plate reconstruction software is described in Torsvik & Smethurst (1999).

The plates and microcontinents involved in the opening of the East-Mediterranean – NeoTethys and relevant for the evolution of the N-coast of Africa are: Africa (Mazzoli & Helman 1994, Torsvik et al. 2001) Europe, Iberia (Olivet 1996), and Greater Italy/Italy (Neugebauer et al. 2001, Stampfli & Mosar 1999).

Different types of models will be discussed: Relative displacement model with Europe fixed and Africa moving towards Europe; and absolute models with all plates moving in an absolute framework based on the hotspot reference model. The hotspot reference frame has been used to make global reconstruction and absolute position of continents through time. Magnetic anomaly data, globally applicable from mid-Jurassic times (ca. 175 Ma), provide relative fits between continents, whereas paleomagnetic data yield paleolatitudal and rotational constraints for continents throughout geological history. Conversely, from Cretaceous times (ca. 130 Ma), hotspot reconstructions provide paleolongitudinal control, absent in paleomagnetic data, for plate positions. A fundamental assumption is that the hotspots are stationary, or move at insignificant speeds, relative to plate-tectonic velocities. Given the similarity between the Early Tertiary hotspot and paleomagnetic reconstructions and recent analysis from the Indian Ocean, argued that the hotspot reconstructions for 130 Ma to present-day by Müller *et al.* (1993) represent viable first-order approximations to use in drift-rate calculations.

latitude	longitude	rotation angle	Age Ma
16.70	322.50	-28.52	130
17.59	321.53	-28.01	125
18.47	320.56	-27.50	120
18.24	320.39	-27.07	115
17.70	320.50	-26.71	110
18.30	319.55	-26.03	105
18.90	318.60	-25.35	100
19.15	318.35	-24.33	95
19.40	318.10	-23.31	90
19.07	318.93	-21.83	85
18.00	321.10	-19.98	80
21.26	320.57	-18.33	75
25.52	319.39	-16.57	70
27.52	318.86	-15.52	65
29.39	318.46	-14.34	60
31.40	318.73	-13.02	55
32.80	319.20	-12.09	50
36.02	318.94	-10.48	45
38.79	318.07	-8.93	40
40.30	317.00	-7.91	35
43.84	316.56	-6.73	30
47.37	316.07	-5.54	25
50.90	315.50	-4.36	20
55.23	321.21	-3.14	15
59.22	328.24	-1.91	10
31.66	348.94	-0.98	5

7.1 Plate Reconstructions

The reconstruction discussed herein cover the Paleozoic period, with few key models relevant to the understanding of the Paleo- and NeoTethys evolution, and the Mesozoic/Tertiary period with a focus on the plate motion of Africa versus Europe.

7.2 Drift curves - velocity curves from real displacements

In order to evaluate the velocity of a plate one can consider a point on a given plate and calculate its displacement in time with respect to a reference frame. One can alternatively choose between the global hotspot reference frame to calculate absolute velocities of any given point on a plate, a relative reference frame, where one plate is fixed and the other moves (mostly based on magnetic anomalies), or else use the virtual geomagnetic poles based on apparent polar wander paths and calculate latitudinal displacements.

- virtual geomagnetic poles: unfortunately the virtual geomagnetic pole curve is not, at present, of good enough quality and precision to be used for precise and detailed reconstructions and velocity measurements for the Mesozoic time period.

- absolute displacement based on the hotspot reference frame for four points on the N-African plate and one on the Apulian promontory/European plate. The points on N-Africa are located to the East of the Gulf of Sirte near Banghazi-Cyrenaica (latitude 32°N; longitude 20°E), one along the coast at the Tunesia-Libya border (latitude 33°N; longitude 11.5°E), one point on the Hyblean Plateau in Sicily (latitude 37°N; longitude 15°E), and one point on the Egyptian coast in Alexandria (latitude 31.5°N; longitude 30°E). The point on the Italian promontory-European plate is on the SE tip of Italy (latitude 40°N; longitude 18°E). The velocity of Africa generally increases up to 80-75 Ma (Campanian) with a maximum velocity around 4-4.5 cm/year. Overall the velocity is increasing from West to East, which is due to the rotational component of the Africa plate motion. Generally the velocity decreases in the Tertiary with one noticeable increase in Lutetian times (40-50 Ma – up to 3-3.5 cm/year). The lowest velocities, with the exception of the present period, are recorded in Late Eocene (40-35 Ma – 1.5 cm/year). A slight recovery is registered between 35 and 5 Ma with velocities around 2.2 cm/year. The velocity of Europe also shows strong increase during the Campanian-Maatrichtian period (3.9 cm/year), but unlike Africa already had a high-velocity peak in Albian. Like Africa the velocity strongly drops in the Early Eocene and recovers in the Middle Eocene. Unlike Africa Europe does not recover as fast from the velocity drop in the Early Oligocene and only recovers velocities above 2 cm/year in the Burdigalian.
- relative motion of Africa with Europe fixed is given for the same three points along the N-Africa coast as for absolute velocities (Hyblean Plateau; Cyrenaica, and Libya-Tunesia border). The relative motion of Africa towards fixed Europe has a strong rotational component. Several periods with distinct displacement directions can be differentiated: westward movement from 190 to 175 Ma (Toarcian-Bajocian); an important displacement to the SE between 175 and 135 Ma (Bajocian-Hauterivian), followed by more E directed displacement from 130 to 100 Ma (Hauterivian-Late Albian); and subsequent displacement to the NE between 100 and 70 Ma (Cenomanian-Early Maastrichtian). A second short-lived motion to the E between 70 and 60 Ma (Maastrichtian-Danian) preceded the final overall movement towards the N between 60 Ma and Present. Smaller changes in this overall north-directed displacement occur around 45 Ma (Lutetian), 20 Ma (Early Burdigalian), and 10-5 Ma (Late Miocene).

7.3 Absolute present plate motion

Determination of absolute plate motion for the present-day can be done using different methods. Using all available information on tectonic features related to plate movement one can determine a best fit solution for the plate motion as has been done in the NUVEL-1 model (DeMets et al. 1990, Gordon 1995). Other techniques are based on direct displacement measured and are based on reference frames external to the earth, such as the Satellite Laser Ranging (SLR; Bianco et al. 1998), the Very long Baseline Interferometry (VLBI), and the Global Positioning System (GPS). In our analyses a combination of data from NUVEL-1 as well as data from SLR and the global GPS receiver network (143 sites – Ref frame ITRF96 epoch 1.Jan. 1996 - data collected by International GPS Service for Geodynamics-IGS and analyzed by Jet Propulsion Laboratory: http://sideshow.jpl.nasa.gov/mbh/series.htm) were used to determine the present-day absolute motion of tectonic plates. To compare past plate motion and compare motion directions with the present –day motion direction I used data models HS2-NUVEL1 (Bianco et al. 1998, Gordon 1995) for present-day motion and future motion as well as data from SLR and VLBI.

Motion directions shown from plate modeling (Nuvel1) are in good agreement with motion detected from space technology. Absolute velocities, however, still show significant discrepancies between the two approaches. It is also noteworthy that absolute displacements of plates are not reflected in a simple way nor by the global stress pattern nor by the relative displacement of plates with respect to each other.

To evaluate a possible future position of the continents one can project the present plate motion into the future (see table where the positions of the major plates have been projected 10Ma into the future according to Euler rotation poles and angular velocities obtained from plate reconstruction, modeling and space geodetic measurements).

Gmap files	latitude	longitude	ang rot	Plate name
anatolia	33	32	13.8	***anatolia***
arab_plate	45.2	355.6	5.7	***arabia***
arabia	45.2	355.6	5.7	
aus_plate	33.8	33.2	6.8	***australia***
austral	33.8	33.2	6.8	
eur_plate	50.6	247.6	2.4	***eurasia***
eurasia	50.6	247.6	2.4	
_europe	50.6	247.6	2.4	
_adriat	50.6	247.6	2.4	
italy	50.6	247.6	2.4	
_iberia	50.6	247.6	2.4	
eurcoc	50.6	247.6	2.4	
nam_plate	-2.5	274	2.2	***north america***
Nam	-2.5	274	2.2	
glakes	-2.5	274	2.2	
con4	-2.5	274	2.2	***greenland-ellesmere***
pac_plate	-63	107.4	6.7	***pacific***
pacific	-63	107.4	6.7	
sam_plate	-25.4	124.6	12	***south america***
SAM	-25.4	235.4	12	
afr	50.6	286	3	***africa***
con3	50.6	286	3	
ant_plate	63	244.1	2.5	***antarctica***
con6	63	244.1	2.5	
car_plate	25	93.1	2.2	***caraibes***
coco_plate	24.5	244.2	15.8	***cocos***
ind_Plate	45.5	0.4	5.7	***india***
_gindia	45.5	0.4	5.7	
juan_plate	-27.4	58.1	6.4	***juan de fuca***
naz_plate	47.8	259.8	7.8	***nazca***

ang rot = angular rotation clockwise (degrees) with projection of 10Ma into the future

Present-day plate velocities; angular rotation is for 10Ma into the future from NNR-NUVEL1

8. PLATE-TECTONICS RELATED TO MAJOR TECTONIC EVENTS IN AFRICA

8.1 Paleozoic plate tectonics of Africa: view from the northern margin

During the Late Cambrian and Early Ordovician the northern margin of Gondwana, enclosing Africa, has seen a succession of collisions of arcs and microcontinents and subduction-related events. In the Ordovician/Silurian times the Hun terranes (future Hercynian terranes in Europe) broke off along the Northern margin of Gondwana and formed the PaleoTethys.

- 490 Ma E. Ordovician: during this period there is south directed subduction under northern Gondwana. This results in the backarc rifting of a small chain of terranes, including Avalonia. Avalonia is going to pursue its drift towards the N to collide in the Silurian with N-America. The terranes east of Avalonia along the African margin will again collide and merge with Gondwana. This rifting and collision event is most likely to be recorded in the sedimentary history (possibly as a peak on subsidence curves).
- 430-420 Late Silurian: continued subduction under northern Gondwana towards the south led to the break-away of the Hun terranes (Spain, Corsica, Sicily, France etc – the future Hercynian terranes) and formed the PaleoTethys ocean with two passive margins developing. After initial rift shoulder uplift this event will be followed by thermal subsidence on the North Gondwana (Africa) passive margin.
- 400 Middle Devonian (410-380): beginning of the collision of the Hun Terranes with Europe and formation of the Hercynian mountain-belt in Europe. This may temporarily have slowed the spreading rate of PaleoTethys and might be reflected in the thermal subsidence curves from N-Africa (f.ex. a break in the slope of the thermal subsidence portion of the curve). Subsequent to the collision of the Hun terranes subduction of PaleoTethys was induced towards the N under Europe and was thus triggering the convergence of Africa with Europe and North America (Laurussia).
- This convergence culminated in the amalgamation of Gondwana with Laurussia – especially North Africa colliding with North America to form the Pangea supercontinent in Late Carboniferous 320 Ma (end of the Hercynian deformation phase). An important erosional event is known in North Africa and might be associated with this plate tectonic event.

8.2) Late Paleozoic, Mesozoic and Tertiary plate tectonic evolution

The continuing subduction of PaleoTethys under Eurasia south and east of the Hercynian mountain belt keeps pulling Africa (Gondwana) closer to Eurasia. This will eventually lead to the opening of the NeoTethys in Early- Late Permian times. From Permian onwards the major basins developing in and around Africa are related to the openings of the NeoTethys along the northern margin of Gondwana, the Atlantic to the west and the Indian Ocean to the east.

- 250-260 Ma - NeoTethys passive margin: rifting started in Late Permian and is responsible for the basin development in the North Africa region. The demise of the PaleoTethys - subducting under Eurasia - is at the origin of the break-off of the Cimmerian terranes and the formation of NeoTethys. Rift-shoulder uplift along the Northern African margin is expected, followed by thermal subsidence during the Triassic and Jurassic. Rim basins will develop further inland Africa (possibly the Murzuk and Kurfa basins, but also in the Arabic peninsula SE of Israel and into the Palmyra region).

- The opening of the Atlantic: the drifting of the Central Atlantic started around 180 Ma and sees the beginning of the overall eastwards drift of the African plate. It is from this time on the African plate has a motion to the E-SE. The Alpine Tethys opens during this period and expands until 130-120Ma. Around 140 Ma India breaks away from E-Africa (first magnetic anomalies – rift-drift transition).

- 130 Ma Hauterivian: This is the period of Early Cretaceous intracontinental basin development: roughly E-W trending basins form as a consequence of submeridional extension in Central Africa; the N-S trending Transahara fault zone acted as a sinistral wrench zone. This period coincides with the rift-drift transition (beginning of formation of oceanic crust) in the South Atlantic. The West African Craton and Amazonia are still attached and there is a strong rotational component to the movement of Africa: the West African Craton tries to break from S-America, from S-Africa (Congo Craton) and the Arabic-Nubian Craton to the East. The strong rotational component supports a tectonic event which results in basin formation along the potential break zone.

