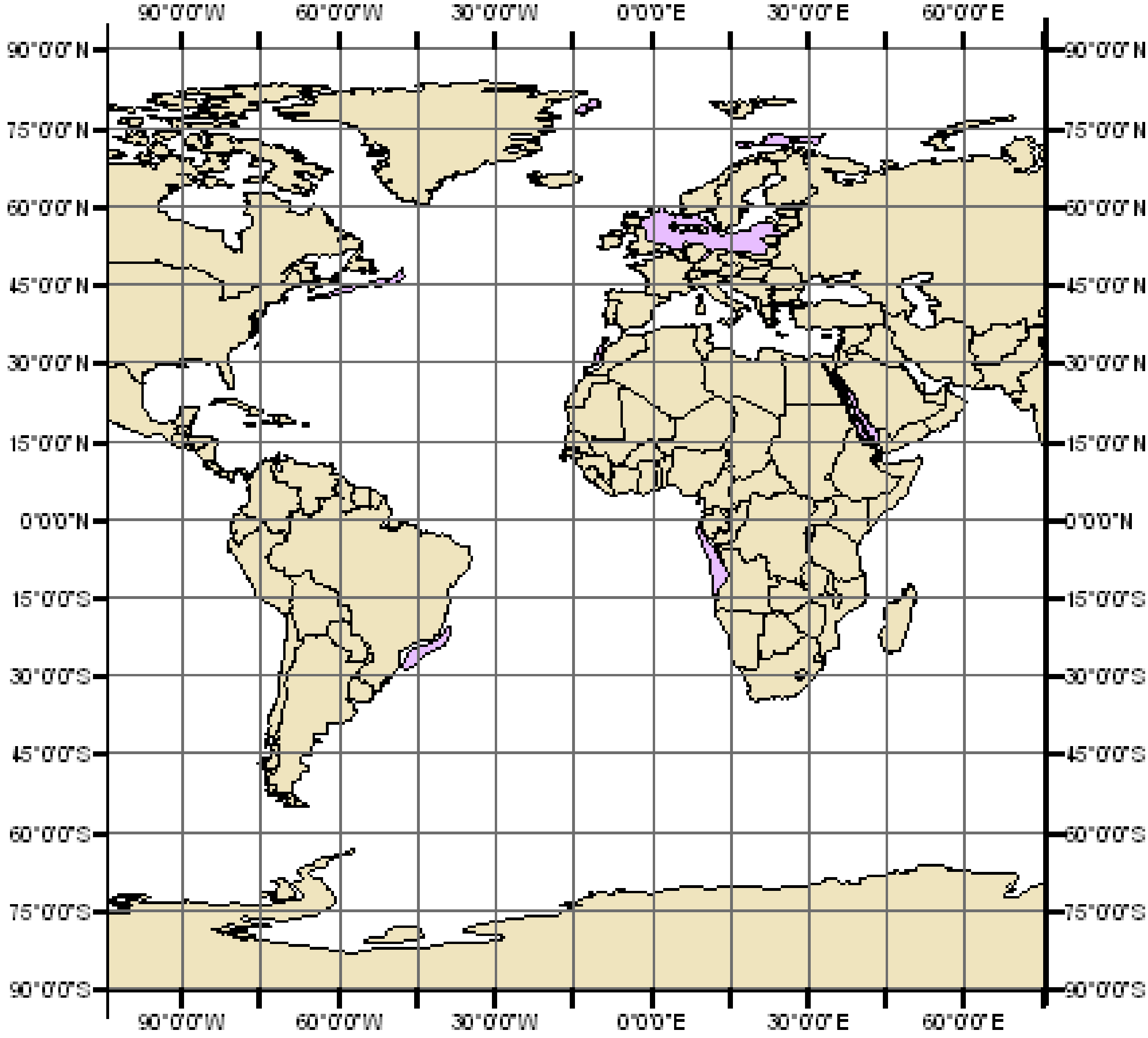

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Salt basins in the Atlantic and Red Sea-Afar

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REPORT

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Summary:

Evaporite basins in the SW Barents Sea, Southern North Sea, Central Atlantic, South Atlantic, and Red Sea-Afar largely formed at subtropical latitudes. This is consistent with classic views on development of evaporites. The SW Barents Sea and the Southern North Sea evaporite basins formed in abandoned continental rifts. The Central Atlantic salt basins (Newfoundland, Nova Scotia, eastern US, Morocco, and Senegal/Guinea Bissau) formed in relatively confined intrarift basins, prior to continental breakup. The South Atlantic salt basins (Brasil and West Africa) formed in a wide basin during the late stage of rifting, shortly before breakup. Only minor magmatism is documented for the Late Carboniferous SW Barents Sea evaporite basin. The Permian North Sea evaporite basin overlies a three-armed Permian rift, which was associated with considerable magmatism. The magmatism preceded the evaporite sequence by c. 50 Ma, and the relationship between magmatism and the evaporites is indirect through thermal subsidence. Neither the SW Barents Sea nor the North Sea developed into oceanic basins. The northern Central Atlantic evaporites formed in Late Triassic – Early Jurassic rifts along the continental margins approximately synchronously with a very widespread pulse of dyke swarm injection at c. 200 Ma. However, the evaporites predate opening of the Central Atlantic by c. 20 Ma. None of the mentioned evaporite sequences can thus be related to oceanic seafloor spreading. The southern Central Atlantic salt basins have not been dated and their age is inferred from the north basins. If the age is similar, the southern basins formed at anomalously equatorial latitudes. In the South Atlantic, the continent-ocean boundary (COB) is not well established, and the relationship between the outmost limit of evaporites and the COB remains somewhat unclear. However, the preferred plate reconstructions performed in this study indicate that the Aptian salt basins formed on rifted continental crust prior to breakup. The Red Sea – Afar evaporites formed within a continental rift in Middle to Late Miocene time. Salt deposition postdated magmatism in the Afar triangle by approximately 20 Ma and postdated dyke swarm injections along the rift shoulders by c. 10 Ma. The salt formed synchronously with minor bimodal rift magmatism but predates breakup by approximately 2-6 Ma. Overall, there is a strong relationship between the timing of salt mobilization (diapirism etc) and tectonic pulses. Except for the Senegal/Guinea Bissau and Carolina Trough salt deposits (southern Central Atlantic), the studied basins were located at subtropical latitudes during the times of salt deposition and therefore, comply with traditional evaporite models. The anomalous Senegal/Guinea Bissau and Carolina Trough salt could be an artefact relating to an erroneous age estimate.

Keywords: Salt	Evaporites	Magmatism
Diapirism	Tectonics	Plate reconstruction

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1. LATE CARBONIFEROUS TO EARLY PERMIAN SW BARENTS SEA BASIN

During Late Carboniferous to Early Permian times arid conditions marked the Barents Sea shelf. This occurred during a major regression, resulting in deposition of successive sabkha cycles and culminating in evaporite deposition in more restricted axial portions of the basin (Fig. 1.1) (Gérard & Buhrig, 1990).

The Barents Sea evaporite basin is considered to be analogous to the Permian evaporite basin in the southern North Sea (although older), but was characterized by shallower environments and lower subsidence rates. This resulted in relatively thin evaporite deposits. The thickest evaporate sequences were deposited in basins bound by major faults utilizing the Caledonian structural grain. These occur in the Nordkapp Basin, the Ottar Basin, and the Maud Basin (Gérard & Buhrig, 1990). The initial basin development occurred during middle Carboniferous regional extension (Gudlaugsson et al., 1998, Nilsen et al, 1995). The eastern basin limit has not been possible to define in this study due to extending into the gray zone against Russia.

Regionally, evaporite sequences also formed in the Timan Pechora Basin (Artinskian-Kungurian age salt). On Svalbard (Gipsdalen Group) and in the Wandel Sea Basin, Upper Carboniferous to Lower Permian carbonates, subordinate evaporites, and clastics formed. Carbonates, clastics and evaporites developed during Early Permian time in the Sverdrup Basin. Offshore NE Greenland, the Denmarkshavn Basin contains evaporites, but the limit of the evaporates in the basin are only poorly defined by limited seismic data and from gravity measurements. The age of the evaporite sequence in the Denmarkshavn Basin is unknown, but is here assumed to be similar to that of the evaporites on the Barents Sea shelf, i.e. Late Carboniferous to Early Permian (see Section 1.5).

Following the evaporite deposition, open marine conditions resulted in build-up of shallow carbonate platforms. These are mainly developed along the gentle basin margins and on fault-bounded intrabasinal highs (Gérard & Buhrig, 1990).

In Middle Permian time a major transgression drowned the shelf and shut off the carbonate build-up.

1.1 Rifting history

The area containing Upper Carboniferous – Lower Permian salt in the SW Barents Sea lies within a region of Middle Carboniferous rifting (Fig. 1.2). This rift belonged to the Palaeozoic Atlantic rift, which utilized the collapsed Caledonian foldbelt (Gudlaugsson et al., 1998; see Fig. 1.5). The Scandinavia – Greenland Caledonides probably continue NE through the SW Barents Sea, while the Innuitian and Svalbard Caledonides swing to the NW between Svalbard and North Greenland.

Two main phases of Palaeozoic extension are suggested.
1) Middle Carboniferous
2) Permian – Early Triassic

The middle Carboniferous rift was c. 300 km wide and extended at least 600 km in a NE direction, and is characterized by interconnected basins with halfgraben geometries (Fig. 1.2). This phase of extension was followed by regional subsidence and sag-basin development, interrupted by the Permian-Early Triassic rifting that reactivated the N-trending structures in the western part of the rift.

The SW part of the Nordkapp Basin is a large halfgraben dipping against the Nysleppen Fault Complex on the W side. The half graben rolls over from the Finnmark Platform against the Norsel High, and the maximum basin depth is estimated to 12-14 km. The NE part of the basin trends more E-W and its main boundary fault appears to be the Thor Iversen Fault Complex on the south side. While the rift changes strike and polarity, there is no evidence at depth for halfgraben geometries.

1.2 Magmatic history

Carboniferous magmatism is locally found onshore in northern Norway, but of Early Carboniferous age. The best studied dykes (continental tholeiites) occur in different places on the Island of Magerøya and yield $^{40}Ar/^{39}Ar$ ages of 337 $\pm$ 0.4 Ma (reviewed in Roberts et al. 2003). The dykes intrude a NW-SE-trending fault- and joint-system that parallel the Trollfjorden-Komagelv Fault Zone (TKFZ). These faults formed during Late Palaeozoic extension, but were reactivated as rifting occurred during the progressive opening of the Barents Sea to the North. Locally, basaltic magmatism accompanying the rifting led to emplacement of the c. 337 Ma dolerite dykes. The dykes are important markers for the timing of this extensional event and are themselves not deformed by later faulting. The dykes were probably contemporaneous with the earliest syn-rift sedimentation in the Barents Sea (e.g. Nordkapp Basin). The offshore continuation of the TFKZ broadly intersects with the Loppa High where a well penetrated a c. 800 meters thick sequence of volcanoclastics of assumed Early Carboniferous age (Geir Elvebakk, pers. communication 2002). It is, therefore, likely that a volcanic centre existed offshore northern Norway in the Early Carboniferous.

1.3 Salt deposition

The age of the SW Barents Sea evaporites is estimated to latest Carboniferous – Early Permian (Gzelian-Asselian/Sakmarian) (Gérard & Buhrig, 1990). This corresponds to absolute ages of c. 296-282 Ma (Batlas, 2002). The evaporites are thus c. 25-30 Ma younger than the Middle Carboniferous rifting that generated the basin architecture into which the evaporites were formed, and c. 40-50 Ma younger than the dated dykes.

In the Nordkapp Basin, early work suggested that the average original salt thickness was more than 2 km in SW basin, and more than 4 km in the NE basin (Jensen & Sørensen, 1992). Reconstruction by Gudlaugsson et al (1998) yielded estimates of original salt basin depth to 1.6-1.8 km for the SW basin, and 2.7-2.9 km for the NE basin. More recently, Koyi et al., (1995) suggested that the salt volume estimate in a given diapir may be significantly overestimated, perhaps by as much as 200-400%; modern seismic data reveal drop-shaped diapirs with thin stems (Fig. 1.3) rather than cylindrical diapirs, which was the assumed geometry by Jensen & Sørensen (1992). If Koyi et al. (1995) are correct, which seismic data seem to suggest, the original salt thickness was probably in the order of 500-1000 m in the SW basin, and c. 1000-2000 m in the NE basin.

The salt was initially overlain by a uniform thickness of Late Permian carbonates and Lower Triassic siliciclastics.

1.4 Salt deformation

In Early Triassic time (Late Scytian, c. 250-240 Ma) regional extension triggered reactive diapirism due to stretching of the overburden. A low rate of extension caused an effective decoupling between basement and overburden. Thus, the overburden extension (grabens) was not necessarily localized above basement faults. The reactive diapers grew tall enough to cause active piercing of the overburden (Nilsen et al., 1995).

In Middle – Late Triassic time the diapirs grew passively by downbuilding. The passive growth was rapid and salt flow was vigorous. Significant thickening of the Middle Triassic section from the basin margins towards the diapirs dates the passive growth phase. The base of the Anisian (240 Ma) marks a major erosional event, related to depletion of the mother salt source layer. Sediments above the Anisian are of relatively uniform thickness. This indicates that the depo-throughs (areas receiving sediments between salt bodies) had grounded. Paradoxically, diapirs continued to rise passively until the Middle Triassic (Ladinan-Carnian, c. 230 Ma), when salt overhangs formed. Especially in the NE basin overhangs are common. After grounding, the cause of continued diapir growth is suggested to relate to regional extension, causing deepening of the basin and inducing gravity gliding towards the basin centre, in turn compressing the diapirs stems and inducing vertical growth (Nilsen et al., 1995). This interpreted phase of extension is supported by observations of coeval normal faulting along the basin margins (Gabrielsen et al. 1992). By Late Triassic the diapirs stopped growing and were buried.

Despite regional extension in Late Jurassic – Early Cretaceous, the Nordkapp Basin diapirs were not significantly affected. The diapirs were buried by 1000-1500 m of flat-lying Cretaceous sediments.

In Late Cretaceous time diapirs grew and arched the thick roof cover. The cause is suggested to have been gravity gliding, triggered by Late Cretaceous regional extension associated with extension in the Norwegian Sea (Nilsen et al., 1995). Similar age diapirism is reported from the Sørvestnaget and Tromsø Basins (Faleide et al., 1993).

In Eocene – Oligocene time a new phase of diapir reactivation took place, and has been related to compression. This last event pinched off the stems of many diapirs, producing teardrop geometries. The salt diapirs have since been stable. Late Pliocene-Pleistocene regional uplift caused erosion of ca 1 km or more, peneplaning the tops of the diapirs.

1.5 Plate tectonic reconstruction

Plate reconstructions in this report uses a refinement of a global and self-consistent plate tectonic model (Torsvik et al., 2001; Torsvik & Van der Voo, 2002) based on palaeomagnetic data, magnetic anomaly fits (< 175 Ma), and estimates of pre-drift extension.

During the Early Carboniferous, northernmost Norway-Greenland, the Barents Sea and Svalbard were located at tropical to low northerly latitudes. Northward drift during the Carboniferous (5-6 cm/yr), as demonstrated from palaeomagnetic data (Roberts et al. 2003), is reflected in the sedimentary facies in the Barents Sea realm, i.e. a change from tropical (Early Carboniferous) to subtropical carbonates and evaporites in the Late Carboniferous. The Late Carboniferous reconstruction shown here (Fig. 1.4) demonstrates that salt deposits now found in Northeast Greenland and the Nordkapp Basin probably formed within an elongate rift basin that stretched from 16°N (North Lofoten – Troms region) to 23°N (Barents Sea), i.e. within the Carboniferous subtropics. The elongated belt of salt basins broadly follows the collapsed Caldeonian fold belt, as well as the axis of the assumed Iapetus Suture, i.e. a former Silurian (c. 425 Ma) subduction zone between Scandinavia and Greenland-Svalbard (subduction beneath Greenland). Mesozoic extension in NE Greenland has been accounted for in the position of the NE Greenland basin. Late Carboniferous magmatism of the same age as the salt deposits is not known.

2. PERMIAN NORTH SEA BASIN

In NW Europe, two east-west oriented basins formed in Late Permian time (Wuchiapigian-Changhsingian). These are referred to as the Northern and Southern Permian Basins (Fig. 2.1). To the west of these basins, in parts of the Irish Sea and NW England, lie smaller localized salt basins that do not appear to have been linked with the Northern and Southern Permian basins.

Rocks in the Permian North Sea basins belong to two sequences.

1) Rotliegend Group: largely sandstones with basal volcanics
2) Zechstein Group: carbonates and evaporates with local clastic rocks.

The Lower Rotliegend volcanic rocks were mainly concentrated east of the Central Graben and between the Mid North Sea – Ringkøbing – Fyn Highs and Norway. These volcanics extend into the Oslo Graben (Figs. 2.1 and 2.2).

Aeolian sandstones dominate the Upper Rotliegend Formation. A lateral facies shift to silty claystones in the Norwegian Central Graben (the Fraserburgh Fm) indicates that the graben already was weakly expressed at this time. Also the southern Viking Graben appears to contain these facies, indicative of an early phase of graben subsidence.

It is worth noting that much of the Permian deposits lack fossils; the strata consist of desert sandstones or of carbonates and halite that resulted from extreme climatic conditions.

2.1 Rifting history

Permian rifting resulted in a three-armed rift, centred on the Skagerrak area. A NNE-trending rift arm includes the Skagerrak Graben and Oslo Rift, while a NW-trending arm is made up of the Farsund Basin, and a SE-trending arm of the Sorgenfrei-Tornquist Zone (Neumann et al., 1992).

A relatively clear correlation can be seen between the three-armed rift configuration, the distribution of magmatic rocks (Figs. 2.2 and 2.3), and the location of the Permian basins (Fig. 2.1). Thus, the subsidence of the Permian Basins in all likelyhood reflects thermal subsidence over an intracontinental rift.

2.2 Magmatic history

Glennie et al. (2003) subdivided the Permian magmatism into three phases:
1) 300-288 Ma
2) 281 +-8 to 274 ±14 Ma
3) 269 +-4 to 261 ± 4 Ma.

Neumann et al. (1992) on the other hand subdivide the tectonomagmatic history of the Oslo rift into five main phases:

1) > 300 Ma	Subsidence without major faulting
2) 300-295 Ma	Vertical motion mainly along N-trending faults during widespread basaltic volcanism

3) 295-275 Ma	Main rift period during which rhomb porphyries formed
4) 275-240 Ma	Change from basaltic shield volcanism to central complexes of mixed composition, many of which experienced caldera collapse
5) <240 Ma	Dyke intrusions lasting into Triassic time.

In the last two decades, refinements in geochronological studies have changed the geological time scale calibration and the Permian – Carboniferous boundary is now placed at 292 Ma (Fig. 2.4). Moreover, most recent discussions of the classical western European stage names such as Autunian and Rotliegendes now assign the lowermost parts of these stages to the latest Carboniferous (Gzhelian or Stephanian). For volcanic results, modern geochronological work (^{40}Ar-^{39}Ar and U-Pb) has also typically given older ages than the K-Ar (because of argon loss) or Rb-Sr techniques. Recent dating of volcanic rocks in Norway, Scotland, Sweden and Central Europe (summarized in Van der Voo & Torsvik 2004) show that the so-called Permo – Carboniferous magmatism in Europe peaks at around 295-300 Ma, i.e. Late Carboniferous. Younger Permian volcanism (e.g. Torsvik et al. 1998) is more of local nature. In summary, bulk magmatism predates salt deposition (Zechstein) by ca. 50 million years.

2.3 Salt deposition

Menning (1995) placed the base of the Zechstein at 258 Ma (intra Wuchiapigian). If correct, this implies that the Rotliegend Group was deposited in the course of s c. 8 Ma time span and the Zechstein Group during c. 7 Ma.

During initial deposition of the Rotliegend Group, subsidence in the two basins outstripped sedimentation, resulting in basin centres below sea level. This created a desert lake with a sabkha environment along its margins in the southern basin. The distribution of lake or sabkha deposits is less clear in the northern basin and in the Moray Firth, but they exist there as well.

It is believed that at the end of the Rotliegend deposition, the central parts of the desert basins were situated 200-300 m below sea level. Sea level rose and Boreal oceanic waters entered the North and South Permian basins from the north via the Viking Graben, which in turn was connected with rifts in East Greenland. The initial entry of oceanic water was catastrophic, resulting in reworking of the top of Rotliegend, forming the Weissliegend facies (Glennie & Buller, 1983). However, surrounding areas were still in a hyperarid desert environment.

Due to post-depositional salt movement, isopach maps are difficult to draw. However, as much as 1500 m of evaporites appear to have deposited in the basin centre (Glennie et al, 2003). The Zechstein stratigraphy follows a three- or fourfold cyclic pattern from basin edge to centre. Typically the pattern is limestones and dolomites, anhydrite, and halite. In the UK, the following subdivision is established: Kupferschiefer Fm, Halibut Carbonate Fm, Turbot Anhydrite Fm, and Shearwater Salt Fm.

Other than reworking of the Rotliegend sands (forming Weissliegend facies), the first marine deposit of the Zechstein Sea is the Kupferschiefer Fm, which usually is c. 1 m thick. It is widespread, but locally absent over highs such as over the northern flank of the Mid North Sea – Ringkøbing - Fyn Highs.

Subsequently, each Zechstein cycle started with limestone deposition on the shallow shelf areas (basin edge) and thin organic-rich calcareous shale in the basin centre. Evaporite drawdown during lowering of local sea level resulted in deposition of anhydrite at the basin edges. These anhydrites lie over and basinward of the limestones. Upon further evaporation of the seawater, halite and other salts formed above the anhydrite in the deeper part of the basin, and later also over the shelf areas.

Two carbonate sequences of the Halibut Carbonate Fm are separated by an anhydrite, the Iris Anhydrite Member. The carbonates reflect deposition during free circulation of seawater, while the anhydrite reflects restricted marine conditions. The mentioned sequences correlate with the classic Zechsteinkalk, Werraanhydrite, and Innes Carbonate Member in the Southern Permian Basin.

The Turbot Anhydrite Fm consists of two anhydrite sequences separated by the Turbot Carbonate Unit. The formation was deposited in a basin-margin sabkha environment.

The Shearwater Salt Fm combines the Z2, Z3, and Z4 salt cycles of the Southern Permian Basin. In the Southern Permian Basin Z2 is thickest, while in the Northern Permian Basin the equivalent to Z3 (Leine) is thickest. This has been suggested to reflect different degrees of subsidence.

The Shearwater Salt Fm formed as the salinity in the Zechstein Sea reached high values. 80% of the brine is believed to have evaporated. The thickness of the formation suggests that the influx of seawater from the Boreal Ocean never was completely cut off. It is believed that salt crystals formed both in shallow water and close to the surface in more basinal areas. The rate of salt precipitation was likely greater at the basin margins, and salt may also have been redistributed basinward by density currents.

2.4 Salt deformation

Late Permian was a significant period of salt redistribution. Throughout the area of Permian salt tectonics, the entire Zechstein interval consists of km-scale pods, separated by salt walls (Fig. 2.5). These salt pods are called mini-basins. Sediments in the mini-basins were deposited during salt withdrawal. The salt activity is not related to basement extension, but is believed to reflect thin-skinned gravity gliding into the subsiding basin. Modest extension of the cover is believed to have generated differences in accommodation space, which set up differential loading. Once achieved this would continue. Well 20/19-1 drilled into a mini-basin encountered 38% net siliciclastics in the Zechstein interval. It seems that the minibasins captrured fluvial systems (Stewart & Clark, 1999).

Triassic deposition is characterized by generation of km-scale mini-basins in the North Permian Basin (same scale as Late Permian mini-basins) (Fig. 2.6). Differential loading became the main driving force during mini-basin development. Triassic basement extension clearly took place in salt-free areas near the Central North Sea (CNS) (e.g. Danish sector, Viking Graben, Moray Firth), and extension is therefore commonly assumed to have taken place in the CNS as well. Stewart & Clark (1999) suggest a pulse of mild mid-Triassic basement extension. This extension triggered thin-skinned extension on the CNS platforms and generated accommodation space in the Central Graben.

In the Forth Approaches Basin, downlapping geometries within the Triassic pod sequence (Smith Bank Fm) suggest that progradation may have been a driving force for mobilizing the salt. The oldest downlapping sequence is probably of Early Triassic age.

Above marginal and thin deposits of the Zechstein on the West Central shelf and Jæren High, the Triassic section is also thin and relatively unstructured. Where the Zechstein is dominated by salt and at the same time is thick (up to 1000m), the Triassic mini-basins are deep (300-1000m). Salt walls between mini-basins are elongate, suggesting that thin-skinned extension probably initiated the salt redistribution. Differential loading probably drove the mini-basin subsidence until the pods grounded. Over much of the West Central Shelf the mini-basin subsidence took place in Anisian – Norian time. They are post-dated by Late Scythian – Early Rhaetian Skagerrak sandstones.

Accomodation space in the Central Graben may have been enhanced by minor episodes of basement rifting in Early and Middle Jurassic. The Central Graben Middle Jurassic section was probably thinned by erosion during the late Middle Jurassic uplift event (the mid-North Sea High), however the regional tilt was small and apparently did not reactivate salt structures.

The main phase of basement extension took place in Late Jurassic time and accentuated the difference between structural highs (West Central Shelf, Forties-Montrose High, and Jæren High) and lows (Central Graben and Fisher Bank Basin). Depending on initial salt thickness, various salt structures formed during the extension. In areas of thin salt, there was a strong coupling between basement and cover structures (including intervening salt). Diapirism occurred mainly were the initial salt was thick. In such areas, graben extension was often linked via the salt to extension in the overburden over the footwall. Sags of the cover into partly evacuated hangingwall basins often also resulted in extension downdip on the hangingwall slope, with resulting diapirsm. The distribution of Triassic mini-basins with Late Jurassic basement faults sometimes caused coincidence between salt pillows and basement faults. In the Central Graben, continued subsidence into Volgian time is believed to have caused further thin-skinned slip of the cover sequence.

Tertiary inversion occurred in Eocene – Oligocene and locally rejuvenated salt diapirs. However, the salt was less important to the inversion since it had been almost completely redistributed in Late Jurassic time.

