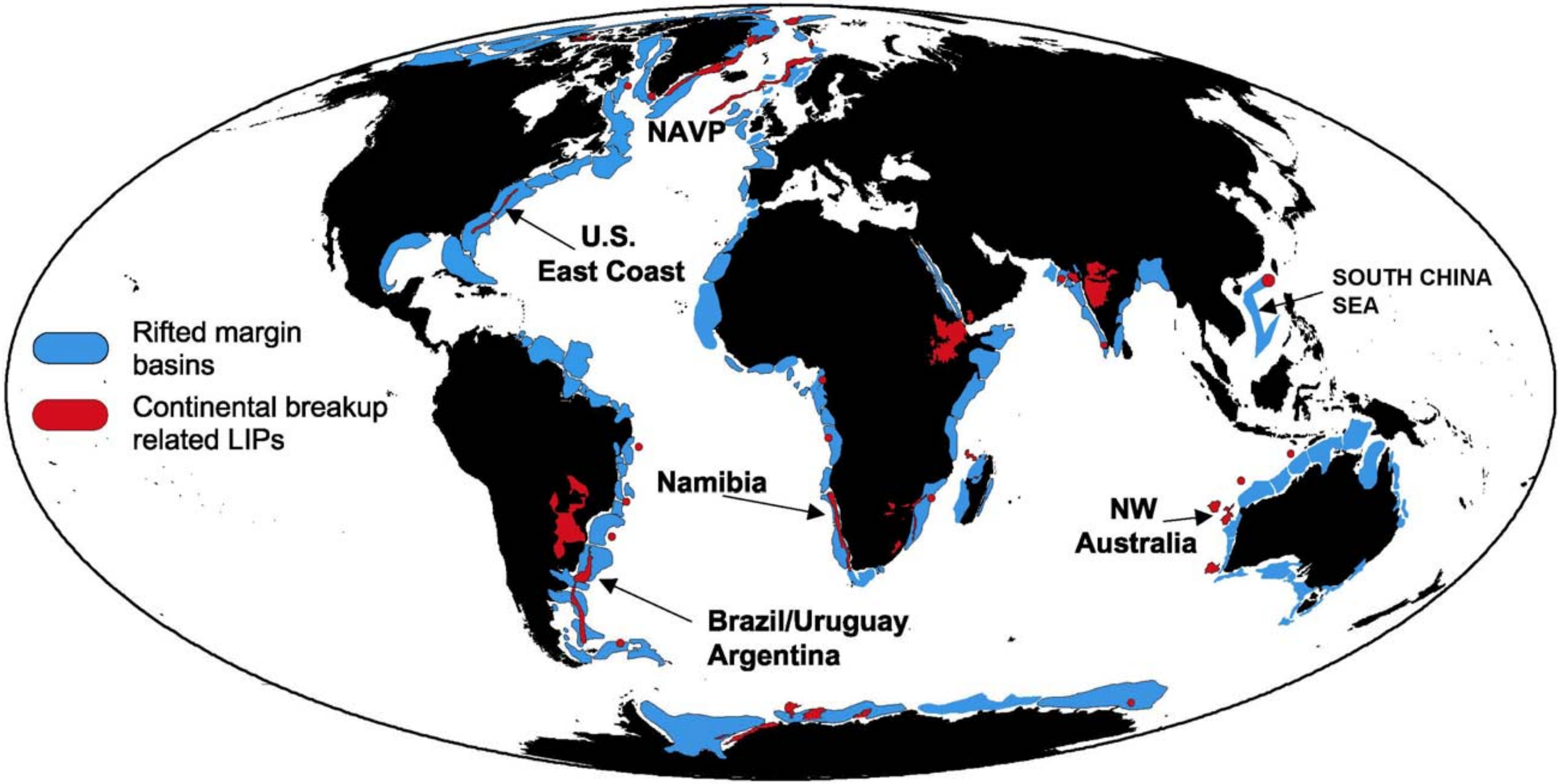
SECTION 2 THE MECHANICS OF SEDIMENTARY BASIN FORMATION

Chapter 3 Basin due to lithospheric stretching



Skogseid (2001), Marine Petroleum Geology, 18.

Reading: Ruppel, C. (1995) Extensional processes in continental lithosphere. JGR, v.100(B12), 24,187-24215.

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

SECTION 2 THE MECHANICS OF SEDIMENTARY BASIN FORMATION

Chapter 3 Basin due to lithospheric stretching

3.1 Introduction to rifts, failed rifts, and passive continental margins
3.2 Geological and geophysical observations in regions of continental extension
- 3.2.1 Rifts
- 3.2.2 Passive continental margins

3.3 Introduction to models of continental extension
- 3.3.1 Active and passive rifting idealizations
- 3.3.2 Postrift subsidence at passive continental margins
- 3.3.3 Dynamical models involving lithospheric strength and rheology

3.4 Uniform stretching of the continental lithosphere
- 3.4.1 McKenzie's (1978a) uniform stretching model
- 3.4.2 Uniform stretching at passive continental margins

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

SECTION 2 THE MECHANICS OF SEDIMENTARY BASIN FORMATION

Chapter 3 Basin due to lithospheric stretching

3.5 Modifications to the uniform stretching model
- **3.5.1 Nonuniform (depth-dependent) stretching**
- **3.5.2 Pure versus simple shear**
- **3.5.3 Protracted rifting and lateral heat conduction**
- **3.5.4 Elevated asthenospheric temperatures**
- **3.5.5 Magmatic activity at passive margins**
- **3.5.6 Induced mantle convection**
- **3.5.7 Radiogenic heat production**
- **3.5.8 Flexural compensation**
- **3.5.9 The depth of necking**

3.6 A dynamic approach to lithospheric extension
- **3.6.1 Generalities**
- **3.6.2 Modes of continental extension: narrow and wide rifts**
- **3.6.3 Forces on the continental lithosphere**
- **3.6.4 Rheology of the continental lithosphere**
- **3.6.5 Numerical experiments on strain rate during continental extension**
- **3.6.6 Insights from analogue models**

SECTION 2 THE MECHANICS OF SEDIMENTARY BASIN FORMATION

Chapter 3 Basin due to lithospheric stretching

3.7 Mantle plumes and igneous activity associated with continental extension

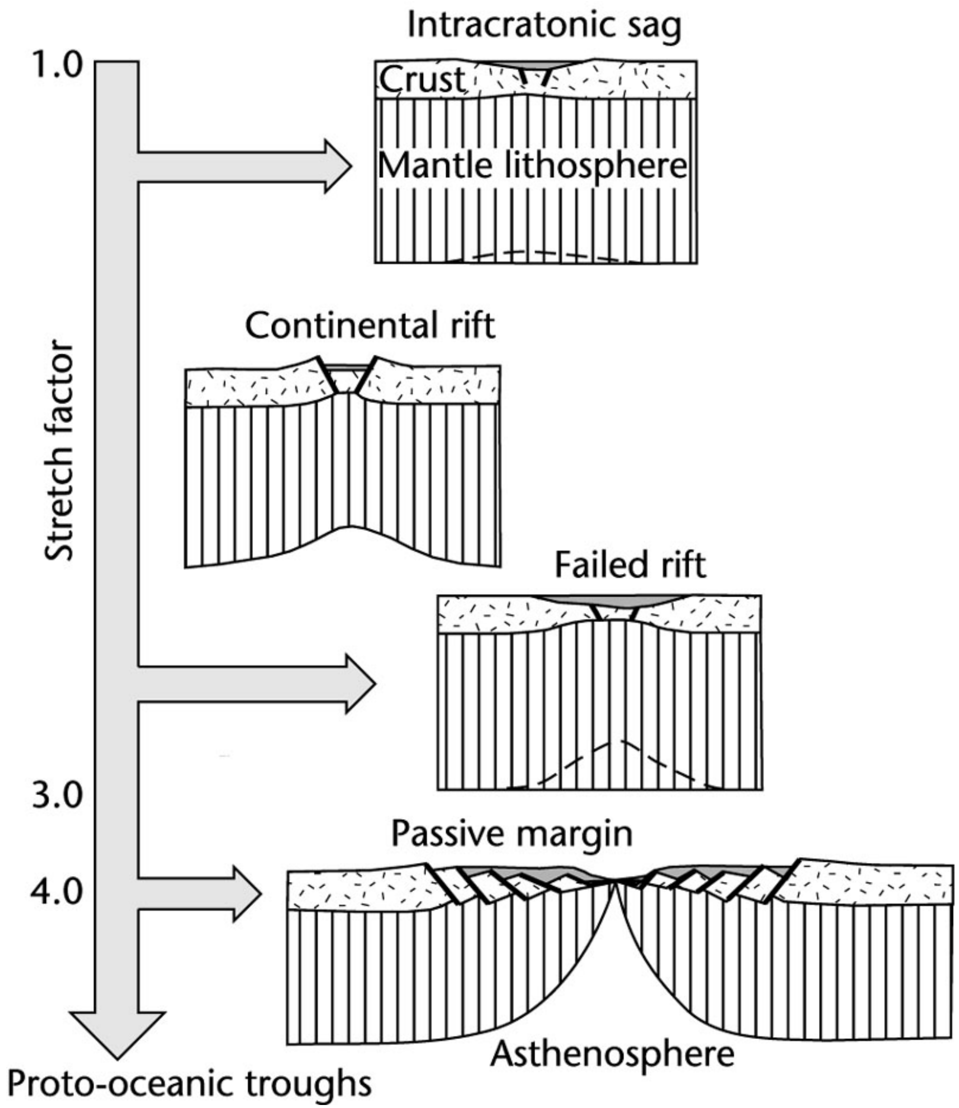
- **3.7.1 Plumes**
- **3.7.2 Melt generation during continental extension**
- **3.7.3 The northern North Atlantic and the Iceland Plume**
- **3.7.4 Low-volume melts in the Mesozoic North Sea Rift**

3.8 Estimation of the stretch factor and strain rate history

- **3.8.1 Estimation of the stretch factor from thermal subsidence history**
- **3.8.2 Estimation of the stretch factor from crustal thickness changes**
- **3.8.3 Estimation of the stretch factor from forward tectonostratigraphic modeling**
- **3.8.4 Inversion of strain rate history from subsidence data**

3.1 Introduction to rifts, failed rifts, and passive continental margin

All rifts, failed rifts, and passive continental margins experience: (1) brittle extension of the crust, causing extensional fault arrays and fault-controlled subsidence, and (2) thermal relaxation following ductile extension of the lithosphere, leading to regional postrift subsidence.



Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

Fig. 3.1 Basins in the rift-drift suite as a function of increasing amounts of continental stretching.

3.2 Geological and geophysical observations in regions of continental extension

3.2.1 Rifts

- **Heat flow**: Rift zones, in general, have heat flows of 90-110 mWm^{-2}. This is a factor of 2 higher than in surrounding unstretched terranes.

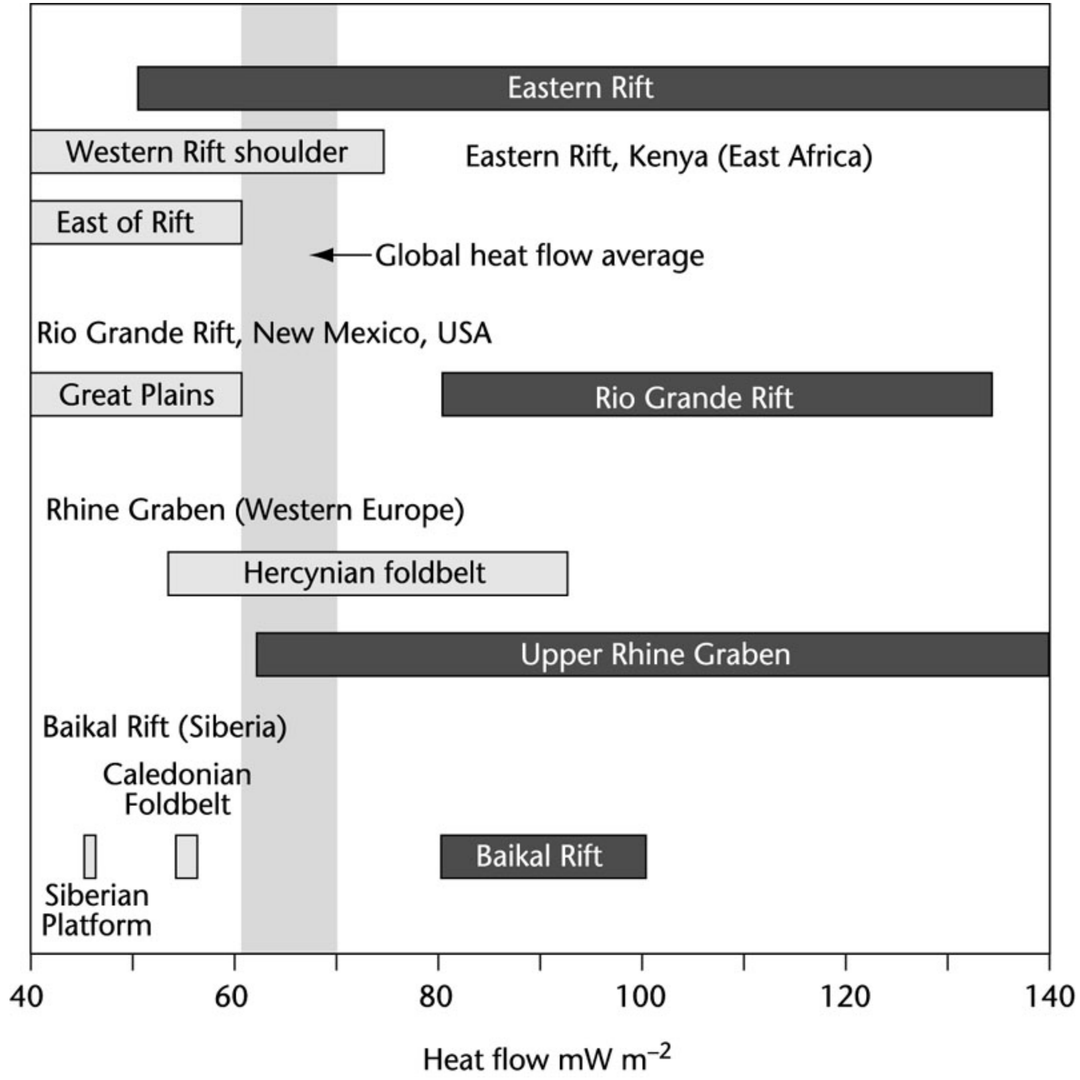


Fig. 3.2 Heat flows in some continental rifts and surrounding regions, compared to the global heat flow average. Dark boxes are rift zones; light boxes are rift flanks or adjacent unstretched regions.

- **Seismicity**: Rift zones are characterized by high levels of earthquake activity. Earthquakes typically have moment magnitudes of up to 5.0 (Rhine Graben) or 6.0 (East African Rift), with shallow focal depths of < 30 km, indicating that the earthquakes are located in the brittle mid-upper crust.

- **Crustal thickness**: Moho is elevated beneath rift zones. Some regions of extensive, diffusive extension such as the Basin and Range, SW USA, are located on previously thickened crust. Another example is the Tibetan Plateau, which is undergoing active extension and overlies crust as much as 70 km thick.

