Teufelsmauer ('The Devil's Wall'). Steeply-dipping Santonian sandstone near the Harz Northern Boundary Thrust, SW of Quedlinburg, Germany. The Teufelsmauer is the oldest natural monument in Germany, legally protected since the middle of the 19th century. The wall-like formation extends for a distance of 20 km just north of, and parallel to, the North Harz Boundary Fault, which lies approximately 2 km south of this locality. The Teufelsmauer is composed of Santonian sandstones, overturned by northward upthrusting of the Harz basement massif, and locally silicified along bedding and joint planes. Despite this deformation and alteration, cross-lamination is well preserved. Note the figures on the left hand side, and foreground, for scale. On the right hand side of the photograph, the Sub-Hercynian Basin comprises a succession of Permian to Late Cretaceous age, up to 3 km thick. The inversion is attributed to the Sub-Hercynian (Late Cretaceous) Phase of inversion by Kockel (2003).

Chapter 3 Tectonic evolution

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1 Introduction

1.1 Summary tectonic history of north-west and central Europe

The crust of the SPB is a mosaic of orogenic terranes accreted first to Baltica, then to the East European Craton during the Caledonian and Variscan orogenic cycles (Ziegler, 1982a, 1990a). High-precision radiometric techniques have revealed that each 'orogeny' comprised several deformation phases, including the closure of oceanic basins, terrane docking (or 'soft collision'), post-collisional crustal shortening and post-orogenic collapse (e.g. see Krawczyk et al., 2008a; Kröner et al., 2008). Some of these phases overlapped, particularly during Paleozoic times when rifting of Gondwana proceeded almost without interruption. Following the Variscan Orogeny, Permian to Cretaceous basin development was punctuated by phases of crustal extension and subsidence associated with breakup of the Pangea Supercontinent. In Late Cretaceous to Cenozoic times, pulses of intraplate compression related to the Alpine Orogeny resulted in widespread basin inversion. These tectonic phases are summarised in **Figure 3.1**.

1.2 Paleozoic continental accretion to form the 'United Plates of Europe'

The majority of the terranes that form the crust of the SPB were rifted off the northern margin of the Gondwana Palaeocontinent, which lay at low southerly latitudes for much of the Paleozoic (Torsvik, 1998). The Iapetus Ocean opened during late Neoproterozoic times and separated Gondwana from other large relics of the Rodinia-Pannotia Supercontinent (Dalziel, 1991, 1997) such as Laurentia and Baltica (**Figure 3.2a**). The Gondwana-derived terranes were transported northwards, pulled by subduction of Iapetus and pushed by the new oceanic basins opening behind them. Accretion to the margin of Baltica (Pharaoh, 1999; Pharaoh et al., 2006) was followed by strike-slip displacements along major crustal lineaments (Nawrocki & Poprawa, 2006). Further Gondwana-derived terranes were accreted to Laurussia during the Variscan Orogeny in Late Carboniferous times.

1.3 Permo-Carboniferous magmato-tectonic activity and the birth of the Southern Permian Basin

The SPB crust, which lay mainly in the foreland of the Variscan Orogen (**Figure 3.3**), was destabilised at the end of the Variscan Orogeny by wrench-induced collapse and widespread alkaline and calc-alkaline magmatic activity (Ziegler, 1990a; Wilson et al., 2004) accompanied by profound thermal thinning of the lithosphere (Ziegler et al., 2004). These events took place in response to dextral translation of Africa relative to Europe during the latest Carboniferous (Gzhelian) and Early Permian final suturing phases of Pangea (Arthaud & Matte, 1977; Ziegler, 1989; Ziegler & Stampfli, 2001; Stampfli & Borel, 2002). Following thermal doming and significant erosion, the crust of the Northern and Southern Permian basins started to subside during the late Early Permian in response to the decay of the earlier thermal anomaly (Bachmann & Hoffmann, 1997). The SPB belongs to the type of intracontinental or cratonic basins (Bachmann & Grosse, 1989) similar to the Michigan, Illinois or Amazon basins and many others (e.g. Bally & Snelson, 1980). A land-locked depression developed as a result of subsidence exceeding sedimentation rates, which was flooded during the Late Permian by Arctic seas.

1.4 Mesozoic rifting and breakup of Pangea

During Late Permian to Mid-Triassic times, the development of the Northern and Southern Permian basins was controlled by thermal subsidence. Development of the Arctic-North Atlantic rift system between Greenland and Scandinavia had commenced in the Late Carboniferous. This rift system propagated southwards during Late Permian and Triassic times as indicated by the Triassic development of the Viking and Central grabens that transected the Northern and Southern Permian basins. In the SPB area, Mid- to Late Triassic east–west extension controlled the accelerated subsidence of the north-trending Central, Horn and Glückstadt grabens, and of the north-west-trending Mid-Polish Trough. Syndepositional normal faulting in these troughs was accompanied by salt mobilisation along the graben margins. The Mid-Polish Trough, which is superimposed on the crustal-scale Teisseyre-Tornquist Zone, remained intermittently active during Jurassic and Early Cretaceous times. During the Late Triassic, the Arctic-North Atlantic rift propagated southwards into the Central Atlantic domain where crustal separation was achieved towards the end of the Early Jurassic. Mid-Jurassic crustal separation and opening of the Alpine Tethys Ocean entailed a reorientation of the stress field of north-west Europe. The Horn and Glückstadt grabens became inactive during Mid-Jurassic uplift of a large thermal dome straddling the Central Graben.

During the Late Jurassic to Early Cretaceous, accelerated crustal extension across the North Sea rift system caused north-west-trending transtensional basins to develop along the southern margin of the SPB, large areas of which became exposed and subjected to erosion. Following a period of intense Early Cretaceous rifting, the North Atlantic Ocean started to open during mid-Cretaceous times, whereas the North Sea rift system became inactive and rifting activity concentrated on areas between Europe and Greenland (Ziegler 1988, 1990a). Meanwhile, the Tethys Ocean was opening to the south of Europe.