2.5 Plate tectonic reconstruction

Permian salt deposits in Europe are located to a region where basically three plates collided during the Palaeozoic. In Late Ordovician times (ca. 450 Ma), Baltica (Scandinavia and Russia west of the Urals) collided with Avalonia (southern England and parts of Europe) along the Thor Suture and with subduction beneath Avalonia. Subsequently, Baltica and Avalonia collided with Laurentia (including North America, Scotland and Greenland) along the Iapetus Suture (subduction beneath Laurentia). Finally, a number of European terranes collided with Avalonia along the Hercynian suture in the Late Carboniferous (formation of the Pangea Supercontinent). During the Late Carboniferous and Early Permian, Central Europe was located at low to equatorial latitudes (humid and wet climate) and widespread coal deposits were at their Phanerozoic maximum at that time. During the Permian, Central Europe drifted northward and by the Late Permian was located within the subtropics. On our Late Permian reconstruction (Fig. 2.7), salt deposits are confined to latitudes of 18-27^{o} N which suggest a strong climatic control (arid climate). Major magmatism is not known for this time, but the salt basins are encircled within three major Palaeozoic suture zones.

3. LATE TRIASSIC- EARLY JURASSIC NEWFOUNDLAND, NOVA SCOTIA, AND US BASINS

Salt basins along the western North and Central Atlantic margins are well developed in the Grand Banks tectonic province off Newfoundland (Balkwill & Legall, 1989; Welsink Srivsatava & Tankard, 1989), and offshore Nova Scotia (Welsink, Dwyer & Knight, 1989). Limited amounts of salt are reported as far south as the Carolina Trough off the US east coast (Dillon et al. 1983).

The Grand Banks tectonic province (Fig. 3.1) includes the Grand Banks, Orphan Knoll, Flemish Cap and SE Newfoundland. The Grand Banks area is limited southward by the Newfoundland Fracture Zone (NFZ) and northward by the Charlie Gibbs Fracture Zone. The NFZ is continuous with the Azores Fracture Zone on the east side of the mid-Atlantic Ridge, and together they form the somewhat arbitrary boundary between the Central and North Atlantic. The Charlie Gibbs Fracture Zone, on the other hand, lies within the North Atlantic realm, forming a boundary between the Labrador Sea/NE Atlantic and the southern North Atlantic. Since the Grand Banks borders against both the Central and North Atlantic, which have different rifting and opening histories, the timing of structuring and basin development is somewhat complicated. Within the Grand Banks province lie several salt basins, such as the Whale Basin, South Whale Basin, Horseshoe Basin, and Jeanne D'Arc Basin (Fig. 3.2). These basins are typically characterized by half graben rift geometries, or by grabens and horsts. The salt was deposited in intra-rift basins separated by footwall crests or horst blocks. Hence, the salt in this region was not a wide blanked deposit, but was deposited laterally discontinuous due to the intervening positive structures during salt deposition. While the salt in the basins is of the same age, the age of salt diapirism varies in accordance with the timing of basement structuring.

The Nova Scotia margin is the northernmost part of the western Central Atlantic proper. Compared with the Grand Banks province, the Nova Scotia shelf is narrow (Fig. 3.3). Much alike the conjugate Morocco margin, the Late Triassic – Early Jurassic extensional basins along the entire eastern US and Canadian margin are NE-trending half grabens that utilized the crustal fabric inherited from the Palaeozoic Alleghanian – Variscan Orogenies.

3.1 Rifting and breakup history

Rifting along the US and maritime Canada margin occurred between Middle Triassic and Early Jurassic time (e.g. Schlische, 1993). Along the northern part of this margin, synrift strata are of the above age, whereas synrift strata along the southeastern US portion of the margin are exclusively Late Triassic in age (Withjack et al., 1998). Schlische (1993) reported that the northern basins (from Fundy Basin to Culpeper Basin) do contain Lower Jurassic lacustrine rocks intercalated with basalt flows, while basins to the south lack those strata. In addition, the southern basins are cut by NW-trending margin-perpendicular dyke swarms. Although being of the same age as the margin-parallel dykes further north, the coast-perpendicular dykes are interpreted to post-date breakup at this location. Such an interpretation hinges on a diachronous margin development.

A synthesis of the US and maritime Canada margin development suggests that the transition between rifting and drifting (onset of seafloor spreading) is time-transgressive in the Central Atlantic, older in the south and younger in the north (Withjack et al, 1998). The transition along the southeastern US margin occurred after deposition of Upper Triassic synrift strata and before extrusion of postrift basalts and post-breakup dyke swarms. Along the NE US and maritime Canada margin, the transition to breakup is placed in time between Lower Jurassic synrift strata and late Early Jurassic to early Middle Jurassic postrift strata. Late Triassic (Carnian) also marks the onset of rifting in the Grand Banks area off Newfoundland (Balkwill & Legall, 1989).

According to Withjack et al. (1998) the southern Central Atlantic started to open shortly after the well-dated NE American magmatic activity (at c. 200 Ma), while the northern Central Atlantic is estimated to have opened near 185 Ma. The Grand Banks, situated against the southern North Atlantic opened progressively from Hauterivian to Albian time (Balkwill & Legall, 1989 and references therein). Thus, the well-known general pattern of northward propagating seafloor spreading in the North Atlantic can be argued to have begun in the southern Central Atlantic.

This contrasts with the work by e.g. Klitgord & Schouten (1986) who interpreted more or less synchronous onset of seafloor spreading in the Central Atlantic, at c. 175 Ma. The so-called East Coast Magnetic Anomaly (ECMA) along the US/Canadian margin (Fig. 3.4) relates closely to the seaward dipping reflector sequence (SDRS) (Holbrook & Keleman, 1993). By analogy with the NE Atlantic, the lower part of the SDRS is commonly assumed to lie near the continent-ocean boundary (COB). If the ECMA does lie near the COB, Central Atlantic plate reconstructions (Klitgord & Shouten, 1986) suggest breakup near c.180 Ma.

3.2 Magmatic history

Major dyke swarms developed in the continents surrounding the Central Atlantic before breakup. A single, short-lived magmatic event of only a few million years, focussed in time at approximately 200 Ma, appears to have caused dike swarms, sill complexes and continental flood basalts along the Central Atlantic margins. According to McHone (1996, 2000), the most authorative age for flow basalts and dykes along eastern N American margin is confined to between 202 and 200 Ma.

The presence of seaward-dipping reflectors, and anomalously high-velocity lower crust (7.2-7.5 km/s) permits defining the Central Atlantic as a volcanic passive margin. The high-velocity lower crust is typically thought to represent underplated material added during the final phase of rifting and early breakup. Volumes estimated for the added underplated material are 3.2×10^6 km^3 (Holbrook & Keleman, 1993). In addition, an area in the order of 500 000 km^2 is estimated to have been flooded by continental flood basalts (McHone, 1996). Considering that these estimates only involve one side of the Central Atlantic, and that the African side was considerably more affected by dyke swarms, there is little doubt that the Central Atlantic margins were at least as volcanic as those along the NE Atlantic (cf. White, 1987, 1988). The view presented here contrasts with White & McKenzie's (1989) conclusion that the Central Atlantic was far less magmatic than the NE Atlantic.

Interestingly, dyke swarms along the southeastern US margin (approximately between Virginia and Georgia) trend NW, i.e. almost margin perpendicular (Fig. 3.5). Wilson (1997) proposed that this pattern was part of a radial pattern, centred on a plume located beneath the NE African craton near Mali. In contrast, Hill (1991) suggested a plume located near Florida while White & McKenzie (1989) proposed that the responsible plume was the Cap Verde hotspot, which they at 180 Ma located c. 1000 km NW of the NE US present coast. Thus, these workers suggest plume positions as far apart as c. 3000 km. Interestingly, several authors (e.g. White & McKenzie, 1989, Sebai et al., 1991, Keleman & Holbrook, 1995) point out that this magmatic event does not appear to have been associated with significantly elevated temperatures.

Withjack et al. (1998) correlated the approximately coast-perpendicular dykes in southeastern US to reflect a post-breakup compressional regime related to incipient ridge push. Reverse faulting and inversion of the former rift architecture is also described as part of the same event. Since these authors interpret diachronous opening of the Central Atlantic, the northern part of the margin could still have been undergoing extension while the southern part of the margin was subject to compression. In such a scenario the different stress patterns may explain the different dyke patterns, forming different trending swarms during the same magmatic event. Hence, the rift-parallel NE dyke trend along the northern US and maritime Canada margin may conceivably be explained as a consequence of diachronous opening.

3.3 Salt deposition

Onset of rifting in the Grand Banks area is marked by a nonmarine synrift sequence dating back to very Early Carnian time (c. 228 Ma). Overlying this is a very thick succession of partly marine red bed clastics and halite, deposited in very Early Rhaetian to intra-Pliensbachian time (c. 200-187 Ma), followed by a sequence of intra Plienbachian to Bajocian/Early Bathonian shallow marine limestones (Balkwill & Legall, 1989). A Bajonian/Early Bathonian to Late Kimmeridgian succession overlying the shallow marine limestones, is characterized by coarsening-upward marine clastics and oolitic limestones. The latter sequence is interpreted to mark breakup between North America (near Nova Scotia) and Africa.

Extension-induced rotation of the strata continued into Early Cretaceous time. The entire synrift sucession is confined to grabens and half grabens and displays wedge-shaped expansion against the main bounding faults. A prominent angular unconformity truncates the rotated and salt-pierced Triassic and Jurassic sequence. The oldest rocks above this unconformity are of Late Barremian or Early Aptian age (c. 115 –113 Ma). However, the hiatus at the unconformity spans back to Tithonian time. Balwill & Legall (1989) interpreted the unconformity to be a compound surface related to the gradual propagation of rifting and spreading, progressing around the Grand Banks from the Central Atlantic to the North Atlantic.

In the Whale Basin, a postrift sequence of remarkable even thickness overlies the mentioned unconformity (Fig. 3.6). Locally, minor reactivation is seen along the main halfgraben bounding faults, but in general the area was structurally quiescent. In contrast, the Late Triassic to Early Jurassic synrift sequence on the Nova Scotia margin is overlain by a very thick postrift succession (expanding basinward to as much as 10 km thickness).

3.4 Salt deformation

Salt diapirs along the NE US and maritime Canada margin tend to be concentrated near the main halfgraben-bounding faults. The reason being that these salt deposits formed as part of the synrift succession, and hence were confined to individual basins, subdivided by intrabasin highs. This contrasts markedly with structuring in the Aptian salt basins of West Africa and eastern South America, where the salt deposits were regionally widespread and acted as a common detachment surface, allowing linkage of post-salt structuring (e.g. Tari et al., 2002).

The Argo salt has pierced the Triassic-Jurassic succession, forming halfgraben-parallel salt ridges that rarely pierce the sub-Aptian unconformity. Rather, the salt is commonly peneplained at the unconformity. The latter may suggest that the mother salt is depleted.

The study by Balkwill & Legall (1989) was one of the first to suggest that basement-involved extension induced the salt diapirism. They proposed that the salt became mobilized up along the major halfgraben-bounding faults. Subsequent work (e.g. Vendeville & Jackson, 1992) has convincingly demonstrated that during regional extension, structural thinning of the overburden reduces the overburden load on the salt, triggering salt mobilization into those areas of reduced load. In essence this is a differential loading phenomenom. It is a very common mechanism for driving salt diapirism.

The fact that vigorous salt diapirism ended when large-scale regional extension terminated in Aptian time is strong evidence that extension drove the salt diapirism. The post-Aptian succession is as much as 1 km thick, but because it was laid down in a tectonic quiescent environment, its thickness was even, causing no differential load.

4. <u>LATE TRIASSIC TO EARLIEST JURASSIC MOROCCO & SENEGAL/GUINEA BISSAU BASINS</u>

Along the NW African margin salt basins are known in two areas, 1) off Morocco (Fig. 4.1), and 2) off Gambia, Senegal, and Guinea Bissau (GSGB) (Fig. 4.2). These salt basins lay conjugate to the northeast US/maritime Canada basins and to southeast US margin respectively. Salt is comparatively limited in distribution and volume off GSGB, much like it is off southeast US in the Caroline Trough area. The age of the salt is considered to be similar along the entire NW African margin (Tari et al, 2003), although the salt age does not appear to have been proven off GSGB or in the Carolina Trough area. The Moroccan salt deposits, like those on the conjugate North American margin, were deposited during Late Triassic to Early Jurassic rifting (synrift) in relatively small grabens and halfgrabens. In contrast to the North American counterparts, the Morocco salt basin does show evidence of linked salt structuring, with an updip extensional zone and a downdip compressional zone. It is not clear if this implies more connected original salt deposits, or later redistribution and/or the influence of a steeper continental margin.

4.1 Rifting and breakup history

During Late Triassic time westward-facing halfgrabens, located west of the Hercynian front, formed as part of Central Atlantic rifting. The common rift depression allowed marine water to enter the rifted area from the Tethys, resulting in hypersaline lagoonal conditions over a large part of the Morocco coast, and in turn in deposition of thick salt deposits (Broughton & Trepaniér, 1993, Morabet et al, 1998). Further extension took place during the latest Triassic – Lower Jurassic, before opening of the Central Atlantic.

When breakup occurred in Early Jurassic time, a major transgression flooded the region, forming a carbonate platform in shallow water (Jarvis et al., 1999). Jurassic rifting associated with the Tethyan extension occurred in the Atlas area (trending NE and thus discordant to the N-S Central Atlantic rift.

In Early Cretaceous time, major easterly- and southerly-derived deltaic deposits drowned the carbonate platform. In middle Cretaceous time (Albian) the shelf was again transgressed, and mudstones filled the deepwater basin topography. These sediments were sourced from the Atlas rift located to the NE. The top of the Cretaceous is bound by a major erosional unconformity, related to uplift of the African hinterland. A second Tertiary erosional unconformity is of mid-Oligocene age and marks the time the Atlas Gulf was inverted to form the Atlas Mountains.

4.2 Magmatic history

Major dyke swarms developed in the continents surrounding the Central Atlantic before breakup (Fig. 4.3). In Morocco such dyke swarms are reported in the Anti-Atlas region, where they are discordant to Hercynian domal uplifts of the Anti-Atlas. The Morocco dykes trend parallel to the proto-Atlantic rift. Associated with the dykes are sills and basalt flows within Triassic red beds. The latter are judged to be similar to the Newark Series of the Appalachian Orogen (May, 1971). Ar/Ar age dating of two basalt flows in Morocco gave plateau ages of c. 200 Ma (201.3 $\pm$ 0.7 Ma and 197.1 $\pm$ 1.8 Ma, respectively) (Sebai et al., 1991). These ages are consistent with dated dykes in Mali and Iberia, (Sebai et al., 1991) as well as along the Eastern North American margin (McHone, 1996, 2000). A single, shortlived magmatic event of only a few Ma, at approximately 200 Ma, appears to have caused dike swarms, sill complexes and continental flood basalts along the Central Atlantic margins. The affected area is c. 5000 km in length and 1000 km in width. Geochemical data indicate that at least the tholeiitic rocks from NW Africa and Iberia represent a remarkably uniform magmatic province. A model of passive rifting, above a mantle with slightly elevated temperature, was proposed by Sebai et al. (1991).

4.3 Salt deposition

Two salt sequences are known from the onshore Essaouira Basin. The oldest salt deposit, being of Rhaetian age (Late Triassic), formed in half grabens that also had accumulated thick red bed sequences. The younger halite deposit is of Lower Liassic age (earliest Jurassic) and records an upward transition from red bed clastics to anhydrite dolomites; to the west the halite grades into massive anhydrites. By middle Liassic time the Essaouira Basin was a carbonate shelf, with intermittent evaporite development (Broughton & Trépanier, 1993).

Comparatively little is known about the age of the salt off the Gambia-Senegal-Guinea Bissau (GSGB) coast, but these salt basins are much smaller and less developed than the Moroccoan counterparts (Tari et al, 2003). The GSGB salt is located basinward of a basement hinge zone, in a setting broadly analogous to the rather small salt deposits located outside the US margin hinge zone in the Caroline Trough area (Dillon et al, 1983).

The age of the GSGB salt is uncertain, but based on structural reconstructions (Fig. 4.4), Tari et al. (2003) interpret the salt to be of similar age as the Morocco salt, i.e. Upper Triassic to Lower Jurassic. Structuring of the salt off GSGB is interpreted to have coincided with Santonian uplift and rapid erosion of the region east of the hinge zone, and associated outpouring of sediments. The prograding sediments presumably loaded the post-salt section and induced folding (Fig. 4.4)

4.4 Salt deformation

Extensionally driven salt movement started already in mid-Liassic (c. at 200 Ma) with the development of grabens and horsts running parallel to the Palaeozoic grain. The timing corresponds closely with the dyke-swarm development seen along both margins of the Central Atlantic (e.g. Wilson, 1997), which most likely signify pre-breakup extension.

In the Essaouira Basin, extension along N-trending major faults produced a syncline (the Neknafa Syncline), which was active throughout most of the Jurassic. Within the syncline evaporates formed (anhydrite) while a carbonate bank was anchored to the bounding high. The syncline widened by salt withdrawal from the underlying Theatian – Liassic salt, probably as a result of redistribution into the syncline flanks where extension was occurring. There, thick salt deposits formed salt rollers although no discrete diapirs. This deformation was active during much of the Upper Jurassic (Hafid, 2000).

While the regional extension diminished significantly in Cretaceous time, some extension is reported, localized to salt structures. Differential loading into the Neknafa Syncline caused salt to be further mobilized into the bounding extensional flanks. Overall, the Cretaceous represented a period of burial with only minor salt deformation onshore.

Close to the Morocco coast lies the Cap Tafelney fold belt, which is regarded to be an offshore continuation of the Western Atlas range. The fold belt, which intersects the Morocco passive margin obliquely, started forming in Cenomanian time in response to the onset of the Atlas Orogeny, and was continuously deformed throughout the Atlas Orogeny (lasting into Early Tertiary). The fold belt soles out on a Triassic salt detachment (Hafid et al., 2000).

The generalized style of salt deformation offshore is characterized by rafting (large-scale gravity gliding) of the shelf and upper slope (Fig. 4.5). Rafting started in Middle Cretaceous time (Tari, et al., 2003), probably in response to general margin subsidence, possibly assisted by loading of the Cretaceous sediments building out from the shelf (so-called gravity spreading). This extensional system is linked via a salt detachment to a compressional regime on the lower slope (Fig. 4.6). This linked system is still active, as the compressional structures are expressed at seabed. Further south along the Morocco margin, the postrift succession is considerably thicker. Also in the Ras Tafelney region, Tari et al. (2003) interpret updip extension (rafting) linked down to compression. In the compressional domain salt canopies, tounges and toe thrusts are interpreted. The latter formed at the outer salt basin limit.

The main phase of inversion during the Atlas Orogeny was post-Eocene in age. Compression caused squeezing of the pre-existing salt walls, resulting in forceful diapirism.

4.5 Central Atlantic plate tectonic reconstruction (200 Ma)

The geological evolution in the Central Atlantic commenced with the suturing of East Newfoundland and Acadia-New England to the North American craton during the Caledonian orogeny (c.420 Ma). Subduction-polarity was mainly below North America. During the Carboniferous (330 Ma and onwards), Iberia, Africa, South America and the Florida Peninsula collided with North America and Central Europe. One Hercynian suture ran down the Central Atlantic (west of Iberia in the north) whilst a second suture ran south of Iberia and into the Tethys. During the Late Triassic and Early Jurassic (Fig. 4.7), the Florida Peninsula and the west coast of Africa was situated at low latitudes. Hence, the Carolina and Senegal salt appears to have been deposited at ‘anomalously’ low latitudes (3-11^{o}N) unless the age of the salt is wrong. Conversely, salt deposits from the Grand Banks and Morocco are confined to sub-tropical latitudes of 18-29^{o}N. The latter two regions belonged to a contiguous rift system before break-up of the Central Atlantic during the mid-Jurassic (c. 175 Ma).

During the Early Jurassic there was considerable volcanism (± 200 Ma) in the Central Atlantic (Central Atlantic Magmatic Province; CAMP). Dykes can be observed as far north as Iberia, and associated flood basalt and sills are observed throughout the Central Atlantic. CAMP magmatism is broadly contemporaneous with salt deposits.

5. APTIAN WEST AFRICA AND SOUTH AMERICA BASINS

5.1 West Africa-South America rifting and breakup history

A very generalized sequence of South America-Africa rifting and breakup history is as follows. Extension between Africa and South America started in the south (at c. 200 Ma) and proceeded north, affecting the Niger region at c. 120 Ma. By c. 112 Ma, sea floor spreading was established throughout much of the South Atlantic (West Africa-South America) margin, although transform motion dominated to the NW of Niger.

Szatmari (2000) suggested the Ponta Grossa dike swarm (Fig. 5.1) c. 130 Ma NW-trending, 200 km wide zone exposed just east of the Paraná basin in southeastern Brasil formed a topographic barrier between a deep, pre-salt (Aptian) depression to the north (see Conceicão et al., 1988) and an oceanic basin to the south characterized by open water. This is similar to the idea of Burke and Sengor (1988), who proposed that pre-Aptian subaerial seafloor spreading occurred north of the proto-Walvis Ridge. In their model, a deep, dry basin formed and later filled (possibly several times?) by Aptian oceanic flooding. Burke and Sengor's model assumed that Africa and South America behaved as two large rigid plates, hence also assuming synchronous opening for the entire area encompassing the Aptian salt basin. They further assumed that the spreading axes formed 3 km below sea level. Thus, their model predicted that the area to become the Aptian salt basin was continuously dammed, subsided rapidly, and experienced subaerial magmatic spreading. Influx of oceanic water over the topographic barrier represented by the proto-Walvis Ridge was thus catastrophic. Burke (pers. com. 2003) relates this event to the marked sea level drop seen on the Haq et al. global sea level curve.

Alternatively, South America and Africa may be divided into smaller sub-plates along sutures defined by old orogenic belts, zones buried under subsequent deposits such as the Etendeka-Paraná continental flood basalts, intersection points between continental crust and oceanic fracture zones, etc. In this model, a very tight Africa-South America fit may be maintained. The latter model is preferred here, and is the basis for our plate reconstruction. The model is not simply based on plate geometries, but is also constrained by ages of rift basins along the margins (e.g. Jackson et al., 2000).

Regardless of model, almost 40 m.y. (from earliest Valanginian to Late Albian time) was required for Africa and South America to completely separate. In both models, the opening of the South Atlantic occurred from south to north during the Early Cretaceous; the Cape and Argentine Basins opened first, followed later by the Brasil and Angola Basins.

To the south of the future Walvis Ridge, global oceanic circulation patterns were unrestricted. The Cape (African) and Argentine basins were characterized by oxygenated terrigenous sediments, clays, and black shales. Under either plate tectonic scenario, the Ponta Grossa dike swarm (and later the Rio Grande Rise - Walvis Ridge) may have created a topographic barrier that restricted (or possibly completely eliminated) oceanic circulation between the Aptian salt basin region and the ocean to the south (e.g. Szatmari, 2000). The Brasilian and Angolan basins accumulated thick evaporite sequences during the Early Cretaceous. Isolation of the northern basins continued until the Late Cretaceous, though sedimentation shifted from evaporites to black shales as the basins expanded and deepened. Eroding landmasses on either side of the expanding South Atlantic delivered terrigenous sediments into all of these basins throughout most of the Cretaceous.

Permanent connection between the North, Central, and South Atlantic commenced about 90 Ma, establishing open ocean conditions throughout the Atlantic. Nevertheless, the deep sea topography created by the mid-Atlantic Ridge, the Rio Grande Rise and the Walvis Ridge continued to promote subtle inter-basin differences in the Carbonate Compensation Depth (CCD) and deep-sea sedimentation, a circulation effect that continues even in modern times.

5.2 Magmatic history of South Atlantic margins

Mantle plumes have often been suggested to have played an important role in the initial breakup of Gondwana and the opening of the South Atlantic. Bizzi (1993) suggested that compositional and isotopic characteristics of basaltic volcanism, that occurred shortly before opening of the new ocean, are explained satisfactorily by asthenospheric plume models. While cause-effect relationships between Large Igneous Provinces (LIPSs) and deep mantle plumes are not fully understood, a clear observation is that many LIPs erupted just prior to and during the fragmentation of Pangaea.

By the Early Cretaceous time, the syn-rift (c.132 Ma) Paraná-Etendeka continental flood basalts (Fig. 5.1) were being erupted, probably synchronously with local rifting and sea floor spreading further south. Other volcanic rocks abound as well: the offshore of nearly the entire South Atlantic margin is characterized by seaward dipping reflectors (SDRs). Some SDRs underlie salt deposits: for example, up to 600 m of volcanic flood basalt rocks along the synthetic (presumed) trail of the (134-122 Ma) Tristan plume (and similar in geochemistry to the late syn-rift Paraná volcanics) have been drilled below Aptian salt in the Campos and Santos basins.

Three episodes of magmatic activity are observed in and near the South Atlantic salt basins: 1) a Late Jurassic – Early Cretaceous event perhaps related to the Paraná-Etendeka flood basalts; 2) thick wedges of volcanic rocks identified as seaward dipping reflector. Commonly, the SDR wedges are locally associated with volcanic intrusions. For example, in the Sergipe-Alagoas region (northeastern Brasil), several volcanic plugs in deep-water region are aligned with oceanic fracture zones that apparently penetrate through the whole crust and reach the upper mantle. (Mohriak et al., 1998); 3) Late Cretaceous – Tertiary events related to hot spot and leaky fracture zone activity (see Mohriak et al., 2002). A poorly dated series of magmatic events occurred in the Niger delta region of Africa (the Cameroon Line volcanics).

5.3 General West Africa-South America evaporite deposition

South Atlantic evaporite deposits of any significant size appear to be confined between the Walvis Ridge/Rio Grande Rise and the Ascension Fracture Zone (Fig. 5.1), between the easternmost tip of Brasil and southern Brasil, and on the African margin between Angola and the Niger Delta. The fully marine Albian carbonate systems of both West Africa (Congo and Kwanza Basins offshore Angola) and Brasil (Santos Basin) overly the Early Aptian salt.