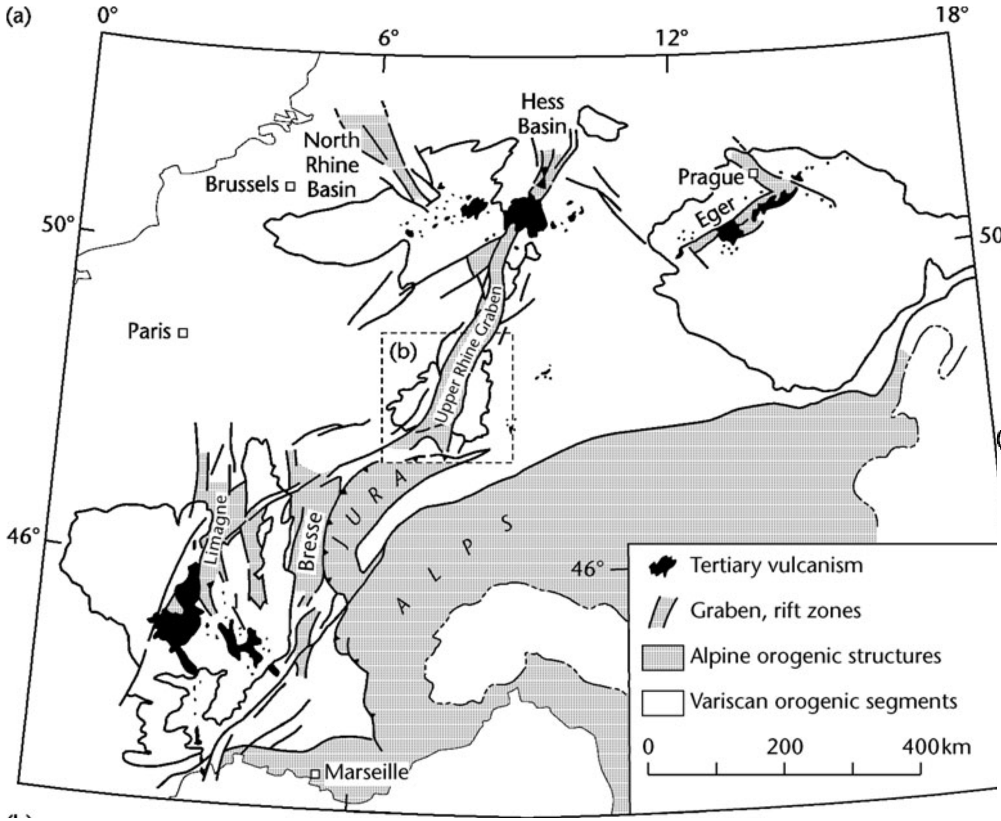
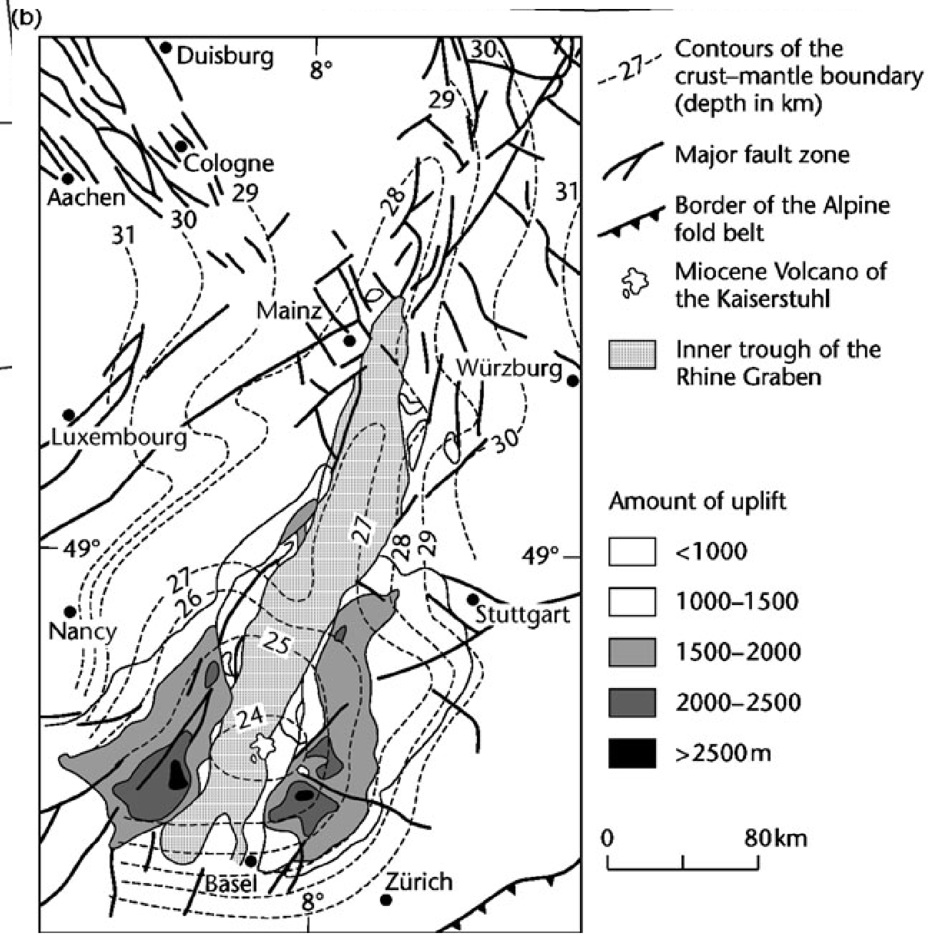


Fig. 3.3 (a) Location of main elements of the late Eocene-Recent Western European Rift System, with sites of Tertiary volcanicity; (b) Depth to the Moho below sea-level (in km), showing a mantle bulge in the southern Rhine Graben centred on the Kaiserstuhl volcano (Illies 1977). The largest amounts of denudation are found on the rift flanks above the shallow mantle.



Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

Reflective lower crust

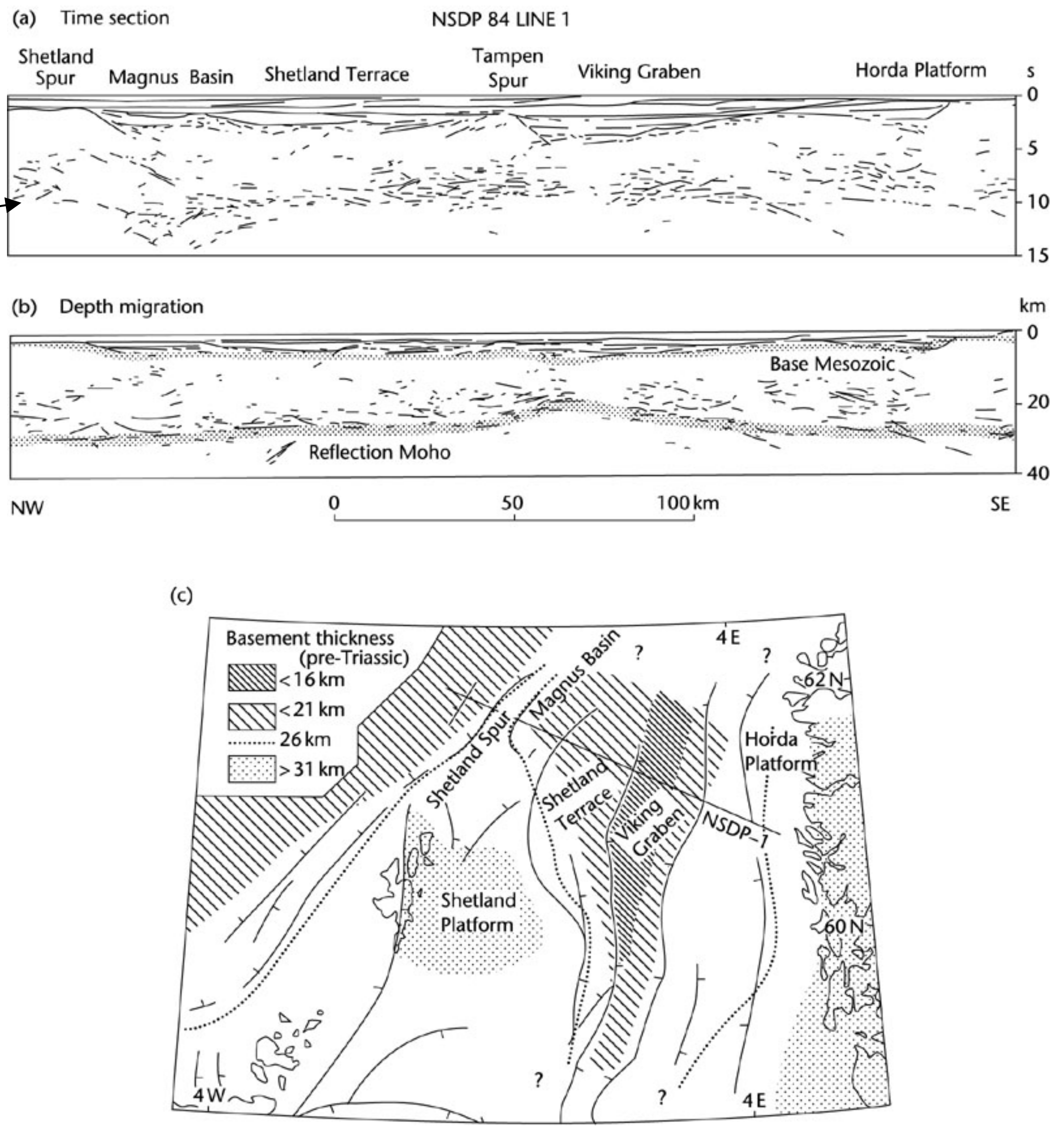


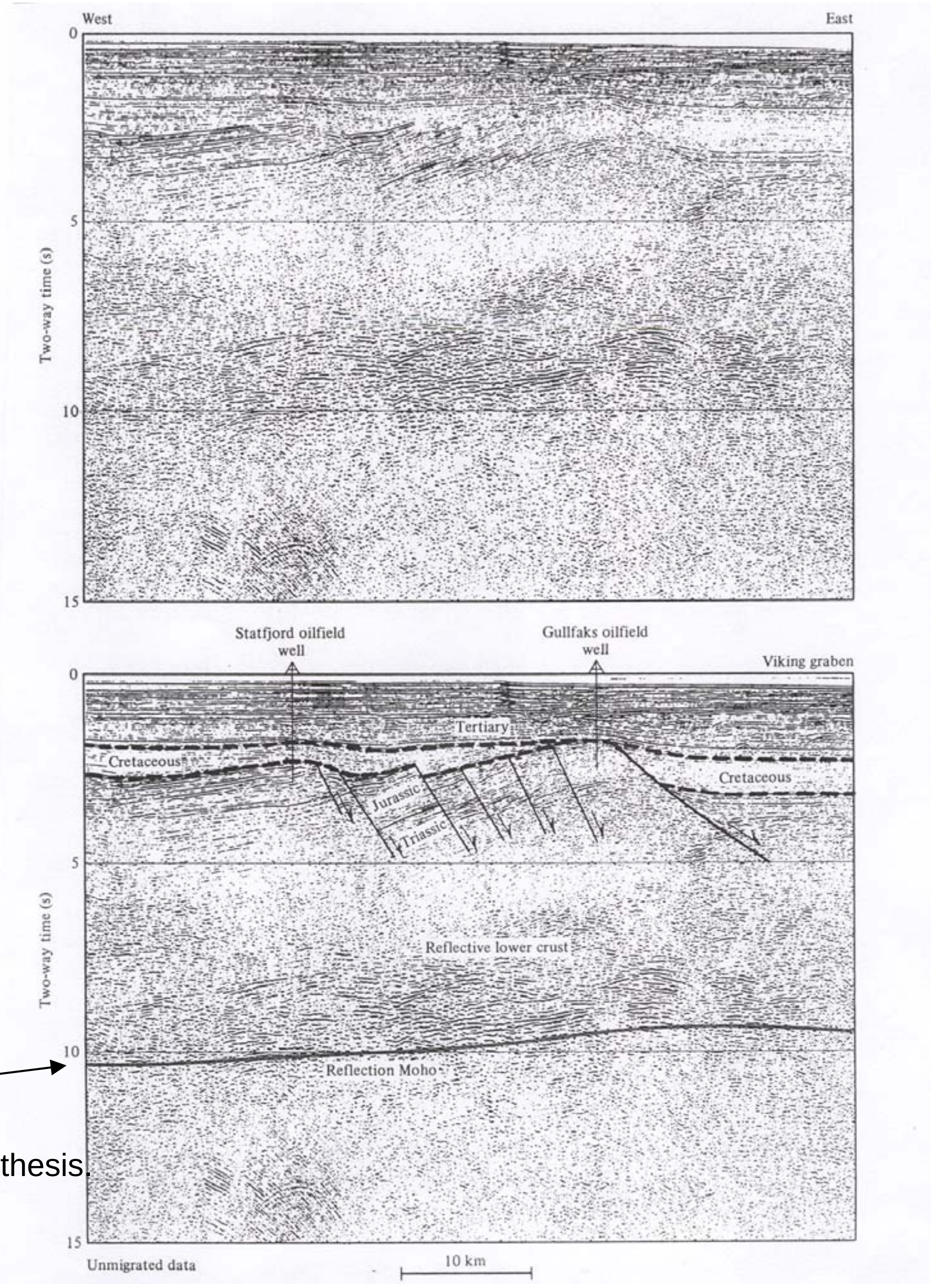
Fig. 3.4 Crustal thickness changes in the North Sea area as a result of Mesozoic rifting (after Klemperer 1988). (a) Unmigrated line drawing in two-way travel time of NSDP line 1 from the Shetland Spur to the Norwegian coast (location in (c)); (b) Depth-migrated version showing the depth of the reflection Moho; (c) Contour map of the interpreted thickness of the prerifting basement (pre-Triassic) showing that the Viking Graben has been stretched by a factor of 2 compared to the Shetland Platform.

Classical example:
Viking Graben
in the North Sea

- Triassic and Jurassic sedimentary strata have been rotated between subparallel high-angle listric faults.

- The faults can be traced deep into the crust and are not bounded by a detachment surface.

- At the time when these faults formed they reached a depth at which the crust was hot enough to deform by ductile flow, causing sub-horizontal reflections in the lower crust.



Moho

Figure 13.10 GECO profile across the northern North Sea, uninterpreted and interpreted data, showing steeply dipping normal faults (domino-style faulting) in the upper crust.

Klemperer & Peddy (1992) in Understanding the Earth: A New Synthesis. Brown, Hawkesworth & Wilson (eds), p.265

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

- **Gravity**: Rift zones typically have a long wavelength Bouguer gravity low with sometimes a secondary high located in the center of the rift zone. The conventional explanation is that rift zones have anomalously hot material in the mantle beneath the rift, producing a mass deficit and therefore a negative gravity anomaly. The subsidiary gravity high is thought to be due to the intrusion of dense magma bodies within the continental crust.

- **Faults**: normal dip-slip faults predominates with a variable number of strike-slip faults depending on the orientation of the rift axis in relation to the bulk extension direction. Most major border faults dip steeply inwards towards the basin center and are planar as far as they ca be imaged. However, some rift bounding faults are low-angle and listric, taking up very large amounts of horizontal extension, such as in the supradetachment basins of SW USA. Metamorphic rocks may be unroofed from < 25 km depth in these “core complexes”.

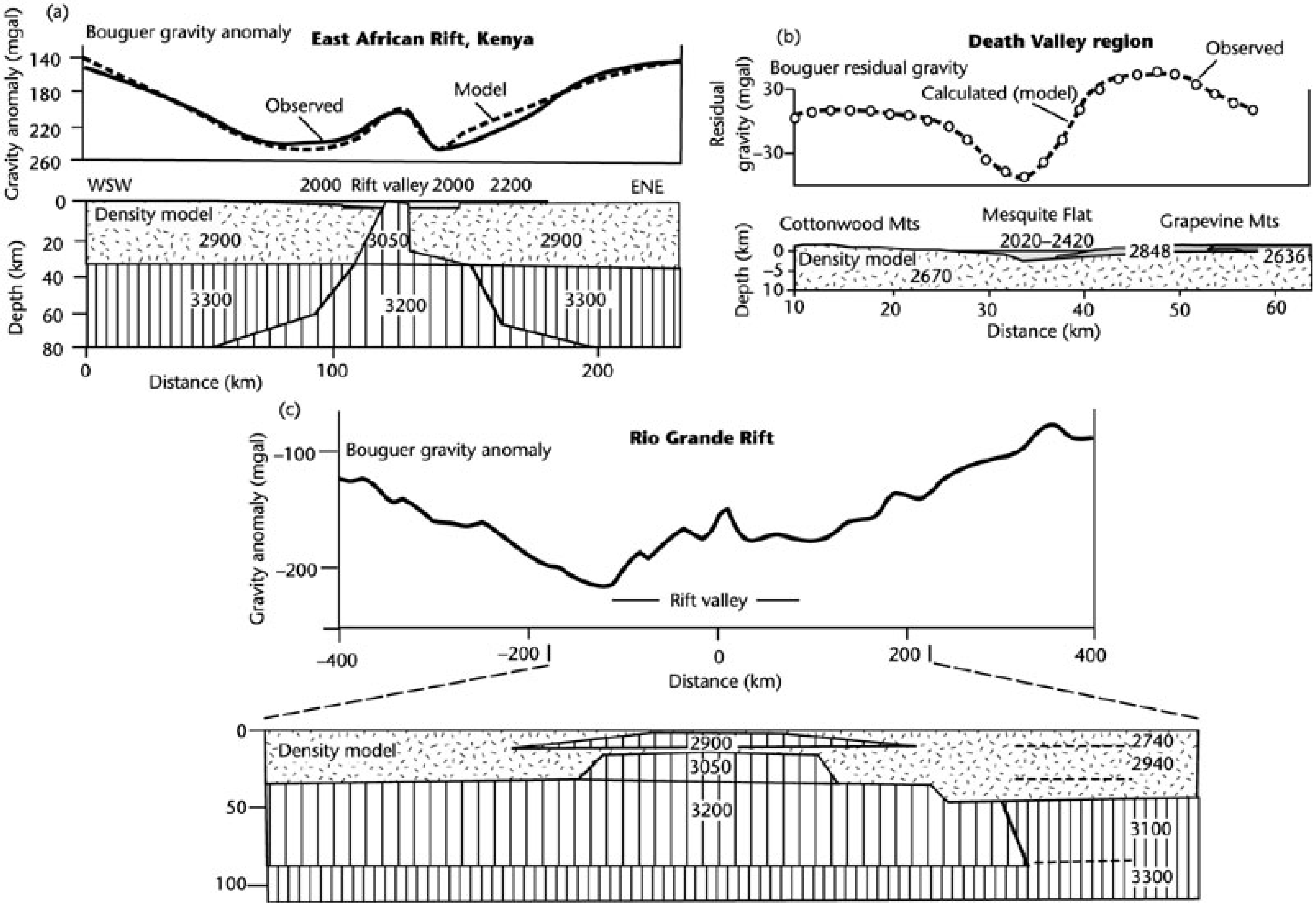


Fig. 3.5 Gravity profiles across rift zones. (a) Gravity profiles and density model across the Gregory Rift, Kenya. The secondary gravity high is modeled as due to the intrusion of dense magma bodies beneath the rift valley (after Baker and Wohlenberg 1971); (b) Gravity profile and density model for a profile across Mesquite Flat, northern Death Valley, California (after Blakeley et al. 1999); (c) Gravity profile (c.33°N) and density model for the Rio Grande Rift of New Mexico (after Ramberg 1978). The secondary gravity high is thought to be due to the presence of dense igneous bodies beneath the rift. Densities shown in kgm^{-3} .