- 110-90 Ma Albian-Turonian: The younger rifting event in Central and North Africa may be related to the final separation of Africa and South America between 100 and 90 Ma. From the Middle Aptian to Late Albian, NW trending rift systems opened in response to a NE extensional regime while along the Central African fault zone pull-apart basins are generated by dextral strike-slip.

- Around 83 Ma a strong inversion (folding) is recorded mainly in North Africa (Syrian Arc-Cyrenaica) and the Peri-Arabic domain (with an ophiolite obduction). This deformational event is related to the intra-oceanic subduction in the NeoTethys in front of the northward moving Indian continent (the rapid northward movement and convergence with Asia began also around 85 Ma) and is concomitant with an overall global change in late motion direction (as seen from the absolute motion trajectories reconstructed based on the hotspot fixed reference frame). The collision of Iberia with Europe (France) and the consumption of the Valais ocean and the Bay of Biscay ocean in a short-lived subduction probably generated inversion along the Africa margin facing Iberia at the time. This is also the time (85-80 Ma) of rift-drift transition between W-Africa and Amazonia.

- 70-60 Ma Maastrichtian: It is during this time period that most of the small oceans between Africa and Europe – with the exception of the NeoTethys-East Mediterranean – were consumed and gave rise to the main Alpine "deformational event". The collision between the terranes is at its climax and this may force Africa to change motion for a brief period of time before resuming N-directed displacement. The Alpine Tethys is not yet completely subducted (see 40 Ma). The India plate motion increases followed by a rapid convergence with Asia. (Along W America the Cordillera sees the important intrusives and may cause a slower opening of the Atlantic as a result?).

- Around 40 Ma: Subduction of the East-Mediterranean-NeoTethys begins under Europe (Greece, Turkey, Italy Corsica-Sardinia). Therefore new oceanic domains develop in the Mediterranean region and are due to suprasubduction related extension (northward subducting oceanic margin of Africa). This may cause inversion along the African margin facing the subduction zone but mainly one should expect uplift along the margin and associated erosion because of the formation of a flexural bulge associated with the subduction of the heavy African oceanic crust to the North. Loading a subducting plate creates a flexure and a flexural bulge that may cause as much as 1km uplift. At the same time the Alpine Tethys finally closes and may add to the inversion. This time period also witnesses the final collision of the India with Asia forming the Himalaya.

- From 30 Ma on we have the formation of the Red Sea.

- Around 20 Ma the Iceland hotspot is captured under the N-Atlantic ridge and there is a change in the global displacement of the Eurasian plate from NE to almost E directed. This change also appears to be recorded in the African relative plate motion; the Atlas in Tunisia and Algeria form leading to more flexuring.

9. KEY REFERENCES TO THE UNDERSTANDING OF THE EAST-MEDITERRANEAN PASSIVE MARGIN EVOLUTION: SEDIMENTOLOGY, TECTONICCS, AND PLATE-TECTONICS

In the following is a list of relevant references suitable for a quick and overall update on the tectonics, plate-tectonics and geodynamic evolution of the East-Mediterranean/North Africa regions.

- An overall view of the North Africa geology is given in:

Guiraud, R., & Bosworth, W., 1999: Phanerozoic geodynamic evolution of northeastern Africa and the northwestern Arabian platform. *Tectonophysics, v. 315,* no. 1-4, 73-108.

Macgregor, D. S., Moddy, R. T. J., & Clark-Lowes, D. D., 1998: Petroleum geology of North Africa, *in* Fleet, A. J., ed., Geological Society of London, *Special Publication, Geological Society of London*, p. 442.

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Stampfli, G. M., & Mosar, J., 1999: The making and becoming of Apulia. *Memorie di Scienze Geologiche di Padova, v. 51*, 141-154.

Stampfli, G. M., Mosar, J., Favre, P., Pillevuit, A., and Vannay, J.-C., in press, Permo-Mesozoic evolution of the western Tethyan realm: the NeoTethys/East- Mediterranean Basin connection, *in* Cavazza, W., Robertson, A. H. F. R., and Ziegler, P. A., eds., Peritethyan rift/wrench basins and passive margins, IGCP 369: Paris, Bull. Museum Nat. Hist. Nat.

- Specific areas are adressed in the following:

Burollet, P. F., Mugniot, J. M., & Sweeney, P., 1978: The geology of the Pelagian block: the margins and basins of Southern Tunisia and Tripolitania. In *The ocean basins and margins*, Edited by A.E.M. Nairn & W.H. Kaines, p. 331-359.

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Grasso, M., Torelli, L., & Mazzoldi, G., 1999: Cretaceous-Paleogene sedimentation patterns and structural evolution of the Tunesian shelf, offshore the Pelagian Islands (Central Mediterranean). *Tectonophysics, v. 315,* 235-250.

Gvirtzman, Z., & Garfunkel, Z., 1998: The transformation of southern Israel from a swell to a basin: stratigraphic and geodynamic implications for intracontinental tectonics. *Earth and planetary Science Letters, v. 163*, 275-290.

Mazzoli, S., & Helman, M., 1994: Neogene patterns of relative plate motion for Africa - Europe: some implications for recent central Mediterranean tectonics. *Geologische Rundschau, v. 83*, 464-468.
Yellin-Dror, A., Grasso, M., Ben-Avraham, Z., & Tibor, G., 1997: The subsidence history of the northern Hyblean plateau margin, southeastern Sicily: *Tectonophysics, v. 282*, 277-289

- General tectonic interpretations are:

Bosworth, B., Guiraud, R., & Kessler, I. L. G., 1999: Late Cretaceous (ca. 84Ma) compressive deformation of the stable platform of northeast Africa (Egypt): Far-field stress efects of the "santonian event" and origin of the Syrian arc deformation. *Geology, v. 27*, 633-636.
Garfunkel, Z., 1998: Constraints on the origin and history of the Eastern Mediterranean basin. *Tectonophysics, v. 298*, 5-35.
Guiraud, R., & Bosworth, W., 1997: Senonian basin inversion and rejuvenation of rifting in Africa and Arabia: synthesis and implications to plte-scale tectonics. *Tectonophysics, v. 282*, 39-82.
-, 1999: Phanerozoic geodynamic evolution of northeastern Africa and the northwestern Arabian platform. *Tectonophysics, v. 315*, 73-108.
Janssen, M. E., Stephenson, R. A., & Cloething, S., 1995: Temporal and spatial correlations between changes in plate motions and the evolution of rifted basins in Africa. *Geological Society of America Bulletin, v. 107*, 11317-1332.
Jolivet, L., & Faccenna, C., 2000: Mediterranean extension anf the Africa-Eurasia collision: *Tectonics, v. 19*, 1095-1106.
Maurin, J.-C., & Guiraud, R., 1993: Basement control in the development of the Early Cretaceous West and Central African rift system. *Tectonophysics, v. 228*, 81-95.
van der Weerd, A., & Ware, P. L. G., 1994: A review of the East Algerian Sahara oil and gas province (Triassic, Ghadames and Illizi Basins). *First Break, v. 12*, 363-373.

- Plate-tectonic papers are:

Jolivet, L., & Faccenna, C., 2000: Mediterranean extension anf the Africa-Eurasia collision: *Tectonics, v. 19*, 1095-1106.
Neugebauer, J., Greiner, B., & Appel, E., 2001: Kinematics of the Alpine-West Carpathian orogen and paleogeographic implications. *Journal of the Geological Society, London, v. 158*, 97-110.
Stampfli, G. M., & Mosar, J., 1999: The making and becoming of Apulia. *Memorie di Scienze Geologiche di Padova, v. 51*, 141-154.
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TABLES