5.3.1 <u>South America Salt deposition</u>

On the Brasilian margin, the South Atlantic salt basins are about c.400 km wide in the Santos Basin, while they taper to c. 100 km in width in the Sergipe/Alagoas Basins. The time of salt deposition is bracketed by Aptian ages of two marine faunas located above and below the salt (Szatmari, 2000).

Along the eastern Brasilian continental margin the salt-bearing Aptian Stage has been subdivided into two sequences. The lower sequence is characterized by continental sediments within a rift setting and only contains signs of sporadic marine incursions. This is related to restricted water circulation conditions at this time in the southern portion of the South Atlantic Ocean (Malvinas area), precluding significant northward movement of water to the Brasilian coast. The Upper Aptian Stage sequence, which was deposited in a sag basin under more quiet and stable tectonic conditions, can be subdivided into two portions with distinct lithologic characteristics. The lower part is dominated by fluvio-alluvial sediments covering most of the proximal areas of the marginal basins. Distal sediments of this unit were, however, deposited under shallow marine conditions. This first ingression is related to an increase of marine water circulation in Southern South Atlantic (Malvinas area). At the end of the Aptian a strong period of aridity associated with a major barrier (possibly created by the Sao Paulo Volcanic Ridge, the Florianopolis High, or the Ponto Grosso dike swarms) led to the deposition of a marine evaporites ("Ibura Event"; Dias-Jefferson, 1999).

Significant thicknesses of Aptian evaporites are known to exist off the Brasilian margin. The thickest portions of the Aptian evaporite sequences occur in the Santos basin. There, the pre-Aptian sequence is largely composed of volcanics (basalt flows and tuffs) that are slightly younger than the hot spot affiliated Paraná Basin continental flood basalts (Szatmari, 2000). The easternmost limit of salt appears to be the submarine São Pãolo Plateau.

The Santos basin salt is thick in the north and thin or absent in the south, roughly following the depositional basin relief. Initial differences in the thickness of the plastic salt sequence apparently resulted in a permanent difference in the mobility of the two areas: in the extreme south of the basin, the sedimentary units deposited after the salt remained stable, whereas, farther to the north, these units glided and spread gravitationally by salt flow from early Albian time onwards (Demercian, 1996). This is broadly similar to the difference in timing between salt structuring in the Kwanza Basin, Angola, and the area of the Lower Congo fan, Angola. Throughout the Late Cretaceous, high sedimentation rates led to massive basinward expulsion of the Aptian salt resulting in a migrating salt wall (Tudoran et al., 2000).

Significant Aptian evaporites are known to exist in the Campos basin, also directly above thick sequences of basalt flows and tuffs (Szatmari, 2000). In the Campos and Santos basins, where total extension was high, the pre-Aptian rift sequence consists mostly of volcanics.

To the north, both the rift and the salt basins taper off near Recife. In the Sergipe-Alagoas basin, the Reconcavo basin (below) and Potiguar basin (below), total extension is moderate.

5.3.2 West Africa Salt deposition

The Kwanza Basin, Angola, is divided into the Inner and Outer Kwanza salt basins (Fig. 5.2 Tari), separated by a chain of basement highs on which salt is thin or absent. Some workers argue that the Aptian salt of Angola was deposited in the post-rift succession (e.g. Tari et al., 2003), while other maintain that the salt was part of the synrift sequence (Karner et al., 2003). Both Cretaceous and Tertiary sedimentation occurred under conditions of continuous salt mobilization and redistribution; in some areas the withdrawal was sufficiently dramatic to allow Miocene sediments to touch down on residual Albian or Aptian carbonates and evaporites, forming so-called salt welds (e.g. Duval et al., 1992).

Aptian salt of the Congo was also deposited in the syn-rift succession. Due to redistribution of salt downslope (mobilization) salt is most abundant in the contractional belt, where it occurs as thick autochthonous salt, salt walls, diapirs, salt tongues, and salt canopies (Fig 5.2).

Aptian salt of Gabon was also deposited in the syn-rift succession. One of the striking features is the anomalously high subsidence rate recorded on the South Gabon platform in the early post-rift stage. Approximately 3 km of sediments were deposited during the first 20 m.y. of rifting, followed by 2 km of post-rift sediments deposited in the next 20 m.y. (c.110-90 Ma. In the Northern Gabon Basin and extending into southern Rio Muni, the rift section comprises lacustrine and fluvio-deltaic faulted and tilted strata of Barremian and Neocomian age. Overlying the syn-rift section is a thick section of Upper Aptian salt and a well-developed succession of Mid to Upper Cretaceous and Lower Tertiary marine limestone and sand-shale sequences.

In northern Rio Muni, the syn-rift section comprises Upper Barremian to Mid Aptian terrestrial clastics and lacustrine shales characterised by extensional rollover structures to mega-scale listric faults updip, and toe-thrust structures downdip. The syn-rift section is overlain by a "transitional" sequence of well-developed salt and good quality marine oil-prone source rock intervals. An Albian (Madiela) carbonate platform developed over the area plus a Cenomanian-Turonian sand-shale sequence, which contains a major source interval. This salt sequence commonly forms extensional rafts detaching on an Aptian salt. A well developed Senonian section onlaps the earlier rafted topography.

5.4 South America salt deformation

Structures related to salt tectonics are well developed in the Campos and Santos basins. During Albian to Campanian time, continuous welding and apparent downlap of sedimentary sequences occurred. The load of prograding sediments began squeezing salt seaward (Fig. 5.3). During Maastrichtian to Paleocene time, discontinuous welding and passive diapirism was initiated. Continuity of the basal salt weld is broken by low-relief diapirs representing remnant pockets of salt trapped after they became rapidly over-ridden by prograding sequence. Between Middle Miocene time and the Present, sedimentation covered distal diapirs and prevented further growth. However, salt remained free to flow along the strike of the salt walls.

5.5 West Africa salt deformation

The West African salt deformation, exemplified by a section across the Kwanza Basin, Angola (Fig. 5.4) is a gravitationally driven linked extensional-compressional system that detaches on Aptian salt, with up-dip extension compensated by down-dip contraction. The African salt deposits have been thoroughly mobilized. Along parts of the West African margin, the gravity gliding started already in Albian time, as a response to subsidence. Areas located inboard of an intrabasin hinge zone, such as the inner Kwanza Basin, were relatively stable until the Middle to Late Tertiary when a second stage of gravity-driven extension occurred in response to Tertiary uplift of western Africa (e.g. Lundin, 1992). This was coupled with extensive Middle and Late Tertiary loading by depositional systems associated with the ancestral Congo and other river systems.

The influence of basement structuring on salt deformation is well expressed in the Kwanza Basin, where northeast-trending structures segment the basin (Fig. 5. 2). These structures probably acted as transfer faults during rifting, and have been repeatedly reactivated since that time. Reactivation has produced three fold-and-thrust belts near basement uplifts, and has also controlled the evolution of salt structures. Salt structural styles change dramatically across the transfer faults.

5.6 Plate tectonic reconstructions of the South Atlantic

South America and Africa are made up of several different cratons that collided with each other at around 550 Ma (forming Gondwana), an event that marked the most spectacular mountain-belt building episode in Earth history. No mountain area as extensive as the Gondwana mountains (c. 20 million km^2) have been constructed on Earth before or since. The South American and African margins therefore represent old sutures that ultimately ruptured during the Cretaceous.

The South Atlantic salt basins are of Aptian age and the salt has received notable attention due to its involvement in hydrocarbon trap development. The Aptian salt substrate (rift basins versus subaerial basalt or oceanic crust), timing of deposition (syn- or post-breakup), mode of formation (seawater versus hydrothermal brines) and their role in deformation of post-rift sediments have gained considerable attention.

By the Early Cretaceous time (c. 130 Ma) seafloor spreading was initiated in the southernmost South Atlantic and this event closely coincides with the Parana-Etendeka (Fig. 5.1) magmatic event. Parana and Etendeka were later broken in two during ensuing opening of the South Atlantic but linked via the Rio Grande Rise and the Walvis magmatic ridges. By Early Aptian time (c. 121 Ma) seafloor spreading had propagated northward to the Rio Grande Rise-Walvis Ridge. The age of the South Atlantic seaward-dipping reflectors is not well known. We also indicate that the Santos and Campos Basins were partly covered by subaerial basalts. Salt accumulated during Aptian time and we show the deposits on a Late Aptian (c. 112 Ma) reconstruction, implying that the salt predates breakup (Fig. 5.5). The Aptian salt basins probably developed as a single basin during the late phase of rifting. The once contiguous salt basin is now broken into two basins - they partly overly subaerial basalt and could also be in part synchronous with magmatism since the volcanic trail of the Tristan plume is recognised just south of the Santos Basin. The main salt basins accumulated between 27^o and 10^o S, i.e. at subtropical to tropical latitudes - seafloor spreading propagated north to Santos-Campos by Early Albian times, and by Mid-Albian times seafloor spreading had propagated north of the Niger segment.

6. NEOGENE RED SEA AND AFAR BASINS

The Red Sea basin is a Tertiary continental rift confined between northeast Africa and the Arabian Peninsula. Exposed areas between the rift escarpments and the shores are very limited and the basin sediments are located essentially completely offshore. Sea floor spreading is ongoing along the axial rift, where basin sediments are restricted. The Red Sea evaporites are restricted to the offshore basin and do not outcrop (Fig. 6.1).

The NW-SE length of the basin is about 2300 km and the width is up to 400 km. The southernmost point, the strait of Bab el Mandeb lies at 12^{o} N and the basin stretches NNW to 30^{o} N in the Gulf of Suez. The main Red Sea basin has experienced active sea-floor spreading over the last 5 Ma, and water depths locally reach more than 2300m. At the northern end lies the Gulf of Suez, an abandoned rift with less than 100m water depths, and the active Gulf of Aqaba transform with a water depth in excess of 1850m.

The mean crustal thicknesses of the central portions of the Nubian and Arabian cratons differ, Arabia being generally some two to four km thicker (Mooney et al., 1989), suggesting the presence of an ancient Nubia-Arabia suture. The Miocene opening of the Red Sea rift system might thus have been partly controlled by pre-existing sutures, as well as by the arrival and continuous presence of the giant Afar plume.

6.1 Rifting and breakup history

Continental rifting between Arabia and Africa (proto-Gulf of Aden) is generally regarded as having propagated toward the Afar triangle. This rift and subsequent breakup predates rifting and seafloor spreading in the Red Sea. Red Sea rifting is estimated to have started about 20-23 Ma ago, based on foraminifera in the oldest sedimentary strata, and dated tholeiitic magmatism along the length of the Red Sea (Bosworth & McClay, 2001). There is a controversy (see Ghebreab, 1998) as to whether Red Sea extension occurred in a single stage since the Oligocene (c. 30 Ma) or in two stages (c. 23.5 to 16 Ma and c. 5 to 4 Ma). However, it is generally accepted that the main Red Sea extensional event began sometime during the Late Oligocene-Early Miocene (c. 28 to 16 Ma; Ghebreab, 1998).

Extension in the Main Ethiopian Rift commenced in the late Oligocene or Early Miocene (c. 26 Ma to 16 Ma) (Woldegabriel et al., 1990). The earliest unequivocal records of sea floor spreading lie in the central Red Sea (Izzeldin, 1987) and are dated to Latest Miocene (c. 5-6 Ma). Older accreted sea floor might exist, but has not been conclusively identified. Oceanic crust of less than 5 Ma age is present mainly along the axial trough, but is also indicated by geophysical data to be underlying portions of the Egyptian and Sudanese offshore, fairly close to the coast.

Break-up began in the south, in the Gulf of Aden, and is generally regarded to have propagated toward the Afar triangle.

The southernmost Red Sea, the Strait of Bab el Mandeb and the adjacent Afar triangle, is an anomalous part of the Red Sea rift, being complicated by the Danakil horst, a microcontinent bridging between Africa and Arabia. The Afar area is being flooded by subaerial rift-related basalt. Redfield et al. (2003) presented a detailed Nubia-Arabia-Somalia kinematic reconstruction based on published magnetic sea floor lineations in the Gulf of Aden and Red Sea, and utilized piercing points along the Red Sea margins. The reconstruction and associated paleontological and (limited) deuterium data from Red Sea cores suggested that land connection between Nubia, Somalia and Arabia over Bab el Mandeb was likely achieved between ca. 4.5 and 3.2 Ma.

6.2 Magmatic history

Regional uplift (dome) of more than 1 km took place in the Oligocene (Pik et al., 2003) and was at c. 30 Ma (Hoffman et al., 1997) followed by emplacement of the Ethiopian flood basalts. Both events are commonly associated with an Afar plume. Today the rift margins remain uplifted, particularly near the triple junction of the Red Sea, Gulf of Aden and the Ethiopian/East African rift. The rift shoulders reach more than 3 km above sea level in this area.

The magmatic history of the Red Sea/Afar region is summarized in the following table.

Series	**Age**	**Volcanics**
Pliocene	5 Ma - present	Oceanic crust development
Mid Miocene	10 Ma - present	Rift volcanics. Basalt – rhyolite
Lower Miocene	24-21 Ma	Tholeiitic magmatism. Rift-parallel dykes
Oligocene	30.6- 29 Ma	Flood basalts, 1-1.5 mill km^3

Oligocene continental flood basalt volcanism is largely limited to the Ethiopian and Yemen and Harar/Somalian Plateaus. The total volume of extruded basalt is estimated as c. 1-1.5 million km^3 (Hoffman et al., 1997; Baker et al., 1996). The lava pile is locally more than 2000 m thick and was emplaced in less than 1 Ma (Hoffman et al, 1997). New high-quality geochronological data combined with paleomagnetic data (Hoffmann et al. 1997, Coulié et al. 2003) have shown the onset of basaltic volcanism occurred between 30.6 Ma and 30.2 Ma in Ethiopia and Yemen, respectively. Hoffmann et al. (1997) suggest that this event had a significant effect on global climate and bio-diversity, triggering cooling and aridity, Antarctic ice-sheet advance and sea level drop.

Sebai et al. (1991) reported Ar/Ar age dates that suggest synchronous dyke emplacement on the Arabian Peninsula over the length of the Red Sea between 21 and 24 Ma.

The active Erta Ale axial volcanic chain along the Danakil depression contains lavas ranging in composition from magnesian basalt to rhyolite. Barrat et al. (1998) demonstrated magmatic differentiation of a basaltic precursor to produce the felsic lavas and concluded that Erta Ale originates from two distinct mantle sources, N-MORB and hot-spot type, none of them containing isotopic signature of continental lithosphere.

6.3 Salt deposition

Fig. 6.2 provides a summary of Red Sea basin development. Sedimentary descriptions are given by Beydoun & Sikander (1992) for the entire Red Sea-Gulf of Aden, Bosworth & McClay (2001) for the Gulf of Suez, Mart & Ross (1987) for the northern Red Sea, and Savoyat et al. (1989) for the Ethiopian Red Sea.

Pre-salt sediments, the Gharandal Group (following nomenclature of the Gulf of Suez) are mostly marine, dominated by marls. During rifting the connection to the Mediterranean was mostly open, while there was no connection to the Gulf of Aden. Biostratigraphic dates place the upper synrift boundary near 14 Ma. This is when extension terminated in the Gulf of Suez and movement began along the Aqaba transverse. Connection between the Red Sea and the Mediterranean was then completely or intermittently blocked, mainly as a result of significant sea-level fall.

The evaporitic Ras Malaab Group is split in 3 formations and is dominated by a thick interval of halite and anhydrite, the South Gharib Formation. The underlying Belayim Formation is a mixed unit with significant anhydrite and halite. Overlying South Gharib is the Zeit Formation. (Zeit means oil in Semitic languages.) The Zeit Formation is dominated by clastics and anhydrite. The Ras Malaab Group is interpreted to end with the Miocene/Messinian, at about 5 Ma. During this period the draw-down of the Mediterranean prevented any communication with the Red Sea. The Red Sea basin was closed until the Pliocene (c. 4-5 Ma) when communication between the southernmost Red Sea (the Strait of Bab el Mandeb) and the Gulf of Aden was initiated (Redfield et al., 2003), resulting in a mostly marine Pliocene succession.

Griffin (1999) presents some interesting interpretations on the late Miocene climate based on the Gulf of Suez and Red Sea sedimentary successions. The Tortonian South Gharib Formation records a prolonged dry period, while the Messinian Zeit Formation records a 2 million year monsoonal wet period.

Formation	**Age**	**Sedimentation rate (Gulf of Suez)**	**Climate**
Zeit	7-5 Ma	55.3 cm/ka	Wet
South Gharib	11-7 Ma	12.7 cm/ka	Dry

The main salt/evaporite formation in the Red Sea is the South Gharib Formation (named Amber in Eritrea), but the overlying Zeit formation also contains significant amounts of evaporites. Locally, associated sandstones and carbonate occur. The salt is generally thicker in the central part of the province and in graben centres. Within the Gulf of Suez the South Gharib Formation averages 200 m in thickness, generally thickening southwards toward the Red Sea. Savoyat et al. (1989) showed variable thicknesses from 200m up to more than 3000m in the southern Red Sea (Figs. 6.3 and 6.4), where the thickest successions relate to thickening in diapirs.

The Red Sea salt deposits seem to consist of up to 5 cycles of halite and anhydrite (Bosworth & McClay 2001, Savoyat et al. 1989). The subaerial salt deposit in the Danakil depression in the Afar was drilled in the 1960s and described by Holwerda & Hutchinson (1968). They showed a succession dominated by bedded halite with two subsurface potash layers with sylvite and other potassium-bearing salts. Norsk Hydro is presently holding a mining concession on these deposits. The evaporite thickness estimates vary from about 1000m

(Holwerda & Hutchinson 1968) to more than 2.2 km (Behle et al. 1975), based on seismic reflection data).

The age of the Afar salt is not well constrained. The time of isolation from the Red Sea is generally regarded to be late, less than 5 Ma, and thus Afar salt is potentially younger than the salt deposits in the Red Sea. However, Red Series sandstones and shales were deposited in a developing basin in the Danakil Depression. These marine clays, limestones, and sub-aqueous lavas (Tiercelin et al., 1980) indicate the Danakil depression was in existence by the Early Miocene and the northern part of the graben below sea level for at least part of Miocene time (Tiercelin et al., 1980). The Afar salt may also have originated within the same general time period as the Red Sea salt.

Along the axial trough of the Red Sea a number of depressions occur, filled with highly saline brines. The Atlantis II Deep is the best known of these. It is an active hydrothermal system discharging fluids with temperature and salinity of 155-310^{o} C and 270-370‰ respectively (Anschutz et al. 2000). Halite precipitates together with metal oxides, sulphides, carbonates, sulphates and silicates in the sediments.

6.4 Salt deformation

Locally, salt has been mobilized into massive domes and walls, yielding total removal in other areas. Mart & Ross (1987) describe the distribution of rift faulting and diapirism in the northernmost part of the Red Sea, suggesting a genetic connection (Fig. 6.5). Subsequent to burial under Plio-Pleistocene sediments, diapiric growth may have been triggered by high heat-flow and extension.

Savoyat et al. (1989) described the distribution of diapirs and salt walls in the southern part of the Red Sea, but did not attempt to associate the diapirism directly with extensional structures (Fig. 6.6). Subsequently, Jackson and Vendeville (1994) point to the Red Sea as a very clear example of recent (<5 Ma) salt diapirism triggered by extension in the overburden. The salt basin offshore Eritrea has at the western edge a marked NNW-trending salt wall, which Savoyat et al (1989) interpreted to have resulted from gravity gliding from the coastal basin monocline. Interconnected diapiric salt masses with post-salt depocenters and turtle-back structures characterise the rest of the basin. Progressive formation of diapirs occured during Upper Pliocene time before the deposition of the Pleistocene carbonate shelf formation (Dhunishub Fm.).

In the Danakil depression an interesting salt mound ("volcano") is present at Dallol (Figs. 6.7 and 6.8). Bedded halite is uplifted, tilted and eroded by seasonal rain (Fig. 6.9). In the central part brines with temperatures up to 130^{o}C reach the surface, depositing salt (Fig. 6.10). In places, e.g. "Black Mountain", the hot brine contains mainly $MgCl_2$. Talbot (1978) suggested this to be the surface expression of ongoing diapirism.

The active volcanism along the Danakil depression axis provides a heat source for these hot springs. In the southern part of the depression the Erta Ale volcano is an impressive topographic feature, about 100 km along axis, 40 km across, and reaching nearly 1000 m above the floor of the depression. A long-lasting lava lake occurs in the main summit caldera (Fig. 6.11).

7. PLATE TECTONIC OVERVIEW OF THE SELECTED SALT BASIN DEVELOPMENT

Except for the Senegal/Guinea Bissau and Carolina Trough salt basins of the southern Central Atlantic, the salt basins analyzed in this study formed at subtropical latitudes (Fig. 7.1), conforming to the generally accepted view of evaporite development (Scotese & Barrett, 1990 and references therein). However, the age of the Senegal/Guinea Bissau and Carolina Trough salt is not certain, and the age is only interpreted to be of the similar to that of the salt basins further north along the Central Atlantic margins (Nova Scotia/Newfoundland and Morocco salt basins respectively). It is possible that the anomalous Senegal/Guinea Bissau and Carolina Trough salt is considerably younger, e.g. similar in age to the Louann salt in the Gulf of Mexico (c. 170 Ma). Such a Jurassic age would make them conform with the rest of the basins. Notably, the NW-trending dyke swarms in Liberia, dated to c. 185 Ma, are c. 15 Ma younger than the dominantly N-trending dyke swarms parallel to the Central Atlantic margins. Speculatively, a Jurassic rift system containing salt deposits may have extended from the Liberia region, via Senegal/Guinea Bissau and the Carolina Trough into the Gulf of Mexico.

This study has only been able to document some relationships between significant magmatism and salt deposition (Fig. 7.2). In the abandoned rifts of the SW Barents Sea and the southern North Sea, an association between seafloor spreading and salt deposition is obviously impossible to invoke. While magmatism was profound in the southern North Sea, this occurred c. 50 Ma prior to Zechstein salt deposition. Thus, the associated heat pulse should have dissipated by the time the salt formed. Furthermore, the Zechstein salt formed directly over desert dunes, strongly suggesting that classic evaporation was the process. Likewise, the SW Barents Sea salt formed in a sabkha environment.

Central Atlantic salt (Morocco and Nova Scotia/Newfoundland) does appear to be approximately of the same age as the major dyke swarms that intruded the margins (Fig. 7.2). Salt in these areas developed over continental deposits in constricted intrabasin lows during rifting. Thus, a structural control on intermittent influx of oceanic water appears likely, and classic evaporation is the most plausible cause.

The Aptian South Atlantic salt basin was confined between a major volcanic construction in the south (the proto-Walvis Ridge/Rio Grande Rise) and a shear margin in the north. Due to the poor constraints on the continent-ocean boundaries in the South Atlantic, it is more difficult to assess whether the salt basin formed during the late stage of rifting or conceivably postdates breakup. However, our preferred plate reconstructions show no overlap of the South American and West African salt basin outlines, suggesting that the salt formed in a contiguous basin during the late stage of rifting. In the Red Sea area, major flood basalt magmatism in the Afar triangle occurred c. 20 Ma before the salt deposition, dyke swarms were intruded c. 10 Ma before salt deposition, and onset of seafloor spreading began c. 2-6 Ma after salt deposition (Fig. 7.3). Salt in the Afar region can be explained as a result if intermittent influx of oceanic water from the Gulf of Aden. The Dallol salt mound and the salt brines along the axis of the Red Sea can probably be explained by heating from magmatic activity beneath pre-existing salt deposits.

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9. FIGURES

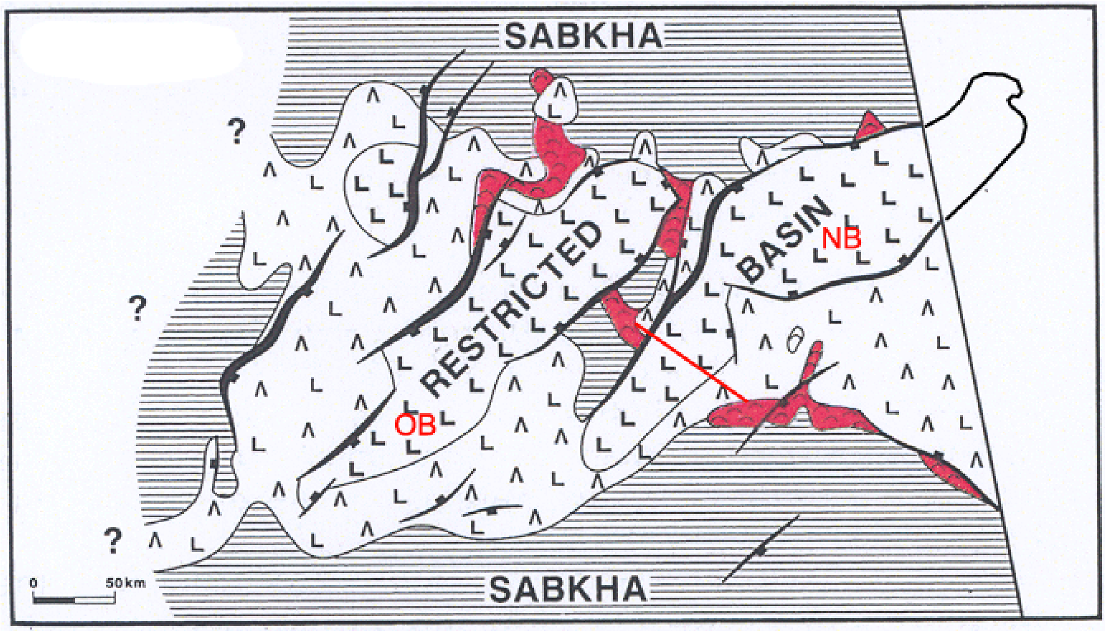


Fig. 1.1 Distribution of the main evaporite facies in the SW Barents Sea. L-symbol = halite, ^ = anhydrite. NB= Nordkapp Basin, OB = Olga Basin. The red line marks the position of a geoseismic profile, shown in Fig. 1.3. Red areas are local carbonate build-ups, generally interpreted as bioherms/reefs. (From Gérard & Buhrig, 1990).