1.5 Late Mesozoic-Cenozoic closure of the Tethys Ocean, inversion of the Alpine foreland and formation of the Cenozoic graben system

During the Late Cretaceous, regional thermal subsidence combined with eustatic sea-level rise resulted in a regional transgression and flooding of the SPB. Closure of the Alpine Tethys Ocean started with the Late Cretaceous onset of counter-clockwise rotational convergence of Africa-Arabia with Europe (Rosenbaum et al., 2002) where compressional stresses began to build-up in the Alpine foreland (Ziegler et al., 1995, 1998; Dèzes et al., 2004; Kley & Voigt, 2008). Early-Late Cretaceous subsidence of the SPB area was accompanied by minor extensional faulting and was followed by the build-up of intraplate compressional stresses as indicated by the inversion of Mesozoic tensional basins and upthrusting of basement blocks that started during the late Turonian and intensified during the Senonian and Paleocene (Ziegler, 1990a). The inversion movements were remarkably heterogeneous in nature, with strain localisation in narrow zones separated by undeformed regions. Deformation was tectonically decoupled by the Zechstein salt, with overall compression in the salt cover and localised basement deformation. The latter was concentrated on the north-west-trending Sorgenfrei-Tornquist Zone of crustal weakness (along the northern margin of the Norwegian-Danish Basin), on the Teisseyre-Tornquist Zone (underlying the axial part of the Polish Basin), and on the southern margin of the North German Basin (Krawczyk et al., 1999; Scheck et al., 2002a). The north-westerly trend of early inverted basins and transpressionally reactivated faults indicates a tectonic setting controlled by north to north-easterly directed compressional stresses (Kley & Voigt, 2008).

With the closure of the Alpine Tethys Ocean during the Paleogene, the Alpine-Carpathian Orogen entered the continent-to-continent collisional stage. The SPB was located in the foreland of this orogen and so was affected by the related stress fields as well as by intermittent pulses of basin inversion (Dèzes et al., 2004). The stress fields repeatedly changed in orientation and magnitude and controlled the evolution of the European Cenozoic rift system (Upper Rhine, Roer Valley and Eger grabens). Throughout Cenozoic times, the western SPB was part of the continuously subsiding North Sea thermal sag basin (Ziegler, 1990a). Miocene to Pliocene uplift of the Variscan massifs on the southern flank of the SPB, and uplift of the Fennoscandian Shield, had severe repercussions on the drainage systems of the SPB (Ziegler & Dèzes, 2007). Subsidence of the North Sea Basin and the adjacent North German lowlands was accelerated by the Pliocene to Pleistocene build-up of the present-day stress field caused by lithospheric folding (Van Wees & Cloetingh, 1996; Scheck-Wenderoth & Lamarche, 2005).

2 Early Paleozoic tectonic evolution: assembly of Laurussia

2.1 Provenance of the crystalline basement of the Southern Permian Basin

Information on the pre-Devonian basement beneath the central SPB area is limited (see Chapter 4); it is therefore conjectured from seismic reflection/refraction data and gravity and magnetic measurements (Chapter 2). Precambrian rocks have been drilled on the Fennoscandian Shield north of Rügen (**Figure 3.3**), where they are overlain by flat-lying Early Paleozoic strata. Metamorphic rocks have been found in northern Schleswig-Holstein and the Ringkøbing-Fyn High. The northernmost group comprises amphibolite-facies gneisses yielding metamorphic ages of 800 to 900 Ma (Frost et al., 1981) and represent splinters of southern Fennoscandia incorporated into the northern margin of the North German-Polish Caledonides. These are variably retrogressed close to Caledonian Deformation Front near the Thor Suture. The southern group comprises lower-grade (typically greenschist-facies) metasedimentary rocks of probable (but unproven) Early Paleozoic age, which may have Avalonian affinities (see Chapter 4).

A borehole in the Dutch sector (A17-1) cored a granitic intrusion. An isotopic study based on U-Pb zircon dates puts emplacement of the granite at 410±7 Ma (Early Devonian), but failed to identify older inherited grains (A. Gerdes, pers. comm., 2002). This supports the concept that the crust here is juvenile, accreted to Avalonia during the Early Paleozoic, and that Precambrian crust is absent (Pharaoh et al., 2006). All other occurrences of rocks assumed to be of Precambrian age (e.g. the Mid-German Crystalline Rise; Ecker Gneiss in the Harz Mountains) have recently proved to be of younger, Variscan age (Kröner et al., 2008; Linnemann et al., 2008).

2.2 Shelveian (Ardennian) Orogenic Phase (closure of the Tornquist Sea)

Avalonia rifted away from Gondwana during the Early Ordovician (**Figure 3.2b**) (Trench & Torsvik, 1992), driven northwards by opening of the Rheic Ocean (Cocks & Fortey, 1982) to the south of Avalonia, as well as by subduction of the Iapetus Ocean at a number of southerly dipping subduction systems (**Figure 3.2c**). Closure of the Tornquist Sea (Cocks & Fortey, 1982) involved a significant dextral-oblique component (Trench & Torsvik, 1992) and eventually produced the Thor Suture (Berthelsen, 1998) (**Figure 3.3**) between Avalonia and Baltica. The soft collision ('docking') of Avalonia and Baltica that produced Balonia (Torsvik, 1998) took place during the Shelveian Phase in Ashgill times (Pharaoh et al., 1995; Pharaoh, 1999; Samuelsson et al., 2002) and is associated with amphibolite-facies metamorphism in the mid-North Sea region (Frost et al., 1981; Pharaoh et al., 1995). ^{40}Ar-^{39}Ar plateau ages indicate prograde greenschist-amphibolite metamorphism at 450-425 Ma, with retrogression at 415-400 Ma (Frost et al., 1981). A zone of listric thrust planes has been observed in seismic-reflection data from the Danish sector of the North Sea, indicating the prolongation of the Caledonian Deformation Front westwards to the Central Graben (MONA LISA Working Group, 1997a). A number of boreholes (e.g. P-1 and Per-1) are interpreted to sample the cataclastic fabric of the suture in this region (Pharaoh et al., 2006; see Chapter 4).

2.3 Scandian (Ardennian) Orogenic Phase (closure of the Iapetus Ocean)

A flexural foreland basin developed on the margin of Baltica (Abramowitz & Thybo, 1998; Berthelsen, 1998) following its docking with Avalonia, in response to its loading by the encroaching orogenic wedge and its continued convergence with the foreland. Rapid exhumation (within 10 Ma of collision) of the collisional welt at the Thor Suture in the central North Sea area fed sediment northwards into this foreland basin (Samuelsson et al., 2002), which is filled with thick Silurian (mainly Pridolian) strata that reflect rapid subsidence of the foreland lithosphere (Sliaupa et al., 2006). This also led to high (>4%) coalification of the Middle to Upper Cambrian source rocks of Bornholm, Scania, northern Denmark and offshore of Rügen Island. Continued north-westward subduction of Baltica beneath Laurentia led to final closure of the Iapetus Ocean during Wenlock times (**Figure 3.2d**) (Leggett et al., 1979; Kneller et al., 1993) and consequently the amalgamation of Laurussia during the latest Silurian Scandian Orogeny (Ziegler, 1989; Gee, 2005). The Ardennian Phase (Verniers et al., 2002) is probably coeval in the SPB. Major crustal lineaments such as the Teisseyre-Tornquist Zone were active by this time or perhaps even earlier (Pharaoh et al., 2006); these lineaments subsequently acted as a focus for Carboniferous to Mesozoic strike-slip displacement and Alpine basin inversion.