- **Topography**: Active rift zones typically have elevated rift flank topography bordering a depositional basin.

Two scales of uplift:

(a) Long length scale (several 100s km: Like the > 3 km-high topographic swells of the Ethiopia and East Africa.

(b) **Smaller length scale** (< 100 km): linear rift flank uplifts associated with border fault arrays. Like the < 1 km-high highlands bordering the Gulf of Suez; southern Rhine Graben, where tectonically driven exhumation of the rift flank has resulted in 2-3 km of erosion. Regions of extensive, diffusive extension are associated with plateau-type topography, such as the Basin and Range, USA (rifting due to shallow subduction of hot oceanic lithosphere) and especially Tibet (rifting due to thickening of continental lithosphere).

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

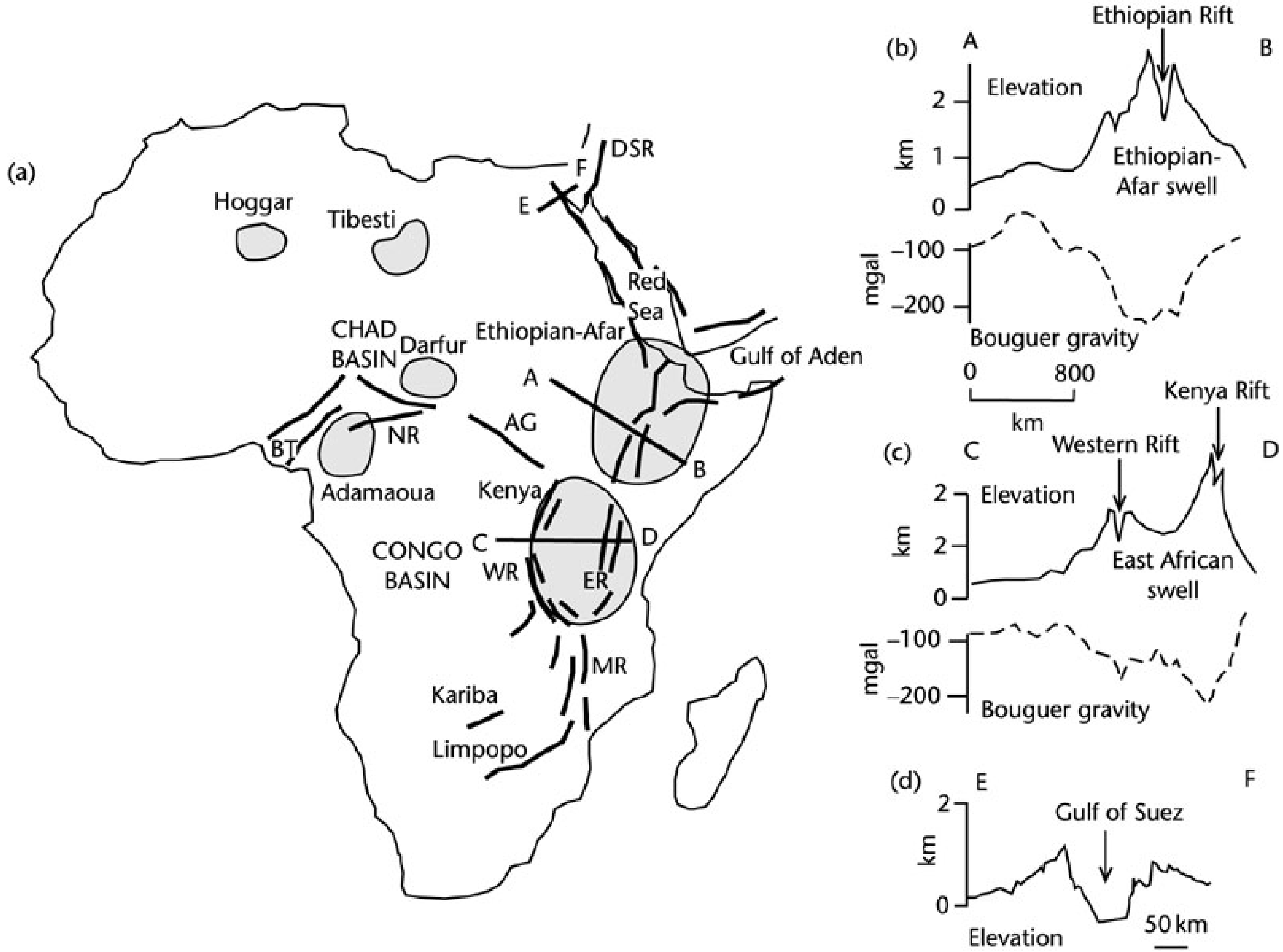


Fig. 3.6 (a) The major domal uplifts of Africa (Afar and East African domes) are due to uplift over hotspots in the mantle. Also shown are the smaller topographic uplifts of central Africa, and the main rift system: AG, Abu Gabra Rift; MR, Malawi Rift; ER and WR, Eastern and Western Rifts; NR, Ngaoundere Rift. Topographic and Bouguer gravity profiles across the Afar (A—B) and East Africa (C—D) swells are shown in (b) and (c). (d) Topographic profile along E—F showing rift flank uplift along the Red Sea and Gulf of Suez. After Edinger et al.(1989).

Time scale and amount of extension

Two families of basins, with different strain rate, total extensional strain (or stretch factor β), and the dip of master faults:

1. **Narrow rifts**: Discrete continental rifts located on normal thickness crust (such as the Rhine Graben, Baikal Rift, Rio Grande Rift) extend slowly (<1 mm yr-1) over long periods of time (10 to > 30 Myr), with low total extensional strain (generally < 10 km). Master fault angles are steep (45-70°). Seismicity suggests that crustal extension takes place down to mid-crustal levels. At higher strain rates, narrow rifts may evolve through increased stretching into passive margins.
2. **Wide rifts**: Supradetachment basins occur within wide extended domains with thickened crust. They typically extend quickly (<20 mm yr-1) over short periods of time (5-12 Myr) with a high amount of total extensional strain (10-80 km). Master faults (detachments) are shallow in dip (10-30°), but may have originated at higher angles. Local anomalies in the ductile lower crust are amplified to produced core complexes.

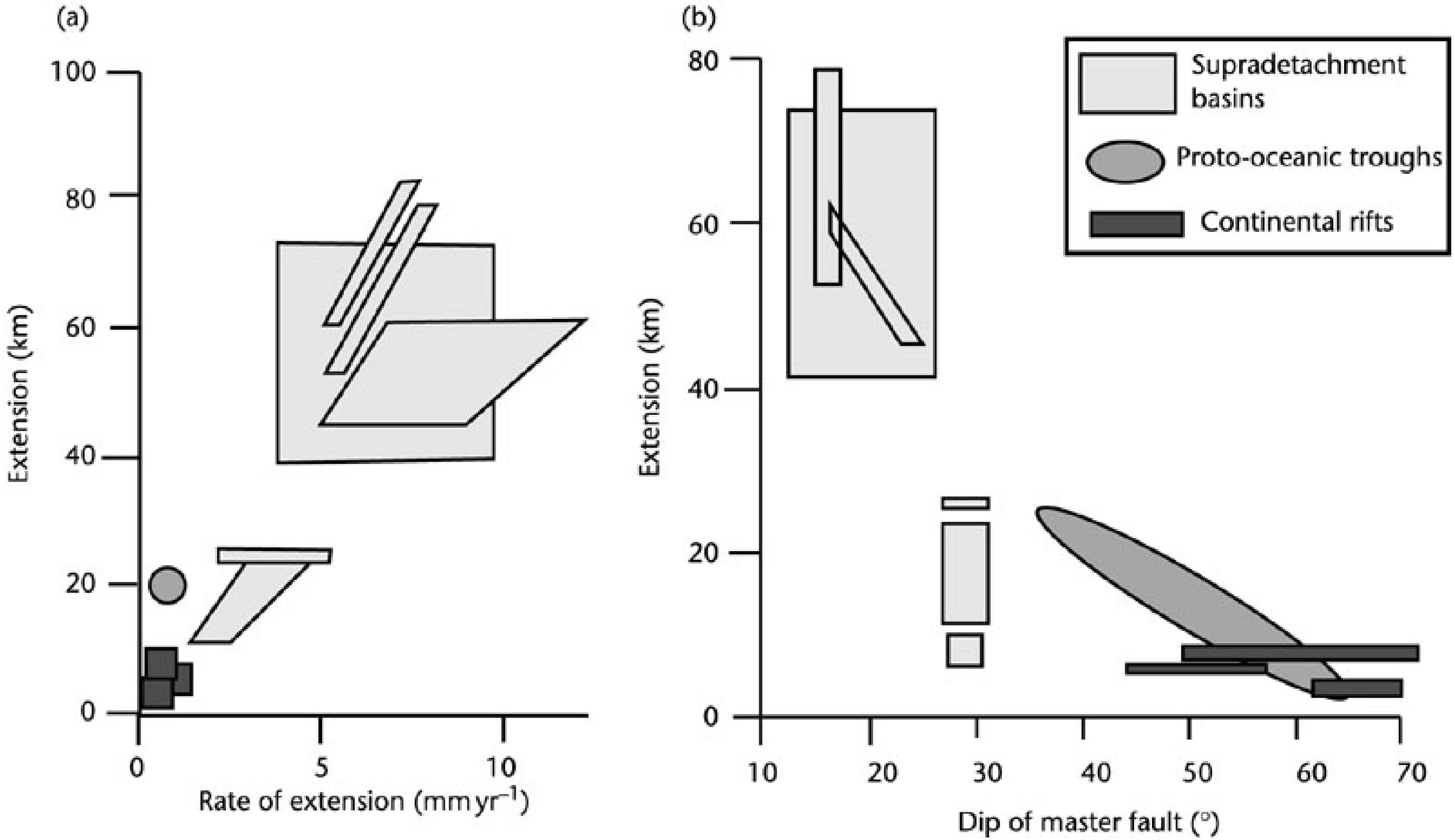
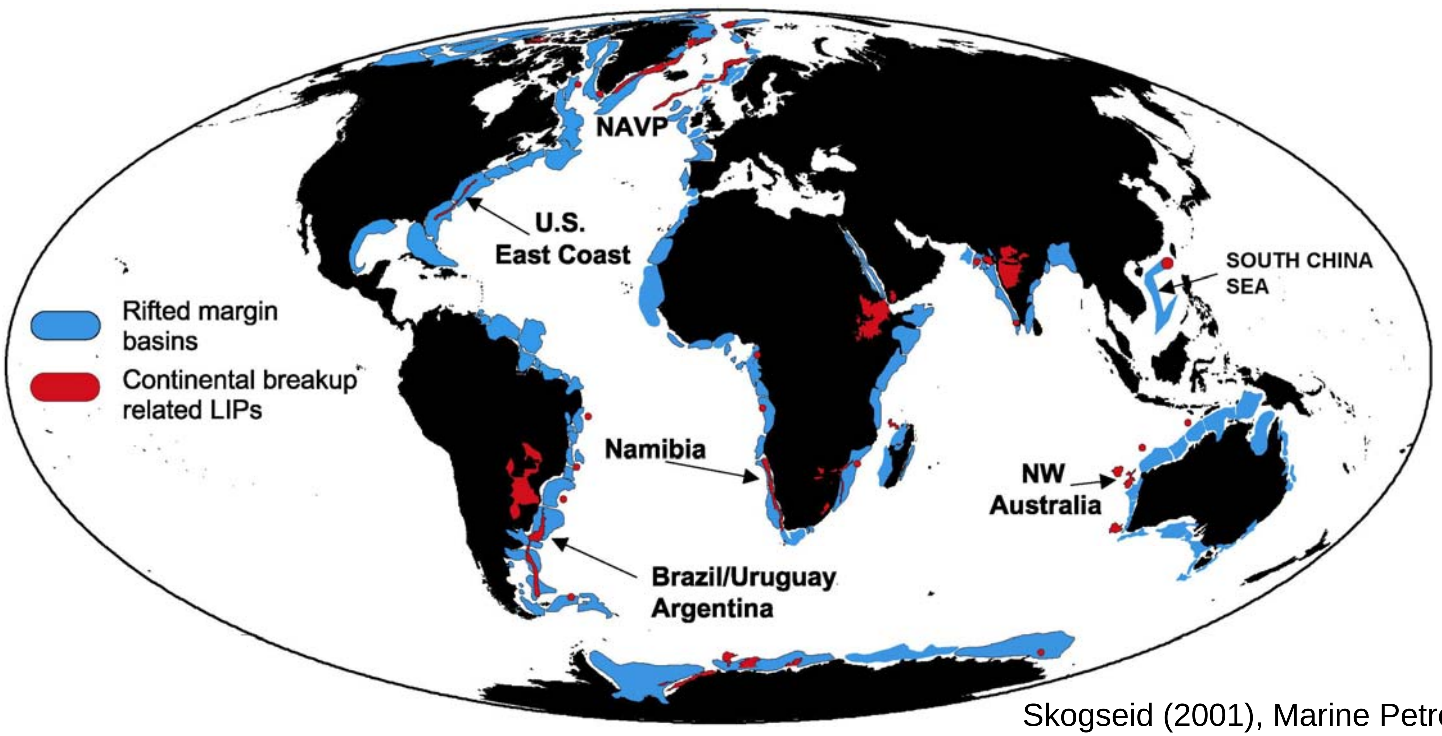


Fig. 3.7 Rift, supradetachment basins, and proto-oceanic troughs in terms of their rate, total extensional strain, and dip of master faults, based on Friedmann and Burbank (1995).

3.2.2 Passive continental margins



Skogseid (2001), Marine Petroleum Geology, 18.

Passive margin: A continental margin which is not also a plate margin. Such margins are also known as “aseismic margins” or “Atlantic-type margins” and are contrasted with active margins.

Table. 3.1 Conjugate margins of the Atlantic

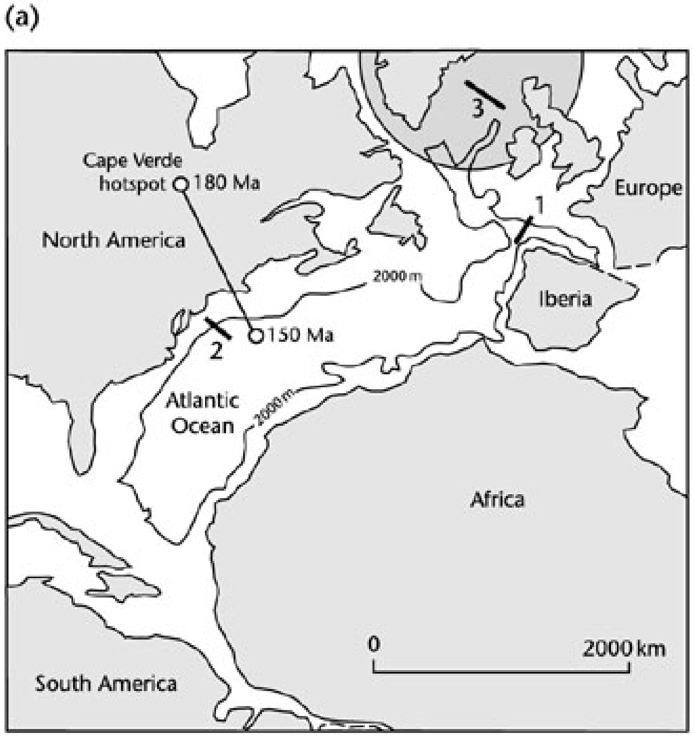
Western margin	Eastern margin	Start of main rifting and duration
Southern Grand Banks	Iberia/Galicia	Valanginian (137 Ma) 15–25 Myr
Flemish Cap	Goban Spur	Barremian (127 Ma) 15–20 Myr
Labrador	SW Greenland	Barremian (127 Ma) 40–65 Myr

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

Two end-members of passive margins based on sediment thickness:

◆ Starved margins (2-4 km thick): North Atlantic European margin.