On computer diskette in Excel spreadsheet format

APPENDIX

Computer diskette in Excel spreadsheet format

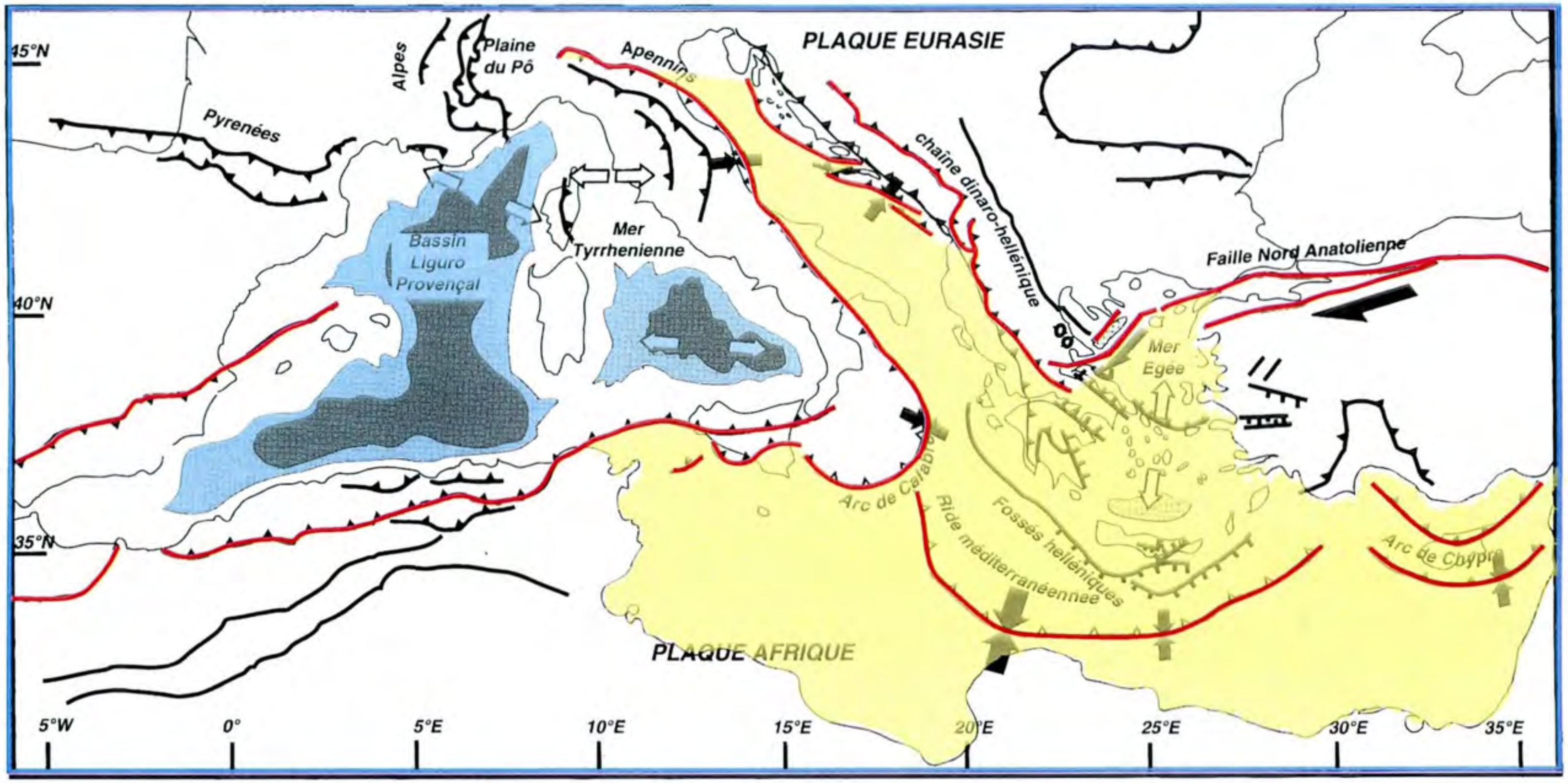
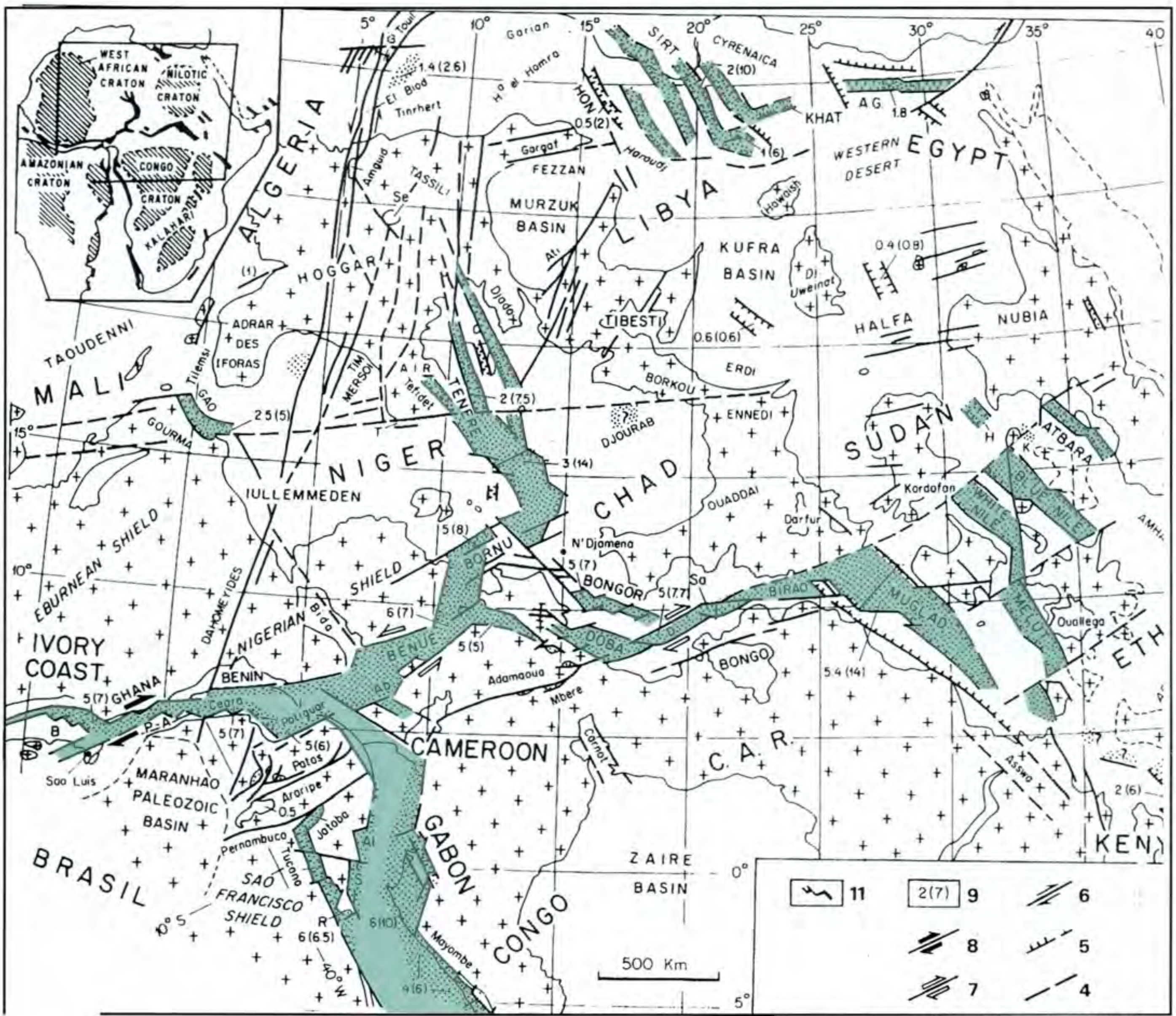
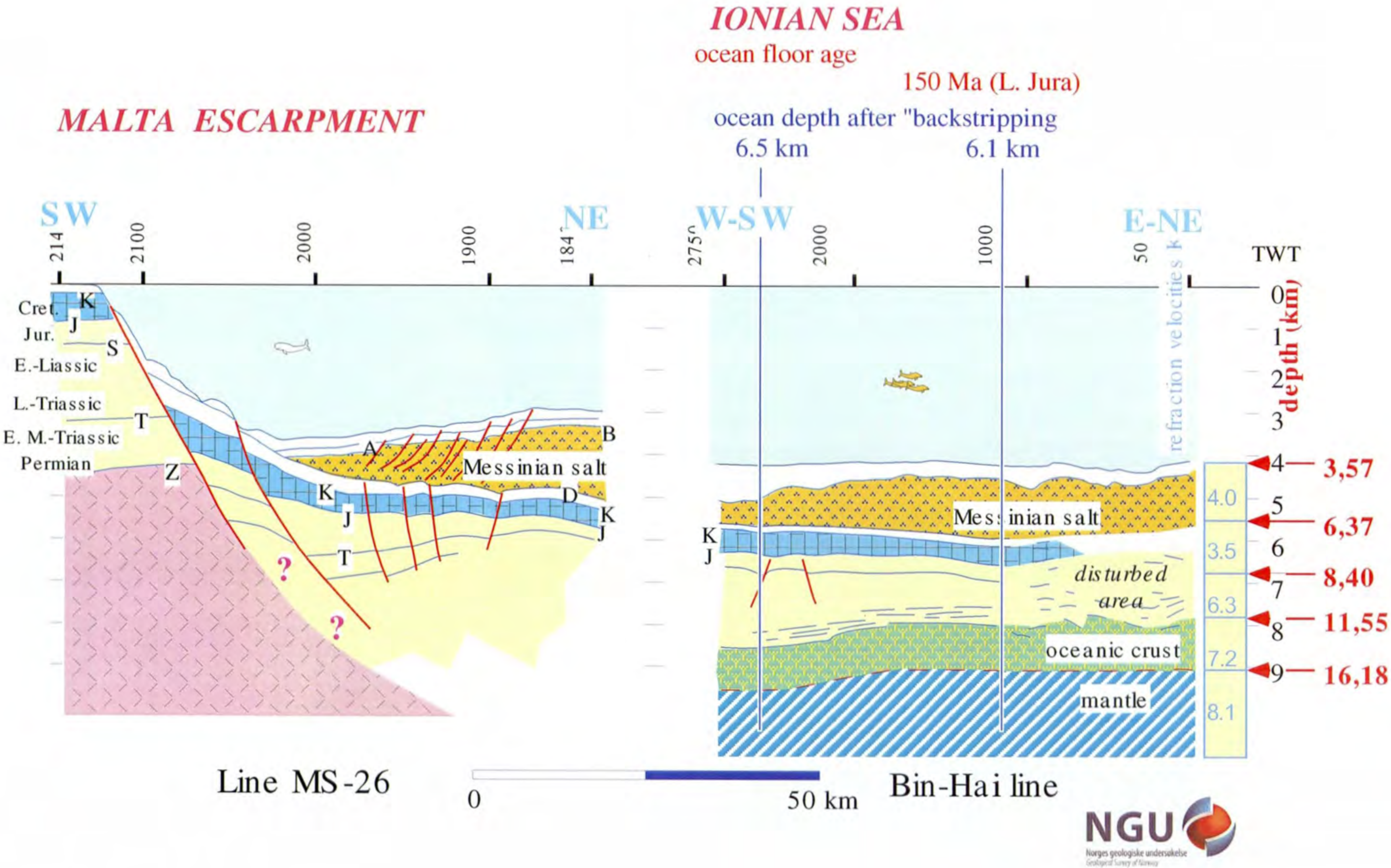

PLAQUE EURASIE
45°N
Plaine
du Pô
Alpes
Apennins
Pyrenées
chaîne dinaro-hellénique
Bassin
Liguro
Provençal
Mer
Tyrrhenienne
Faille Nord Anatolienne
40°N
Mer
Egée
Arc de Calabre
Ride méditerranéenne
Fossés helléniques
Arc de Chypre
35°N
PLAQUE AFRIQUE
5°W
0°
5°E
10°E
15°E
20°E
25°E
30°E
35°E

FIGURE 1



Guiraud & Maurin 1992

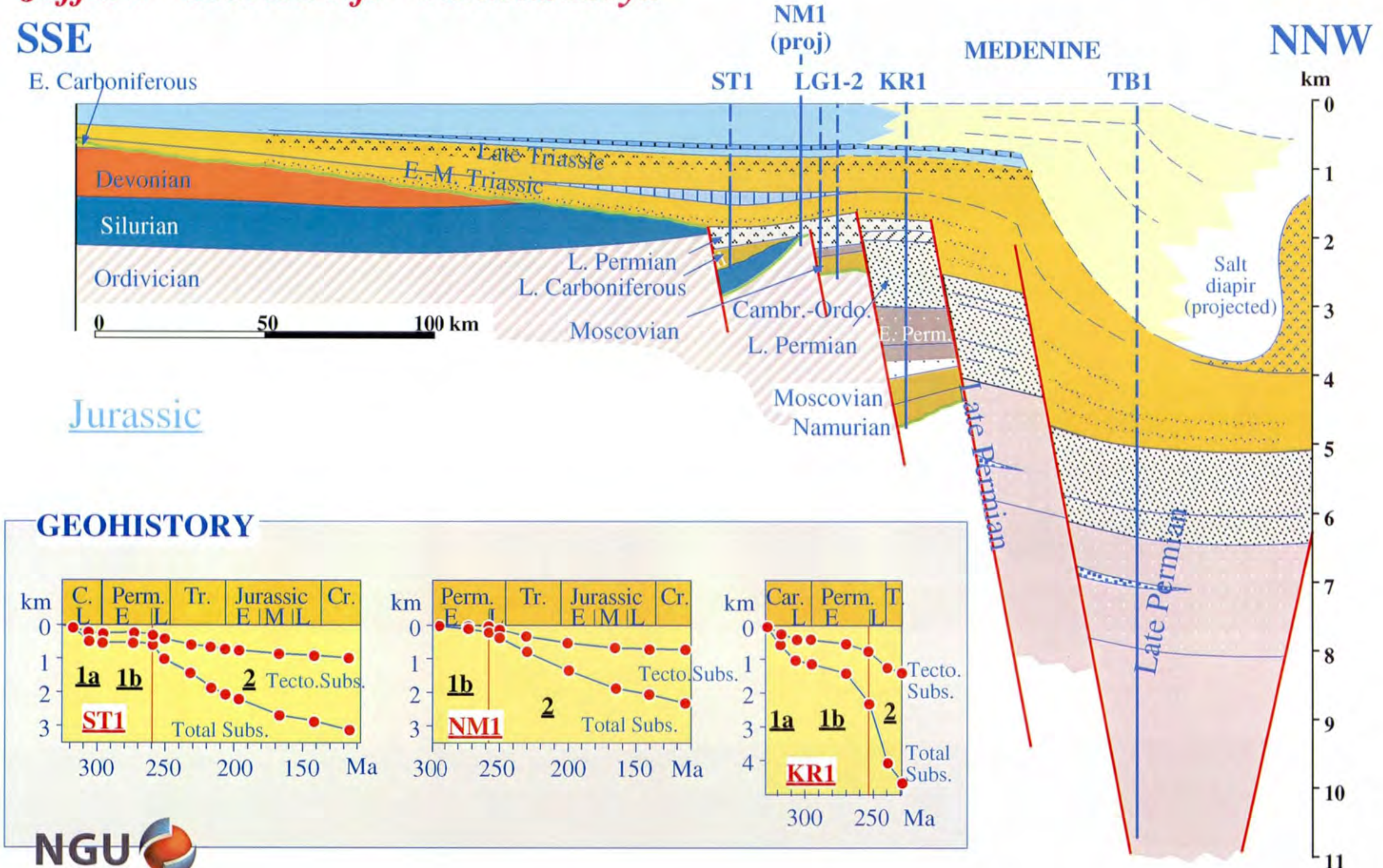
FIGURE 2



Jon Mosar @ 2001

FIGURE 3

Jeffara - aborted rift - Tunisia/Libya



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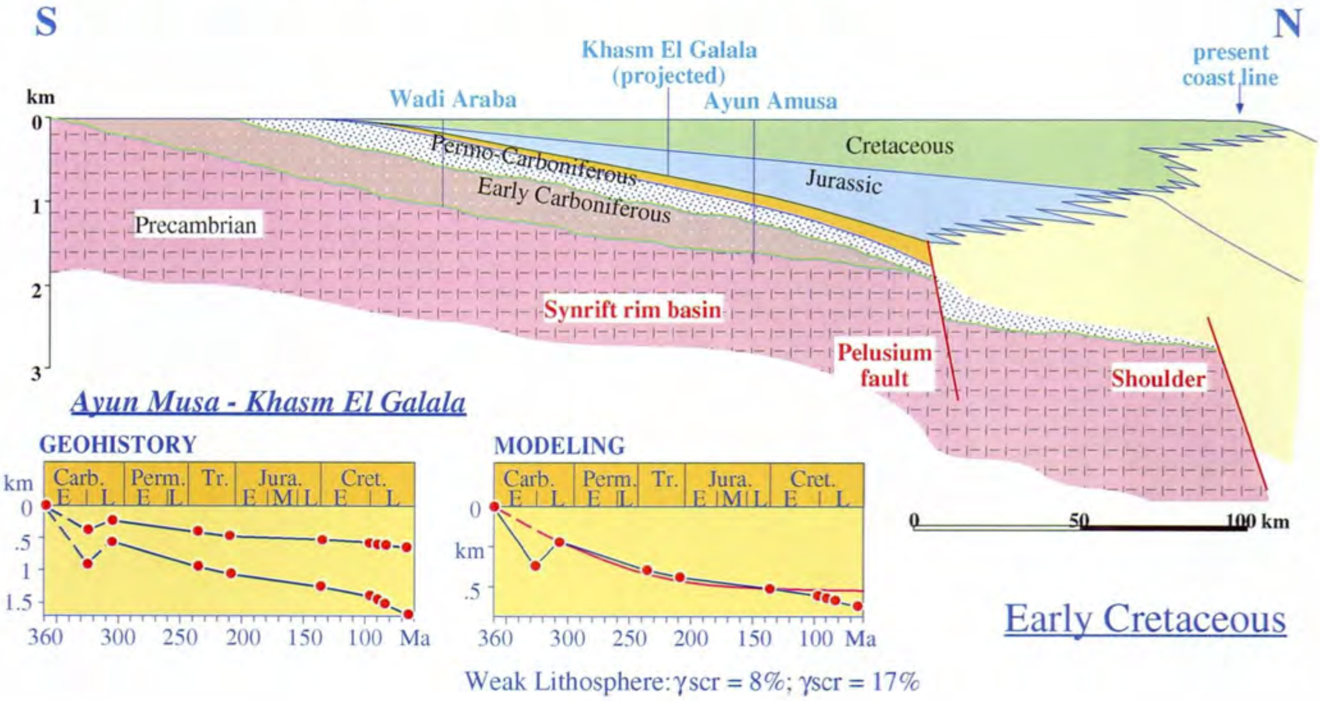
Jon Mosar @ 2001