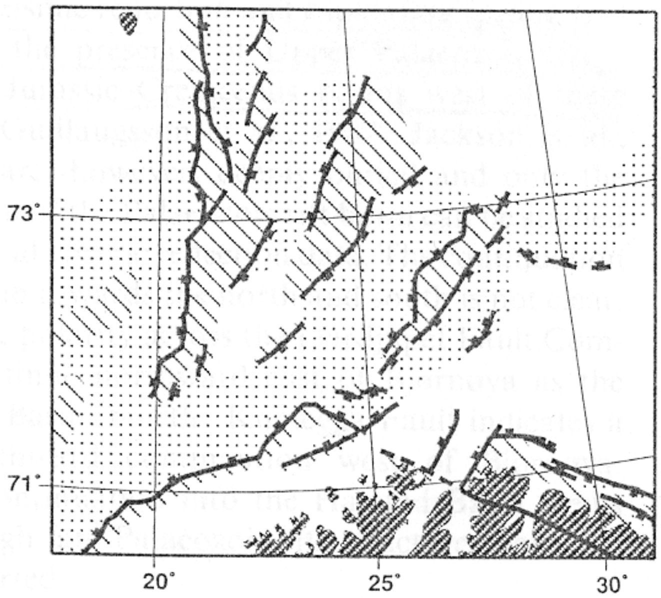


Fig. 1.2. Distribution of basins (dotted pattern) and highs (dashed pattern) of the Carboniferous-Permian rift in the SW Barents Sea (From Gudlaugsson et al., 1998).

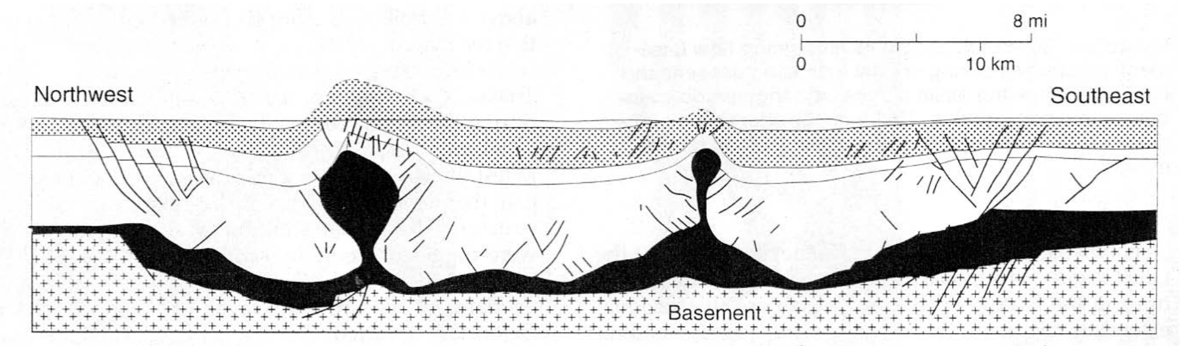


Fig. 1.3 Geoseismic profile across SW Nordkapp subbasin. Note the drop-shaped diapirs. From Nielsen et al (1995).

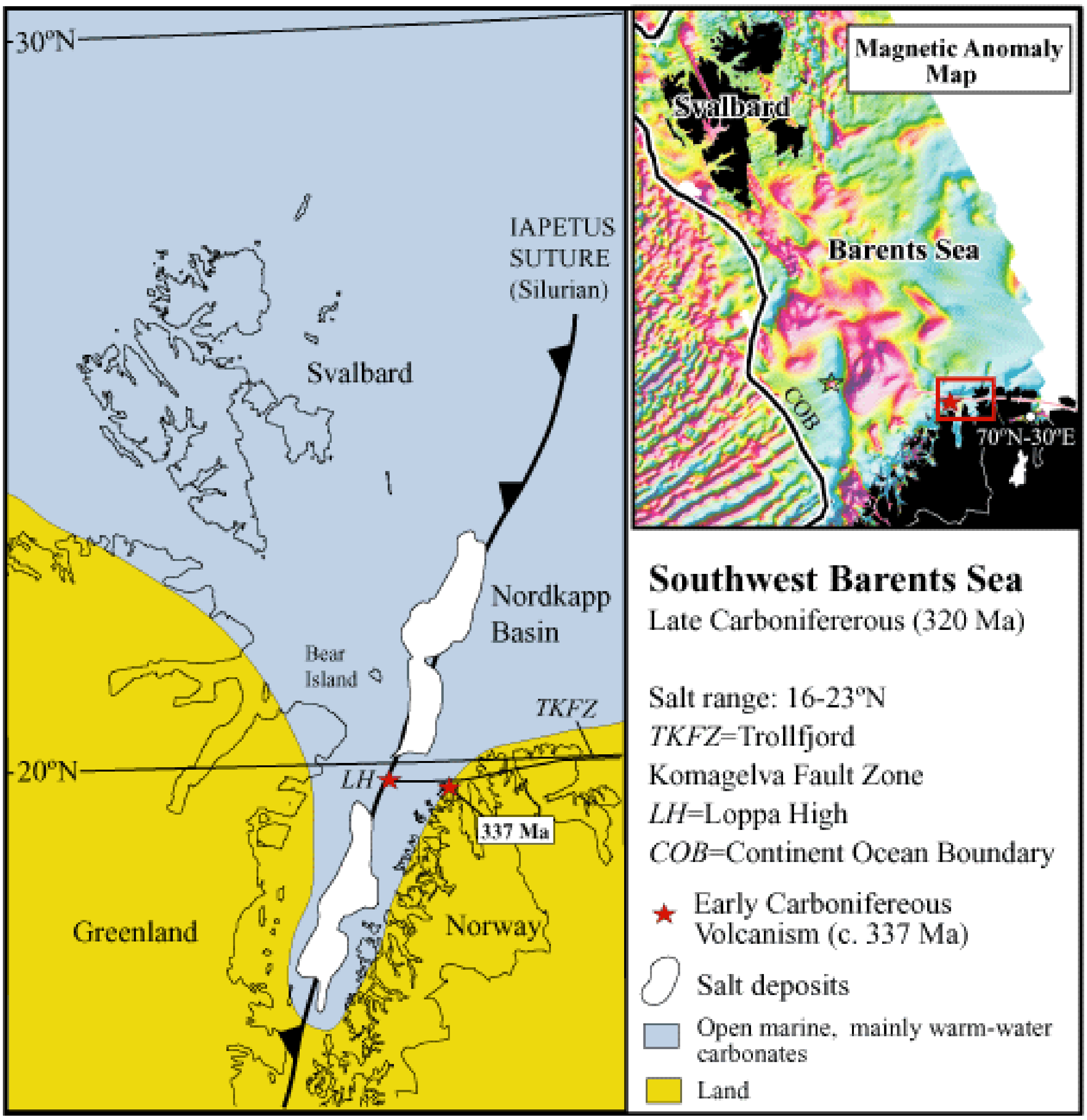


Fig. 1.4 Plate reconstruction to 320 Ma, indicating that the Nordkapp Basin in the SW Barents Sea and the Denmarkshavn Basin offshore NE Greenland probably formed as part of the same rift system. Red stars mark dated dykes along the Finnmark coast, and drilled but undated volcaniclastics on the Loppa High.

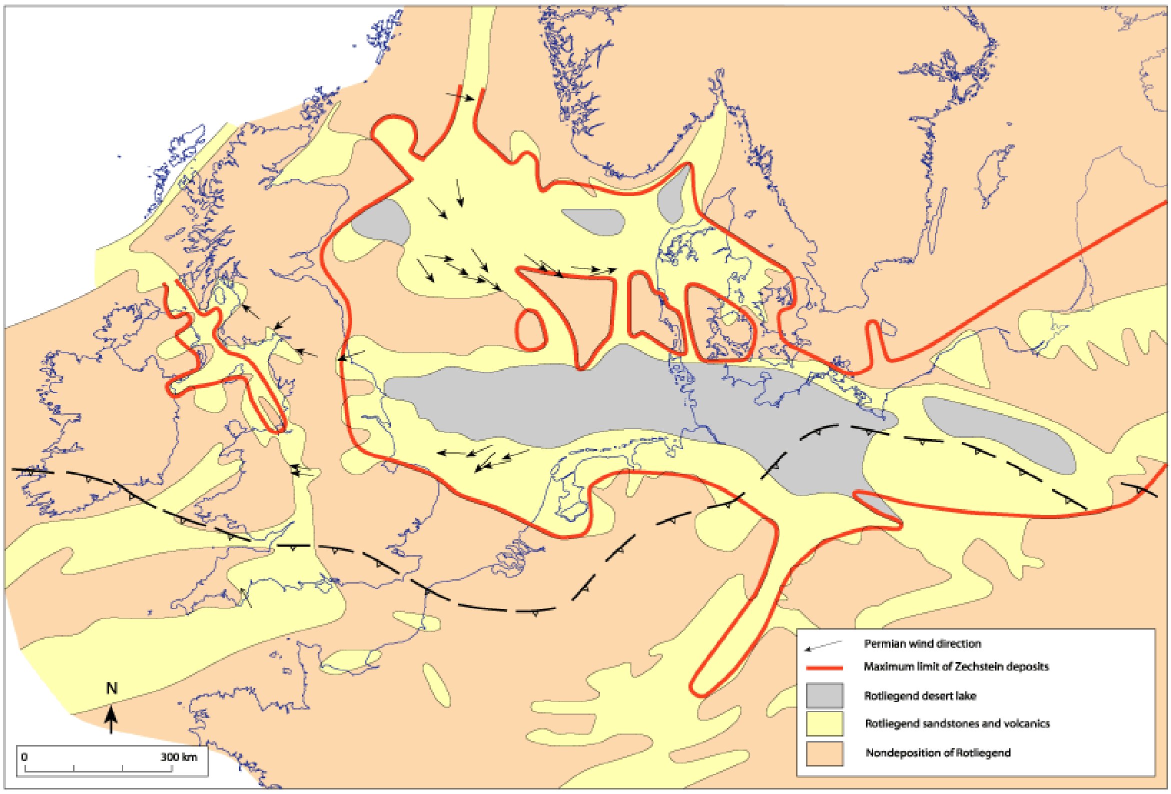


Fig. 2.1 Map of extent of Permian evaporite basin in southern North Sea (red line). Gray area is Rotliegend desert lake, yellow sandstones and volcanics, and light brown areas of no Rotliegend deposition. From Glennie et al. (2003).

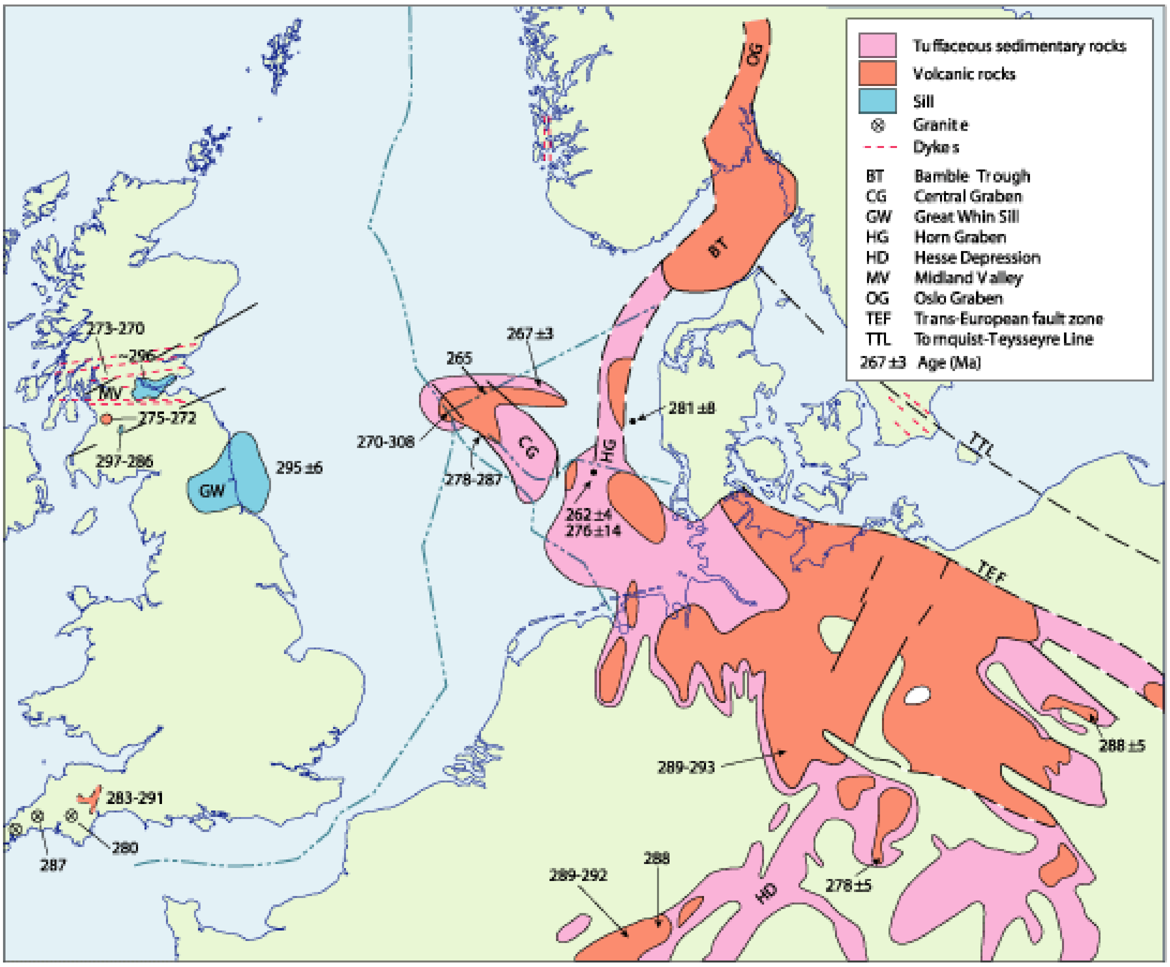


Fig. 2.2. Distribution of Permian magmatic rocks in the Southern Permian salt basin area. Numbers refer to age-dated samples. From Glennie et al. (2003).

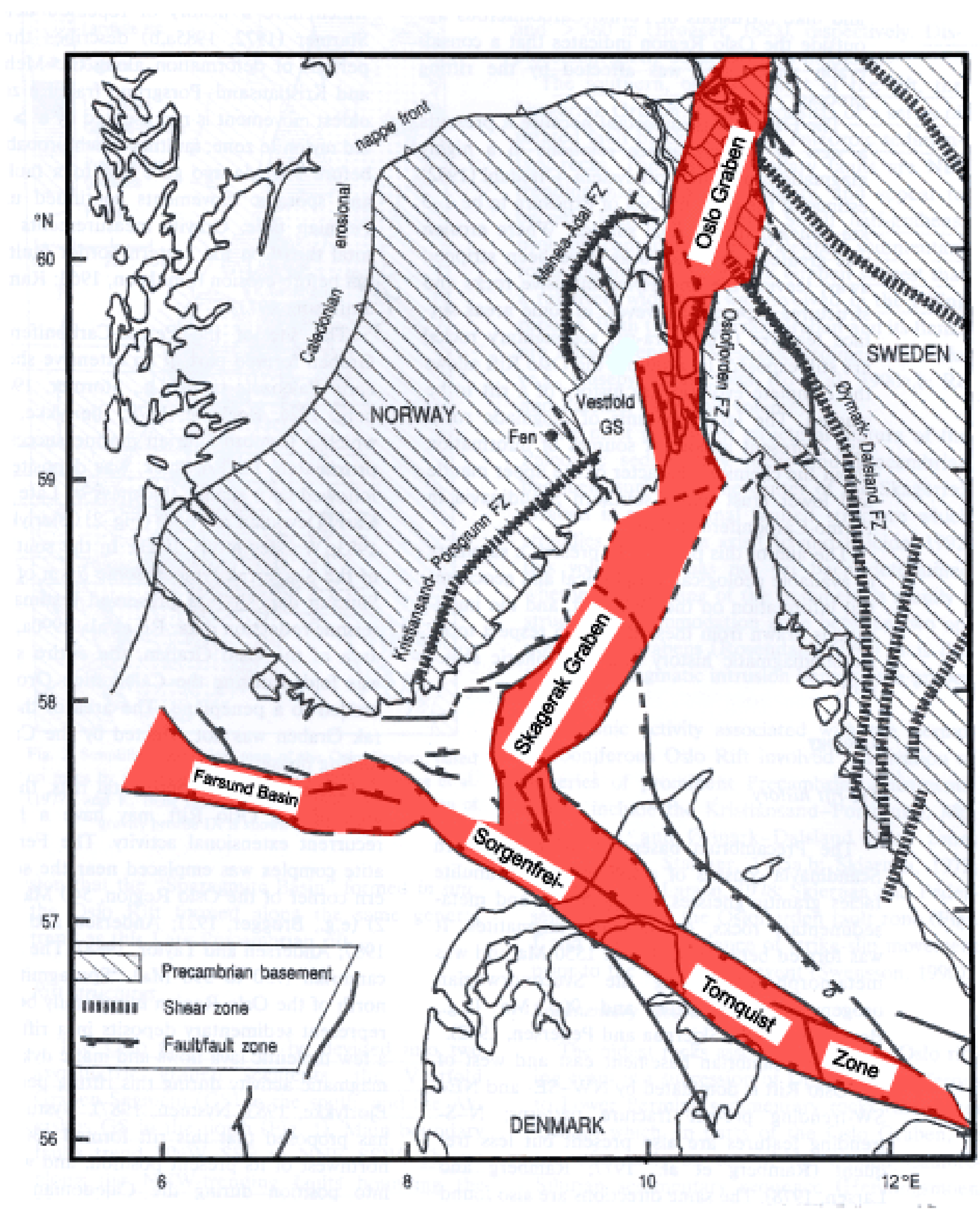


Fig. 2.3. Pattern of Permian rifting in the southern Permian salt basin. From Neumann et al (1992).

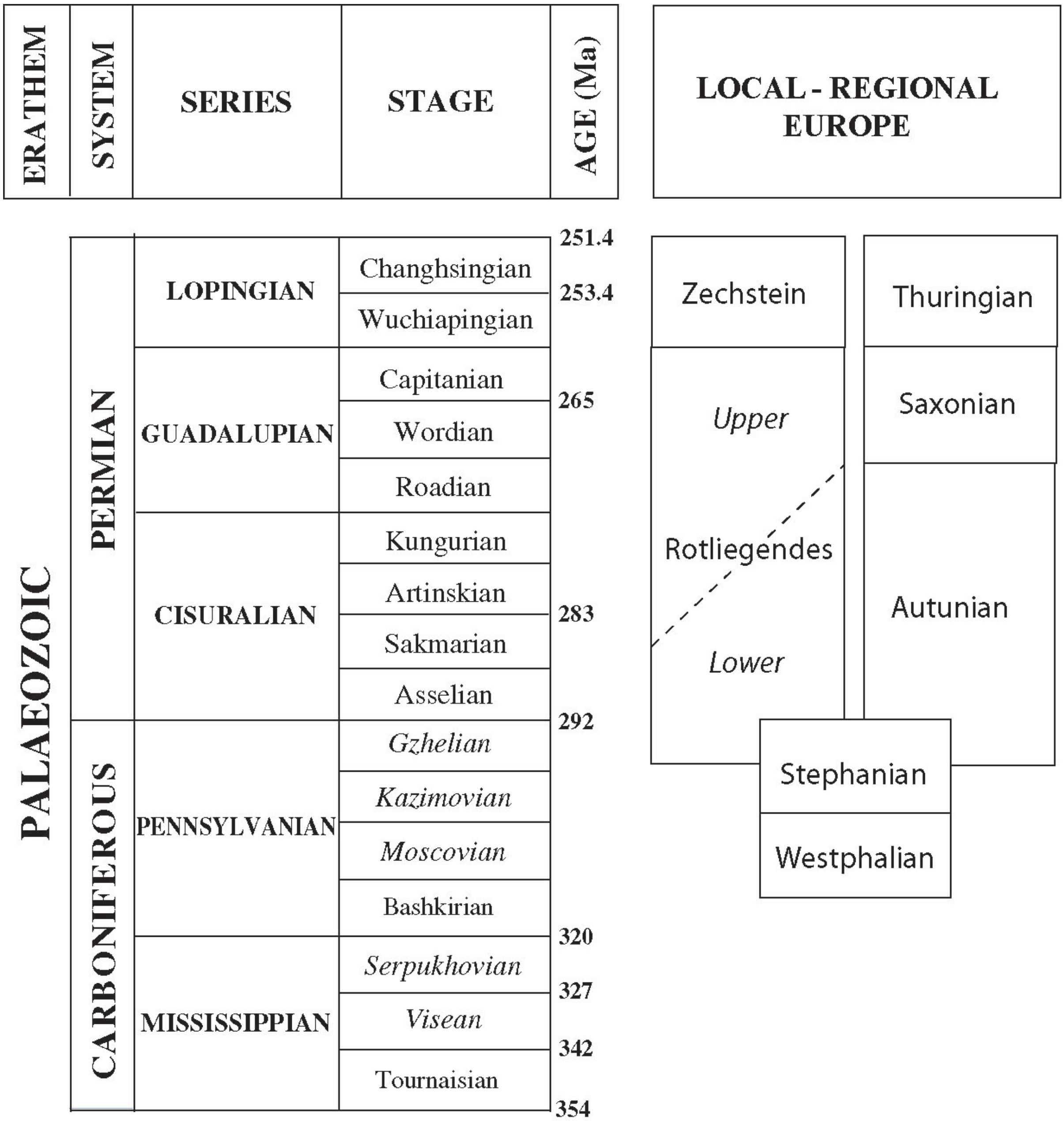


Fig. 2.4. Revised time scale. From International commision on stratigraphy timescale http://micropress.org/stratigraphy/) and tentative correlation with local-regional European stage names (Van der Voo & Torsvik 2004). Note the uncertainty in assigning a numerical boundary (stippled line) for the Lower Rotliegendes (c. Autunian) and the Upper Rotliegendes (c. Saxonian).

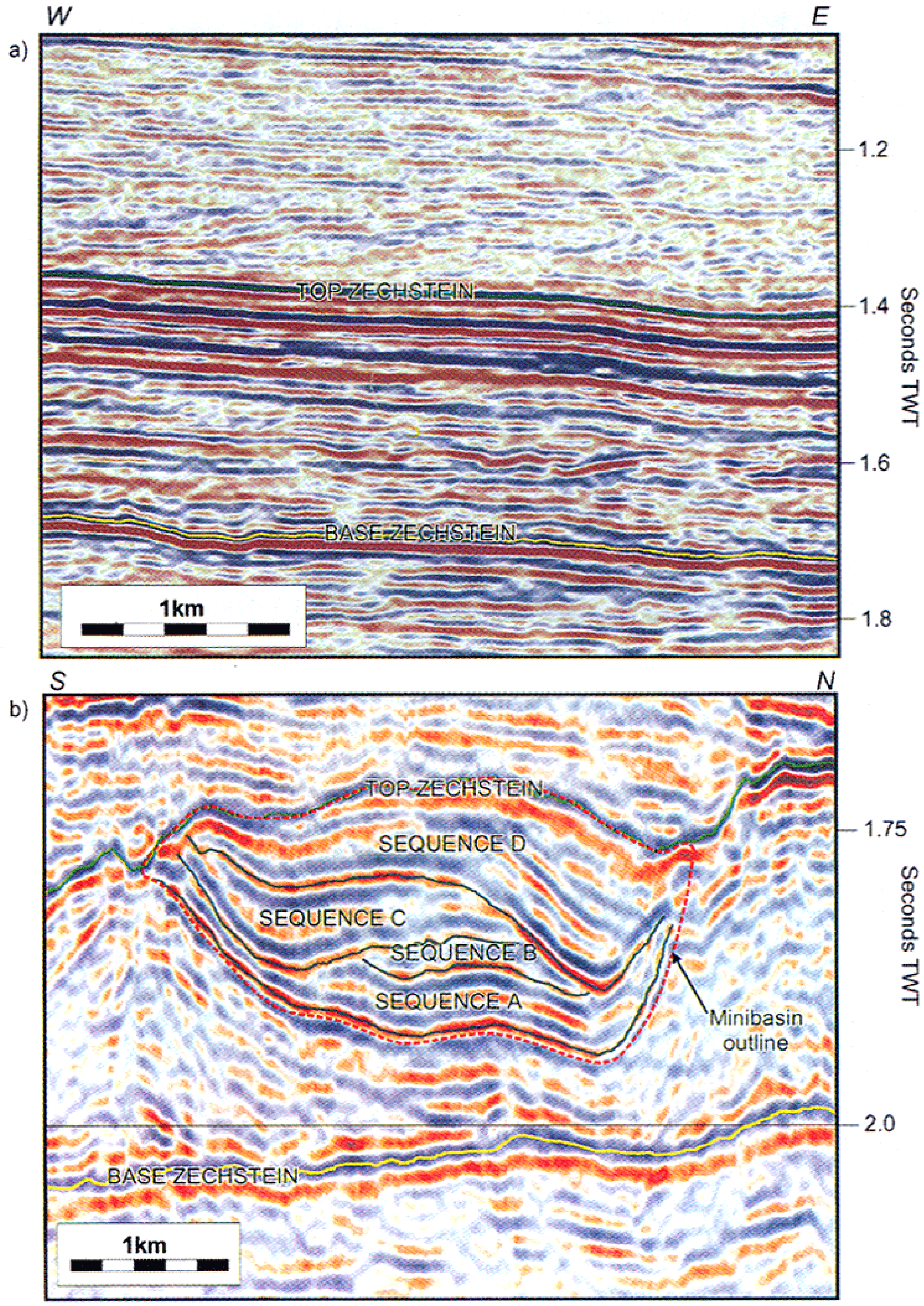
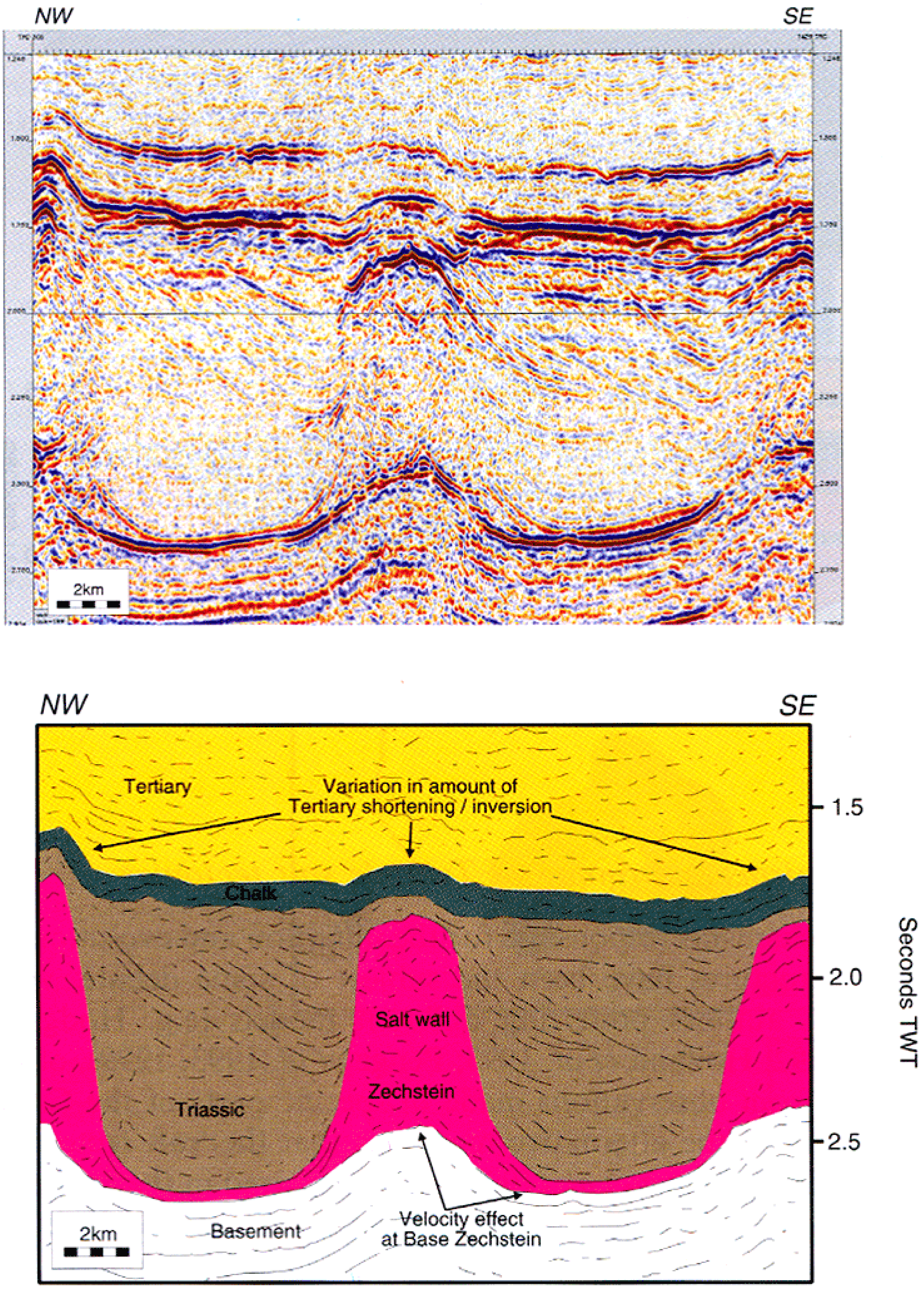


Fig. 2.5. Top section = undeformed example of the Zechstein succession. Lower section = Permian mini-basin. From Stewart & Clark (1999).