◆ Nourished margins (generally 5-12 km): North Atlantic US margin.



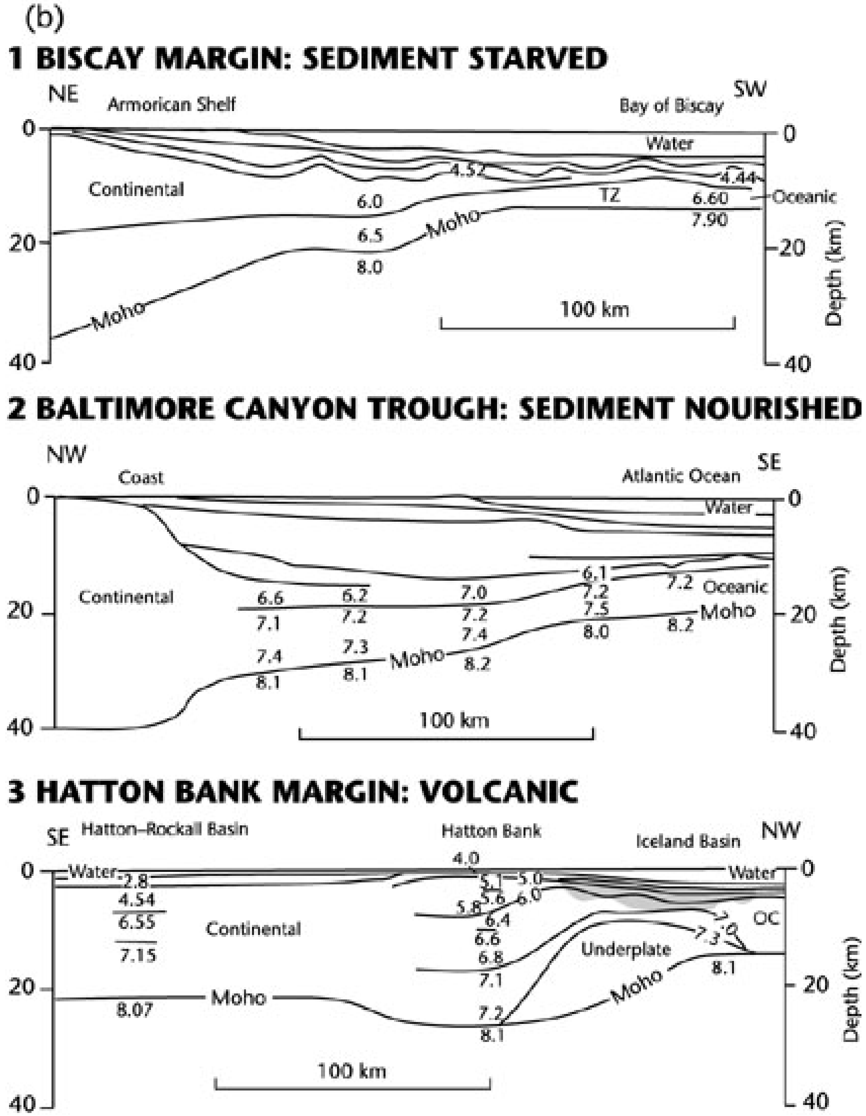


Fig. 3.8 Volcanic, sediment-nourished, and sediment-starved margins (after White and McKenzie 1989). (a) Location of margins in the central-north Atlantic region on a Middle Jurassic reconstruction (170Ma), shortly after the onset of seafloor spreading; (b) Biscay margin, which is sediment starved; (c) Baltimore Canyon Trough margin, which is thickly sedimented; (d) Hatton Bank margin, which is characterized by important magmatic activity. Shaded area shows extent of extrusive basalts. Moho is overdeepened due to presence of igneous underplate. TZ, ocean-continent transition zone; OC, ocean crust.

Passive continental margin of the eastern USA

Flex-up

Flex-down & Thermal subsidence

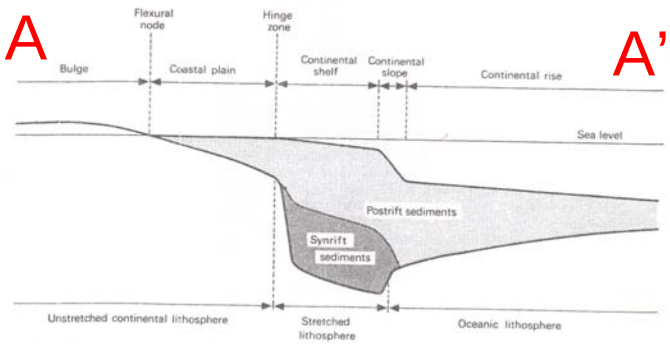


*Fig. 4.14 The major structural and morphological elements of passive (rifted) continental margins. Synrift sediments are those deposited in a rift prior to continental rupture, whereas postrift sediments date from the initiation of sea-floor spreading. The **hinge zone** marks the line along the margin dividing continental lithosphere of normal thickness from lithosphere which has been thinned and stretched during rifting and subsequently subsided under the load of overlying sediments. The **flexural node** separates the (oceanward) portion of the margin subject to subsidence from that experiencing uplift. (Modified from A. B. Watts (1982) Nature 297, Fig. 2, p. 470.)*

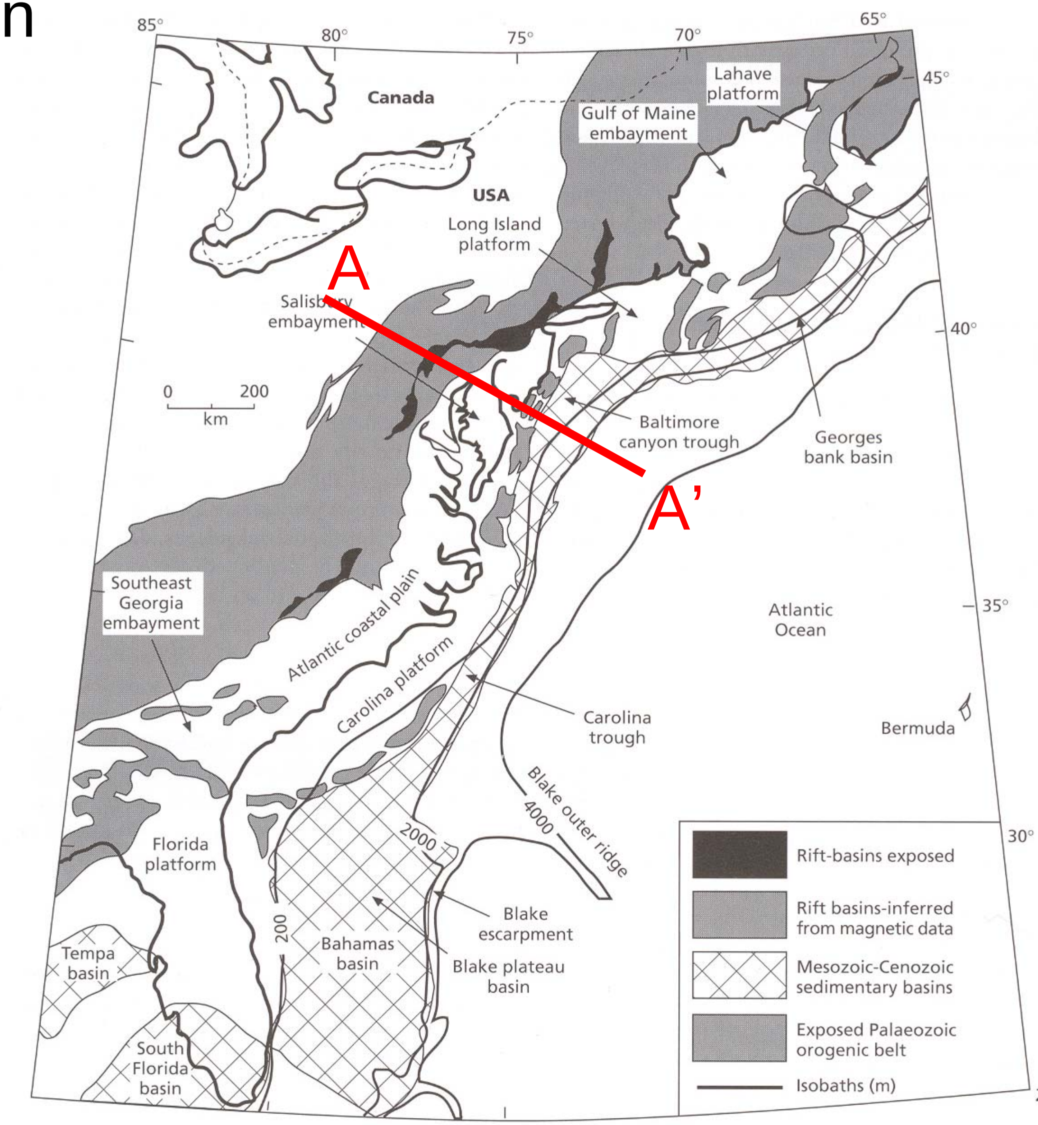


Fig. 28.8 Map of the eastern United States passive continental margin and adjacent coastal plain and ocean floor. This is really a *locus typicus* for the full spectrum of passive-margin sedimentary environments, all juxtaposed over a zonal transect of some 20° of latitude. (From Sheridan & Grow, 1988.)

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

A cross-section of the NE Atlantic margin of the USA

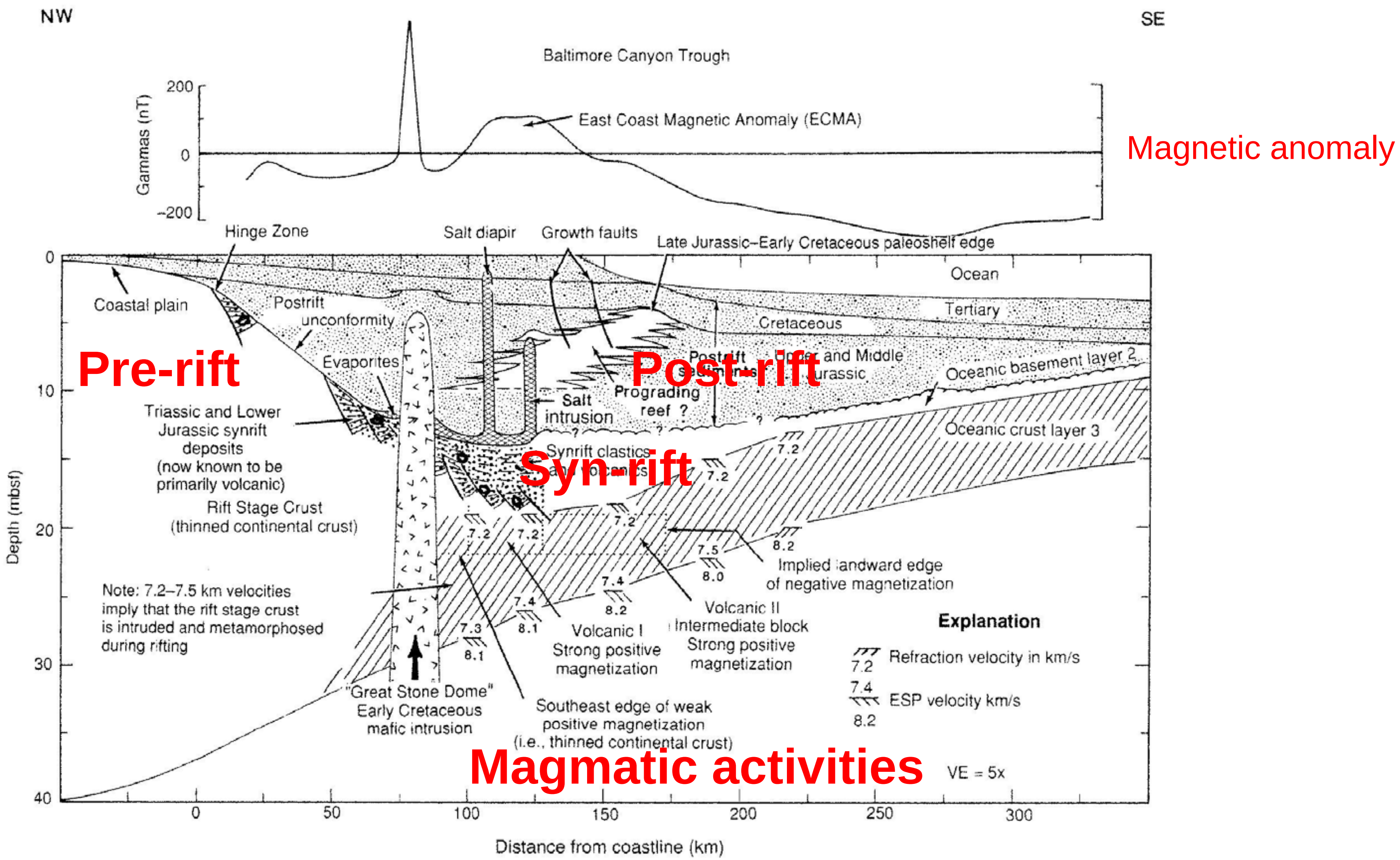
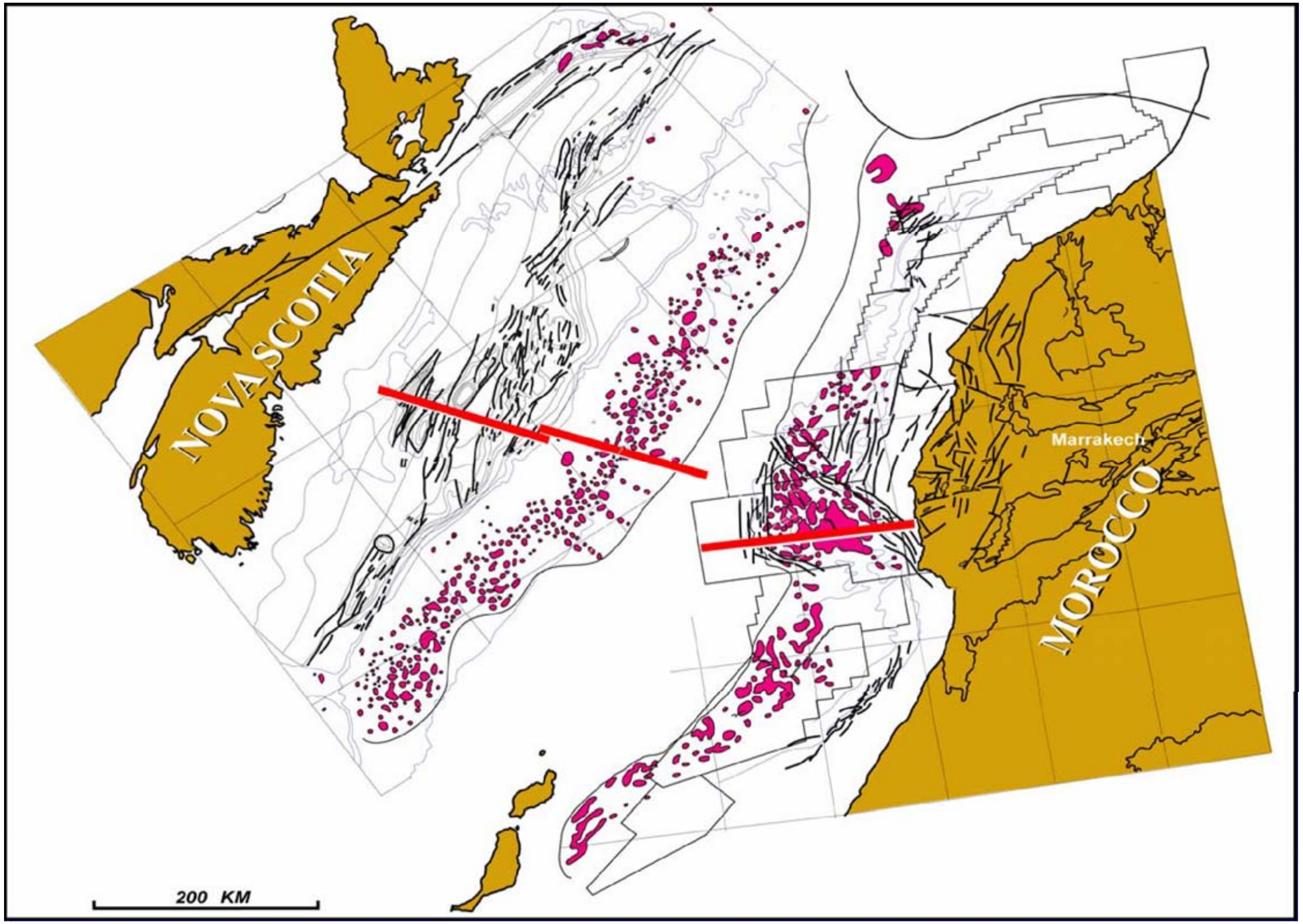


Figure 1. Schematic cross-section from the New Jersey Coastal Plain crossing the continental shelf, slope, and rise in the vicinity of Leg 174A (after Grow and Sheridan, 1988). The wedge labeled šyn rift clastics and volcanics"i s now known to be composed primarily of volcanic rocks (Sheridan et al., 1993).

from Christie-Blick et al. (1998)

Conjugate margins: Original matching margins on either side of the ocean (Nova Scotia vs. Morocco)



Source: Molnar et al. (2002) Correlation of syn-rift structural elements across the central Atlantic between Morocco and Nova Scotia, AAPG Lecture Series.