2.4 Acadian Orogenic Phase

Following the end-Silurian accretion of Avalonia to Baltica, and docking of at least part of the Armorican Terrane Assemblage against the southern margin of Avalonia, the Caledonian Ardennes Fold Belt that marked the Rheic Suture was disrupted by Early Devonian rifting. This rifting controlled the subsidence of the Rheno-Hercynian Basin and the opening of the Lizard-Giessen oceanic basins (Ziegler, 1990a; Franke, 2000). The Rheno-Hercynian Basin developed in a back-arc position relative to the northerly dipping subduction zone that fringed the southern boundary of the Armorican Terrane Assemblage (Ziegler, 1989). This is shown in a generalised way in **Figure 3.4a**; more detailed reconstructions were presented by Tait et al. (2000) and Winchester et al. (2002). Orogen-parallel collapse of the Arctic-North Atlantic Caledonides commenced under a sinistral transtensional setting during the latest Silurian and Early

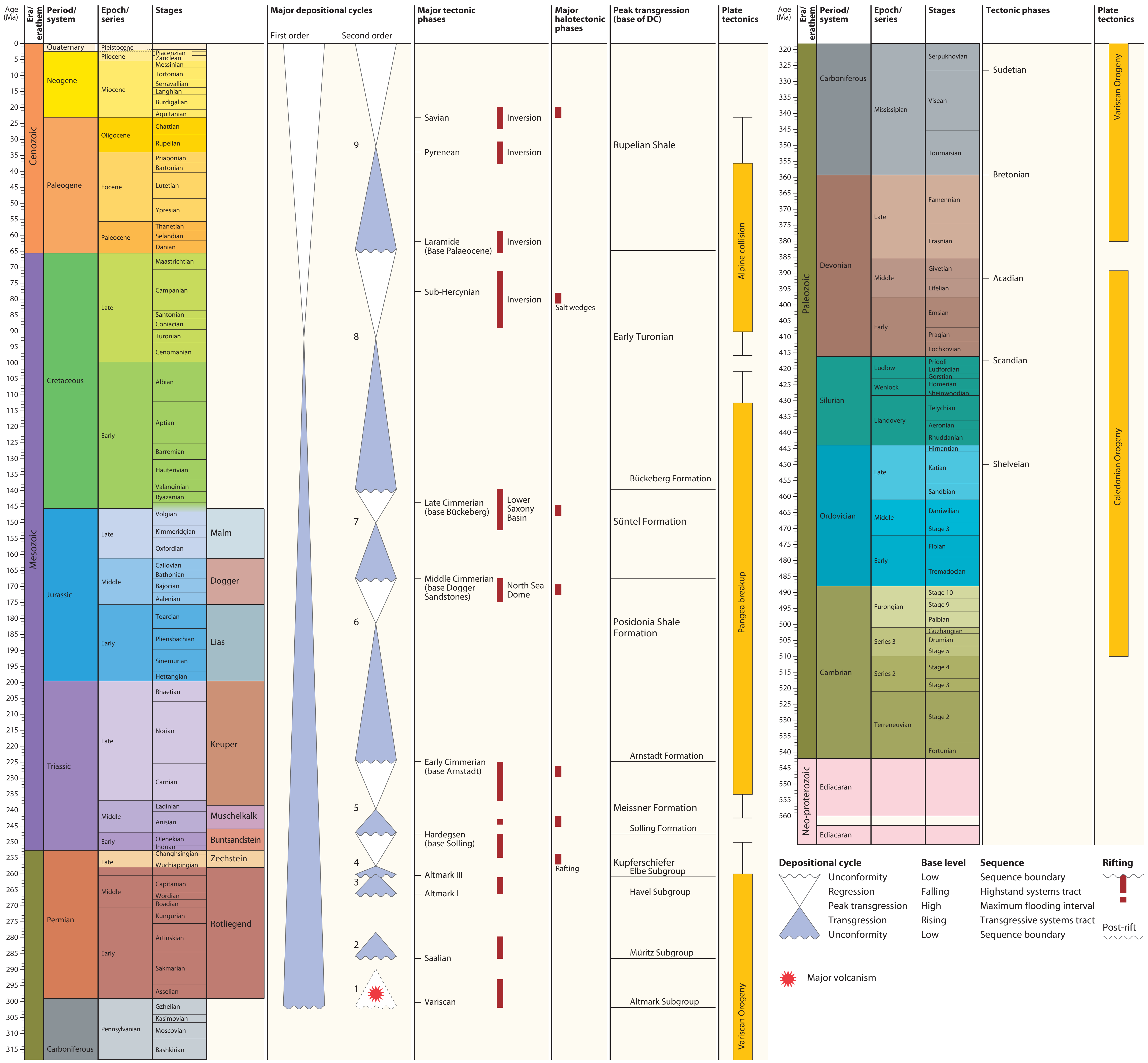


Figure 3.1 Paleozoic – Cenozoic timescale with tectonic episodes, major depositional cycles (DC) 1 to 9 (2nd order) (see Annex 1 in Chapter 6 for explanation), major unconformities, peak transgressions, rifting phases and halotectonic phases of the SPB. The upper, regressive phases of DC4-9 seem to correspond with the rifting and halotectonic phases. DC 8-9 are inversion related. Upper Carboniferous to Quaternary modified after Bachmann et al. (2008). Chronostratigraphy and numerical ages after DSK (2002), Bachmann & Kozur (2004), Menning & Hendrich (2005) and Kozur & Weems (2007).

Devonian, as shown by the development of intramontane Old Red Sandstone basins and the widespread granitic plutonism commonly seen in northern England (Ziegler, 1989; Braathen et al., 2002). At the same time, a first compressional phase affected the Rheno-Hercynian back-arc basin, which was followed by renewed Late Devonian extension and volcanism (Ziegler, 1989, 1990a).