FIGURE 4

Gulf of Suez - Nile Delta, Egypt



Western Desert, Egypt

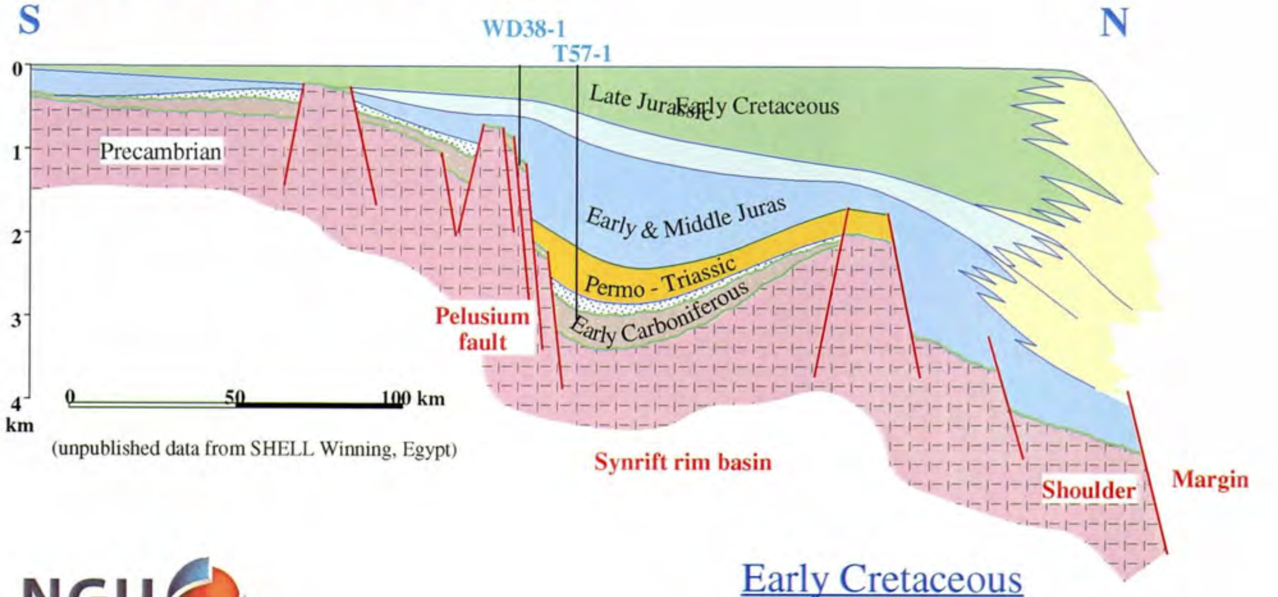


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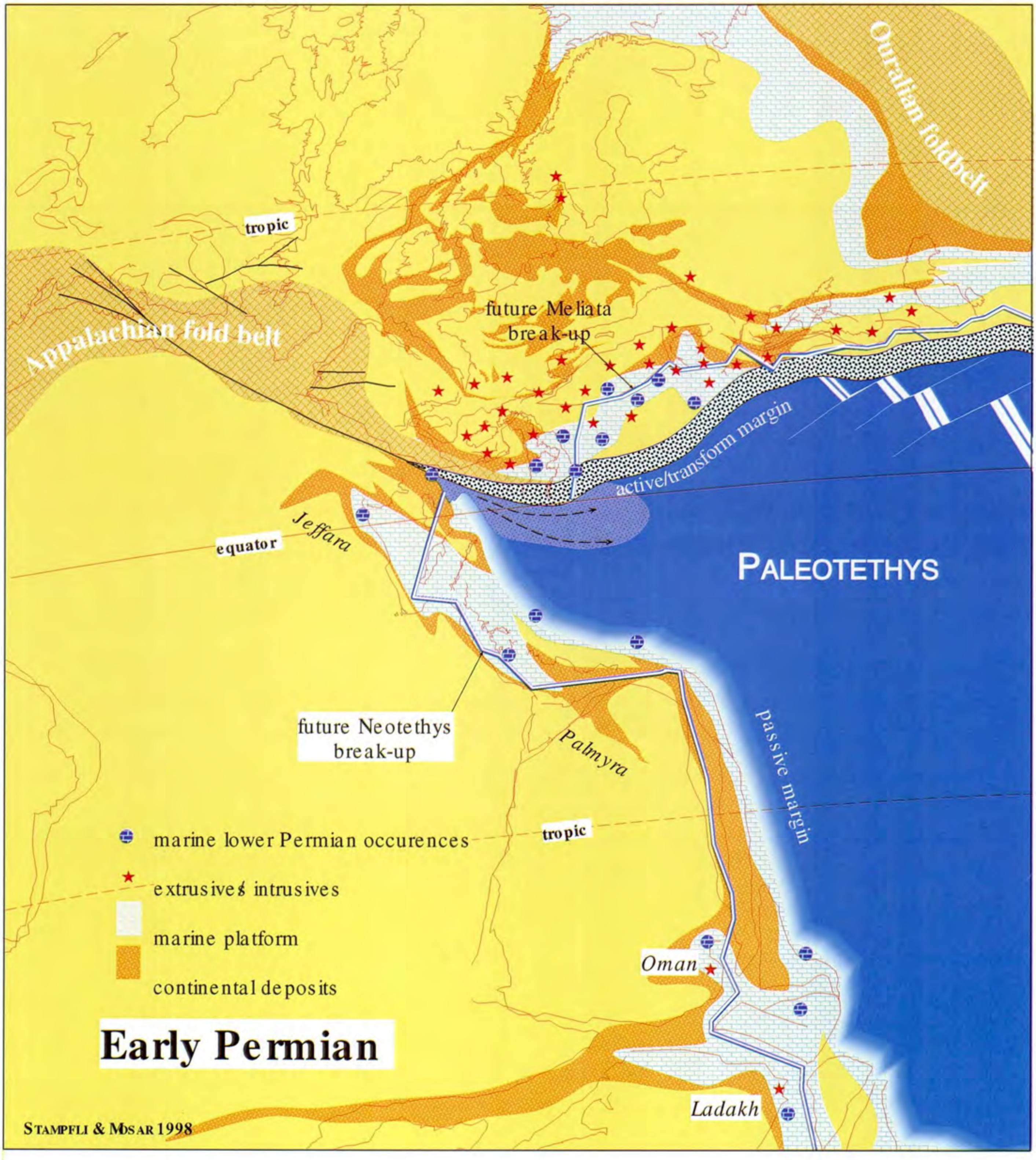
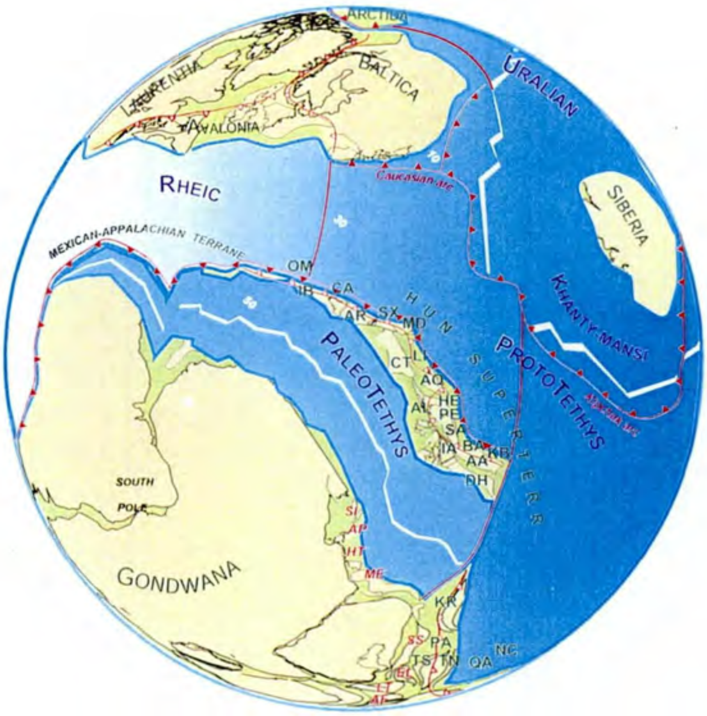
FIGURE 5



Ouralian foldbelt
tropic
Appalachian fold belt
future Meliata break-up
active/transform margin
Jeffara
equator
PALEOTETHYS
future Neotethys break-up
Palmyra
passive margin
marine lower Permian occurences
extrusives/ intrusives
marine platform
continental deposits
tropic
Oman
Early Permian
Ladakh
STAMPFLI & MOSAR 1998

FIGURE 6

Early Devonian (400 Ma)



Late Silurian (420 Ma)

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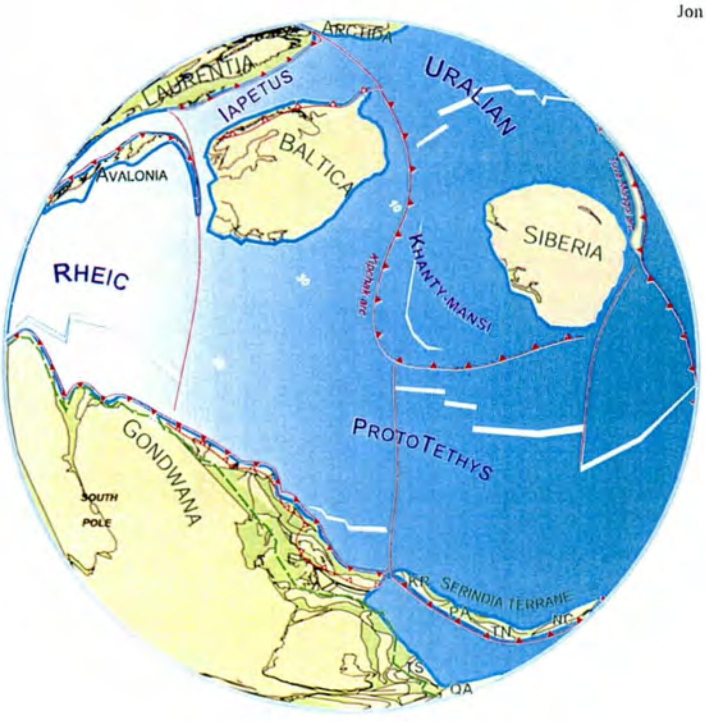
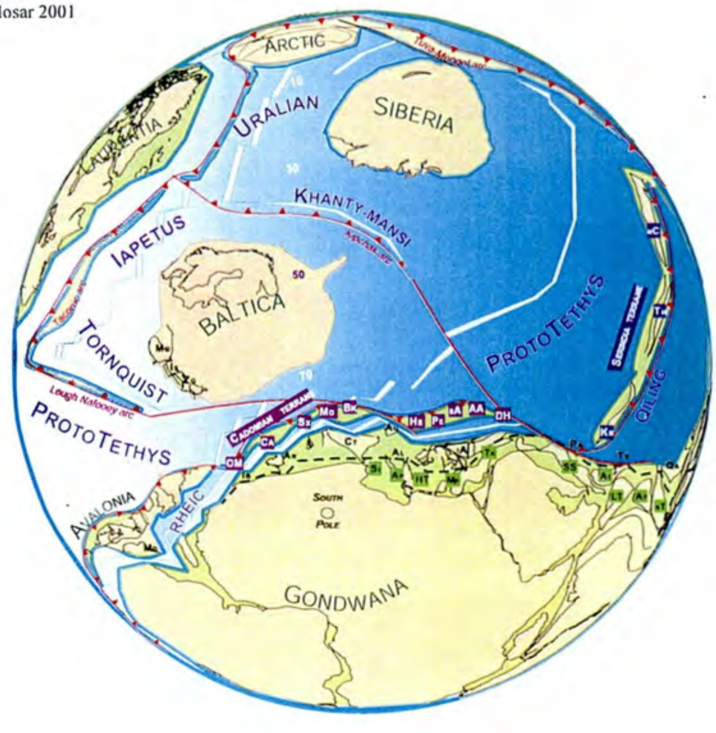


Early Silurian (440 Ma)



Early Ordovician (490 Ma)