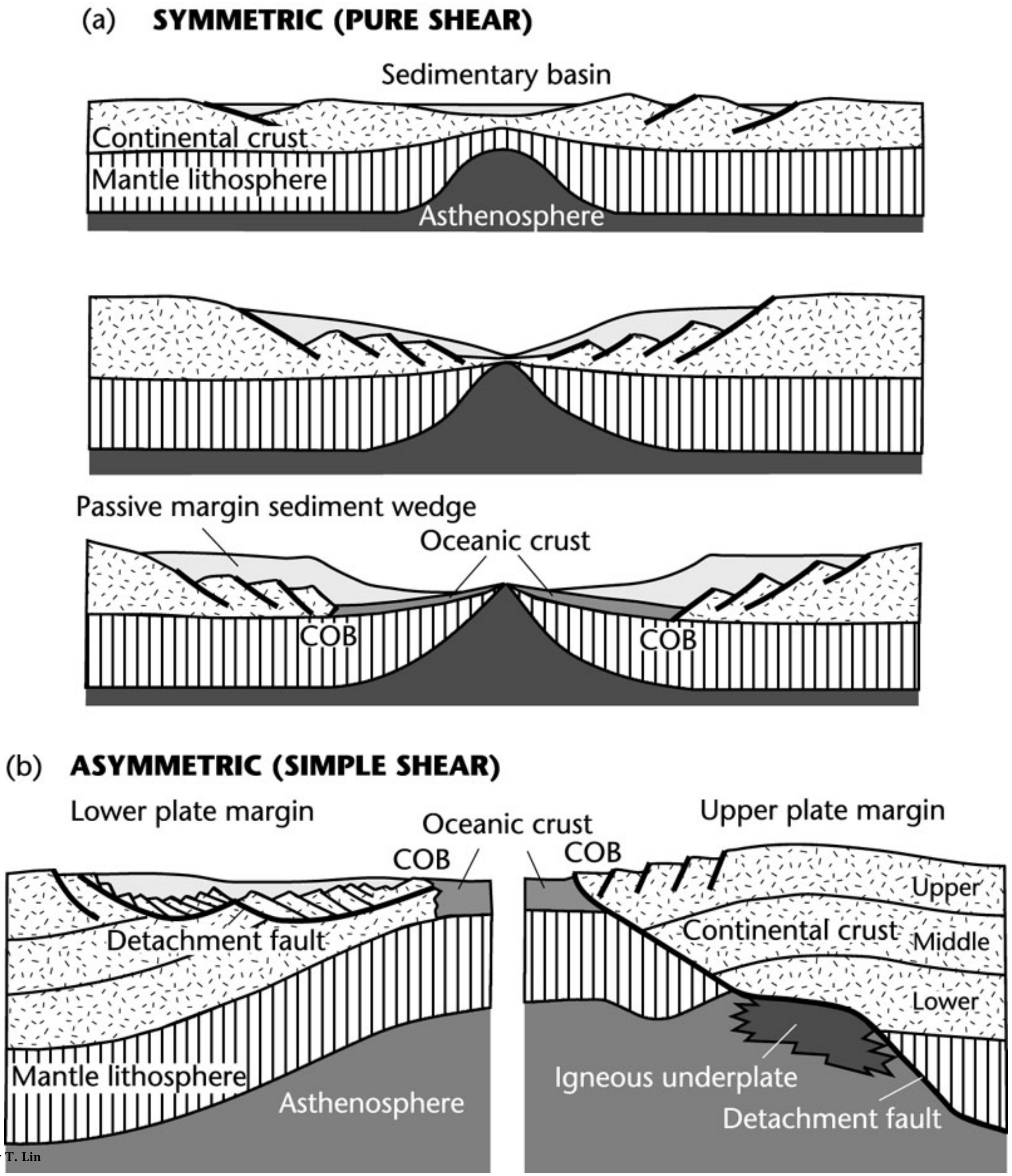


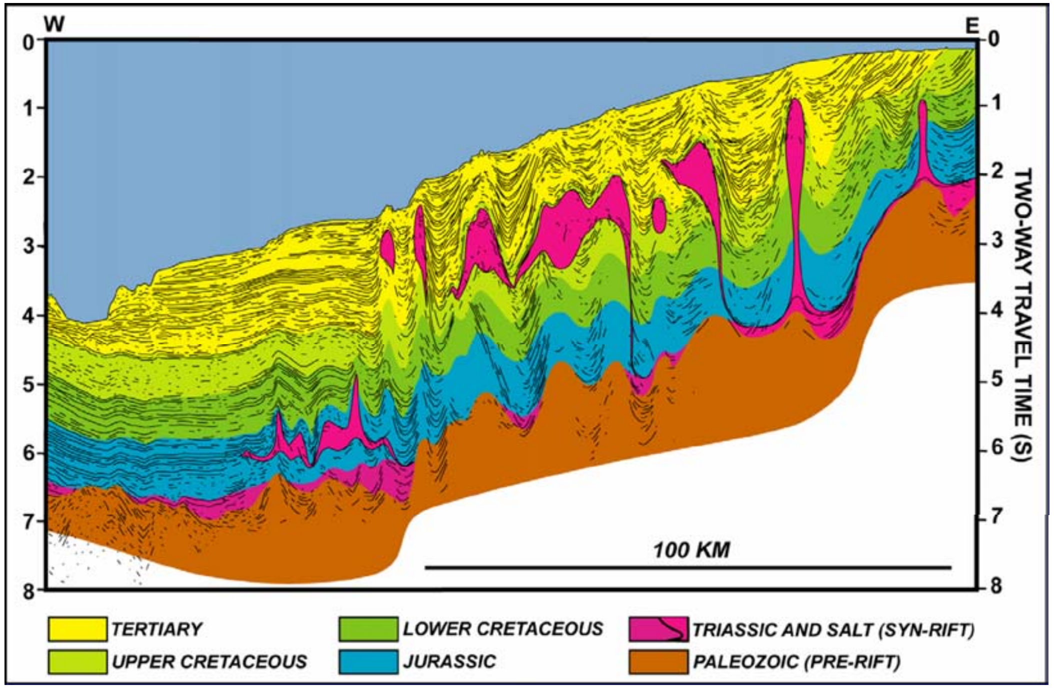
Fig. 3.9 Conjugate margins based on deep seismic information (after Lister et al.1986; Louden and Chian 1999). (a) Symmetric margin (pure shear), and (b) asymmetric (simple shear) with a lithospheric detachment fault. COB is ocean-continent boundary.

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

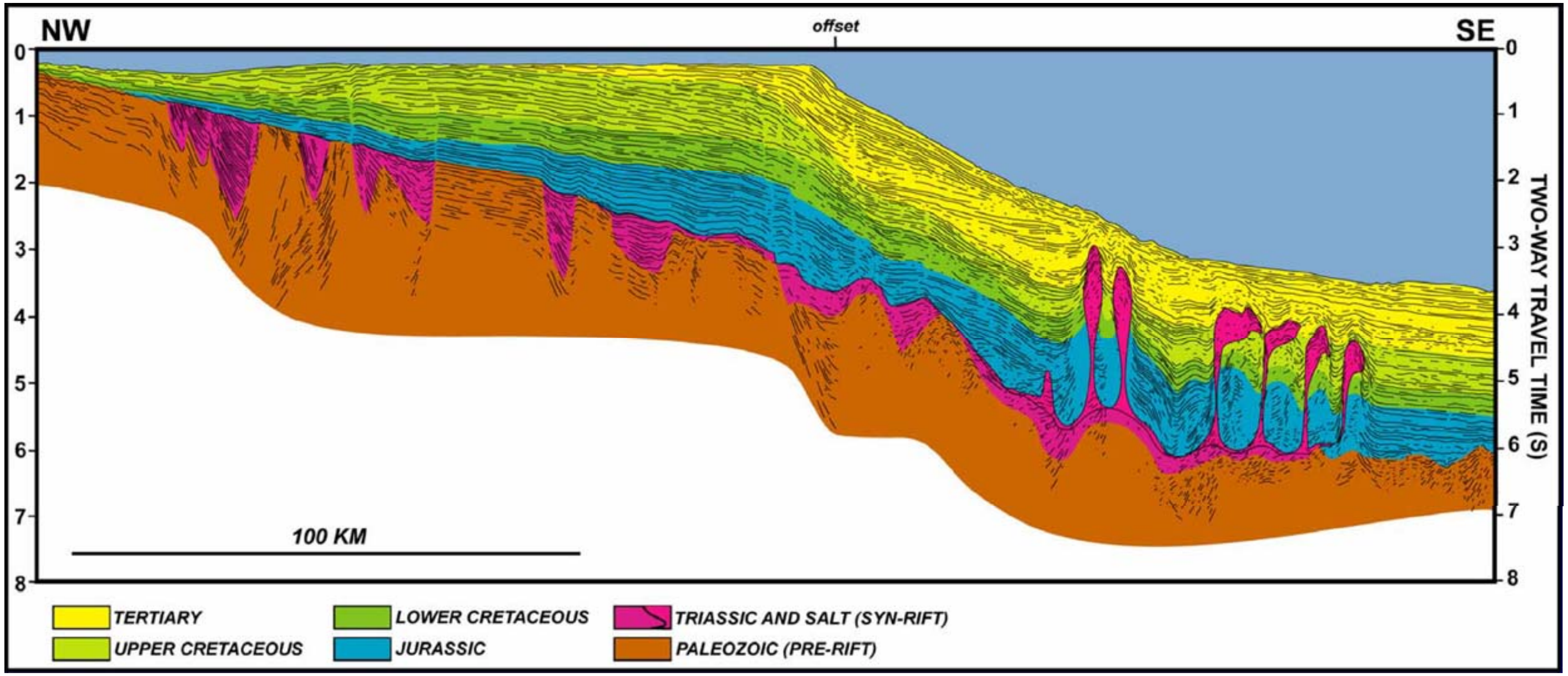
Regional seismic transects

Molnar et al. (2002)

Morocco

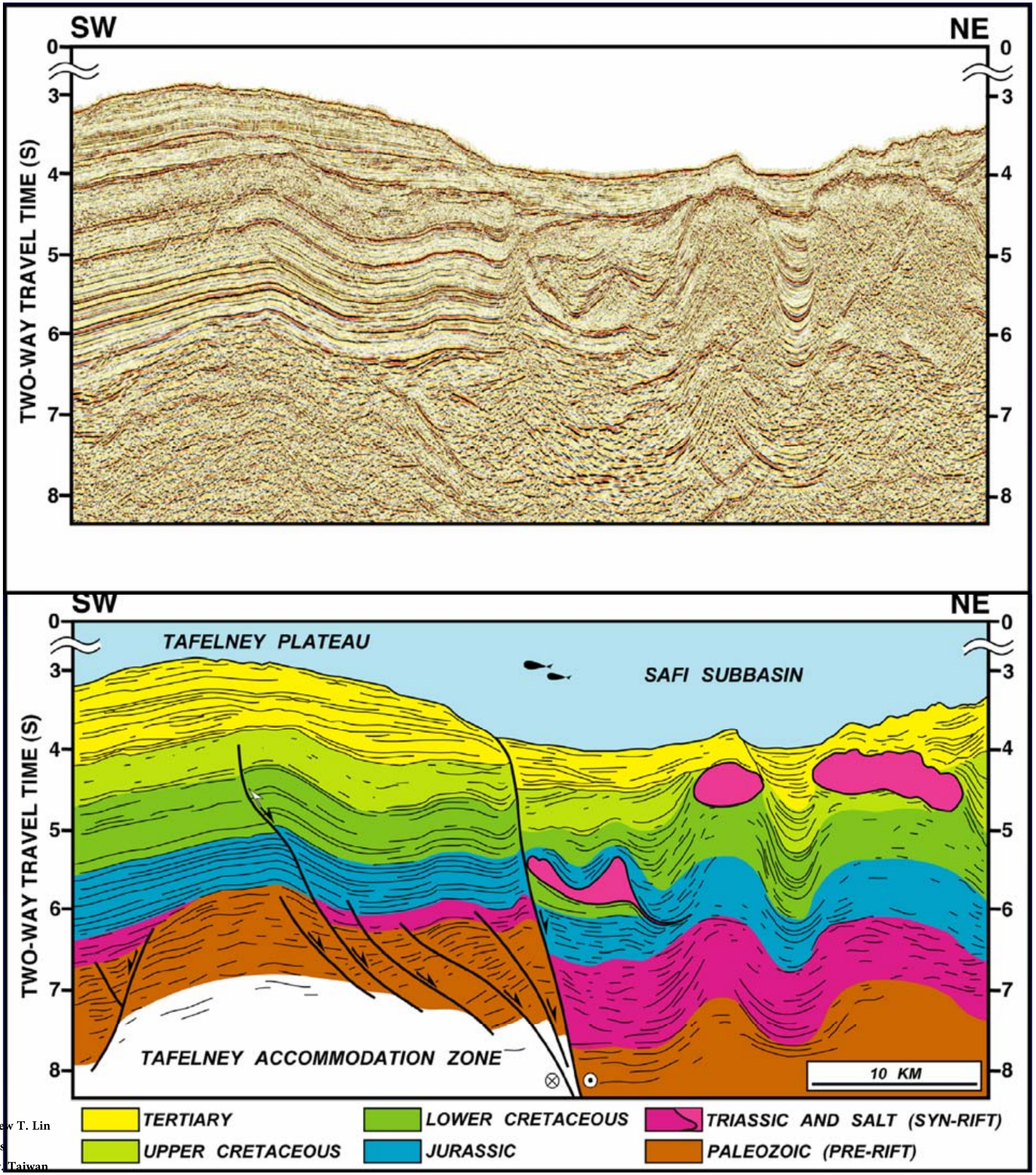


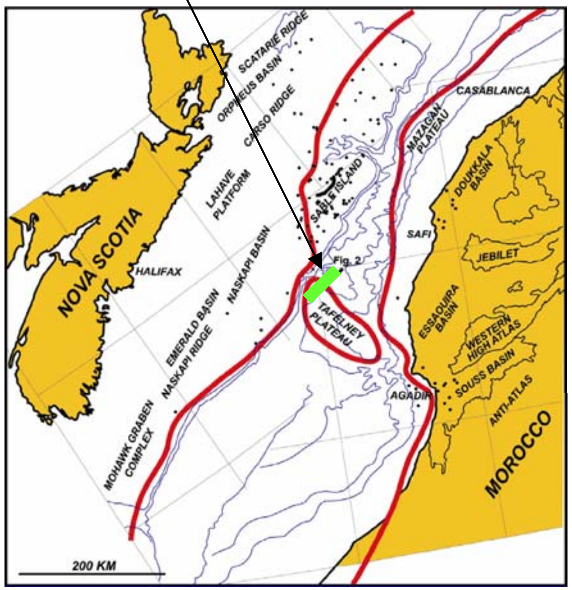
Nova Scotia



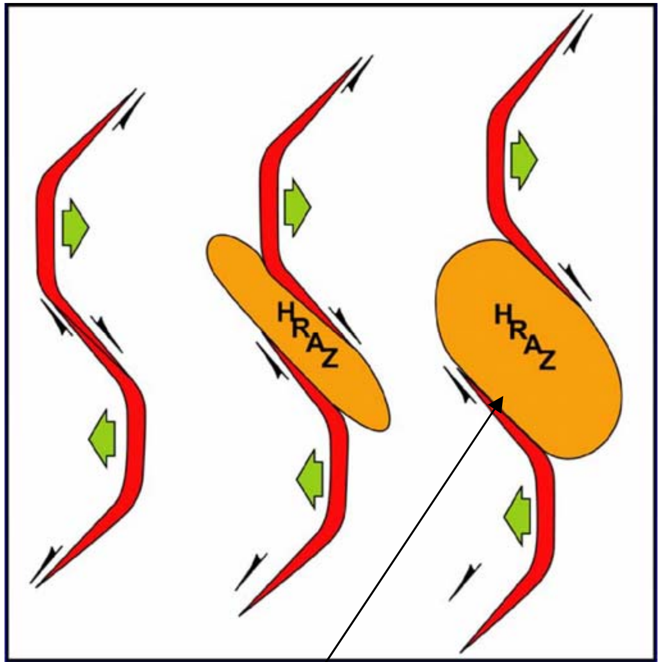
Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