3 Late Paleozoic tectonic evolution: assembly of Pangea

3.1 The margins of Laurussia and the Rheno-Hercynian Ocean

During the Mid- to Late Devonian, sediments of 'Old Red' continental facies accumulated in the area of the future North Sea, which was flanked to the south by the Rheno-Hercynian Basin. Middle Devonian marine intercalations are found in the UK, in the vicinity of the Central Graben, in Schleswig-Holstein, the south-eastern North Sea and southern Baltic Sea. During Givetian and Late Devonian times, the nonmarine facies persisted in the north-west beneath the central SPB area, whereas the southern, central and north-eastern parts as far north as Rügen were occupied by marine-carbonate platforms and reef structures on the northern shelf of the Rheno-Hercynian Basin (**Figure 3.4b**). A mid-Famennian regression led to the development of widespread coastal sand bars of the Condroz facies and evaporitic strata in north-western Poland. Carbonate platforms developed across much of the northern SPB area during Dinantian times. In northern England, grabens with more basinal strata that are up to 4000 m thick (see Chapter 6) are flanked by carbonate platforms (Fraser & Gawthorpe, 1990). Similar features may occur in the southern North Sea (Kombrink, 2008) and beneath Hiddensee and Rügen Island.

Starved anoxic basins became widespread during Dinantian to earliest Namurian times; the Bowland and Edale Shale formations (UK), Geverik Formation (Netherlands) and Chokier Formation (Brabant Massif) include black shales that are important oil and gas source rocks in the UK and Netherlands (see Chapter 6) (Fraser & Gawthorpe, 1990).

3.2 Closure of the Rheno-Hercynian Ocean and development of the Variscan Internides

Gondwana began to converge with Laurussia in a clockwise-rotational mode during the Late Devonian. Initial contacts between Gondwana and the Gondwana-derived terranes of the Variscan Orogen were established in the Iberian domain by the end-Devonian (Ziegler, 1989). Northward subduction of the Paleotethys Ocean was accompanied by progressive closure of the oceanic basins that had separated the different Gondwana-derived terranes. Palaeomagnetic data (Tait et al., 1997, 2000) supports independent motion between distinct Moldanubian and Perunica terranes within the Armorican Terrane Assemblage until at least the Late Devonian (~370 Ma).

After the Mid-Devonian Acadian/Ligerian compressional phase and the subsequent extensional pulse, compressional stresses built up again during the latest Devonian. The 'Bretonian' orogenic pulse spans the Devonian-Carboniferous boundary. Widespread high-pressure metamorphism in the Variscan Internides reflects closure of the oceanic basins separating the Bohemian and Saxo-Thüringian terranes. Closure of the Rheno-Hercynian oceanic basin resumed (Franke, 1995a, 2000) and, during the Dinantian, the starved Rheno-Hercynian Basin was invaded by south-easterly sourced olistostromes and flysch fans and subsequently overridden by the Giessen and east Harz nappes (Franke, 2006). The parautochthonous Rheno-Hercynian nappes of south-west England and central Germany contain fragments of the southern margin of Eastern Avalonia (McKerrow et al., 1997; Verniers et al., 2002). However, widespread Tournaisian to early Visean basaltic magmatic activity in the Rheno-Hercynian Basin testifies to the resumption of extension prior to its final closure during the Sudetic Orogenic Phase, which started during the Visean (Ziegler, 1990a; Franke, 2000; Ziegler et al., 2004).

Following collision of the Variscan orogenic wedge with the passive northern margin of the Rheno-Hercynian Basin, essentially thin-skinned thrusting during late Visean to Westphalian D times accounts for about 300 km of shortening in the Variscan Externides (Oncken et al., 2000). (**Figure 3.4d**). The Rheno-Hercynian Suture can be traced along the periphery of the SPB from the south-eastern Hunsrück via Frankfurt-am-Main to the south-eastern tip of the Harz Mountains (Wippraer Zone) and the River Neisse (**Figure 3.5**). Collision-related compressional stresses exerted on the foreland at the Visean-Namurian boundary caused disruption of the Dinantian carbonate platforms (Ziegler, 1990a) and late Westphalian inversion of Carboniferous rifts on the British Isles, such as the Munster-Dublin-Solway-Northumberland trough and the Midland Valley Graben (Fraser & Gawthorpe, 1990; Ziegler, 1990a; Ziegler et al., 1995).

3.3 Development of the Variscan foreland-migrating fold-and-thrust belt

Following the end-Visean disruption of the Dinantian carbonate platform and the development of a regional unconformity, the Rheno-Hercynian Shelf rapidly subsided during the Namurian in response to tectonic loading by the north-westward advancing Variscan orogenic wedge. Beginning with the Namurian C, the