FIGURE 7

Stampfli & Borel 2000

Late Carboniferous (320 Ma)

Stampfli & Borel 2000

Early Carboniferous (340 Ma)

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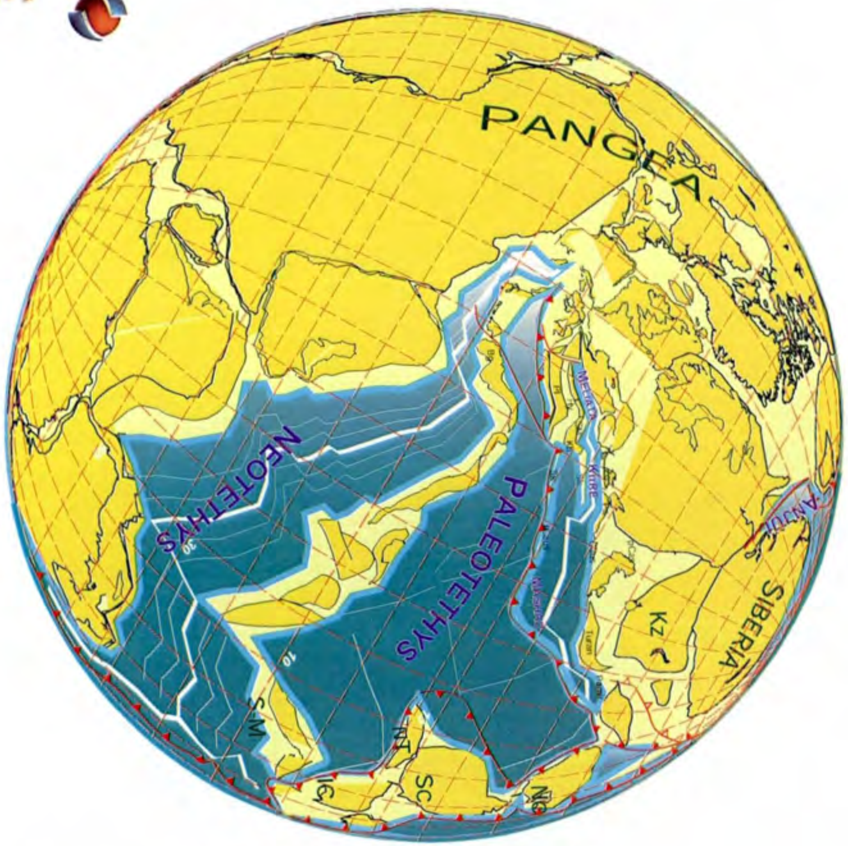
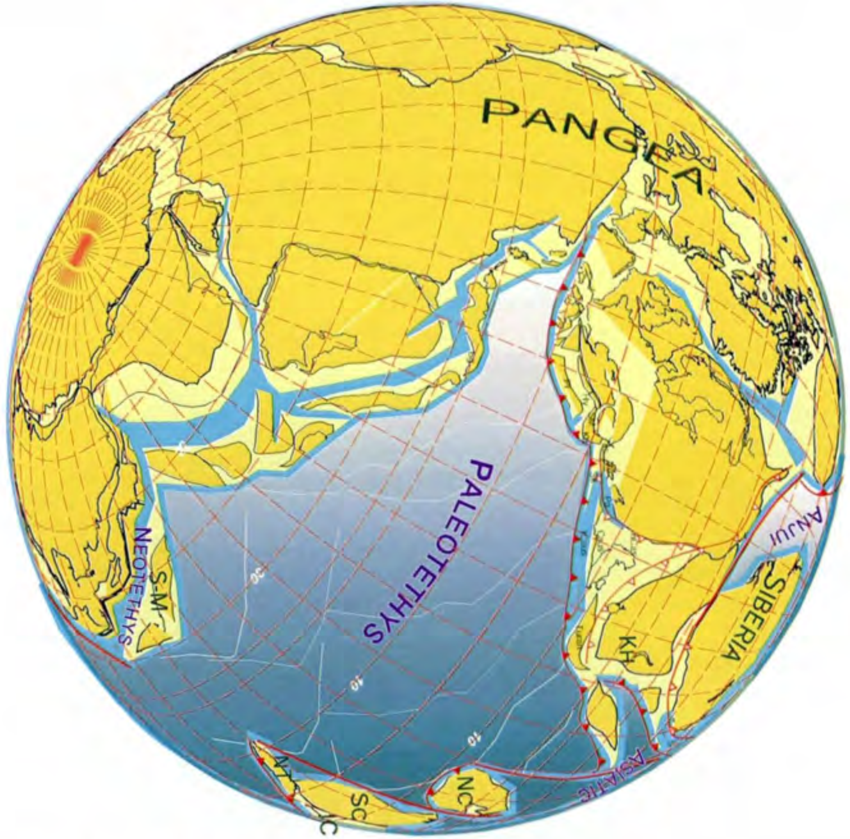
FIGURE 8



Middle Triassic (240 Ma)



Early Permian (280 Ma)

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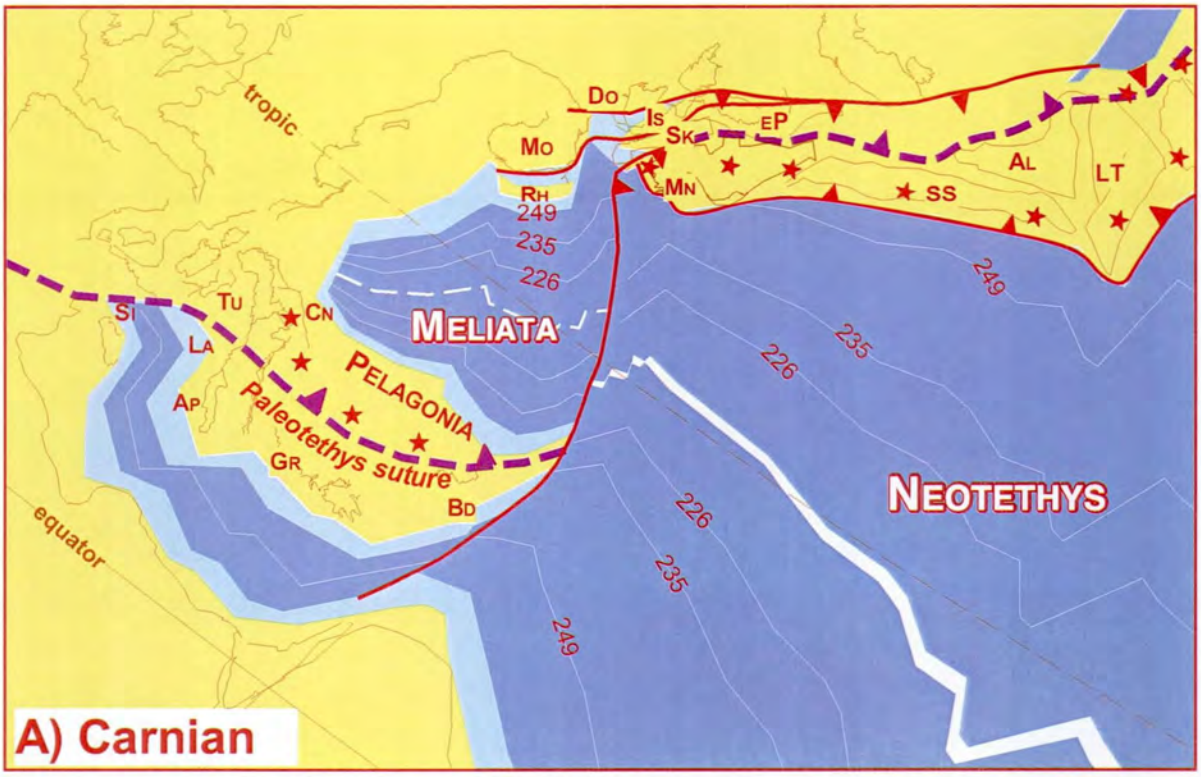
Stampfli, Mosar & Borel 2000

FIGURE 9



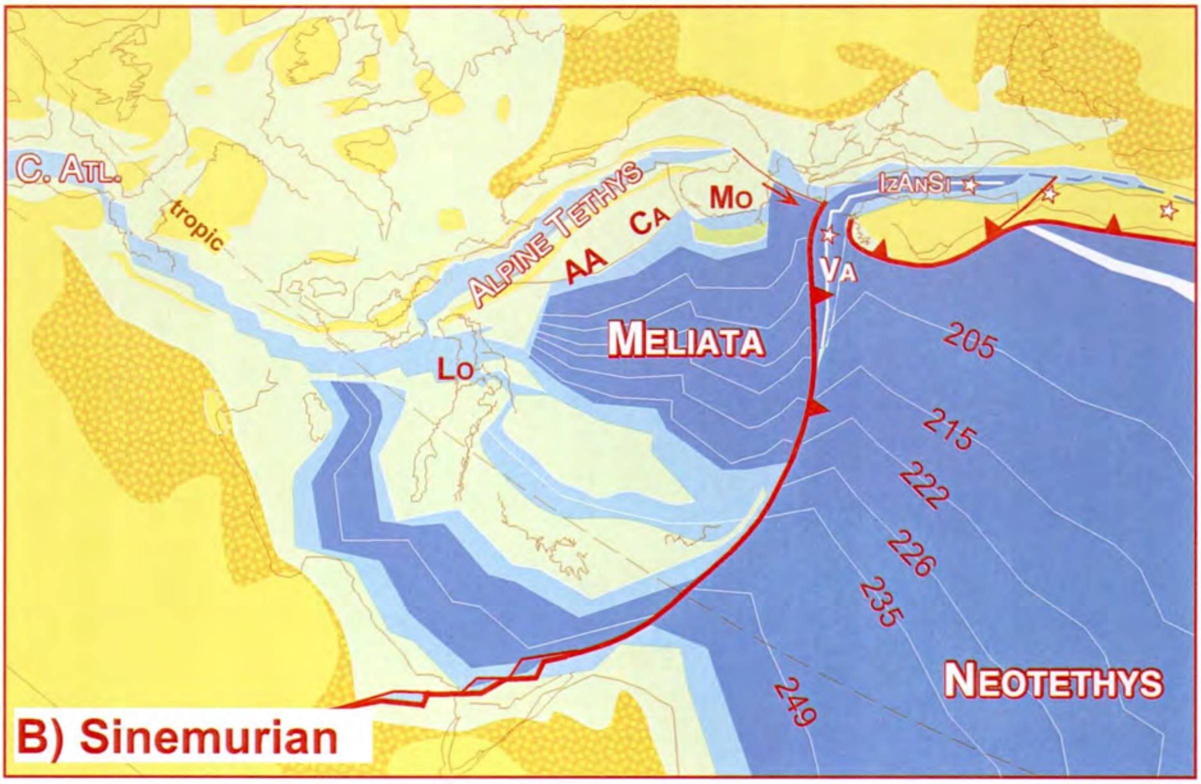


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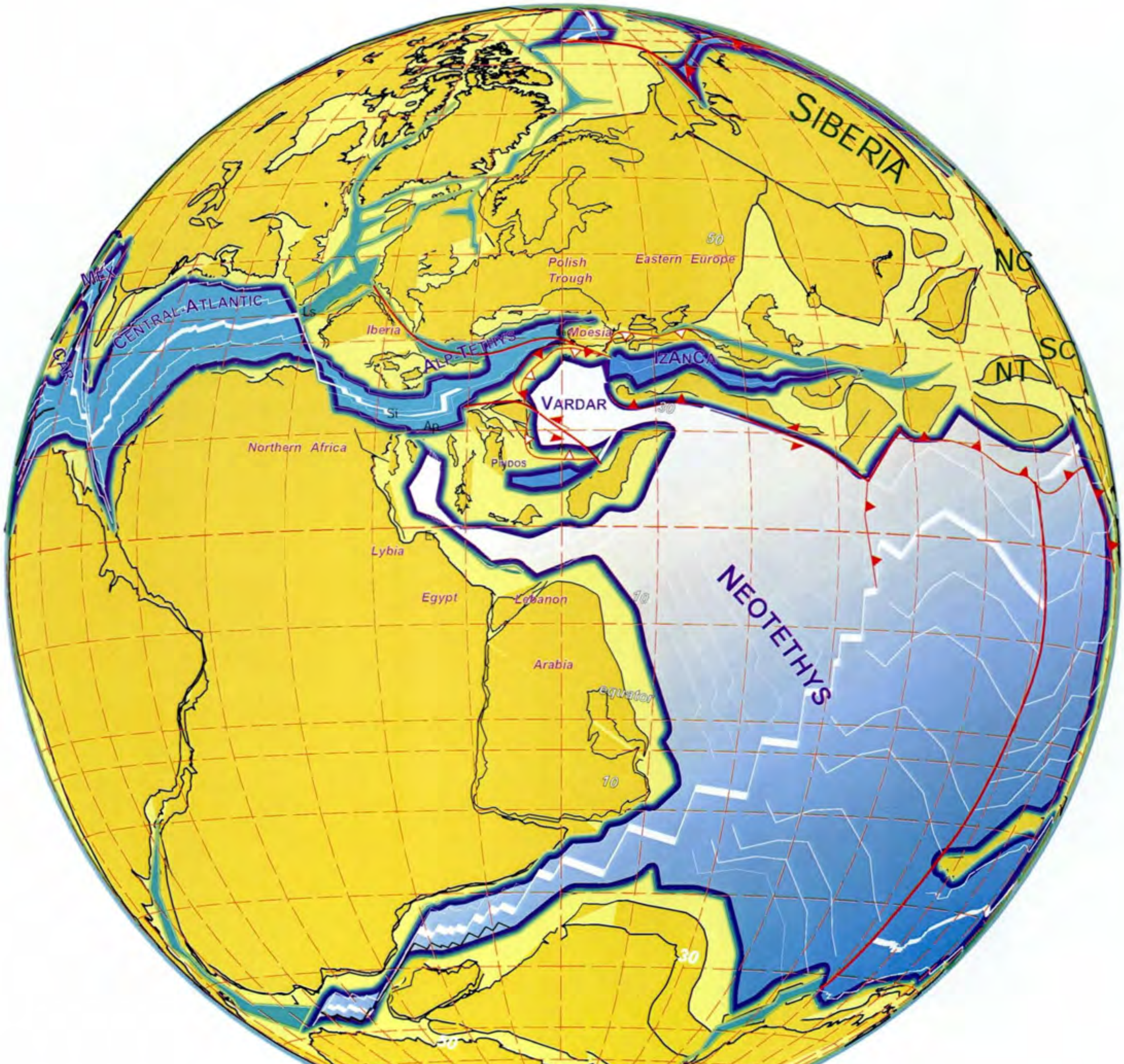
FIGURE 10