Molnar et al. (2002)

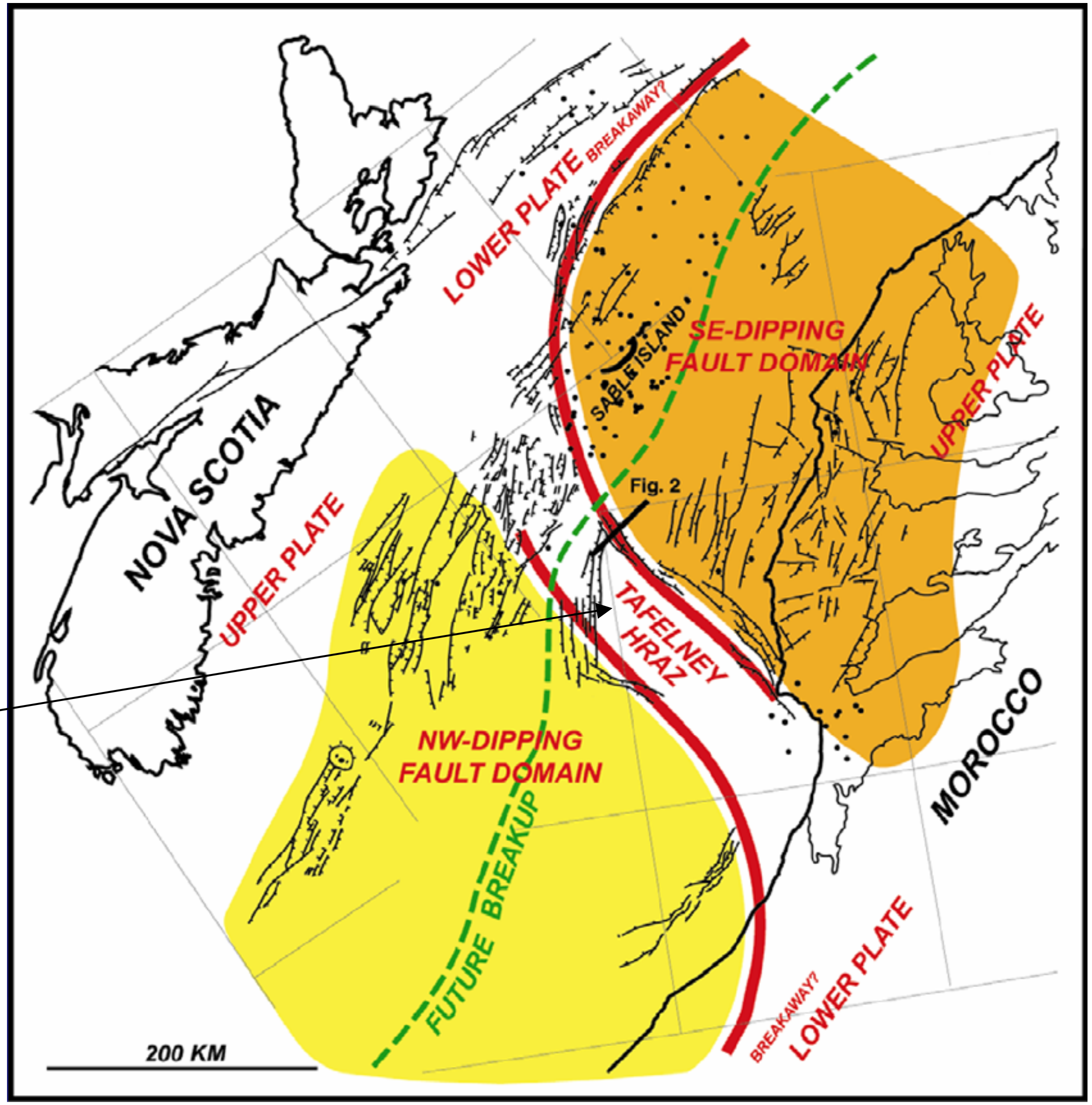




Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan



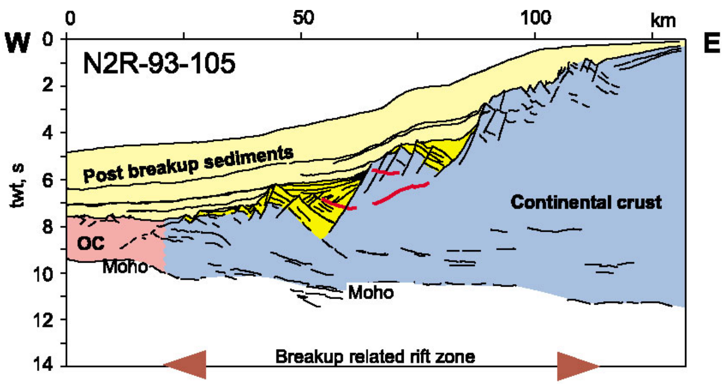
HRAZ: High Relief Accommodation Zone



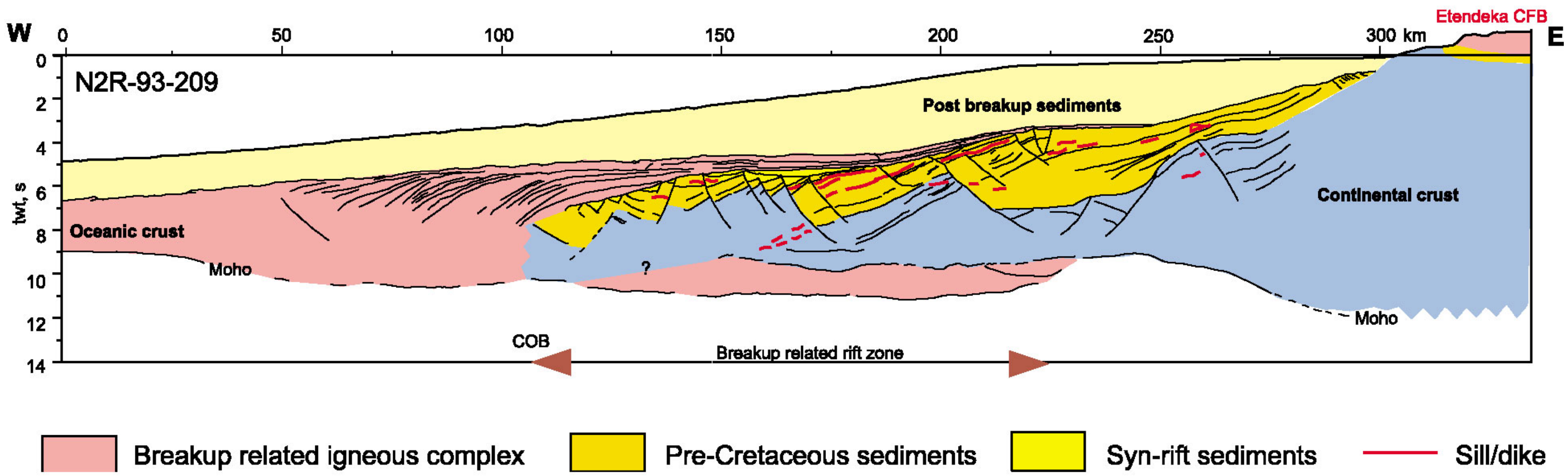
Molnar et al. (2002)

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

Crustal transect across the northern Namibian margin



Skogseid (2001), Marine Petroleum Geology, 18.



Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

Passive margin stratigraphy

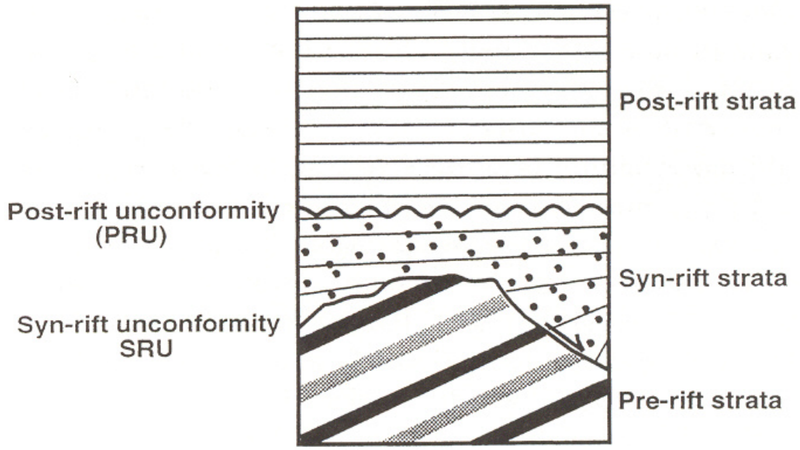


Figure A2.2 Terminology of rift-basin stratigraphy used in this chapter.

Passive margins are characterized by rifted and rotated blocks of usually thick sedimentary sequences.

The stratigraphy of a passive margin consists of pre-rift, syn-rift, and post-rift successions. The pre-rift and syn-rift strata are separated by the syn-rift unconformity whereas the syn- and post-rift strata are divided by the post-rift (or break-up) unconformity.

From: Bosence (1998) in: Sedimentation and Tectonics in Rift Basins. Rurser & Bosence (eds), p. 11.

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

Pre-rift sections

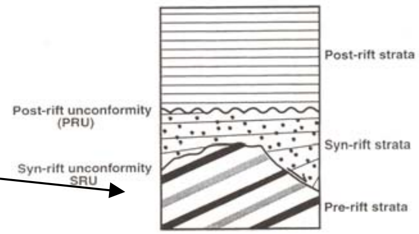


Figure A2.2 Terminology of rift-basin stratigraphy used in this chapter.

- The pre-rift section is totally unrelated to the subsequent rift phase and can be of any lithology.
  The pre-rift section is faulted during rifting and theoretically has a better chance of preservation when the rift basin was initiated as a sag rather than as an arch.
- The top of the pre-rift section is usually marked by an angular unconformity (syn-rift unconformity) that marks the onset of rifting and is visible on seismic sections.

Syn-rift sections

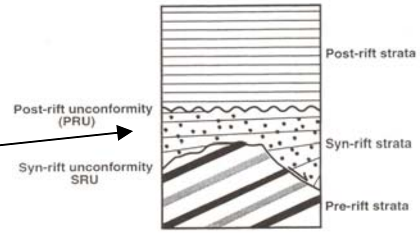


Figure A2.2 Terminology of rift-basin stratigraphy used in this chapter.

• The most common sediments that accumulate during rifting are proximal coarse clastics, including conglomerates and red beds shed from rising fault blocks, and distal finer grained clastics, including lacustrine lithologies, all in a continental setting.

• Several rifted basins have source and reservoir rocks entirely of continental origin with no marine rocks present. Examples include rifted basins in China, Brazil and Sudan.

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

“Frozen” Triassic-Jurassic rift basins around eastern USA

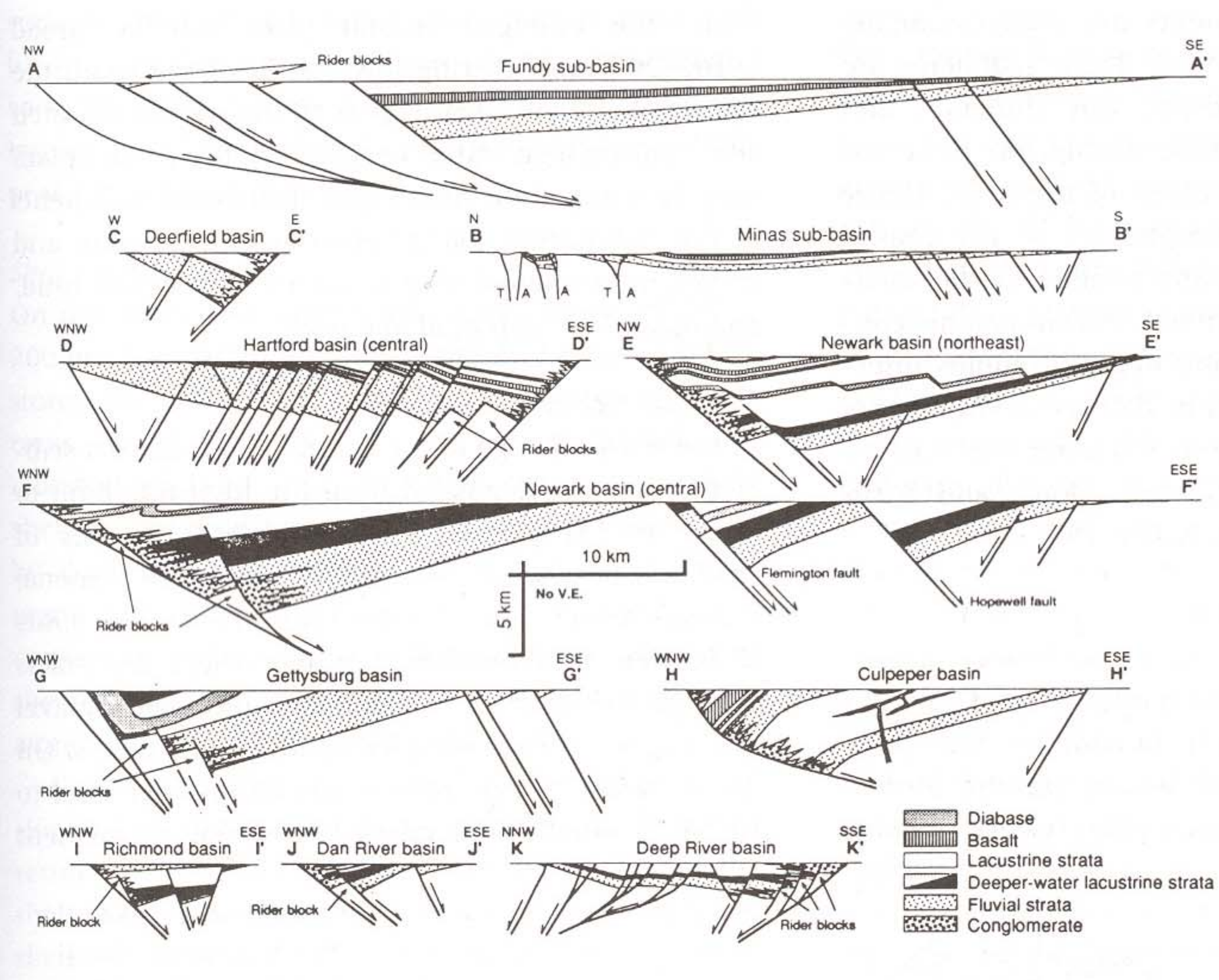


Fig. 4.22 Sealed cross sections of the rift basins of eastern North America, showing the tripartite stratigraphy of fluvial/lacustrine/fluvial successions. Also shown are basalt flows, and diabase dikes and sills (from Schlische, 1992).

Bond et al. (1995) in: Tectonics of Sedimentary Basins, Busby & Ingersoll (eds.), p.169.

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

Post-rift sections

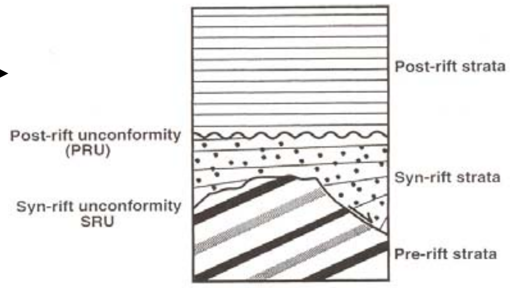


Figure A2.2 Terminology of rift-basin stratigraphy used in this chapter.