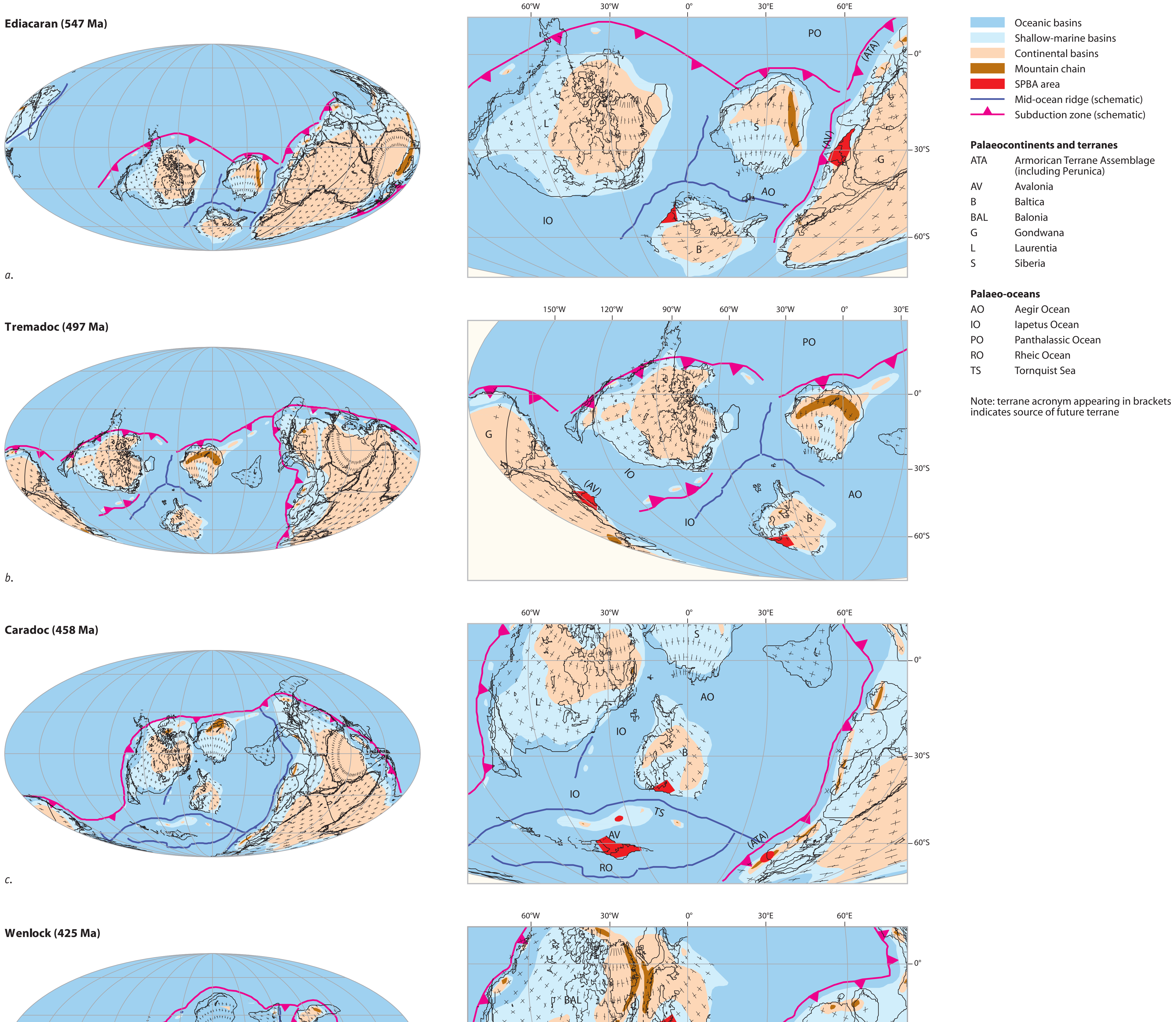


Figure 3.2 Late Proterozoic – Early Paleozoic tectonic evolution. Palaeogeographic maps are shown for the:
a. Ediacaran (547 Ma);
b. Tremadoc (497 Ma);
c. Caradoc (458 Ma);
d. Wenlock (425 Ma).
Reconstructions by C. Scotese, kindly provided by Shell.

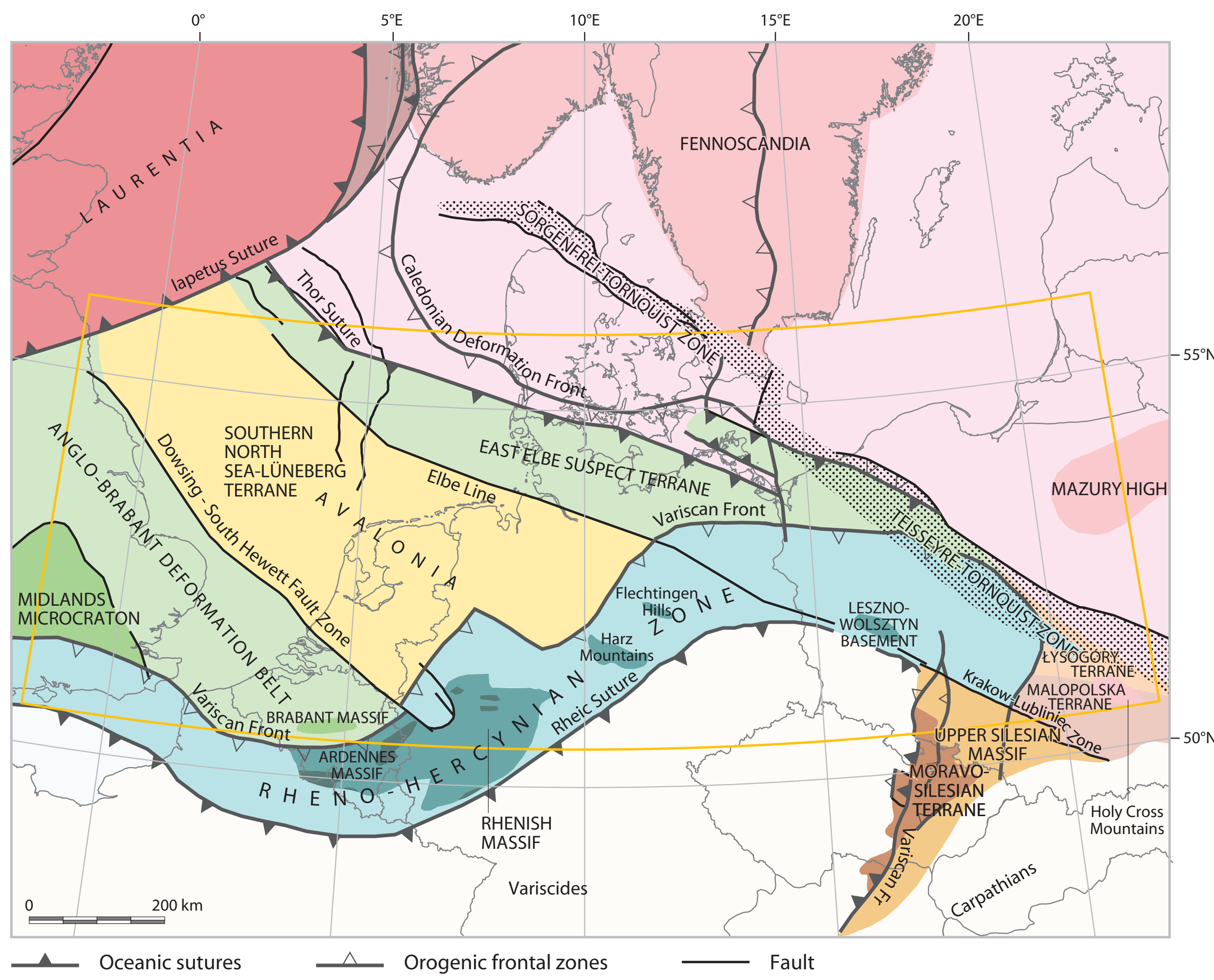


Figure 3.3 Terranes amalgamated to form Laurussia. Non-palinspastic map after Pharaoh et al. (2006) and sources therein. Note that the Rheno-Hercynian Zone is interpreted as the Variscan-deformed southern margin of Laurussia following McKerrow et al. (1997).

north-east-trending flexural foreland basin was filled by orogen-derived deltaic and coal-bearing clastics that contain sporadic marine intercalations. The axis of this foreland basin migrated progressively north-westwards as its internal parts became involved in the advancing fold-and-thrust belt of the Variscan Externides. Westphalian A and B coals are the main gas source rock in the UK, Netherlands and Germany. The foredeep itself became folded during Westphalian C and D times (Corfield et al., 1996; Besly, 1998). Stephanian sediments unconformably overlie the Westphalian in the east Netherlands and Germany (Lower Saxony Basin) and testify to the termination of thrusting activity in the Variscan Externides at the end of Westphalian D times (Ziegler, 1990a); there are also remnants of Stephanian strata in the West Netherlands Basin and Cleaver Bank High. Carboniferous deformation patterns differ on either side of the Variscan thrust front: thin-skinned, short-wavelength folds and thrusts are typical of areas south of this front; long-wavelength folds and tilted fault blocks are typical to the north (Franke et al., 1995, 1996; Franke & Hoffmann, 1997; Oncken et al., 2000). Parts of the Variscan Externides, which are now buried beneath a thick pile of post-Carboniferous sediments in the eastern SPB, are characterised by thin-skinned deformation along gentle southerly dipping fault planes (Franke & Hoffmann, 1997). Organic-rich shales in the Cambrian, Silurian, Dinantian and Namurian successions are the most likely detachment horizons (Franke et al., 1995, 1996; Hoffmann et al., 1996). In contrast, the Variscan Internides at the southern SPB margin are characterised by thick-skinned, basement-involved tectonics (Chapter 2).