(e)

Fig 2.6. Seismic (top) and geoseismic section of Triassic mini-basin. From Stewart & Clark (1999).

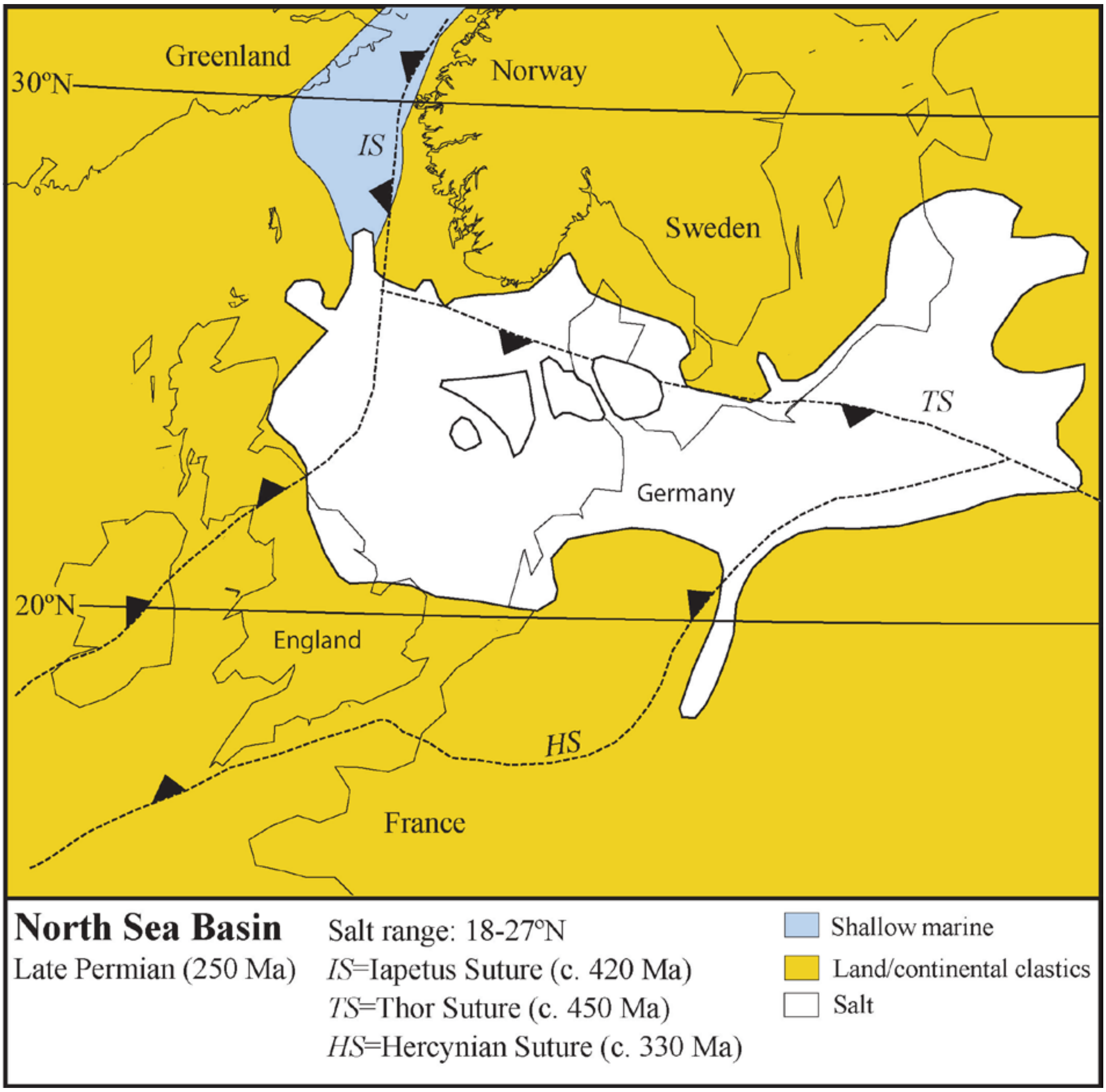


Fig. 2.7 Late Permian (250 Ma) palaeomagnetically constrained plate reconstruction showing the distribution of Zechstein salt deposits (18-27^{o}N) and the location of older collisional sutures.

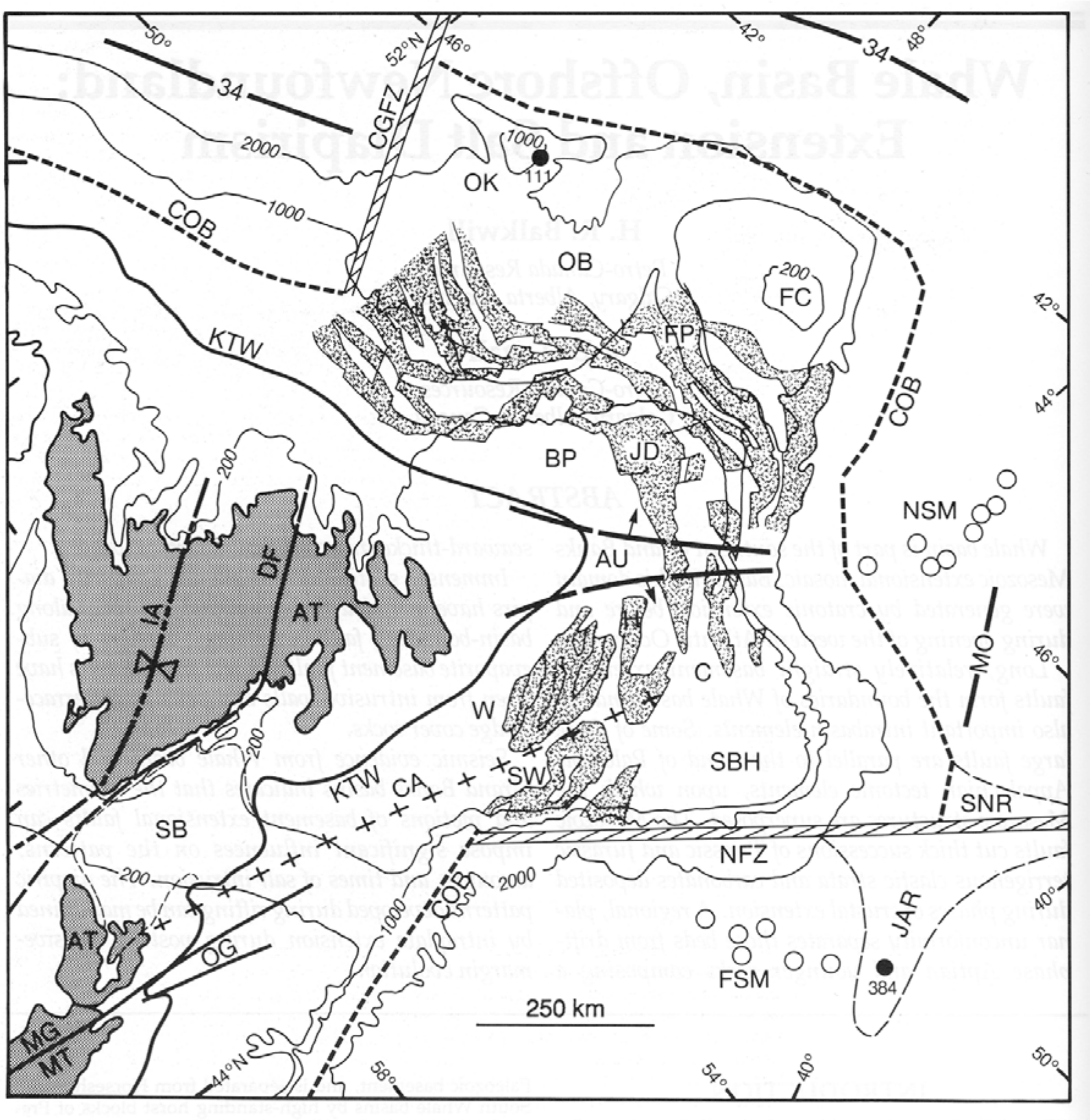


Fig. 3.1. Grand Banks tectonic elements. Abbreviations: COB= Continent-ocean boundary, CGFZ= Charlie Gibbs Fracture Zone, FC= Flemish Cap, IA= Iapetus Suture, NFZ=Newfoundland Fracture Zone. OK=Orphan Knoll, OB=Orpheus Basin. Basins shaded in light grey: H=Horseshoe Basin, JD=Jeanne D'Arc Basin, SB= South Whale Basin, W=Whale Basin. From Balkwill & Legall, 1989.

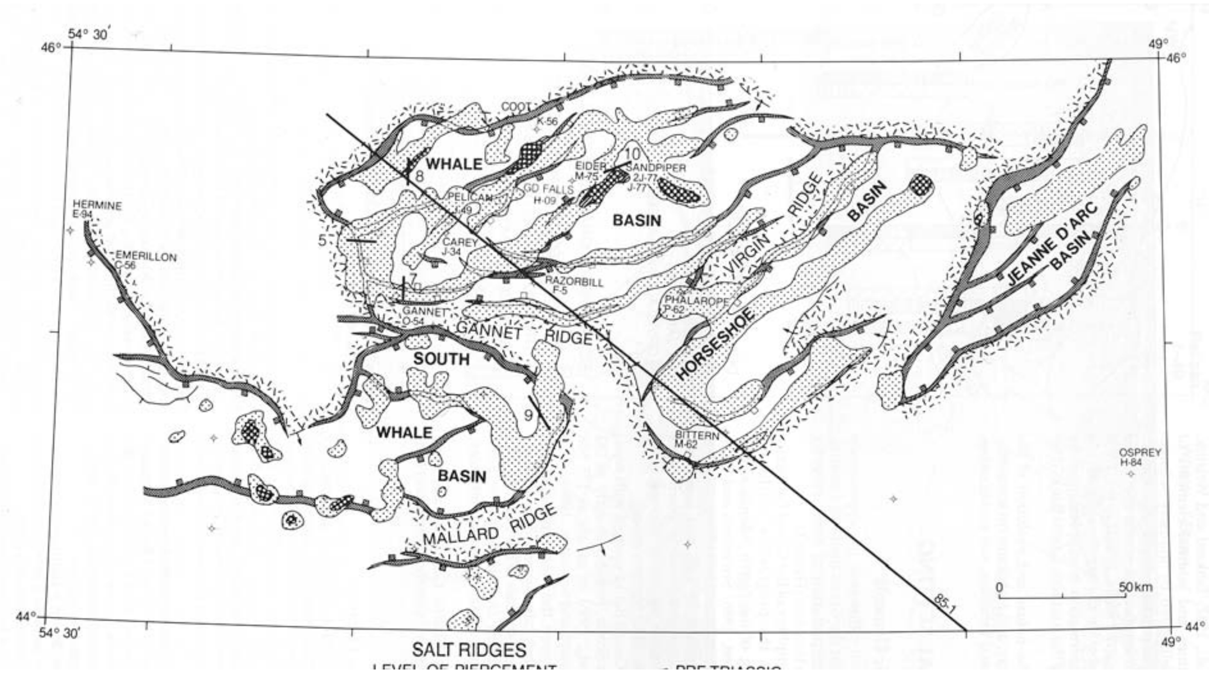


Fig. 3.2. Basins containing salt in the Grand Banks area. Salt basins are marked by dotted pattern. From Balkwill & Legall, 1989.

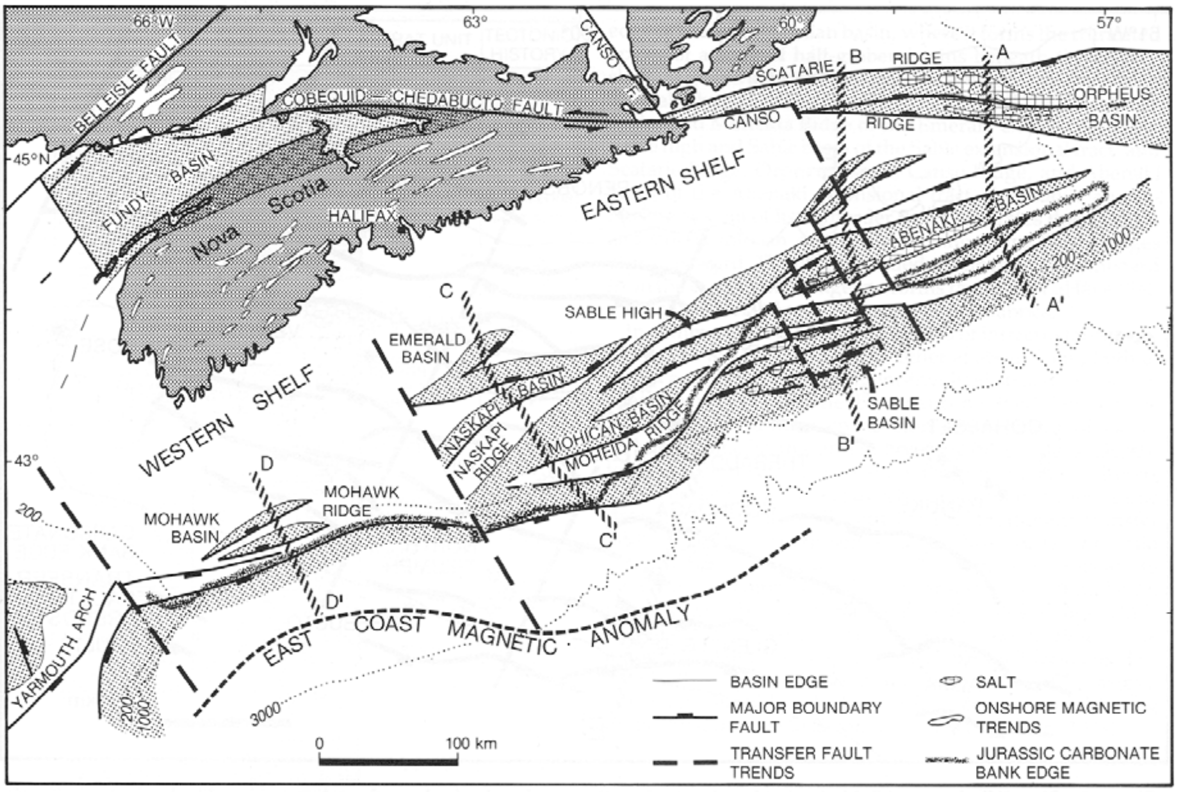


Fig. 3.3 Structural elements map of the Nova Scotia margin. Salt basins are marked with hatchured pattern. From Welsink, Dwyer & Knight, 1989.

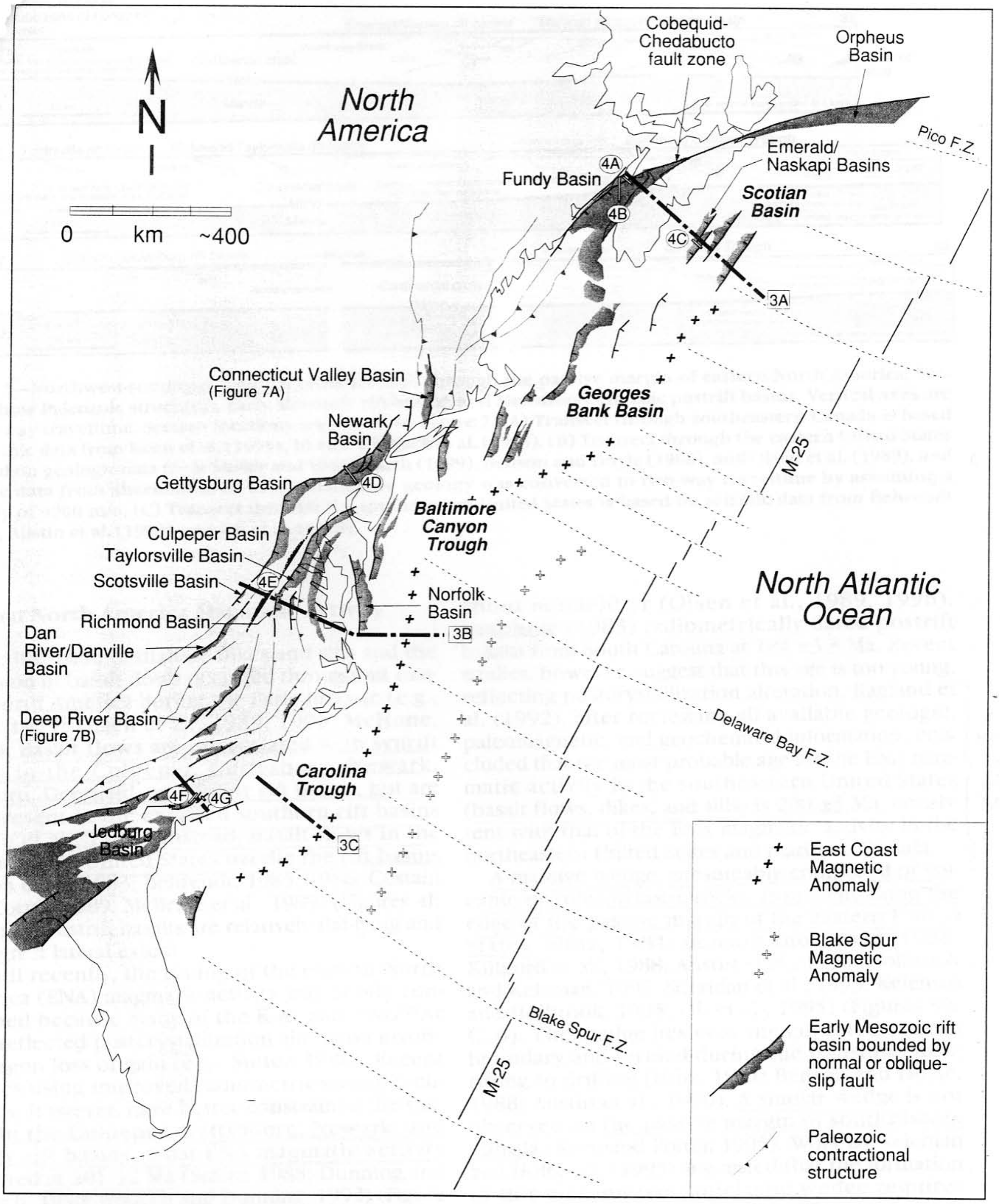


Fig. 3.4 Tectonic features map of eastern US and maritime Canada highlighting major Palaeozoic compressional structures and Mesozoic extensional structures. The heavy + symbols signify the East Coast Magnetic Anomaly, which correlate will the seaward-dipping reflector sequence (Holbrook & Keleman, 1993).

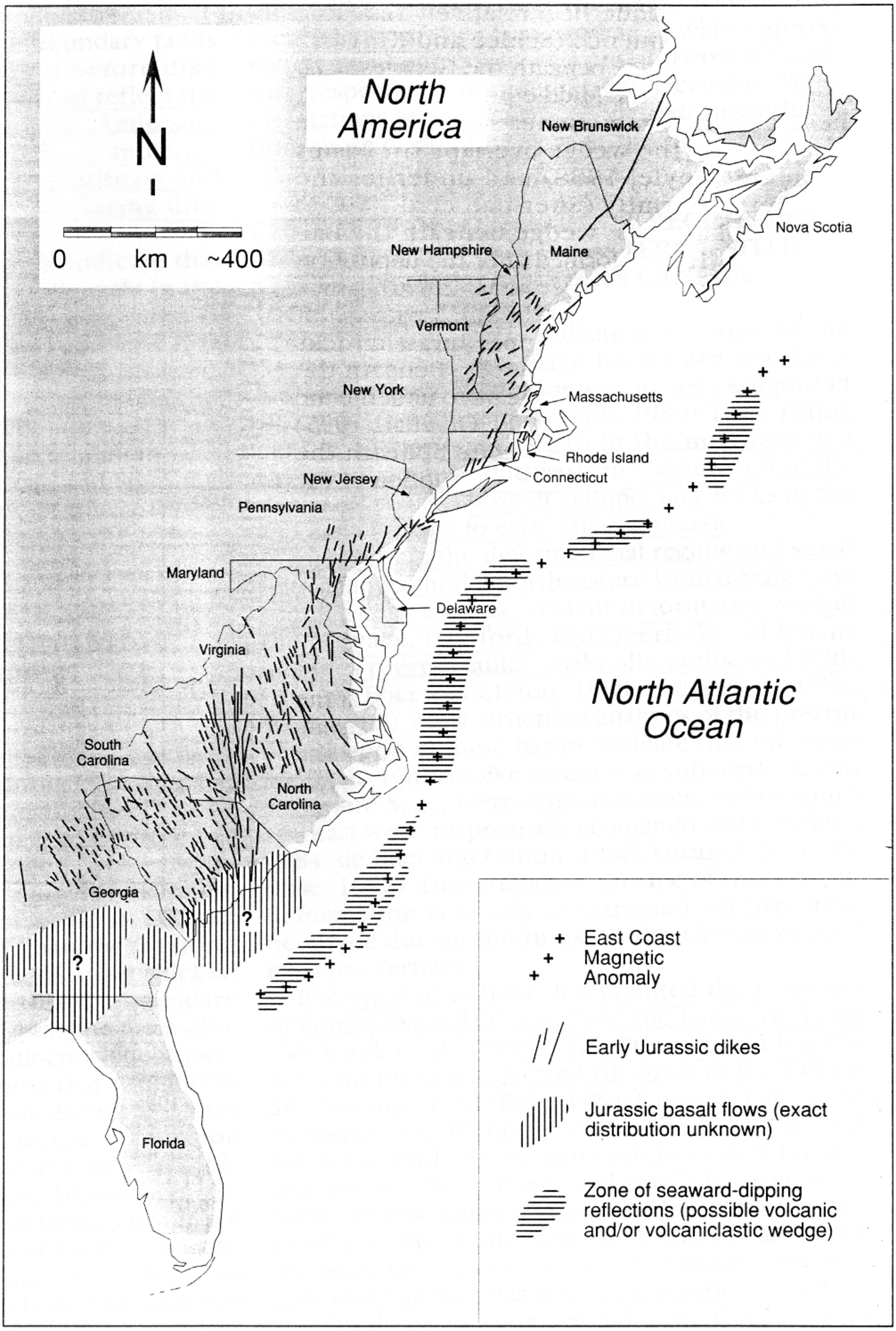


Fig. 3.5 Map of eastern North America highlighting dyke swarms and basalt flows of Early Jurassic age. Note the coast-perpendicular trend in southeastern US in contrast to the coast-parallel trend in northeastern US. From Withjack et al (1998).