- The lower part of the post rift section is characterized by gently dipping reflections that represent the final establishment of a marine transgression.
- The later post rift sequence is often marked by cycles of sigmoidal shape, which progradation of sediments in a seaward direction.
- A thick accumulation of salt, continuous evaporation of normal sea water, which was periodically replenished, is a normal consequence of continental breakup and seafloor spreading at low latitudes.
- Postrift phase (or drift phase) is typically dominated by gravity-controlled deformation (salt tectonics, mud diapirism, slumps, slides, listric growth faults).

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

Melting is common during rifting

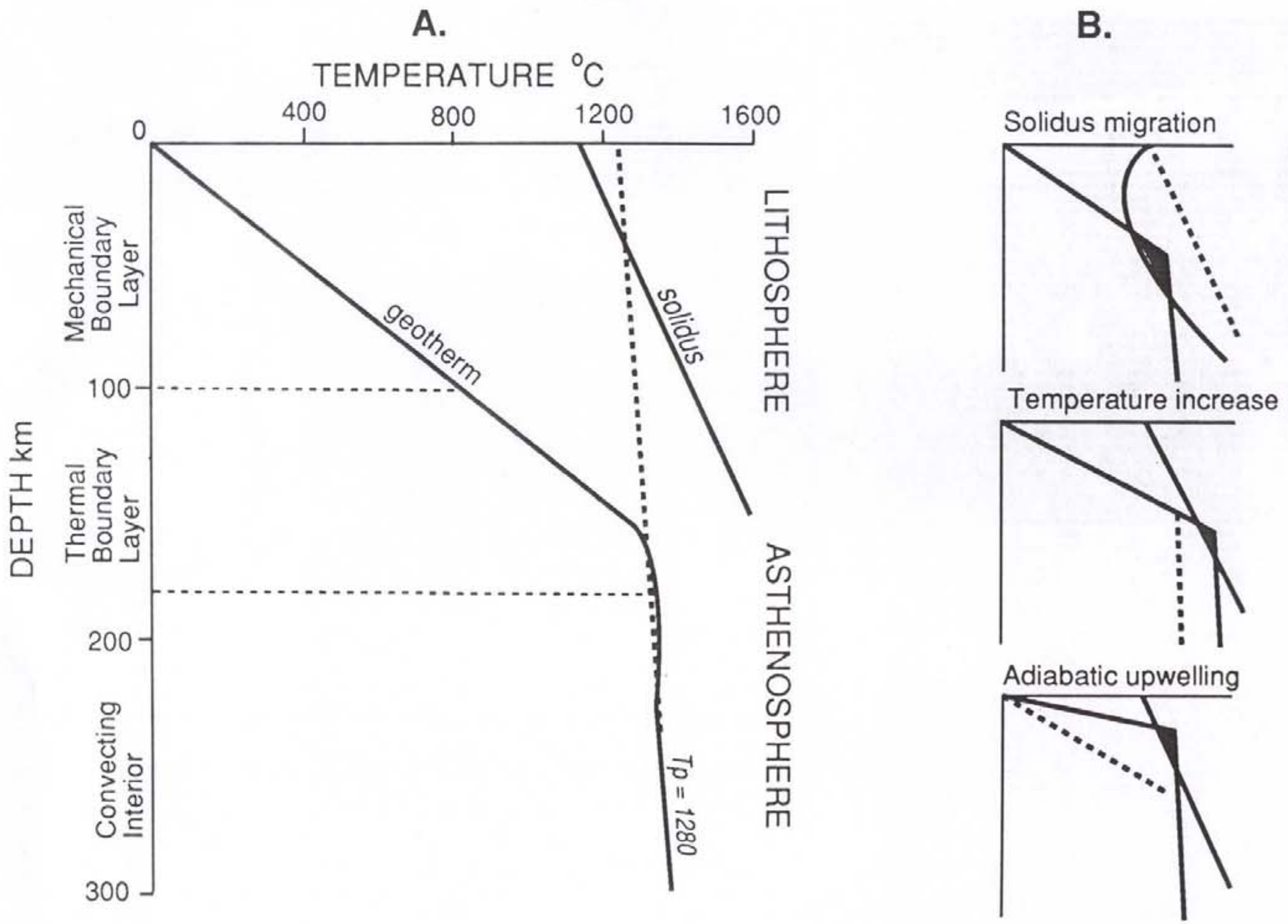
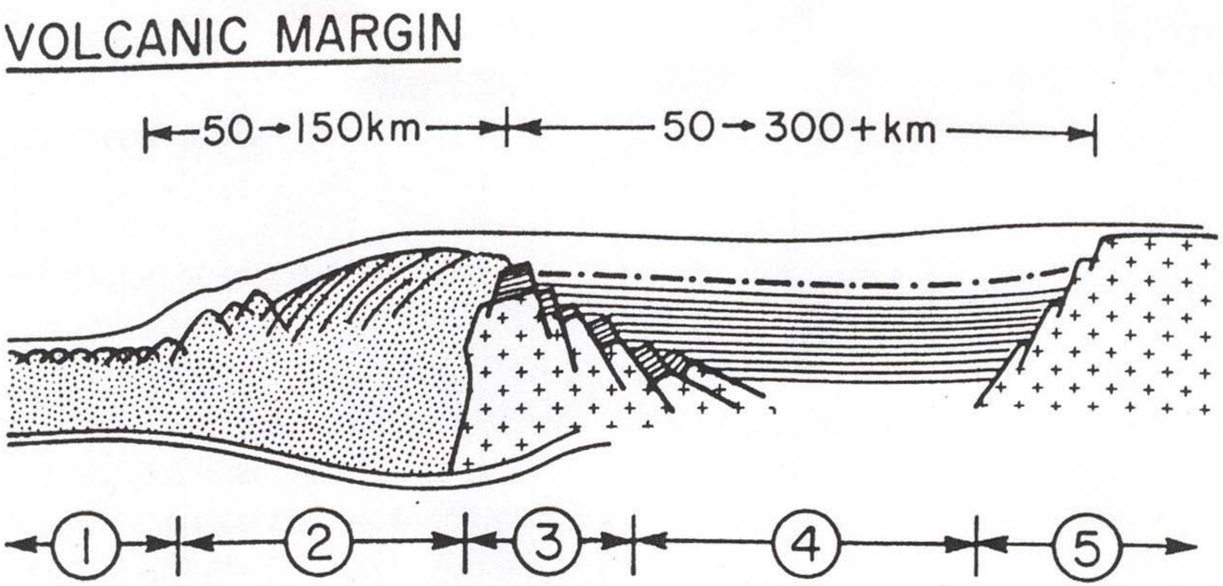


Figure 3.4 Rifting and melting. A. Horizontally averaged thermal structure of lithosphere for potential temperature (Tp: the temperature on adiabatic gradient projected to surface pressure) of 1280°C, mechanical-boundary-layer (lithosphere) thickness of 100 km, and interior viscosity of 2.10^{17} m² s⁻¹ (after McKenzie and Bickle, 1988). B. Sketch graphs (after Latin et al., 1990) to summarize three possible mechanisms for producing melts during rifting from results of Fig. 3.4A. In B1, solidus migrates to left because volatiles like water are added to the system, as in island-arc environments. In B2, potential temperature is raised, causing geotherm to migrate to right, due to rising hot spot or plume (open-system melting). In B3, the lithosphere is thinned mechanically by closed-system stretching, with asthenosphere rising to be partially melted due to adiabatic decompression.

Leeder (1995) in: Tectonics of Sedimentary Basins, Busby & Ingersoll (eds.), p.123.

Volcanic vs. non-volcanic margins



Mutter et al. (1988) JGR,93

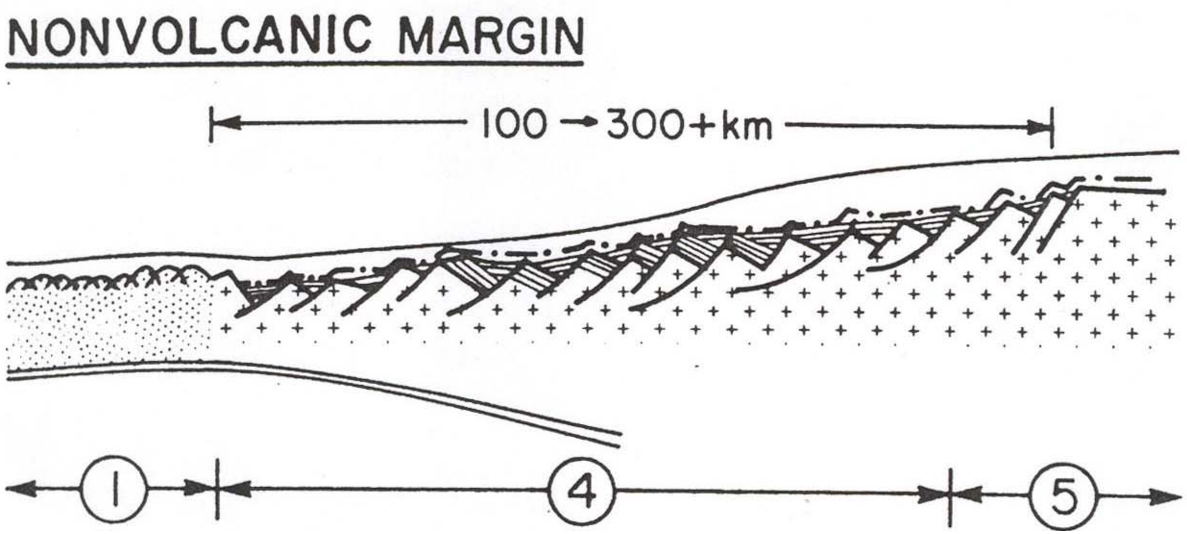


Fig. 7. Comparison of the typical structural elements of "volcanic" and "nonvolcanic" margins. The numbers refer to structures described in the text and are 1, the normal thickness oceanic crust; 2, the thick volcanic succession associated with the continent-ocean boundary of volcanic margins of which the seaward dipping units form the upper sequence; 3, a structural high in continental crust that often occurs adjacent to the thick volcanic succession; 4, thinned, subsided continental crust; 5, unstretched continental crust. The dot-dash line marks the stratigraphic level of breakup. Parallel ruled regions indicate sediments. Other symbols are as for Figures 3 and 6.

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

An idealized volcanic margin

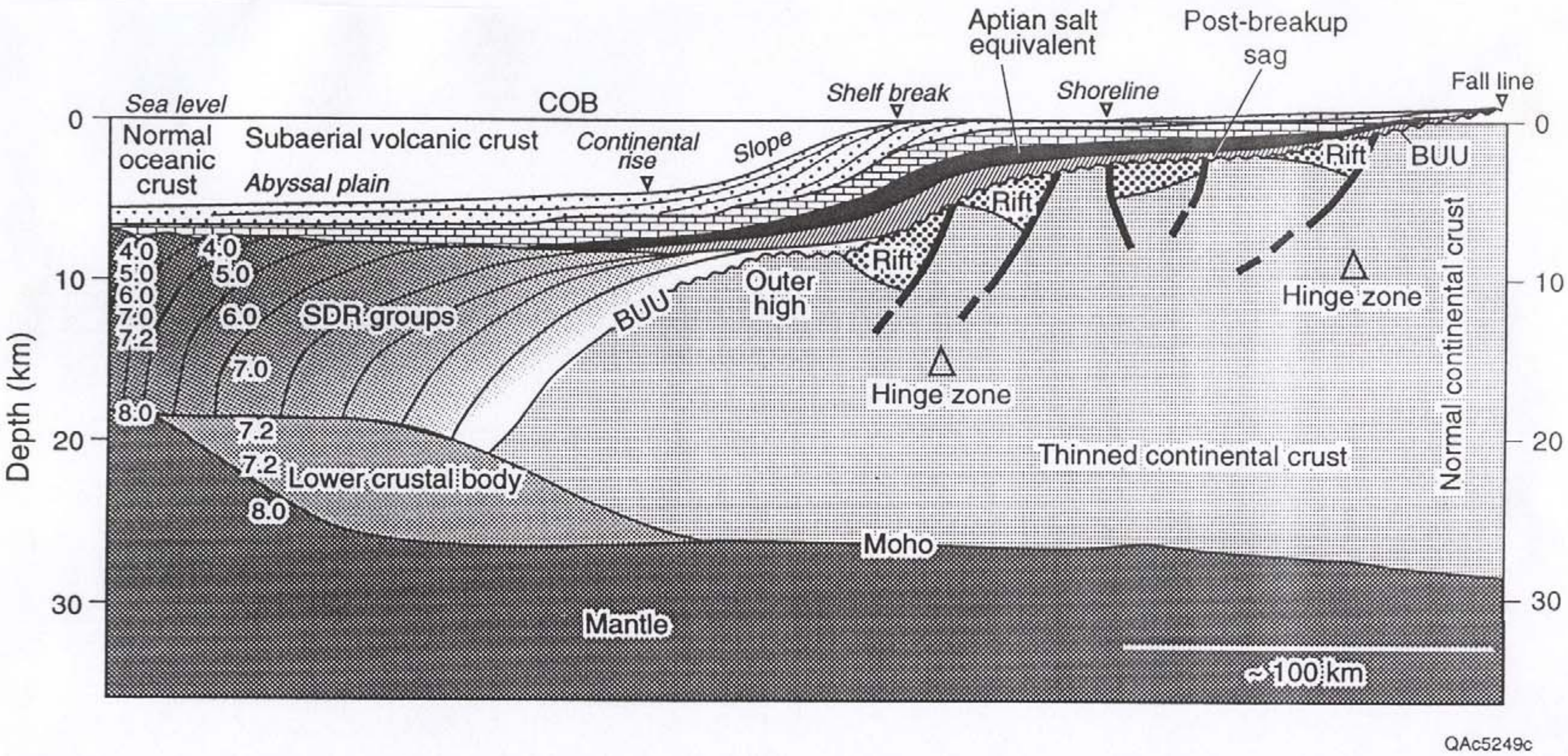


Fig. 4. Schematic cross-section of a generic volcanic rifted margin in the South Atlantic. Crustal units and dimensions are based on North Atlantic volcanic margins and on the Namibian margin (Gerrard & Smith, 1982; Gladczenko et al., 1997). Refraction data are sparse, so the P-wave velocities shown (4.0–8.0 km/s) are speculative. Salt-tectonics effects have been omitted for clarity. BUU, breakup unconformity; COB, the continental-oceanic boundary, varies in position from modern continental rise to shelf. Vertical exaggeration is roughly 4:1.