3.4 Features of the Variscan structural template

The Mid-Polish Trough is an inverted late Early Permian and Mesozoic tensional basin, which is superimposed on the north-west-trending Teisseyre-Tornquist Zone (TTZ) (cf. Kutek & Głazek, 1972; Pożaryski & Brochwicz-Lewiński, 1978; Dadlez et al., 1995; Dadlez, 1997a, 1997b, 1998b; Kutek, 2001), one of the most fundamental lithospheric boundaries in Europe (Thybo et al., 2002). The TTZ is characterised by a very complex internal structure, which has been imaged on deep-seismic refraction data (Chapter 2) (BABEL Working Group, 1993; Grad et al., 2002; Dadlez et al., 2005; Grad & Guterch, 2006a, 2006b). The north-eastern boundary of the TTZ generally coincides with the south-western boundary of the East European Craton (cf. Grabowska et al., 1998; Grabowska & Bojdys, 2001; Grad et al., 2002a; Grad & Guterch, 2006a; Pożaryski & Nawrocki, 2000; Królikowski & Petecki, 1997, 2002) whereas its south-western boundary lies within the so-called Trans-European Suture Zone (Gee & Zeyen, 1996. This is a wide zone encompassing the fronts of the Caledonian and Variscan orogens and various terranes, some of which are only hypothetical, (for details see Thybo et al., 1994; Abramowitz & Thybo, 1998, 2000; Pharaoh, 1999; Grad et al., 2002a; Mazur & Jarosiński, 2006; Nawrocki & Poprawa, 2006; McCann et al., 2006; McCann, 2008c). Pharaoh et al. (2006) suggested that the TTZ may have originated as the rifted margin of Baltica

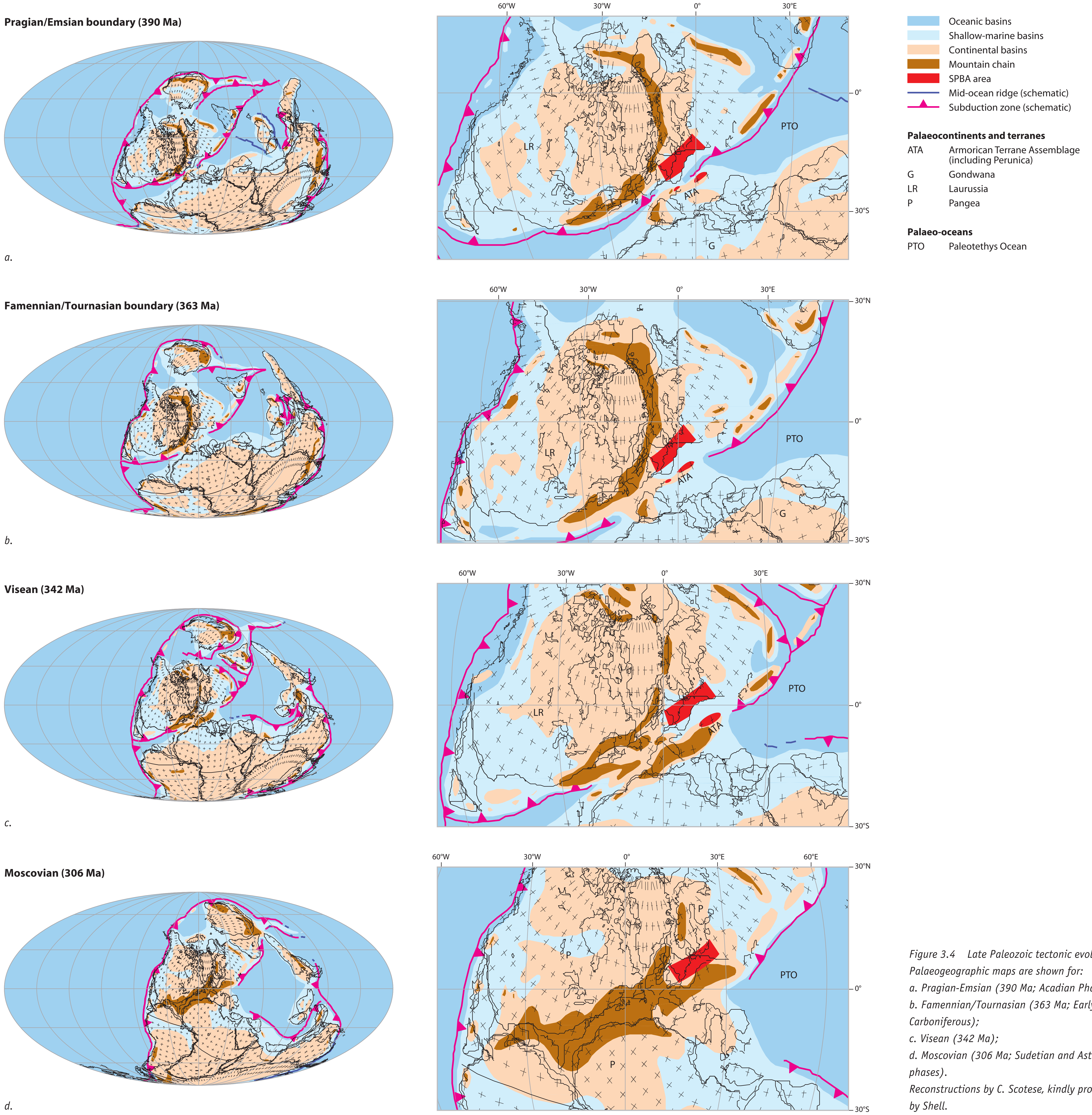


Figure 3.4 Late Paleozoic tectonic evolution. Palaeogeographic maps are shown for:
a. Pragian-Emsian (390 Ma; Acadian Phase);
b. Famennian/Tournasian (363 Ma; Early Carboniferous);
c. Visean (342 Ma);
d. Moscovian (306 Ma; Sudetian and Asturian phases).
Reconstructions by C. Scotese, kindly provided by Shell.