131 Ma - M0 - Valanginian

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FIGURE 11

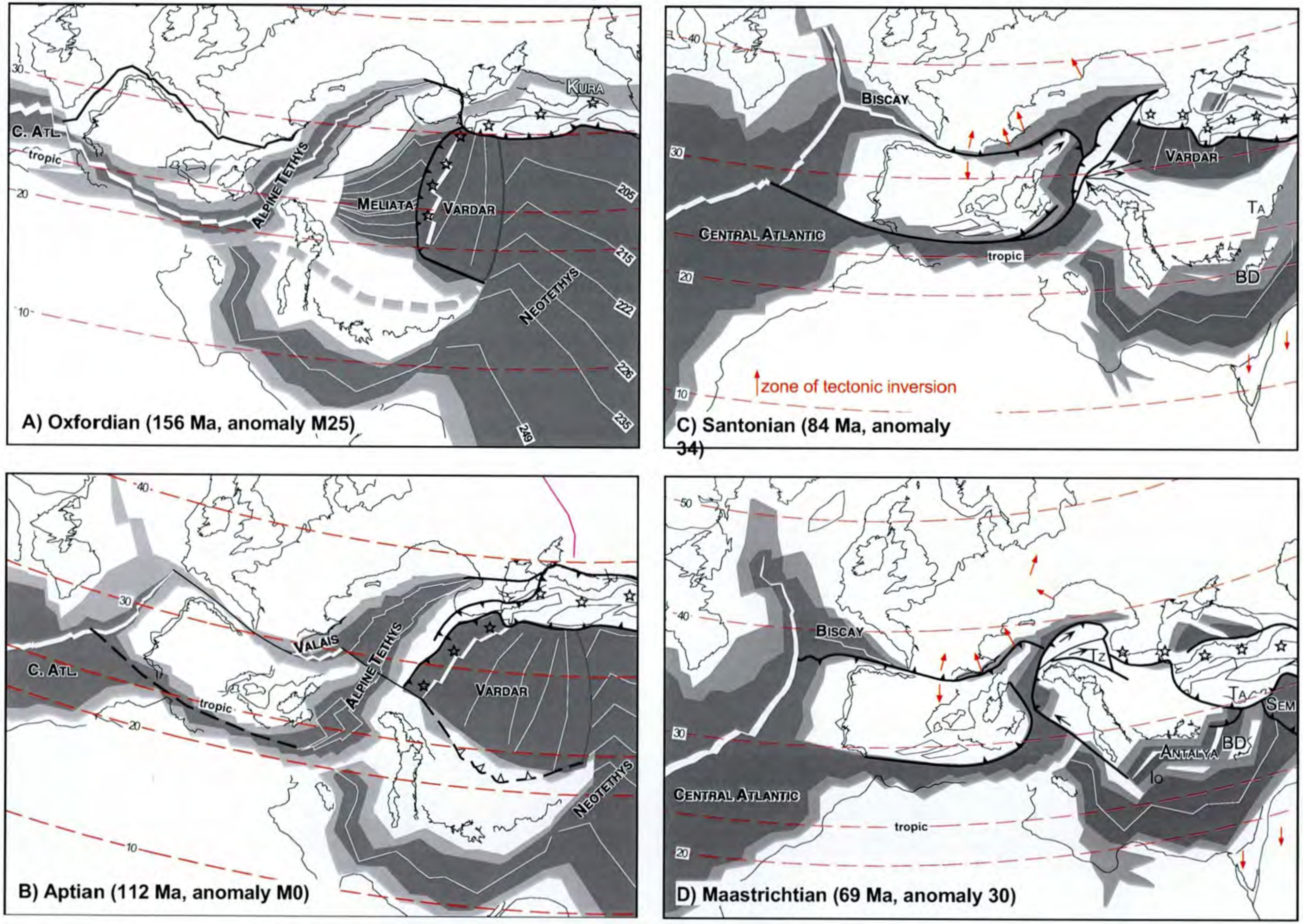
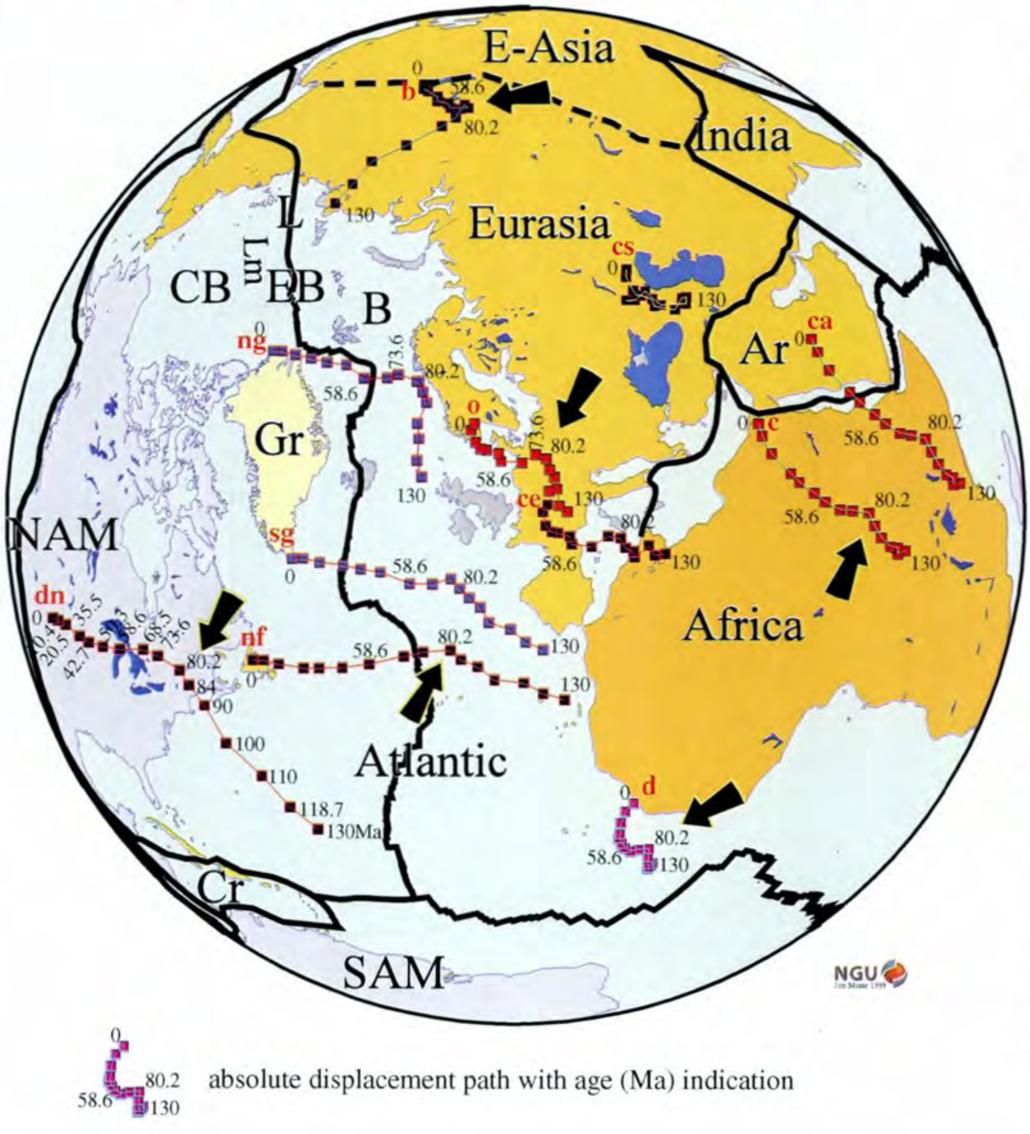

A) Oxfordian (156 Ma, anomaly M25)
C. Atl.
tropic
Alpine Tethys
Meliata
Vardar
Kura
Neotethys
205
215
222
226
235
249
C) Santonian (84 Ma, anomaly 34)
Biscay
Central Atlantic
Vardar
Ta
BD
tropic
zone of tectonic inversion
B) Aptian (112 Ma, anomaly M0)
C. Atl.
Valais
Alpine Tethys
Vardar
Neotethys
tropic
D) Maastrichtian (69 Ma, anomaly 30)
Biscay
Central Atlantic
Tz
Ta
Sem
Antalya
BD
Io
tropic

FIGURE 12

Present-day plate configuration and plate trajectories since 130Ma



absolute displacement path with age (Ma) indication

Coordinates (lat,long) of reference points

Eurasia:
48,005 Central Europe (ce)
60,010 Oslo (o)
48,050 N Caspian Sea (cs)
56,110 N Baikal (b)

Africa-Arabia:
15,343 Dakkar (d)
30,031 Cairo (c)
25,045 Central Arabia (ca)

Greenland:
60, 316 S Greenland (sg)
84,326 N Greenland (ng)

North America:
49,303 New Foundland (nf)
40,255 Denver (dn)

Ar = Arabia
B = Barents Sea
CB = Canada Basin
Cr = Carribean
EB = Eurasian Basin;
Gr = Greenland
L = Leptev Sea
Lm = Lomonossov Ridge;
NAM = North America
SAM = South America

FIGURE 13

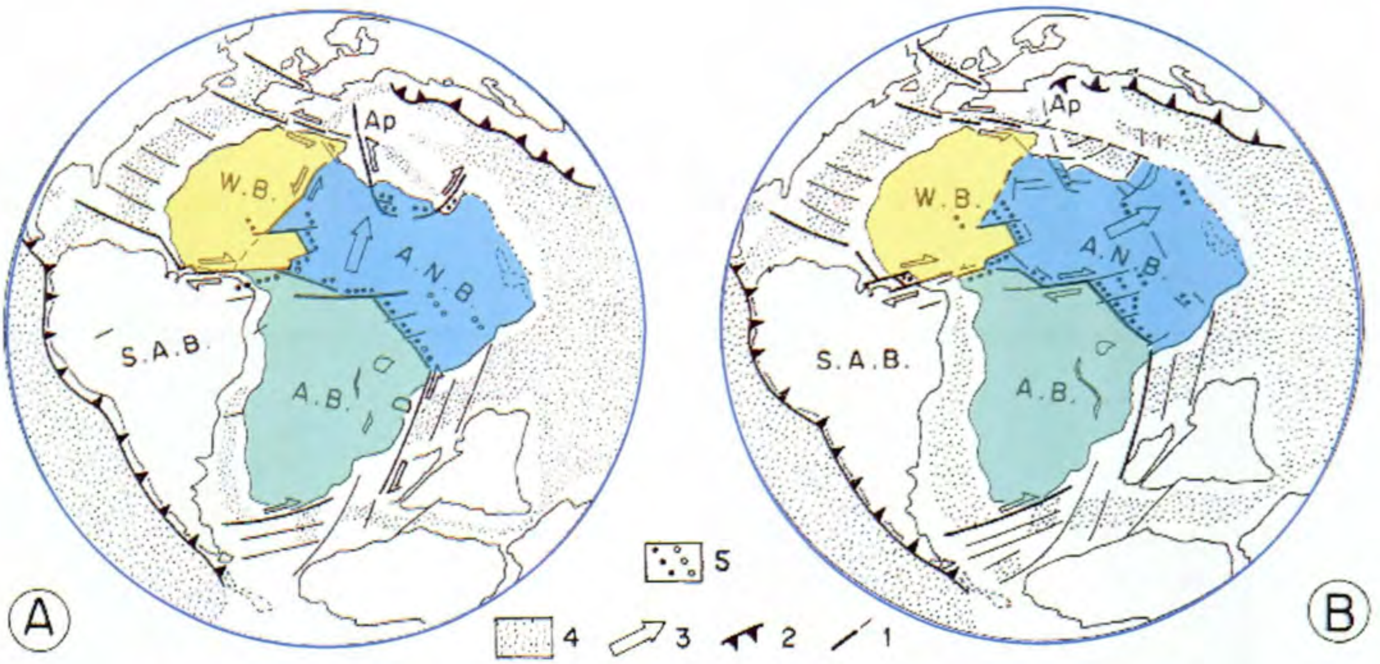


Fig. 6. Late Jurassic–Early Cretaceous break-up of western Gondwana. (**a**) Barremian (reconstruction *c*. 122 Ma). (**b**) Albian (reconstruction *c*. 107 Ma). Slightly modified from Guiraud and Bellion (1995). 1, Fault or strike-slip fault; 2, subduction zone; 3, relative motion of the Arabian-Nubian block; 4, oceanic crust; 5, main northern and central African Late Jurassic (○) and Early Cretaceous (●) rifts. A.B., Austral block; A.N.B., Arabian-Nubian block; S.A.B., South American block; W.B., Western block; Ap, Apulia; D, Davie Ridge.