Magnetic anomaly, main magmatic features in the northeast Atlantic

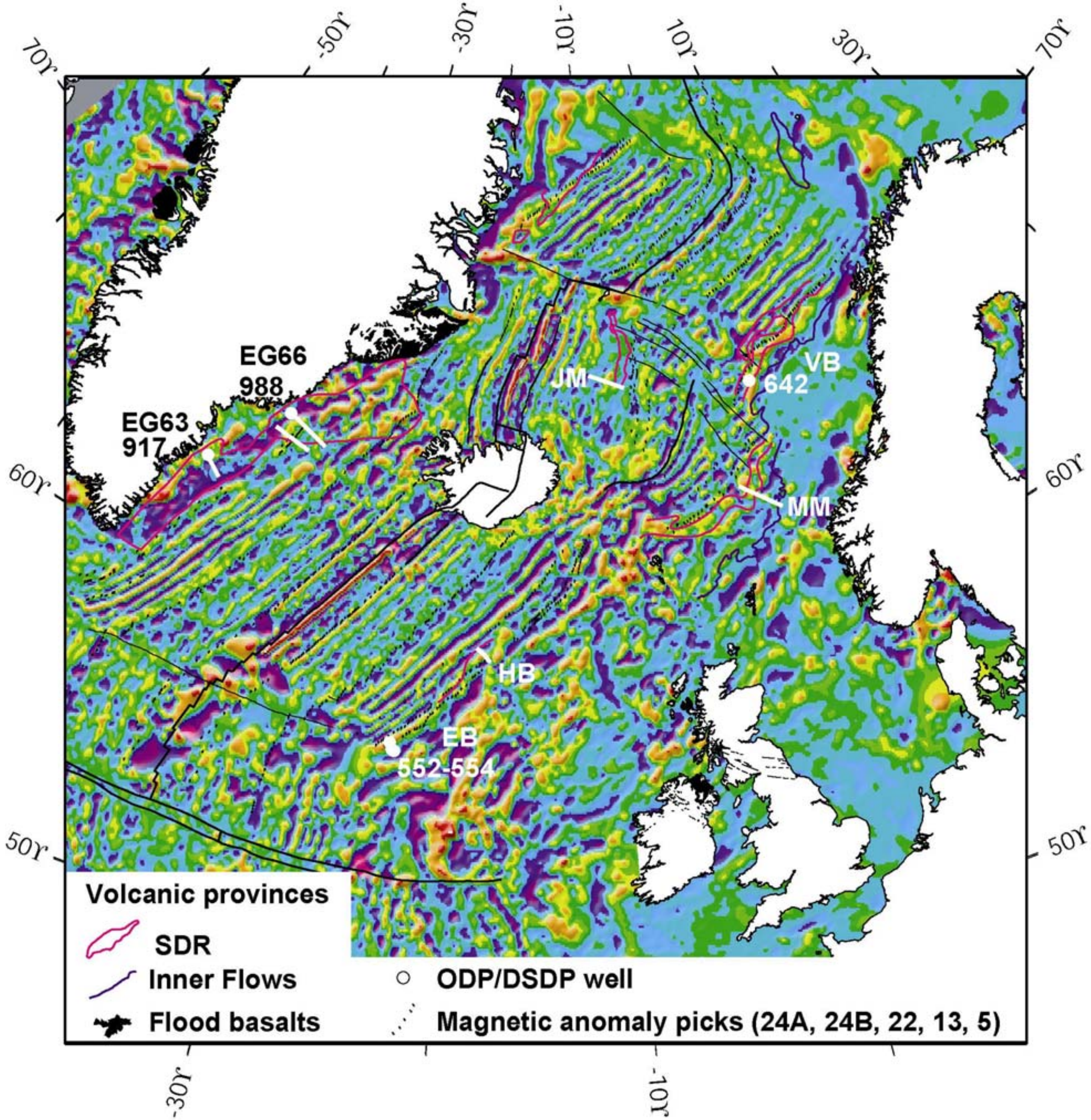


Figure 1. Magnetic anomaly map of the northeast Atlantic, showing main structural and magmatic features, selected drill sites penetrating breakup volcanics, and location of seismic profiles. Shown are DSDP Sites 552–554 (Roberts, Schnitker, et al., 1984), ODP Site 642 (Eldholm, Thiede, Taylor, et al., 1987), and Transects EG63 and EG66 (Larsen, Saunders, Clift, et al., 1994; Duncan, Larsen, Allan, et al., 1996). EB = Edoras Bank, HB = Hatton Bank Margin, JM = Jan Mayen Ridge, MM = M re Margin, VB = V r ing Basin. Magnetic data from Verhoef et al. (1996).

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

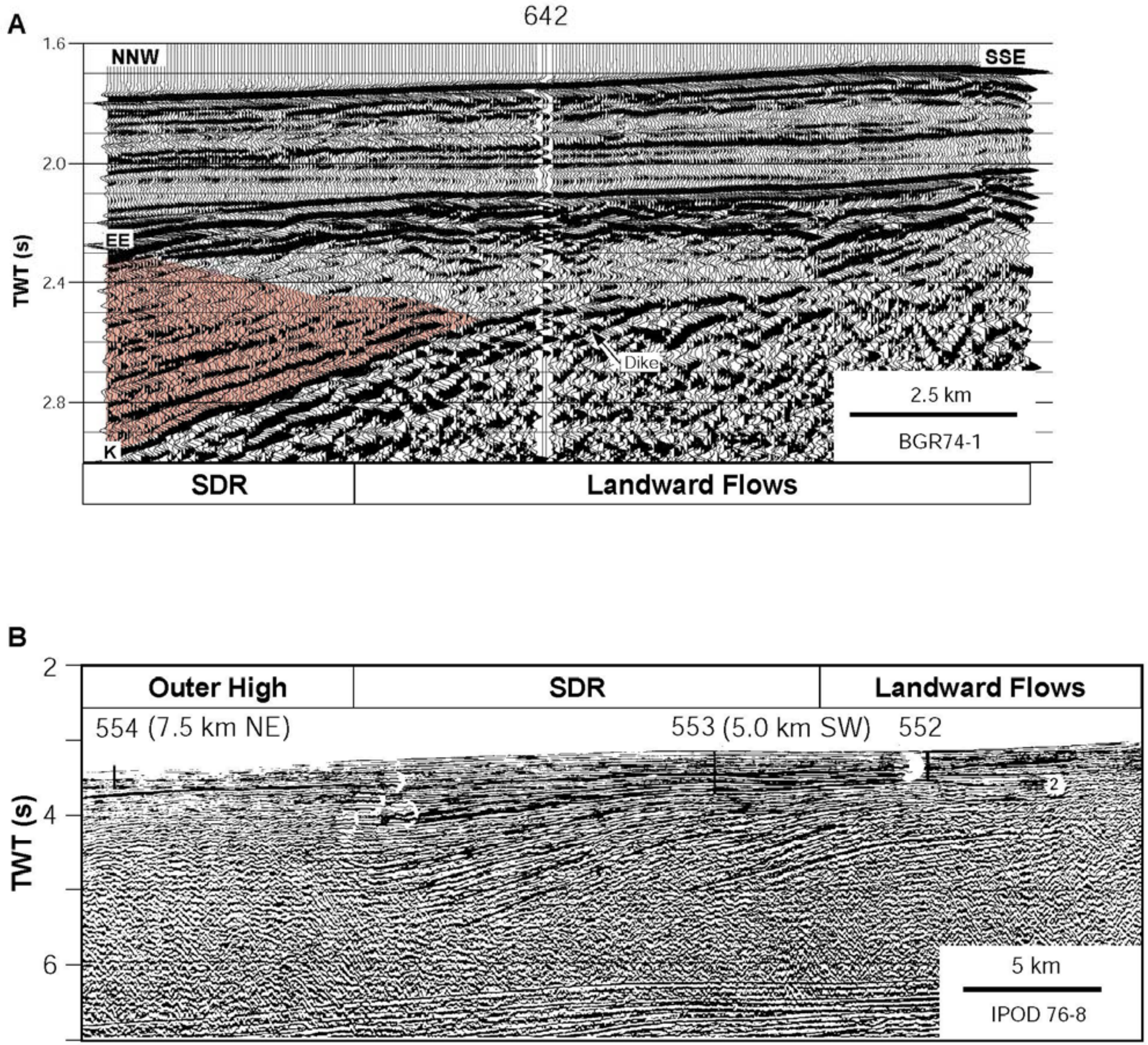


Figure 3. Seismic profiles across boreholes penetrating extrusive volcanics. **A.** ODP Site 642 on the V r ing Margin, modified from Planke and Eldholm (1994). **B.** DSDP Sites 55255 4 on the Edoras Bank Margin, modified from Roberts, Schnitker, et al. (1984). EE = top-basement, K = base-SDR.

East Greenland margin (EG63)

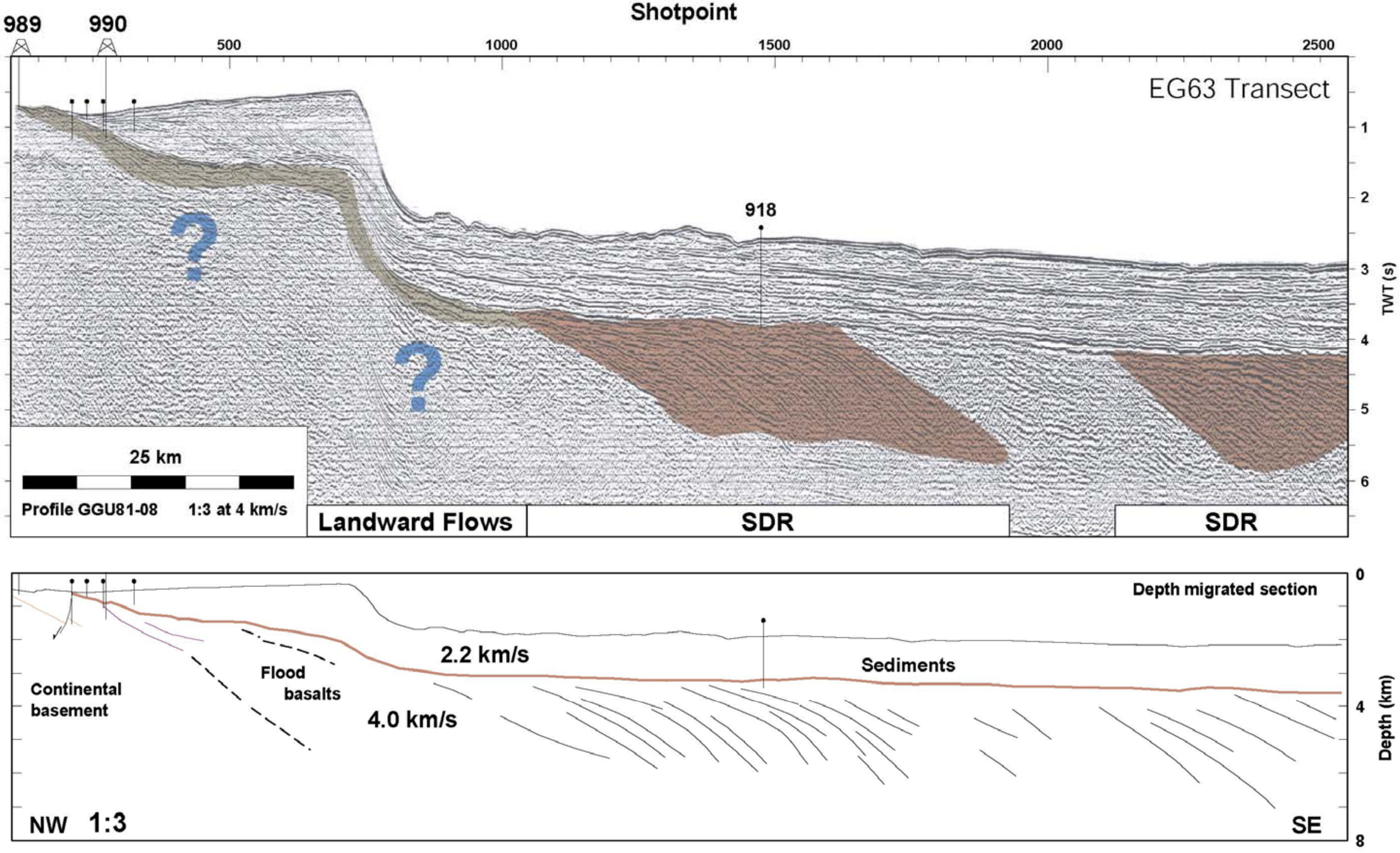


Figure 4. Interpreted and depth-converted MCS reflection profile along the EG63 transect. Ball-and-line symbols = Legs 152 and 163 sites. Based on Duncan, Larsen, Allan, et al. (1996).

Prepared by Dr. Andrew T. Lin
Institute of Geophysics
National Central Univ. Taiwan

How SDR being formed?

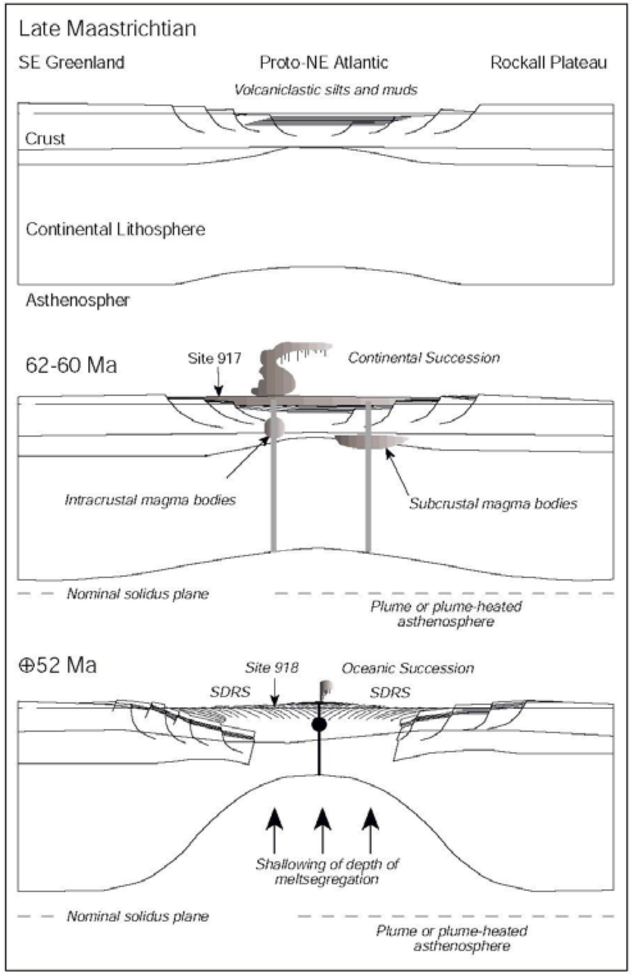
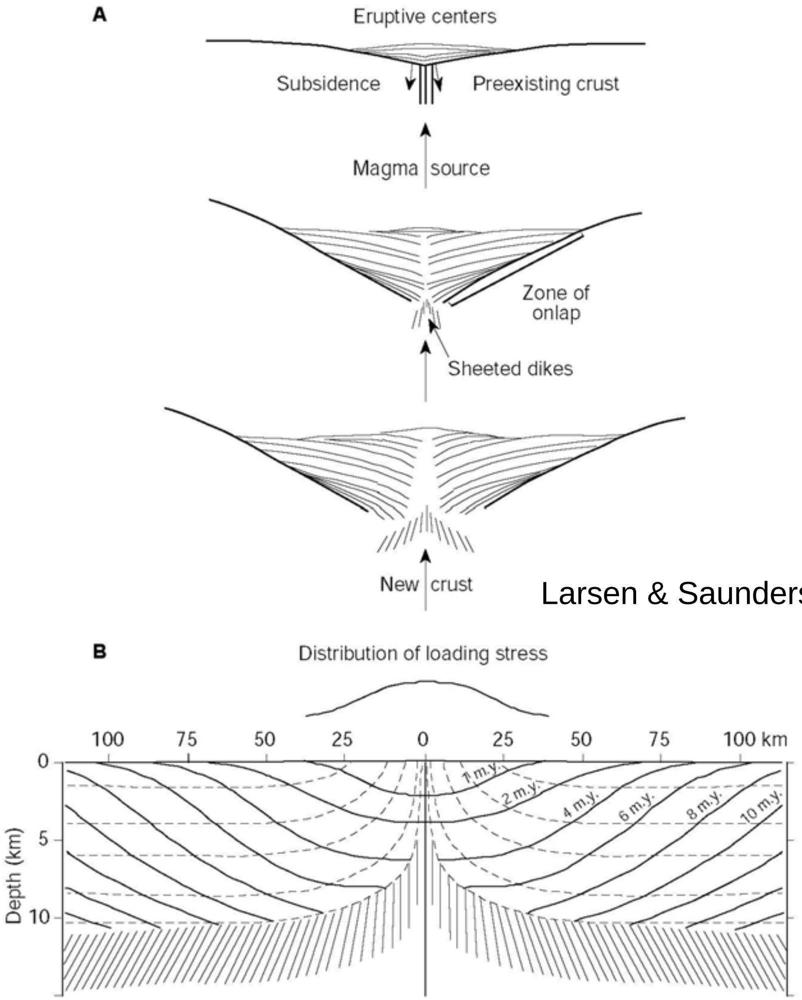


Figure 15. Schematic reconstructions of the southern northeast Atlantic at approximately 63N during the Paleogene, showing the development of the rifted margin and the contemporaneous magmatism.

Saunders et al. (1998)



Larsen & Saunders (1998)

Figure 6. The principle of seaward-dipping reflector sequence (SDRS) formation. The interpretation builds on the model for crustal accretion in Iceland (Pálmason, 1986). **A.** Initiation of SDRS formation during breakup and formation of the featheredge of the SDRS onlapping onto continental crust. In the case of the southeast Greenland COT, the Continental Succession is to be considered part of the continental crust. When spreading continues, a wide zone of SDRS crust forms. The model implies a down-dip, narrow and fairly linear, subaerially exposed volcanic source that stayed above sea level during the entire SDRS formation (Larsen and Jakobsdóttir, 1988). **B.** Kinematic model of formation of SDRS-type crust. Loading stress highest in center. Flow lines in dashed line and resultant stratigraphic structure in solid lines with age progression shown in m.y. Sheeted dike complex at the bottom of the lava pile. Modified from Pálmason (1986).

Divergent Margins and Petroleum Potential