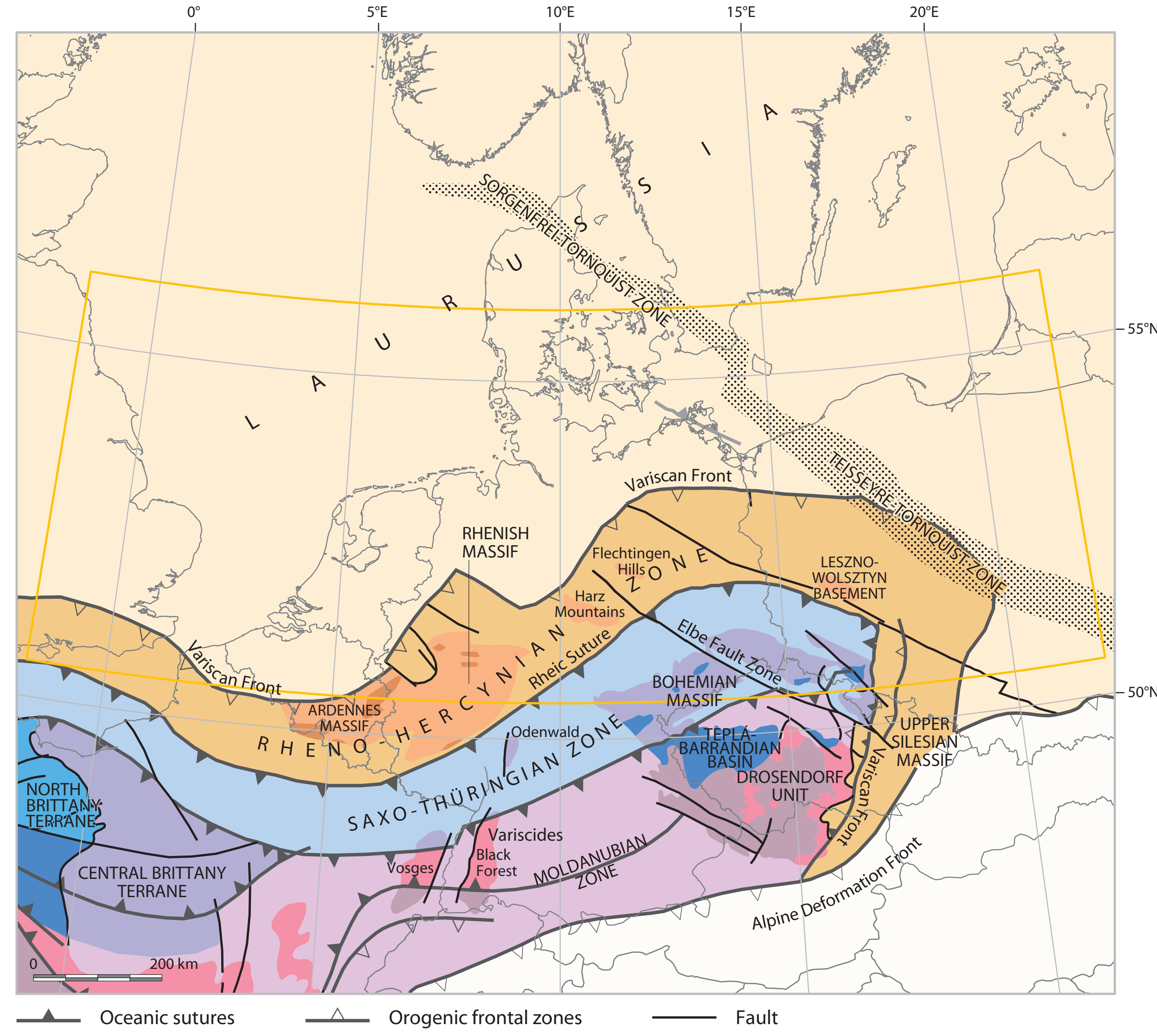
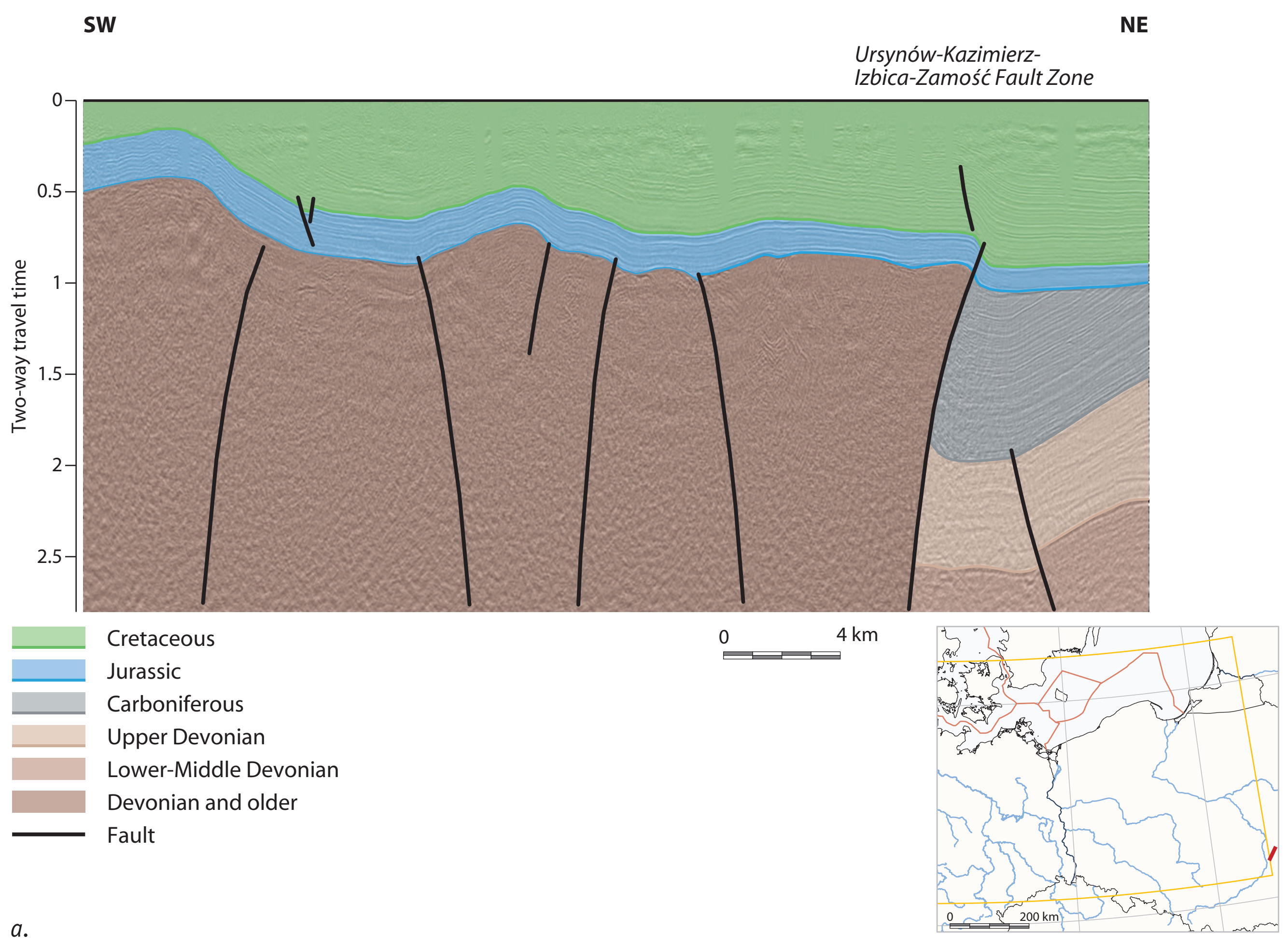