90 Ma Turonian

125 Ma - Barremian

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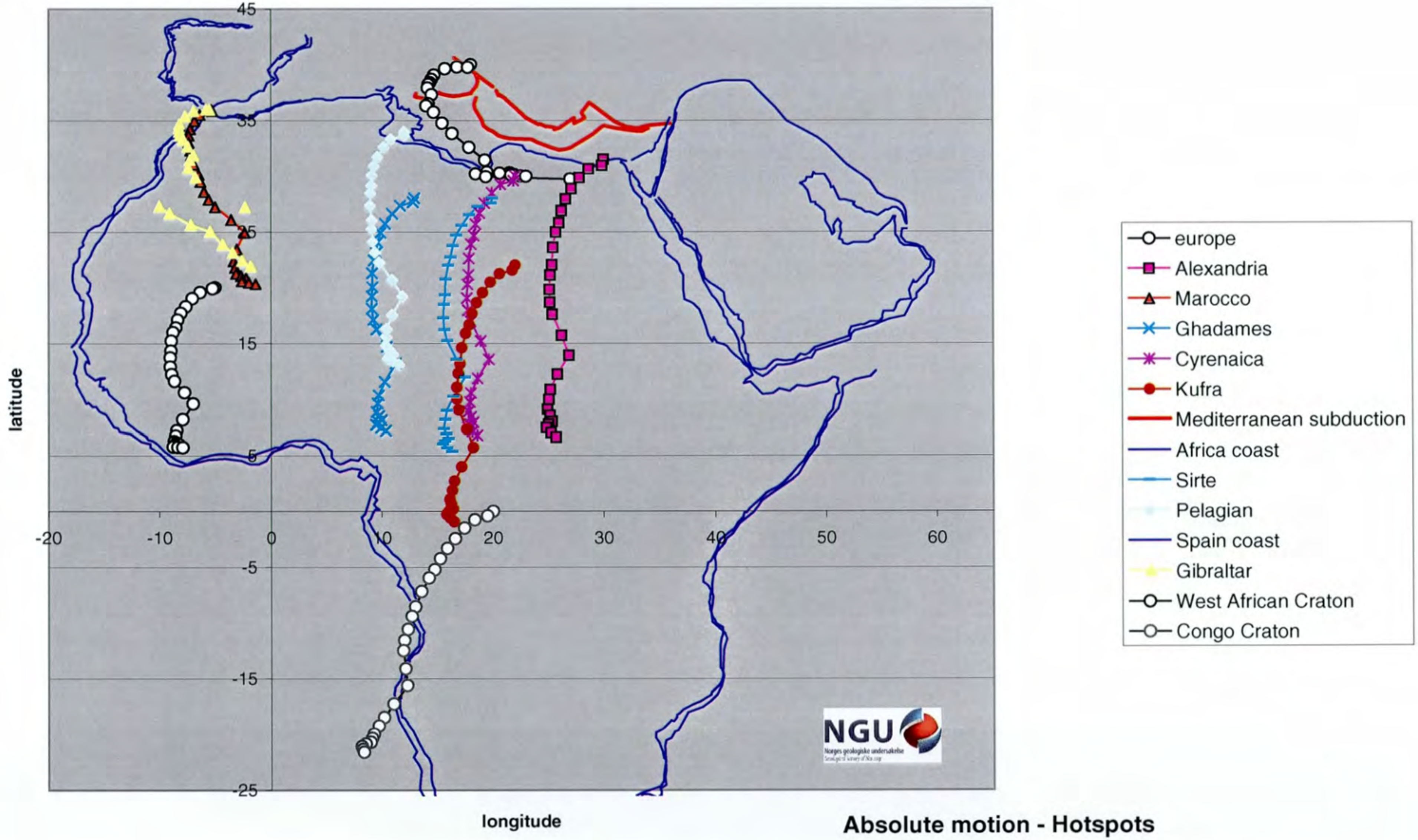
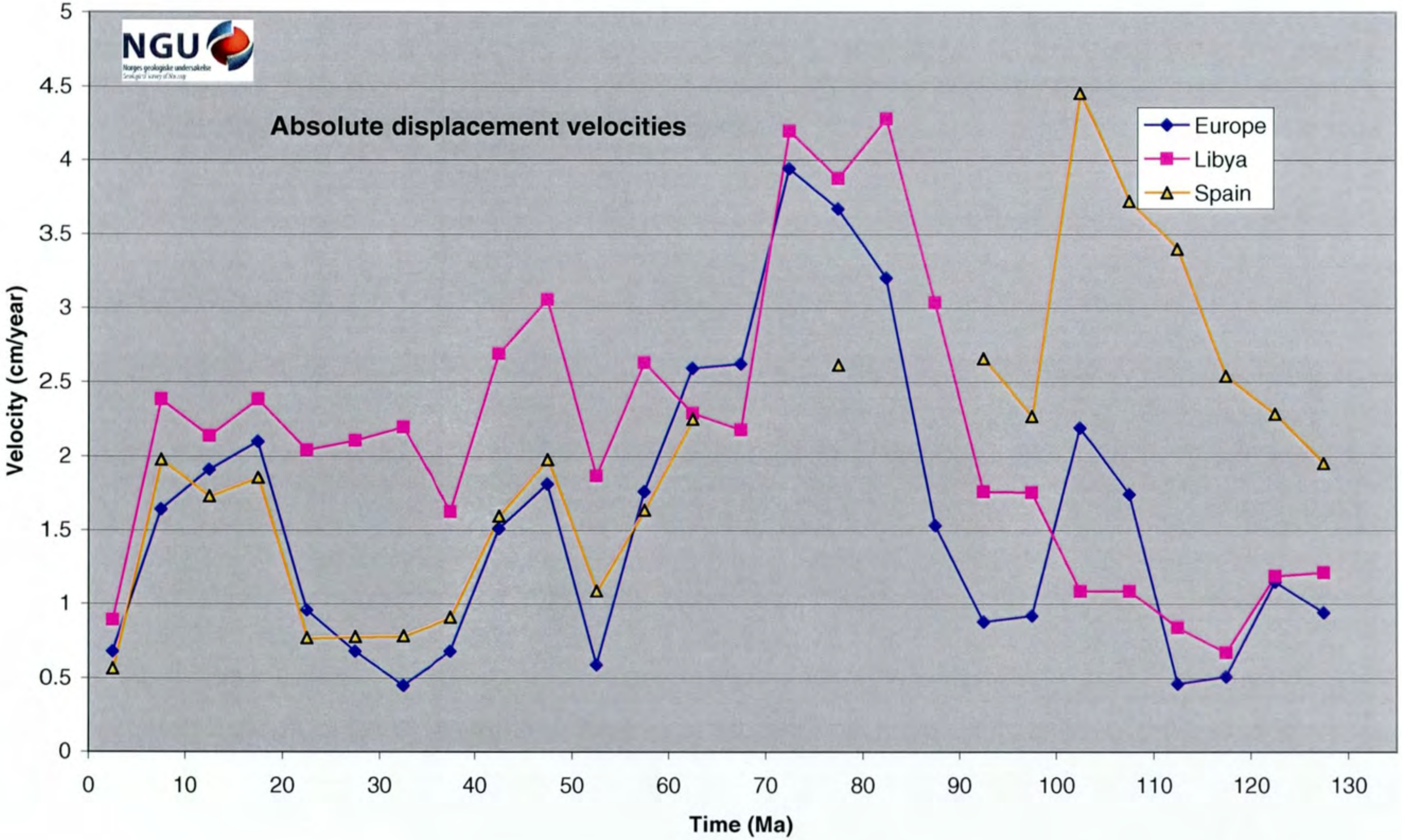
FIGURE 14



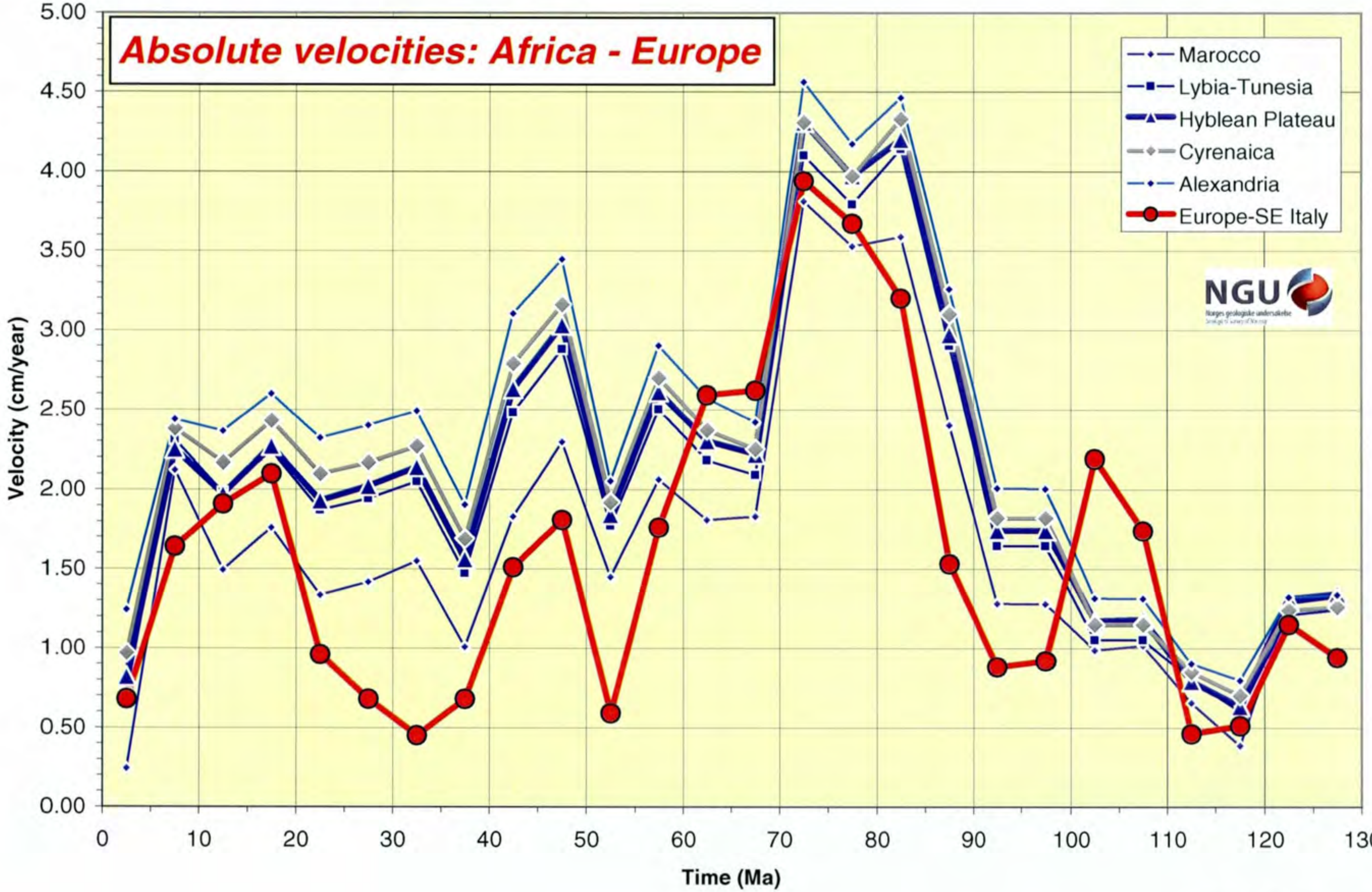
Absolute motion - Hotspots

FIGURE 15



Absolute displacement velocities
NGU
Norges geologiske undersøkelse
Europe
Libya
Spain
Velocity (cm/year)
Time (Ma)
0
0.5
1
1.5
2
2.5
3
3.5
4
4.5
5
0
10
20
30
40
50
60
70
80
90
100
110
120
130