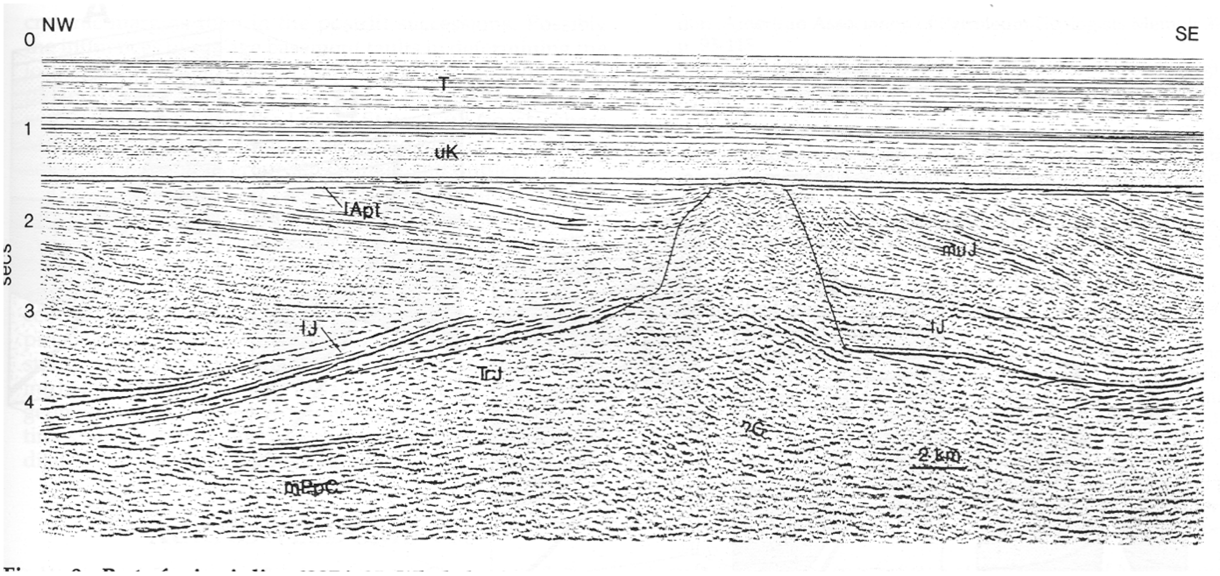


Fig. 3.6 Seismic section across part of the Whale Basin, Grand Banks. TrJ=Argo salt, IJ= Early Jurassic carbonates, Iapt=sub-Aptian unconformity. Note the pronounced rotation of strata below the sub-Aptian unconformity, the truncation of the salt diapir, and the undeformed horizontally bedded Upper Cretaceous and Tertiary strata. From Balkwill & Legall (1989).

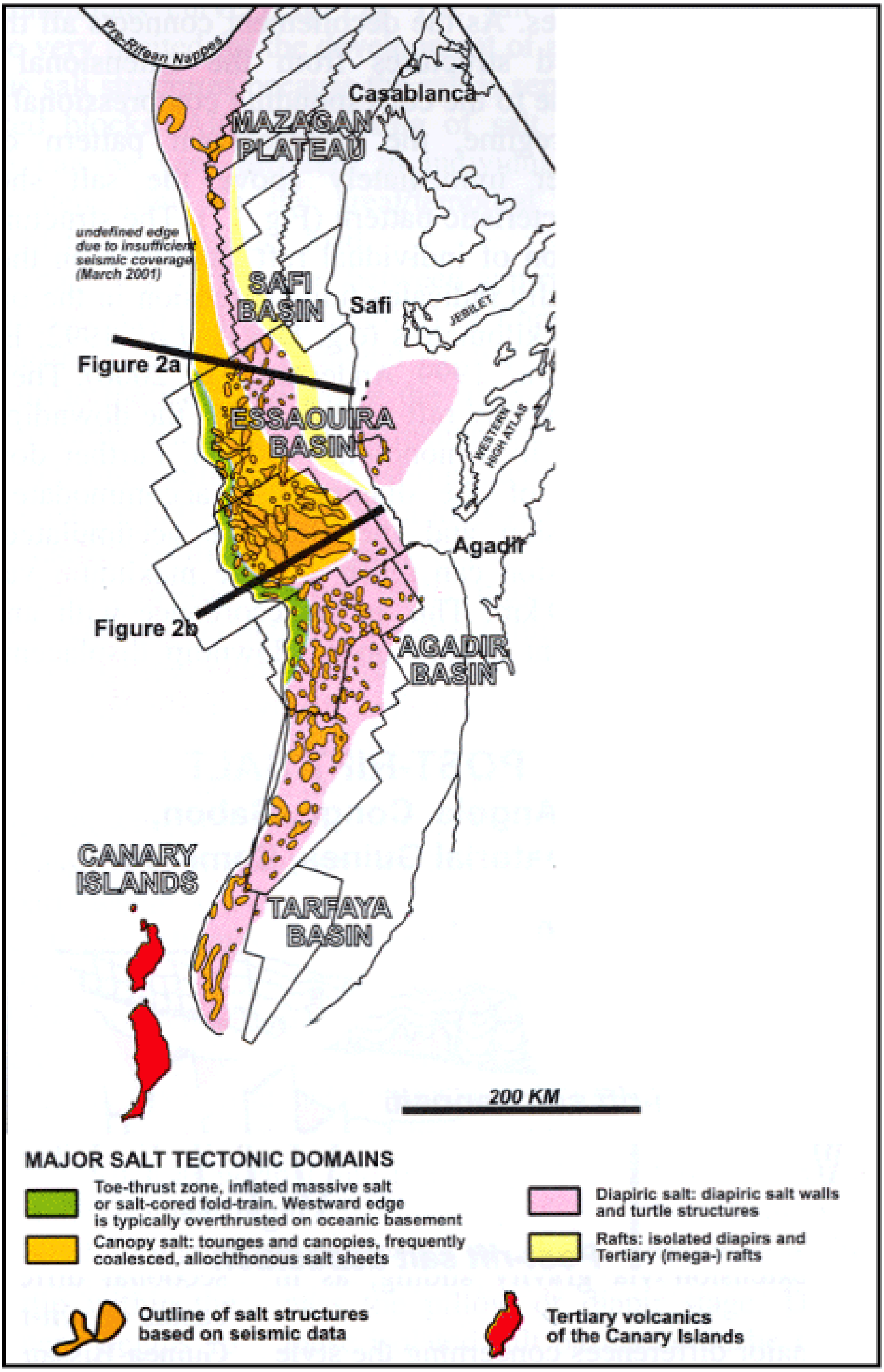


Fig. 4.1 Map of salt distribution and styles of salt deformation off Morocco. From Tari et al. (2003).

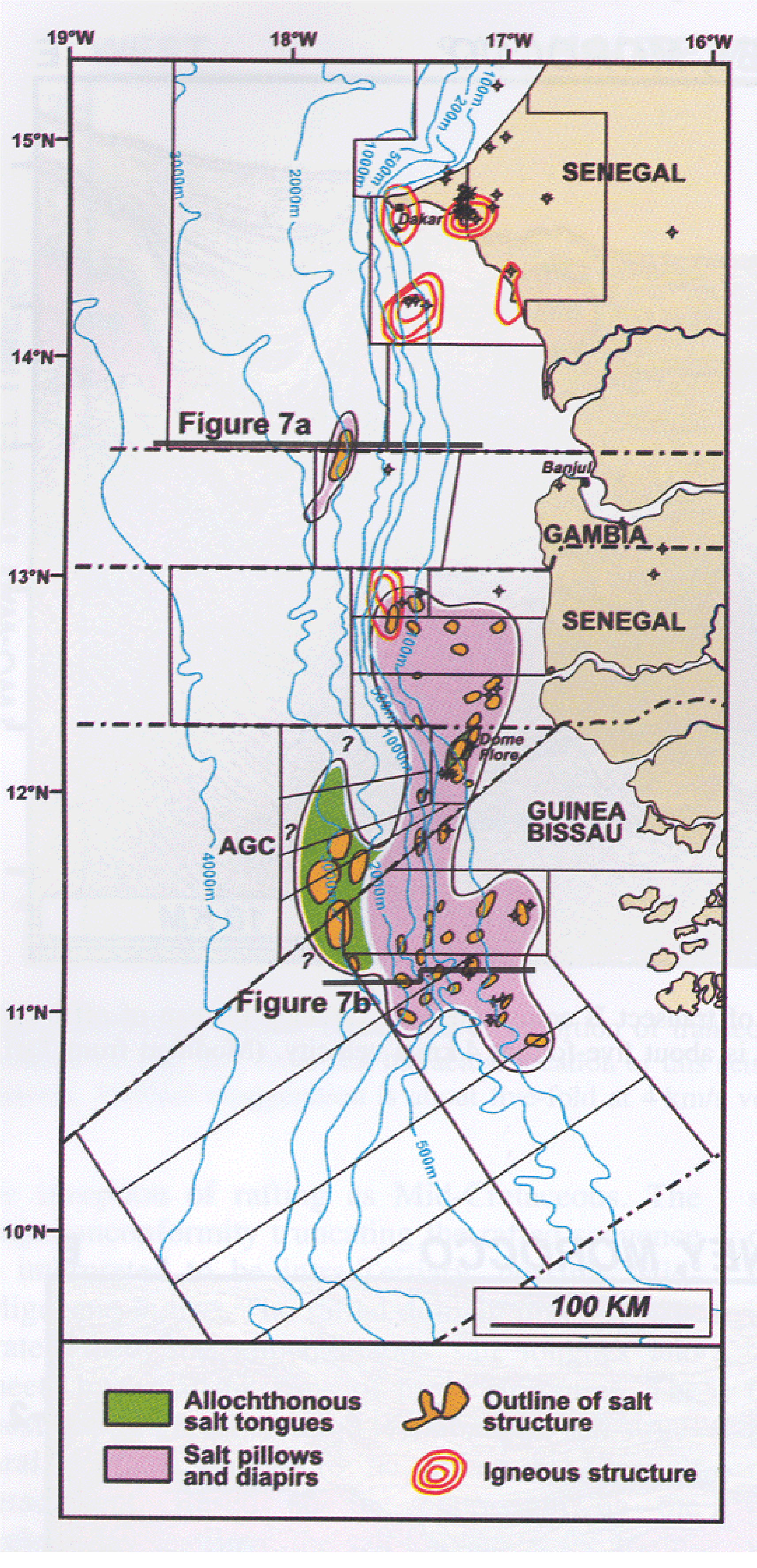


Fig. 4.2 Map of salt distribution and outline of salt structures off Gambia, Senegal, and Guinnea Bissau. From Tari et al. (2003).

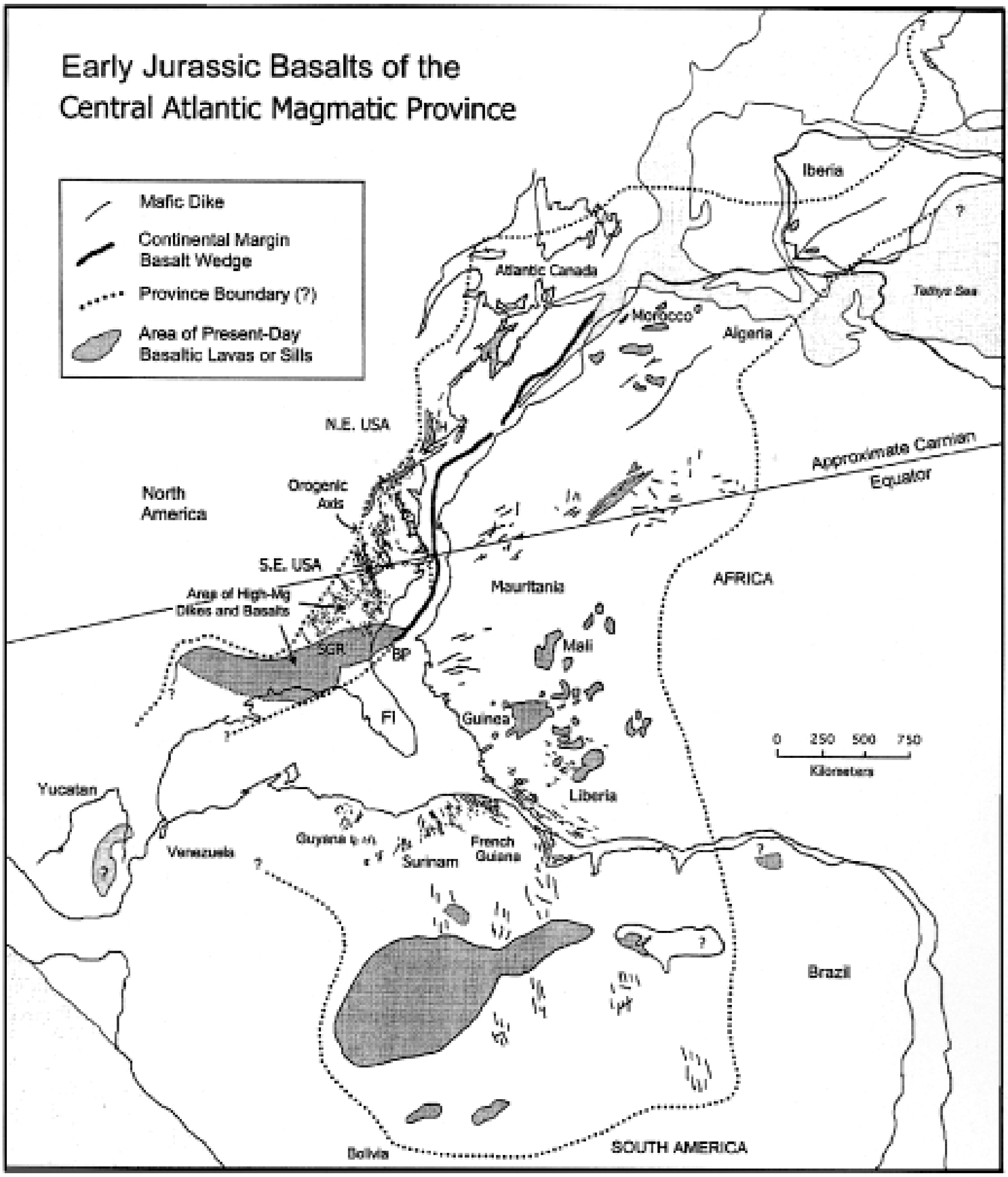


Fig. 4.3 Central Atlantic magmatic features, largely formed around 200 Ma (McHone, 2000). Notably, the NW-trending dyke swarms in Liberia (L) are Ar/Ar age dated to 185 $\pm$ 3-4 Ma (Dalrymple & Lanphere, 1974) and thus do not belong to the suite of 200 Ma dykes.

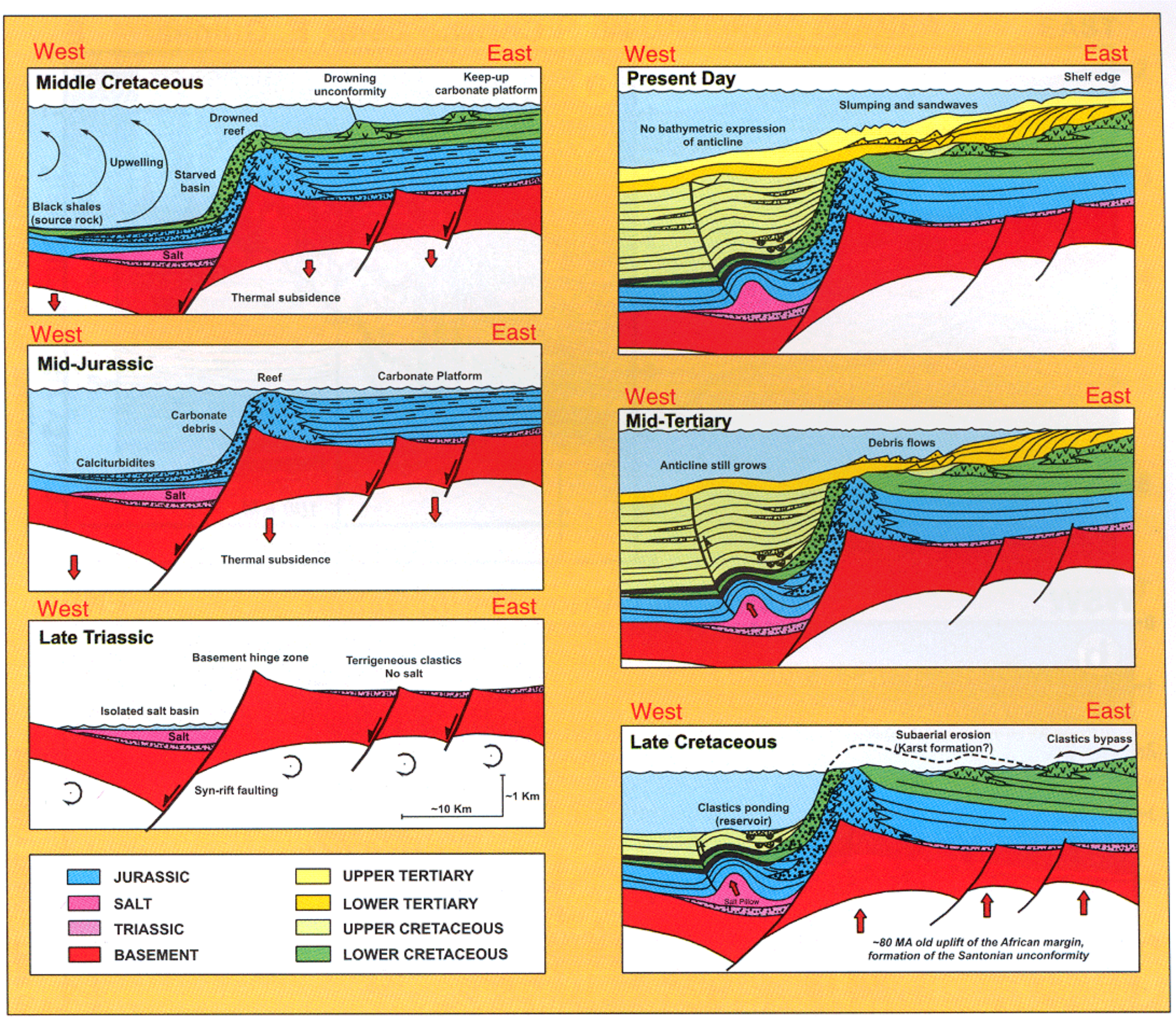


Fig. 4.4. Sequential restoration of a regional seismic profile (transect 7a in Fig. 4.2). Salt was deposited west of the basment hinge zone in Late Triassic – Early Jurassic time, and deformed in Santonian time during uplift and erosion east of the hinge zone. From Tari et al. (2003).

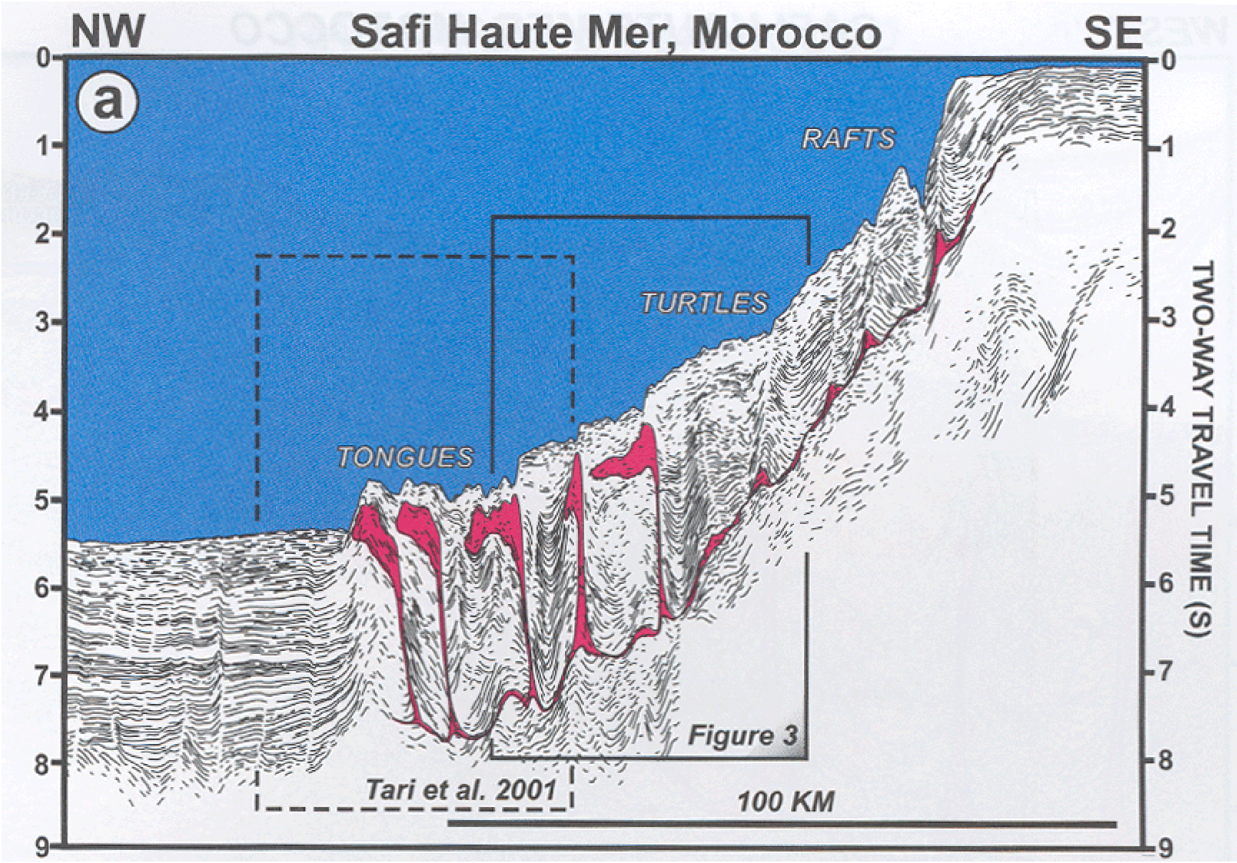


Fig. 4.5 Geoseismic profile across the Safi Haute Mer area offshore Morocco. See Fig. 4.1 for profile location (profile 2a). The profile shows a linked system of post-salt deformation, with updip extension and downdip compression. From Tari et al. (2003)

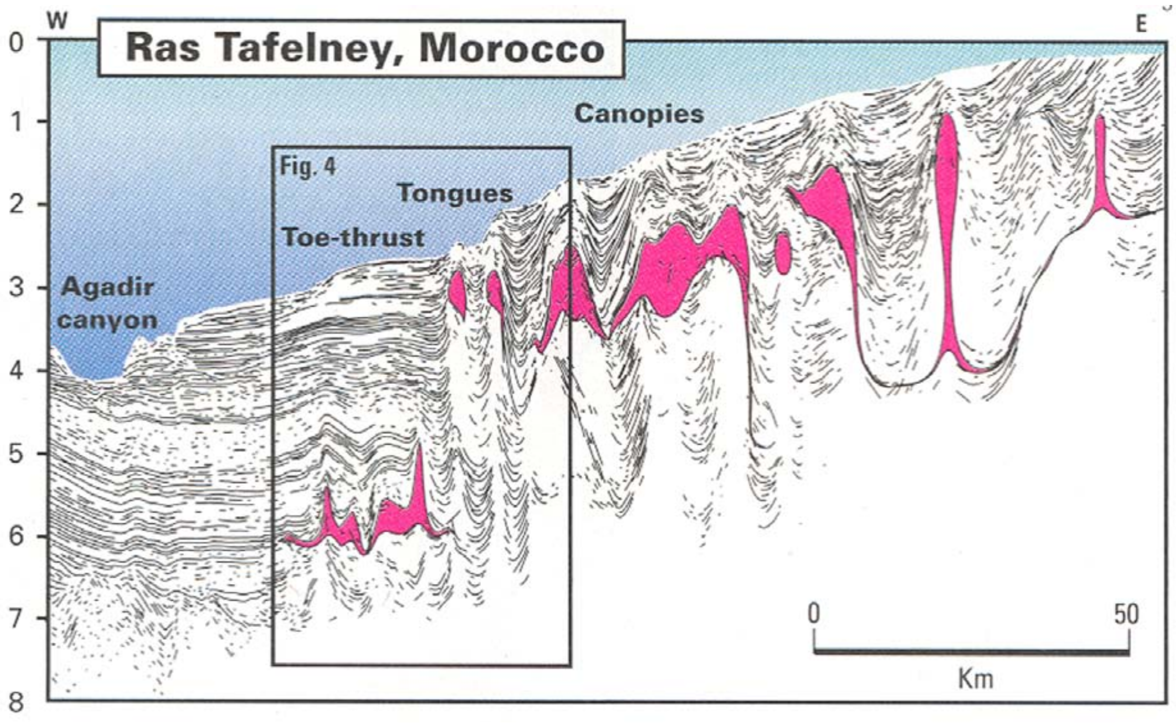


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Fig. 4.6 Geoseismic profile from the Ras Tafelney fold belt, Morocco. See Fig. 4.1 for profile location (profile 2b).