FIGURE 16



Absolute velocities: Africa - Europe
Marocco
Lybia-Tunesia
Hyblean Plateau
Cyrenaica
Alexandria
Europe-SE Italy
NGU
Velocity (cm/year)
Time (Ma)
0.00
0.50
1.00
1.50
2.00
2.50
3.00
3.50
4.00
4.50
5.00
0
10
20
30
40
50
60
70
80
90
100
110
120
130

FIGURE 17

Relative Movement of N-Africa - Europe Fixed

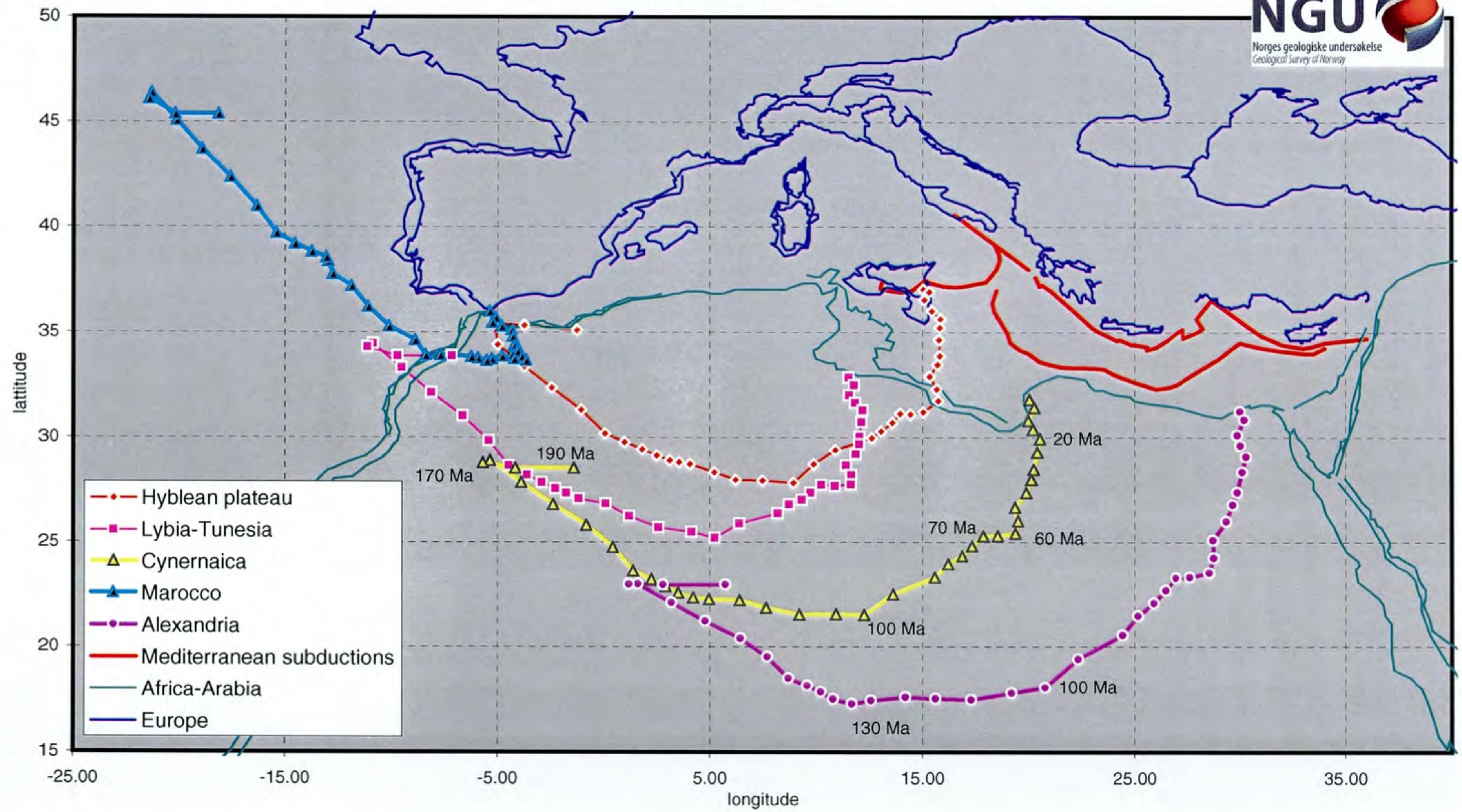
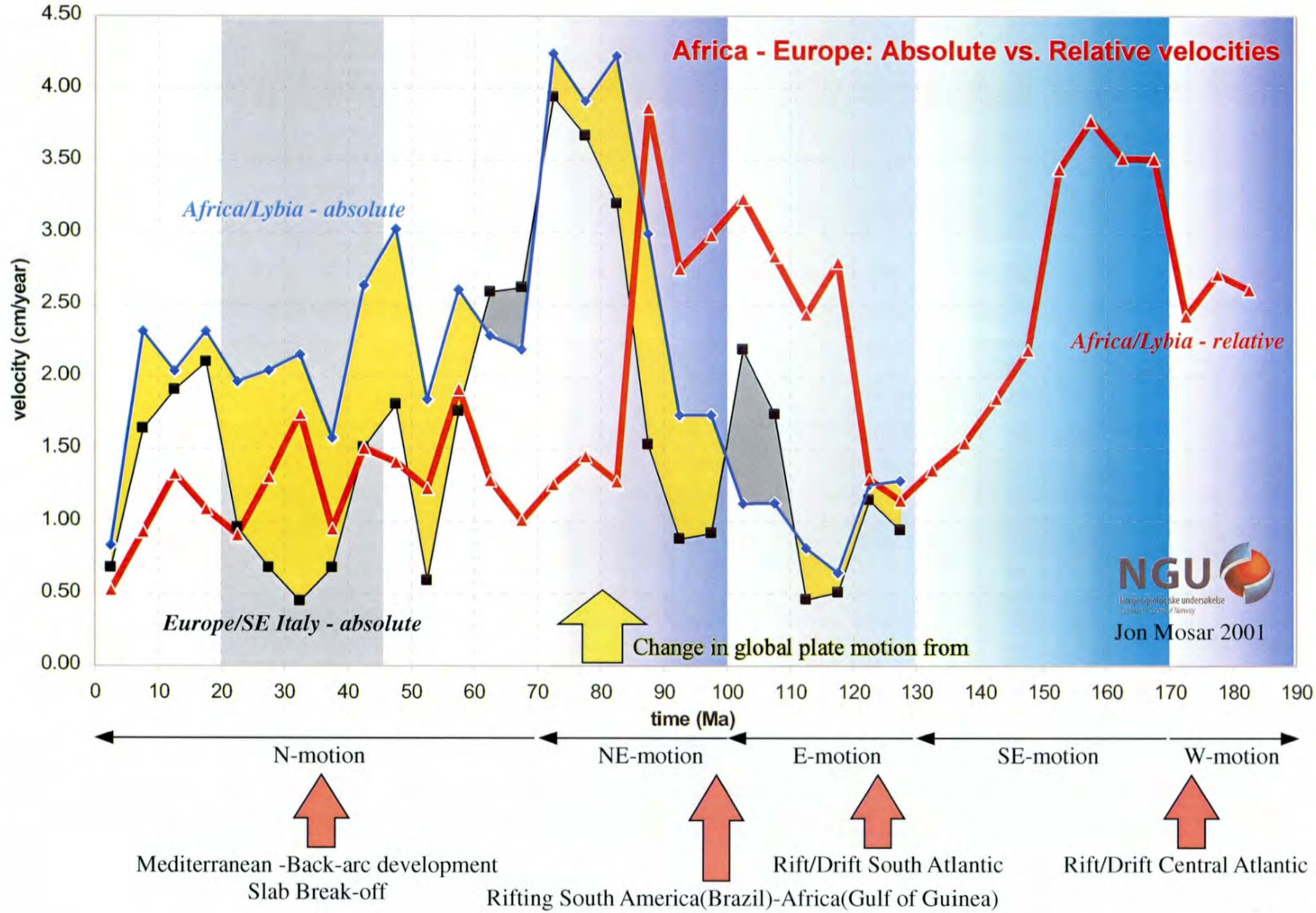


FIGURE 18



Africa - Europe: Absolute vs. Relative velocities
velocity (cm/year)
time (Ma)
Africa/Lybia - absolute
Africa/Lybia - relative
Europe/SE Italy - absolute
Change in global plate motion from
NGU
Norges geologiske undersøkelse
Jon Mosar 2001
N-motion
NE-motion
E-motion
SE-motion
W-motion
Mediterranean -Back-arc development
Slab Break-off
Rifting South America(Brazil)-Africa(Gulf of Guinea)
Rift/Drift South Atlantic
Rift/Drift Central Atlantic

FIGURE 19

Absolute vs. Relative Velocity of Africa - Europe

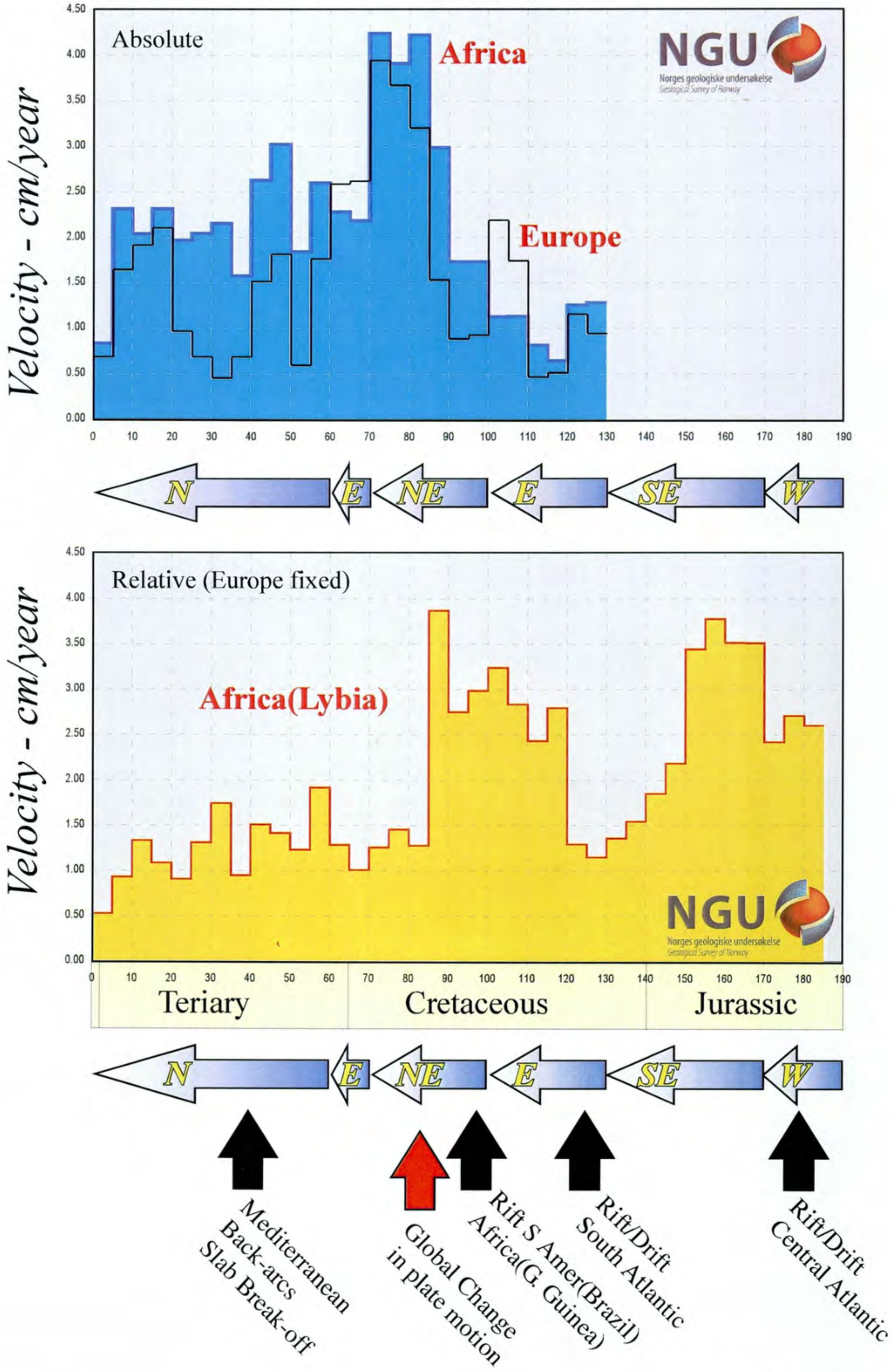


FIGURE 20

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