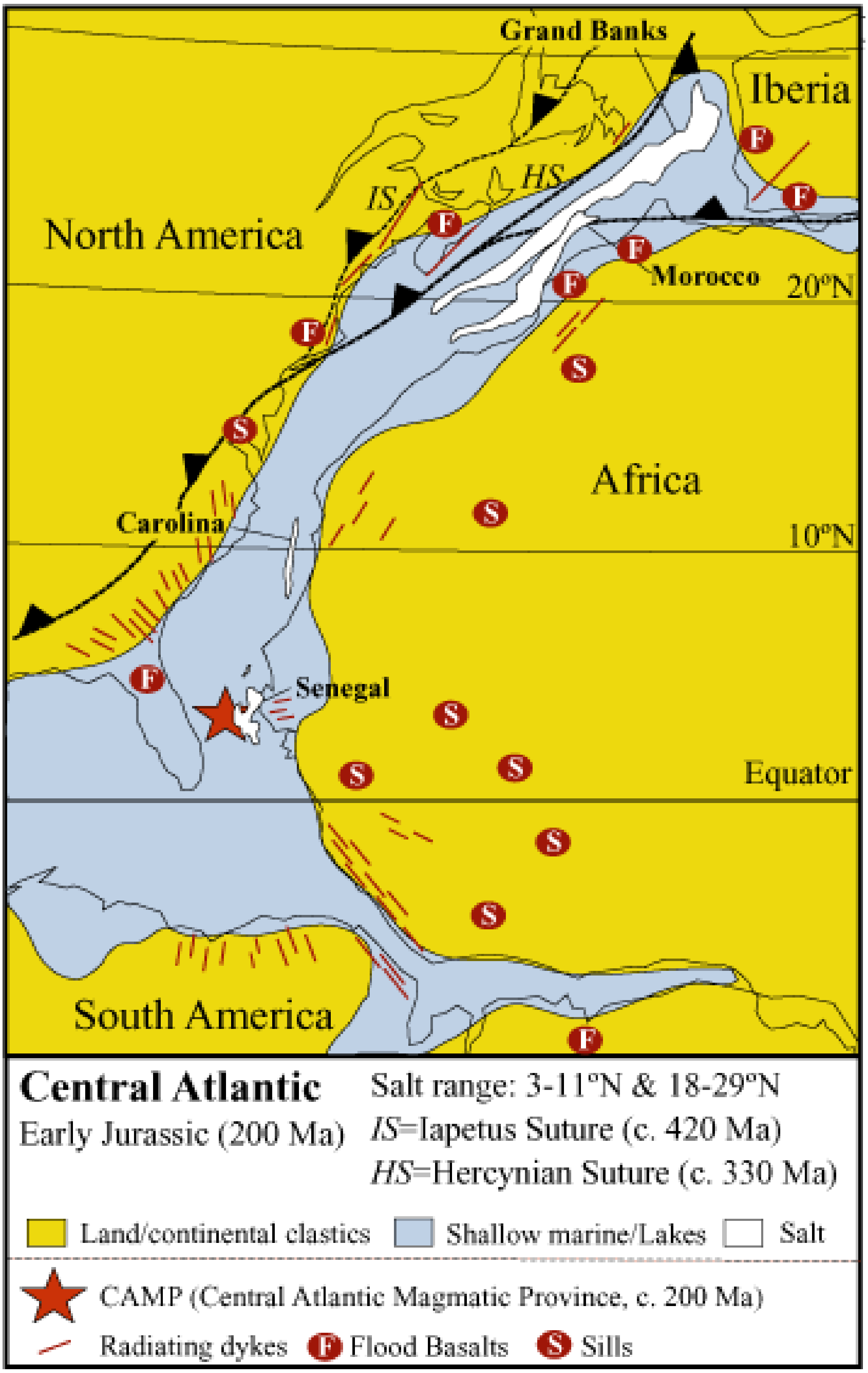

Fig. 4.7. Late Triassic-Early Jurassic (c. 200 Ma) palaeomagnetically constrained plate reconstruction showing the distribution of salt deposits along the Central Atlantic margins, and suture zones. While the Morocco and Newfoundland/Nova Scotia salt basins plot at subtropical latitudes c. 18-29° N), the Senegal/Guinea Bissau and Carolina Trough salt plots anomalously equatorial (c. 4-10° N), suggesting that they may be of a different age. The CAMP star is a proposed position of a plume responsible for the magmatism.

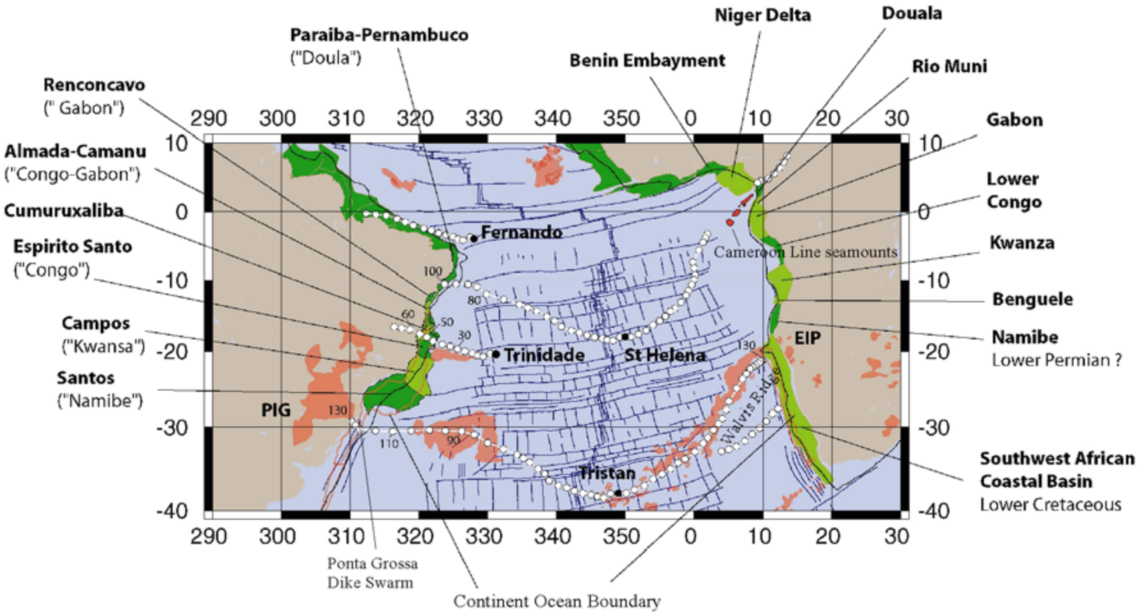

Fig. 5.1. Generalized location map of the South Atlantic region, including the Aptian salt basin, showing geographical and geological features mentioned in the text. PIG and EIP refer to the Parana and Etendeka continental flood basalts (c. 134 Ma).

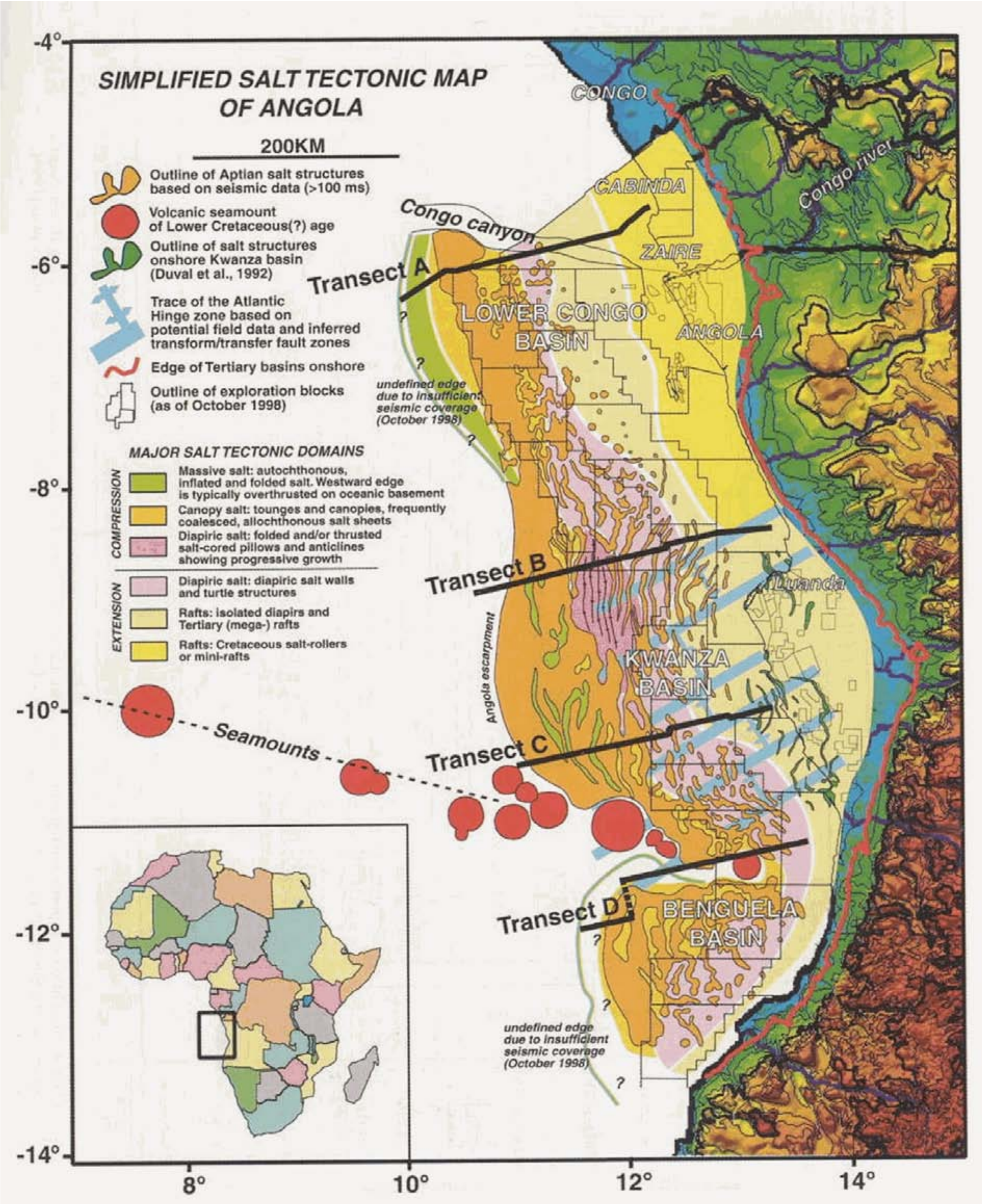

Fig. 5.2. Map of salt features along the Angola margin, West Africa. From Tari et al. (2003).

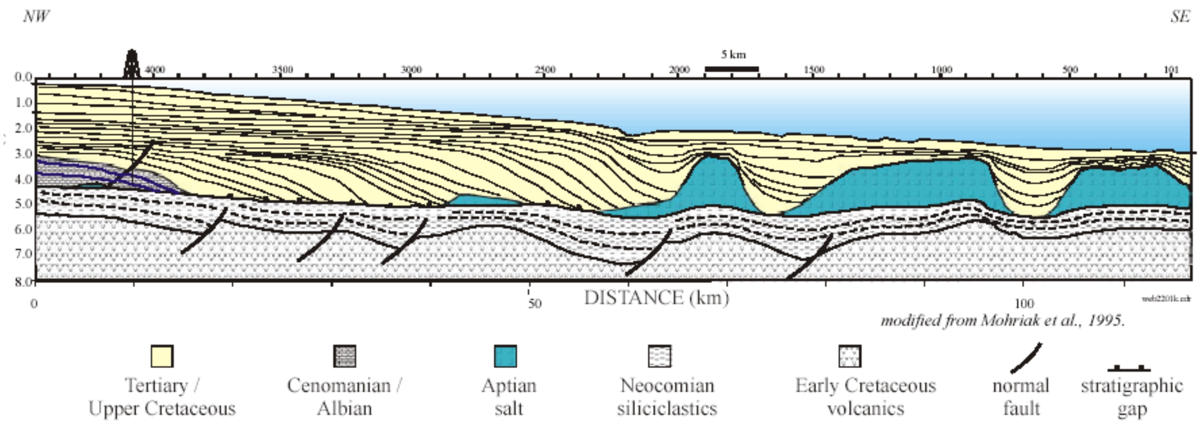

Fig. 5.3. Geoseismic profile from the Santos Basin, Brasil, illustrating how prograding sediments progressively have pushed the salt basinward. From Mohriak et al., (1995).

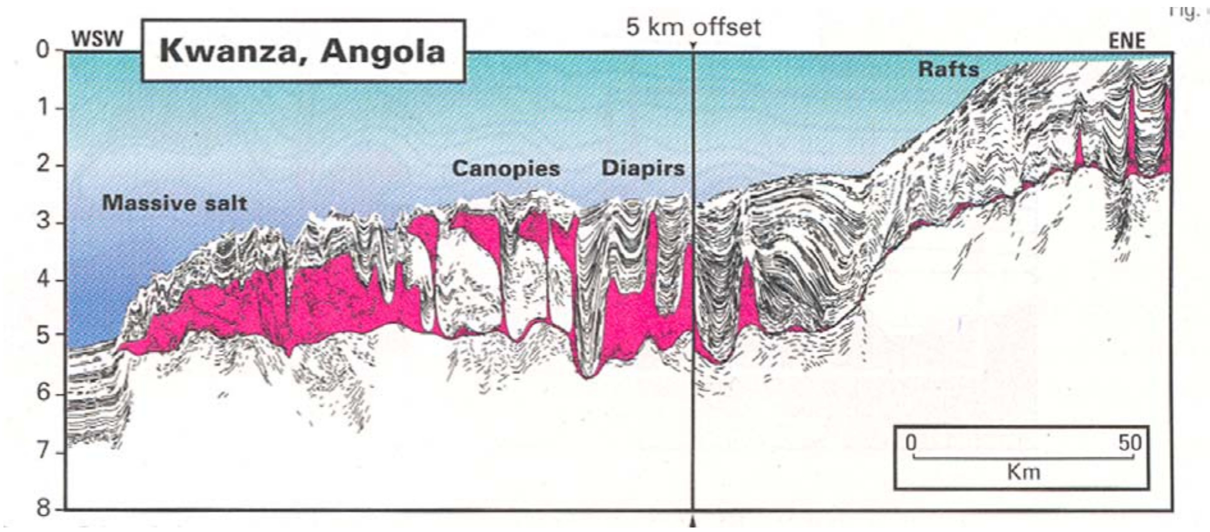

Fig. 5.4. Geoseismic section from the Kwanza Basin, Angola, illustrating the typical domains of salt structuring observed along the West Africa margin. Extensional structures, including large rafts, characterize the updip portion of the margin, while complementary compressional structures mark the lower part of the margin. Salt has been displaced basinward during buildup of the post-salt succession. See Fig. 5.2 for profile location (Transect C). From Tari et al. (2003).

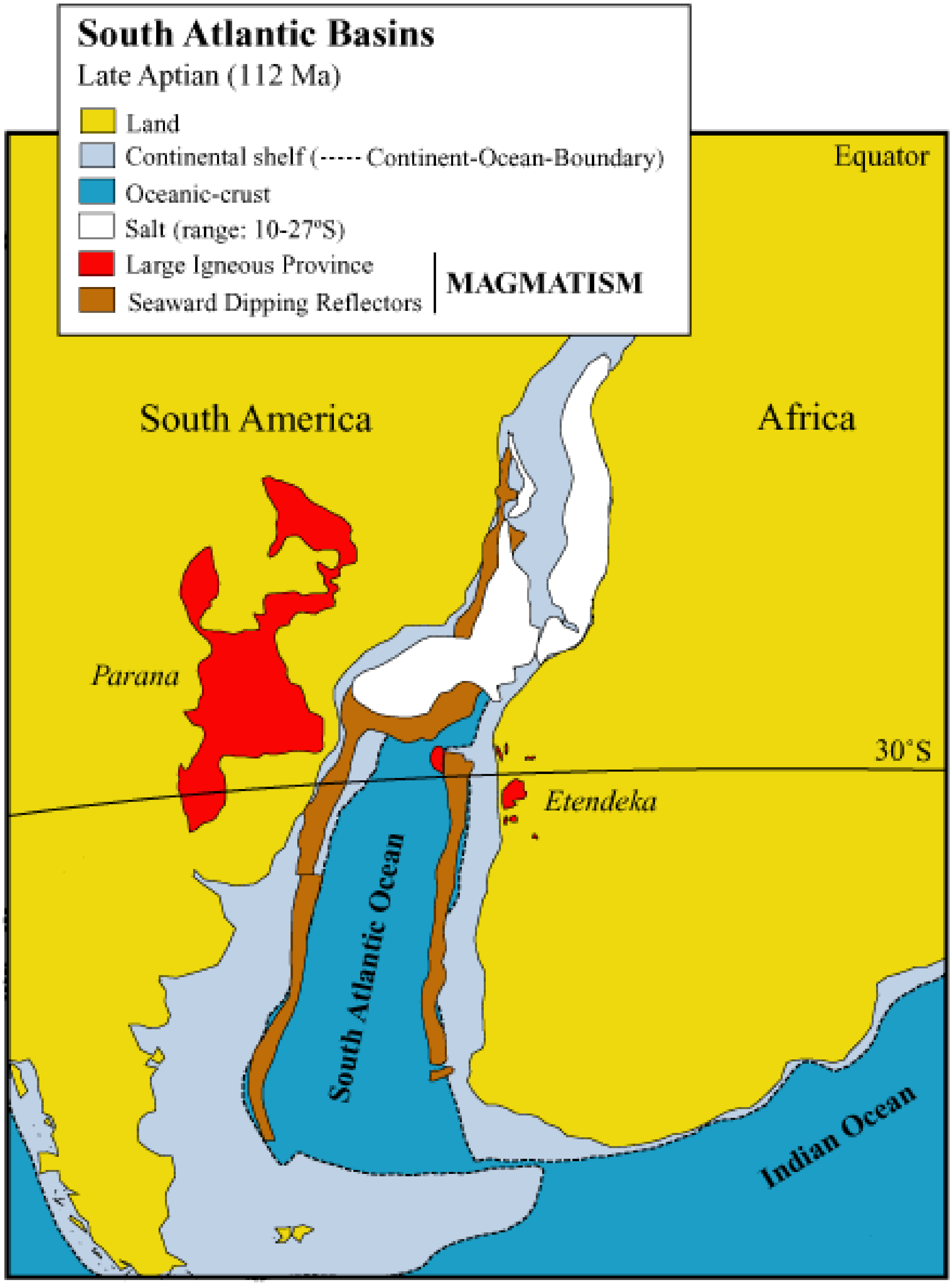

Fig. 5.5. Aptian (112 Ma) palaeomagnatically constrained plate reconstruction of the South Atlantic.

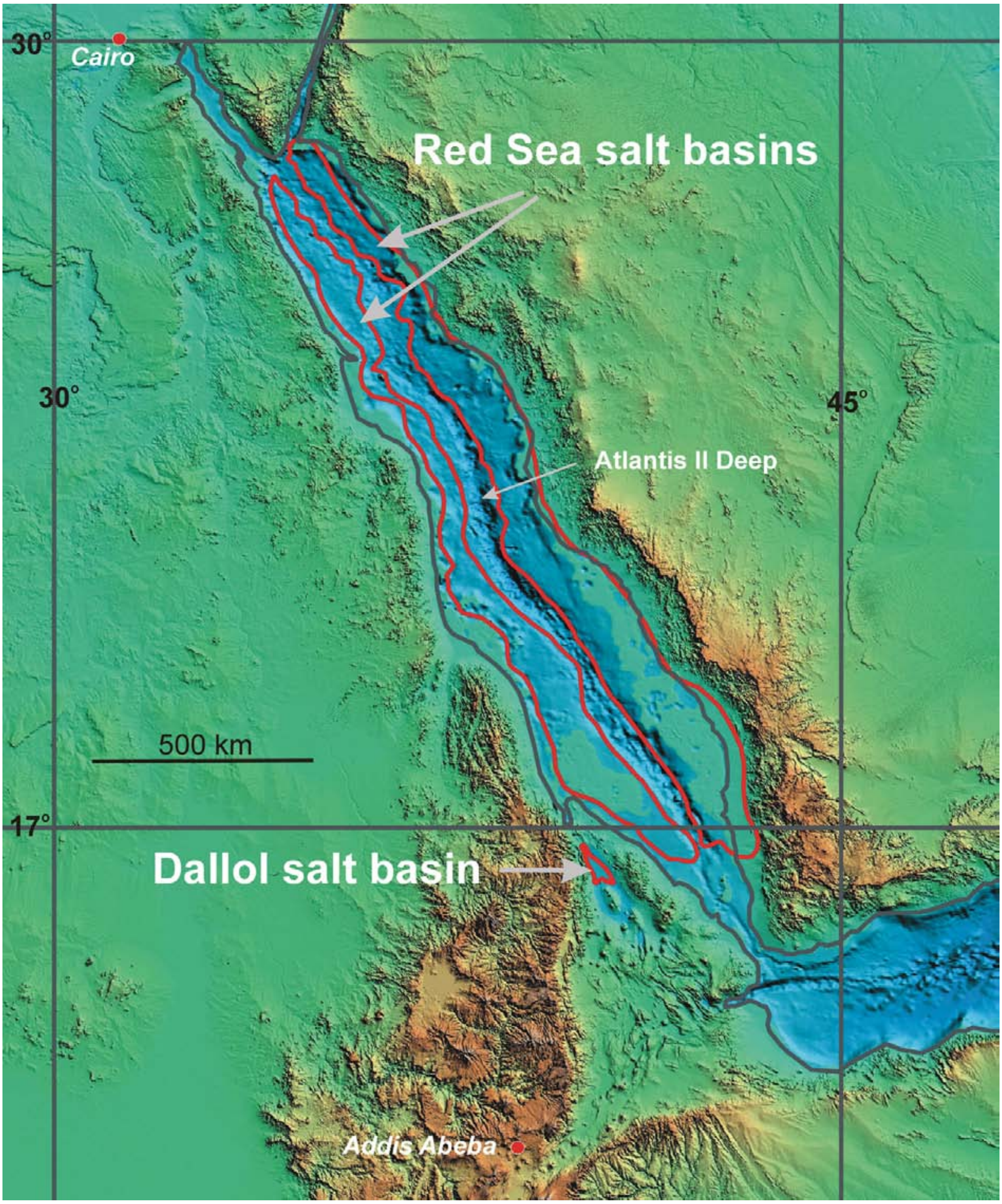

Fig. 6.1. The Red Sea basin with salt basins outlined.

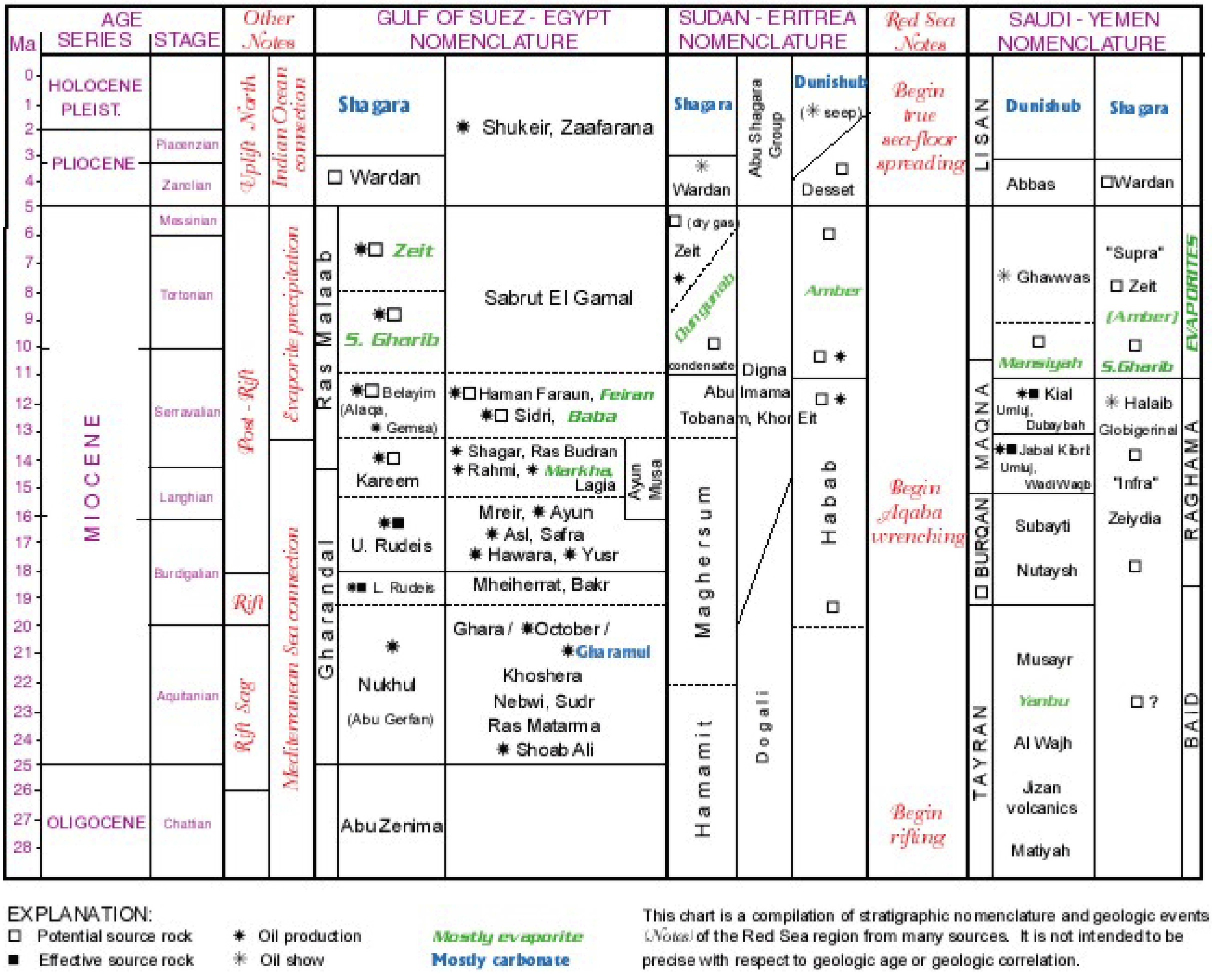

Fig. 6.2. Time-chart/Basin development, Oligocene to Present, for the Red Sea and the Gulf of Suez. From Lindquist (1999).

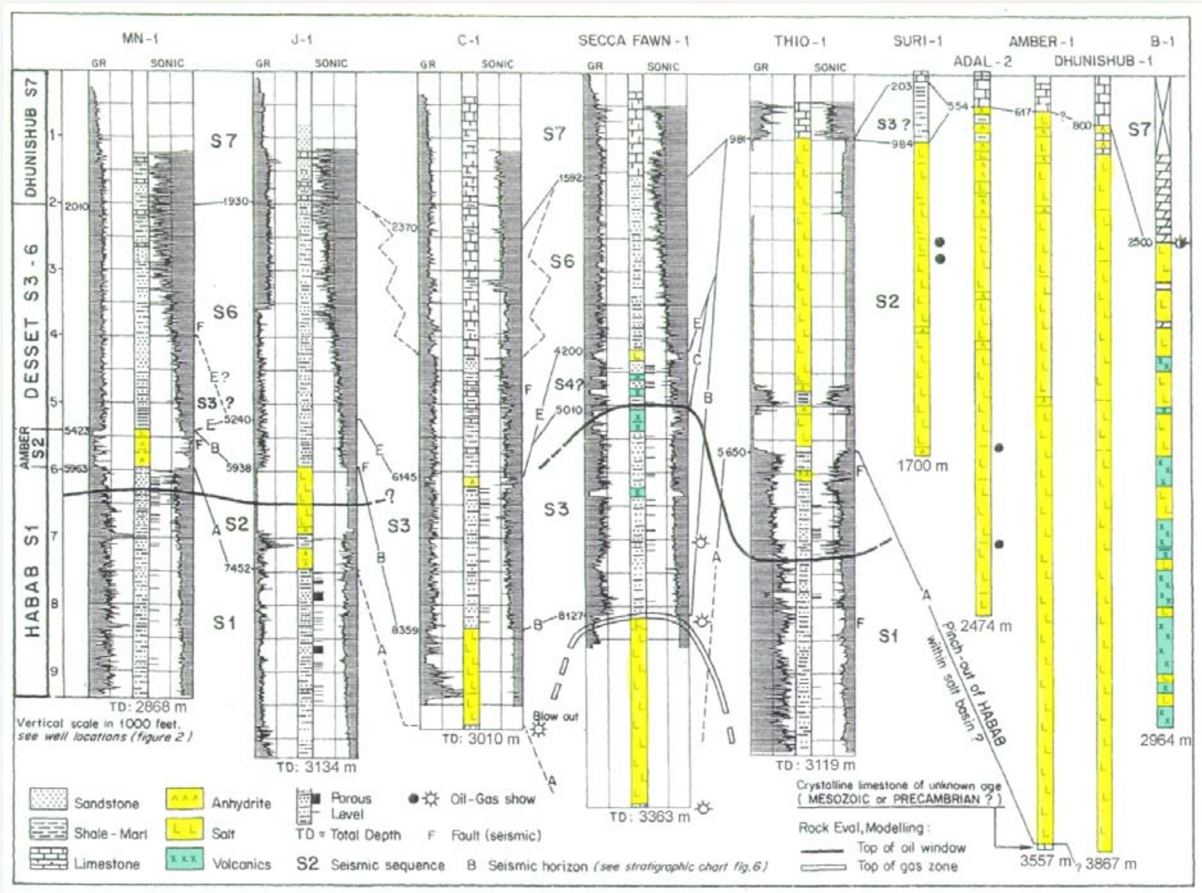

Fig. 6.3. Well logs from the southern Red Sea showing thickness and depth of evaporites (yellow) and volcanics (green). From Savoyat et al. (1989).

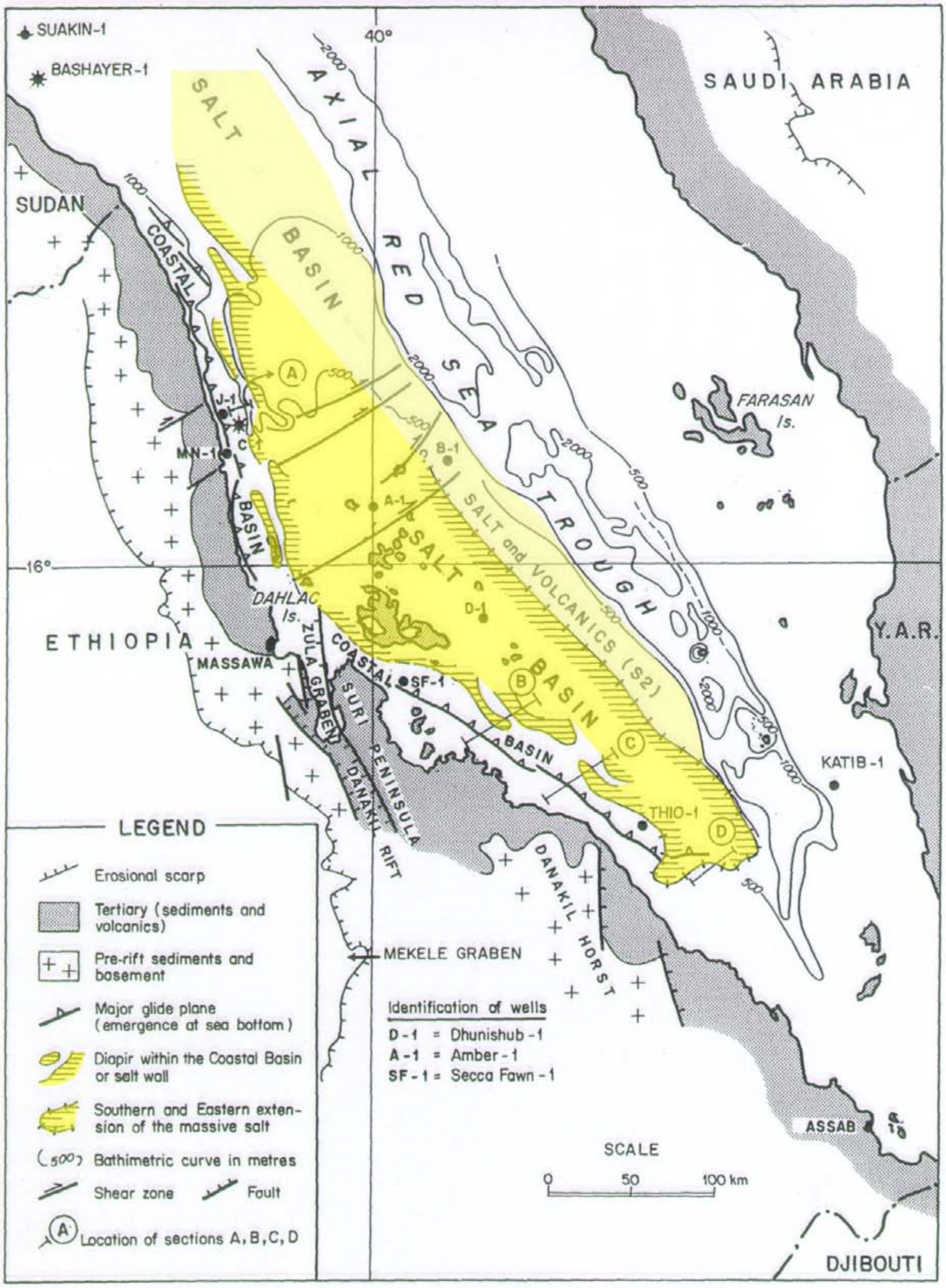

Fig. 6.4. The salt basin off the coast of Eritrea (yellow). From Savoyat et al. (1989).

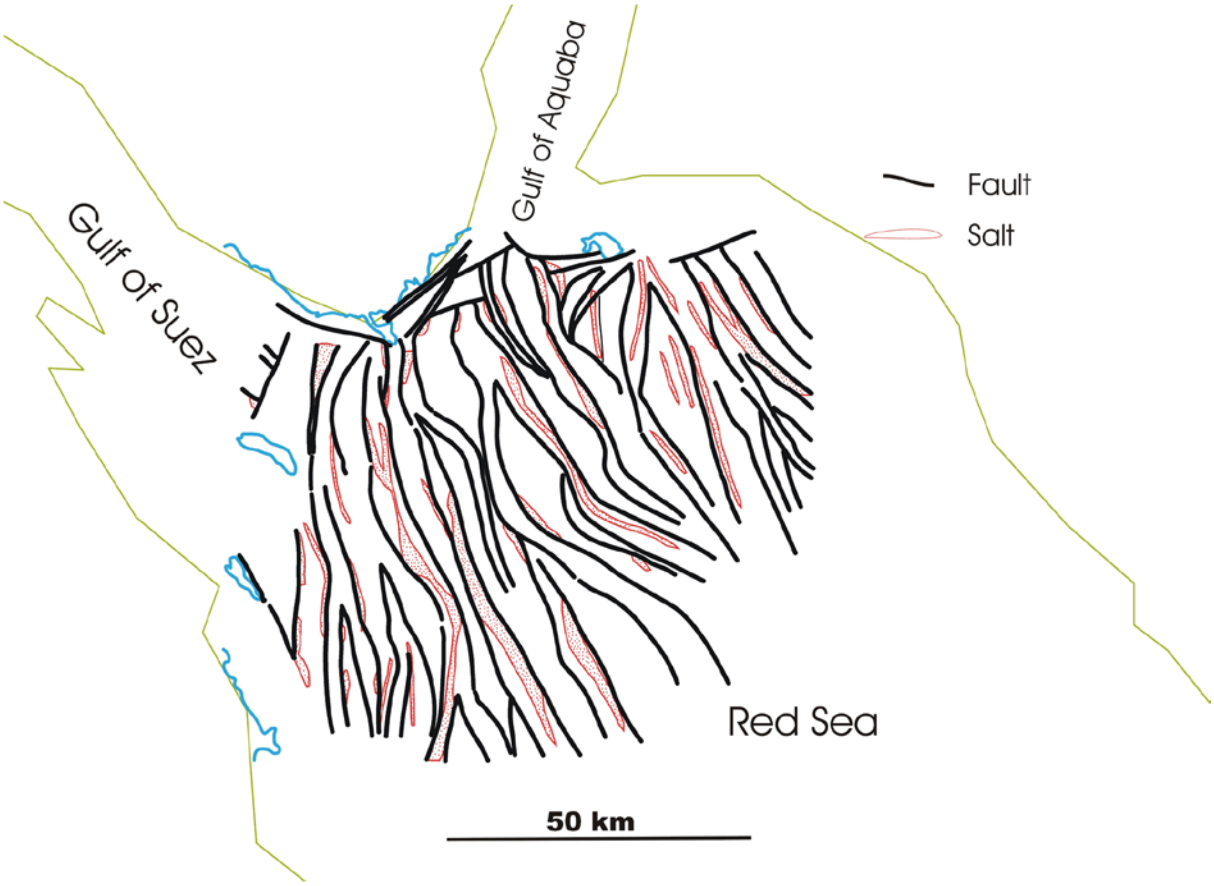

Fig. 6.5. Linear diapiric salt walls parallel to and closely associated with grabens and half grabens produced by ongoing extension in the northern Red Sea. After Mart & Ross (1987).

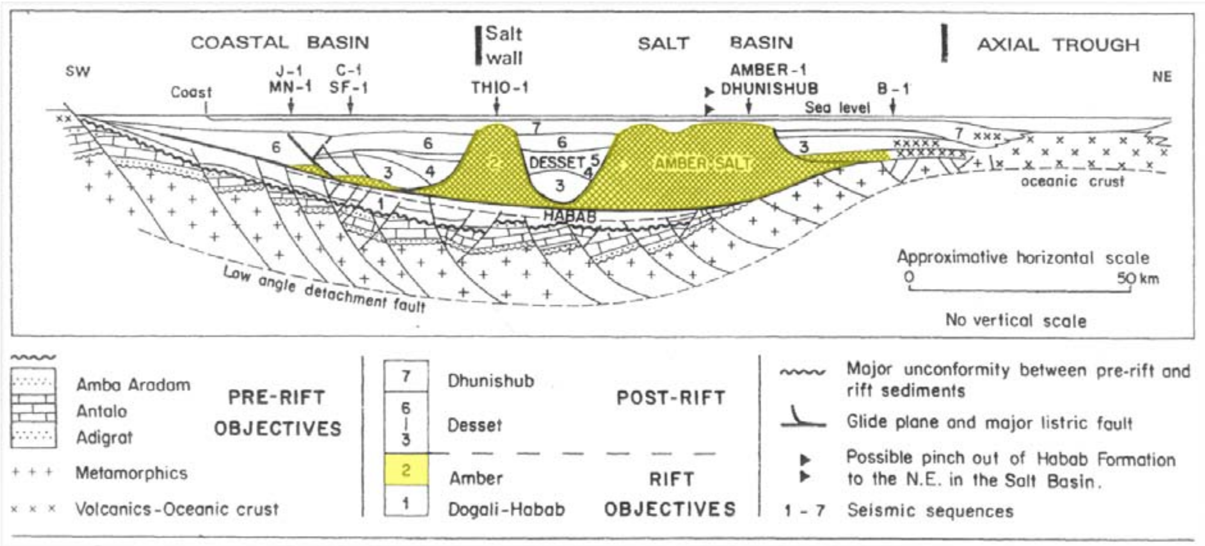

Fig. 6.6. Synthetic section across the Eritrean Red Sea. Salt highlighted in yellow. From Savoyat et al. (1989).

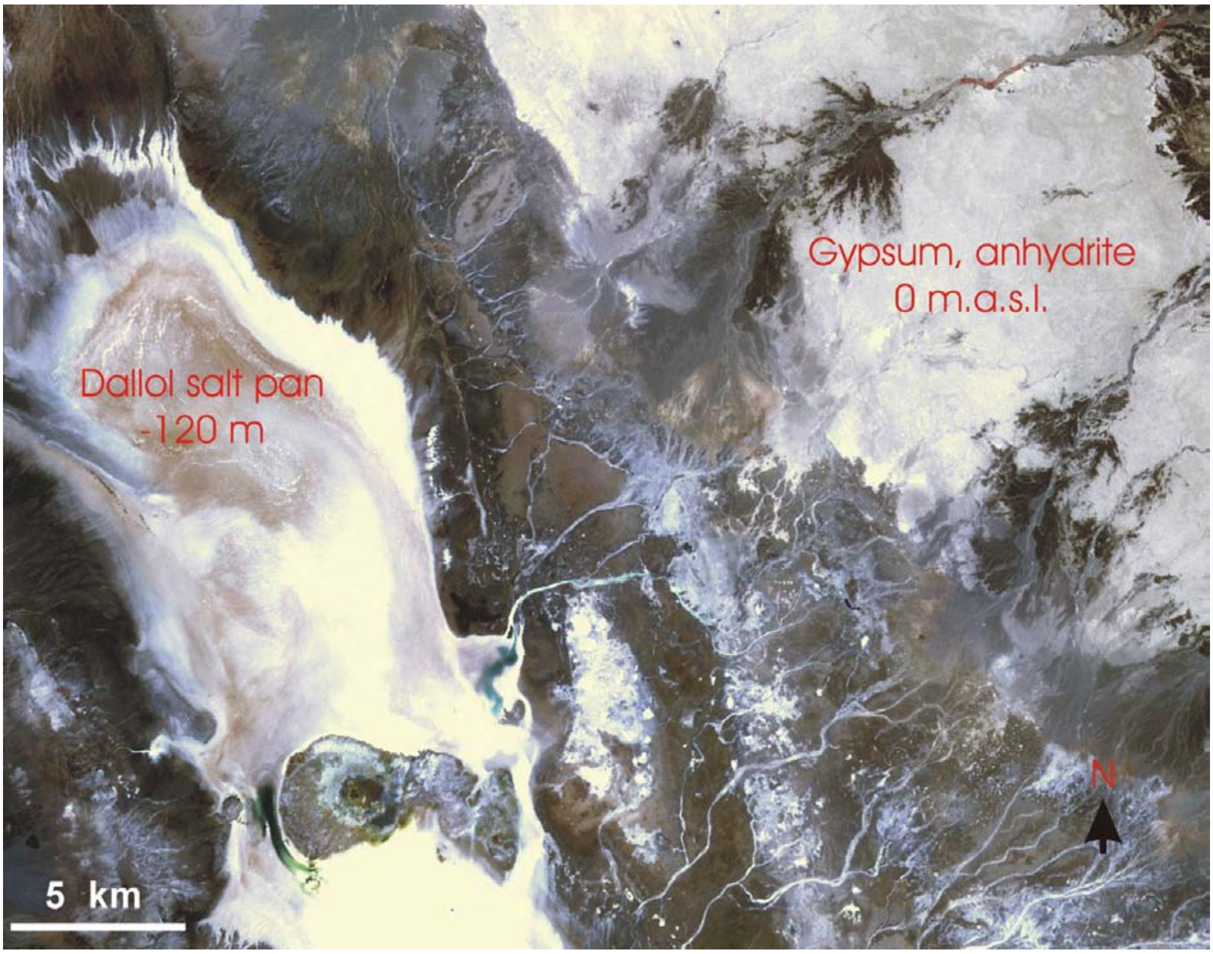

Fig. 6.7. ASTER image of part of the Danakil Depression, NE Ethiopia. The Dallol salt plain (-120m) in the soutwest and gypsum/anhydrite deposits (c. 0 m.a.s.l.) in the northeast. Image parameter: very-near infrared (resolution 15m).

Fig. 6.8. The Dallol salt "volcano", suggested by Talbot (1978) to be formed by a thermal convection cell. Close-up of figure 7.7 to maximum resolution, pixel size 15m. The structure called "Black Mountain" by Holwerda & Hutchinson (1968) is apparently eroded by seasonal waterflow.

Fig. 6.9. Bedded halite in the Dallol "volcano" area. (http://www.educeth.ch/stromboli).

Fig. 6.10. Boiling brine, 100-130^{o}C, spurting from salt mound, precipitating halite and $MgCl_2$. Photo M. Fulle. (http://www.educeth.ch/stromboli).

Fig. 6.11. Active volcanism along the Danakil depression axis. The lava lake at the Erta Ale volcano. 80m deep, diameter 140m. The lava lake is known to have existed for more than 50 years. The Erta Ale volcano reaches nearly 1000m above the Danakil depression floor. (http://www.educeth.ch/stromboli).

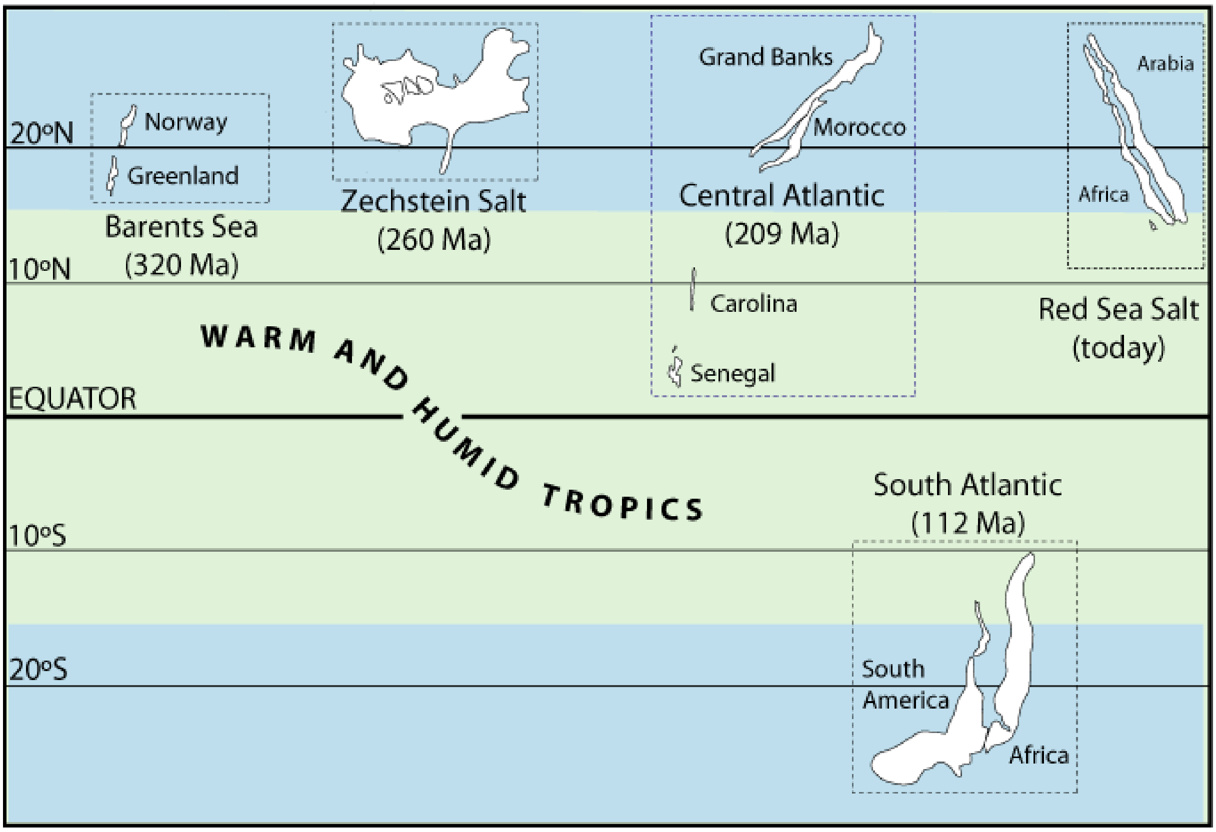

Fig. 7.1 Summary figure, superimposing the palaeopositions of the studied salt basins on single diagram. The salt basin positions are derived from the palaeomagnetically constrained plate reconstructions for each area. Except for Senegal/Guinea Bissau and Carolina Trough salt deposits, all studied deposits formed at subtropical latitudes, consistent with classic views on evaporite deposits.

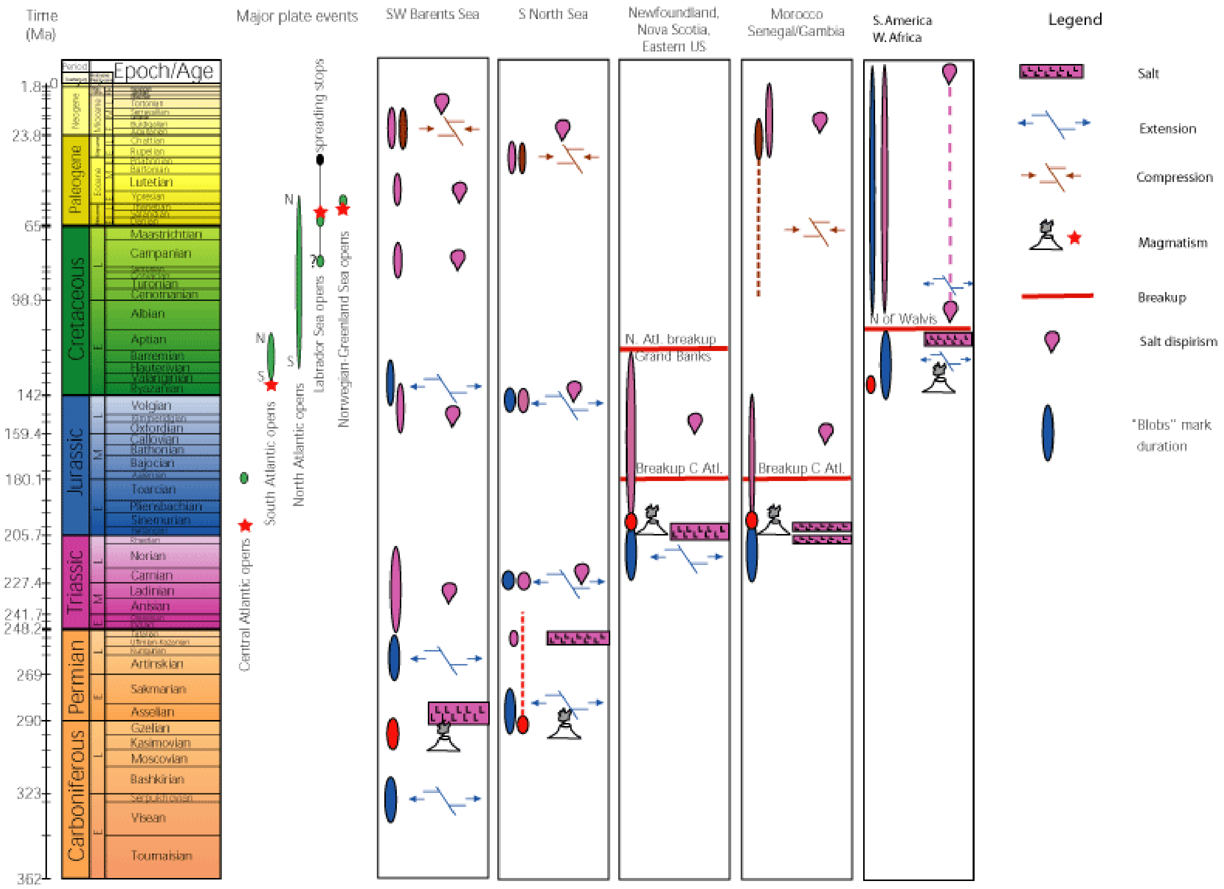

Fig. 7.2. Chronologic diagram, illustrating the relationship between timing of rifting, salt deposition, magmatism, breakup, and salt diapirism.

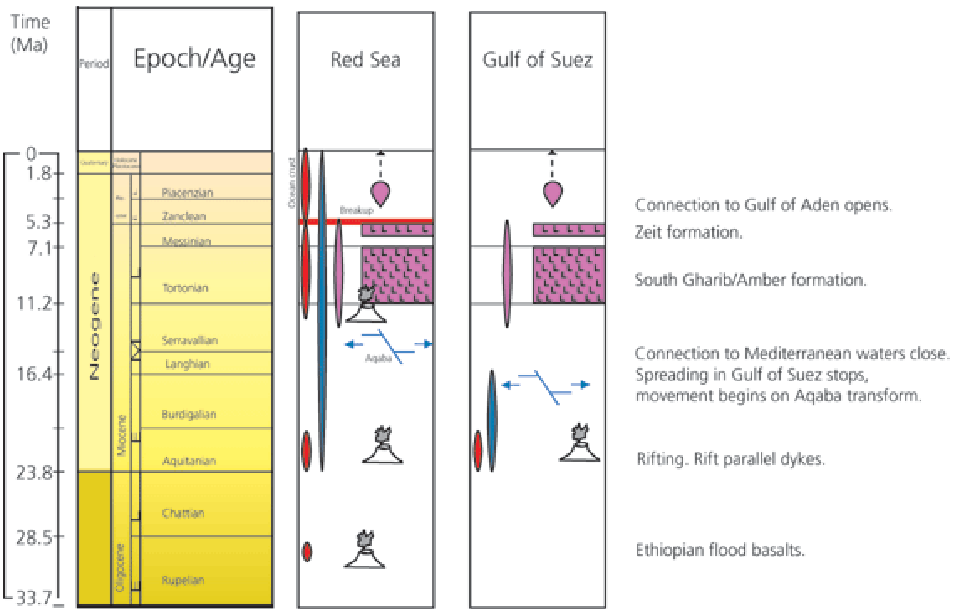

Fig. 7.3. Chronologic diagram for the Red Sea and Gulf of Suez, illustrating the relationship between timing of rifting, salt deposition, magmatism, breakup, and salt diapirism. Symbols as in Fig. 7.2.