Figure 3.5 Terranes amalgamated to form Pangea. Non-palinspastic map after Pharaoh et al. (2006) and sources therein. Note that the Rheno-Hercynian Zone is interpreted as the Variscan-deformed southern margin of Laurussia following McKerrow et al. (2002).

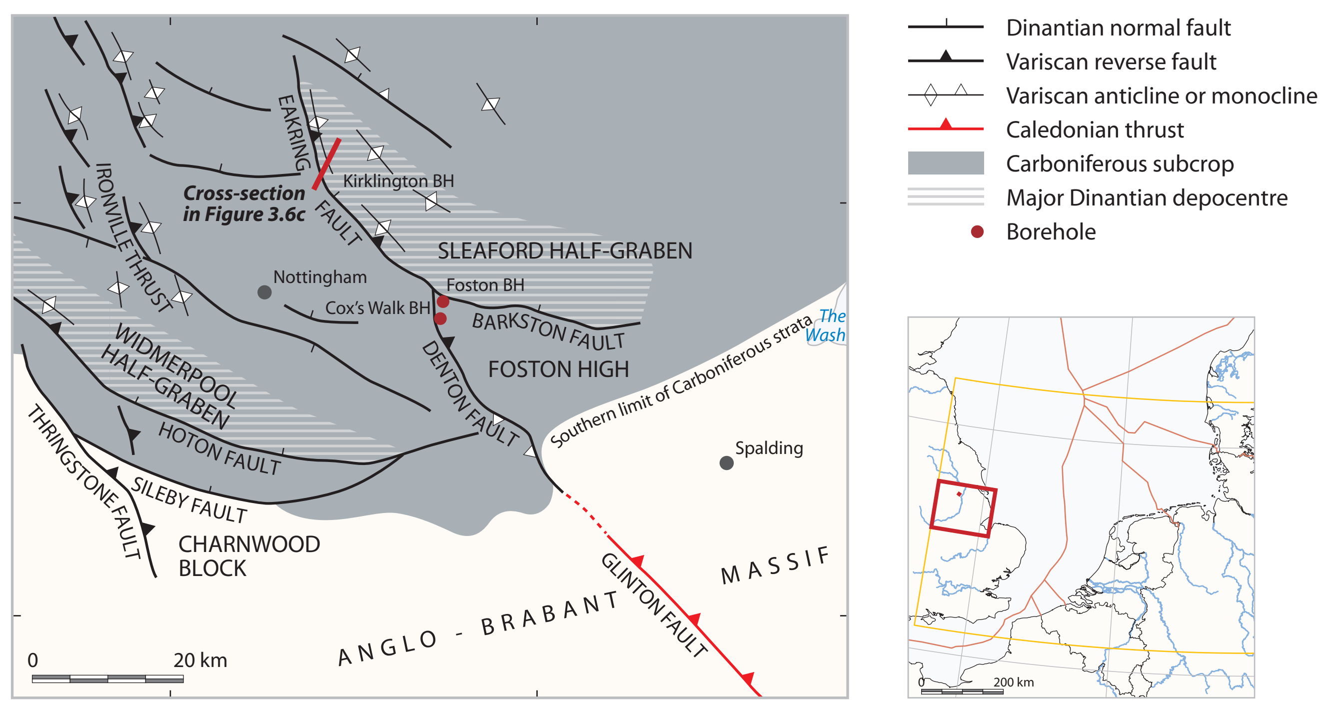
during the late Neoproterozoic breakup of the Rodinia-Pannotia Supercontinent (Dalziel, 1997). Seismic evidence indicates that the TTZ and its north-westward continuation, the Sorgenfrei-Tornquist Zone (STZ) that transects the Precambrian crust, were certainly active by Stephanian to Early Permian times (Priem et al., 1968; Ziegler, 1990a; Thybo, 2000). The geometry of these zones is consistent with dextral strike-slip displacement of up to 20 km (Ziegler, 1990a). During these displacements, the Rønne Graben was located at a releasing bend on this lineament. The Devonian-Carboniferous Lublin Basin straddles the TTZ in the south-eastern corner of the SPB area (cf. Narkiewicz, 2007). The Ursynów-Kazimierz-Izbica-Zamość Fault Zone (**Figure 3.6a**), the present-day south-western boundary of the basin, is a zone of complex reverse faulting that developed during Late Carboniferous inversion.

In the north-west, the northern boundary of the SPB diverges from the Sorgenfrei-Tornquist Zone and is defined by the trend of the Mid North Sea, Ringkøbing-Fyn and Møn highs, which developed during Stephanian to Early Permian times (Ziegler, 1990a). Here, the distinctive characteristics of the crust are the result of emplacement of Caledonian granites adjacent to the Thor Suture. The trend of highs is interrupted by the Central and Horn grabens and Brande Trough. Lower Rotliegend volcanic rocks have been encountered by wells both inside and outside these troughs, supporting the idea that they became significant passageways through the highs only during the Triassic (Best et al., 1983; Vejbæk, 1990; Ziegler, 1990a).

The Pennine Basin of eastern England lies within the Variscan foreland and comprises a number of rhomb-shaped Carboniferous sub-basins, for example the Widmerpool, Gainsborough and Sleaford half-grabens. These basins extend beneath the thin Permian-Mesozoic rocks of the East Midlands Shelf at the western SPB margin. Dinantian extension was largely accomplished by faulting on easterly trends (**Figure 3.6b**). The Eakring Fault (**Figure 3.6c**) shows Dinantian syndepositional growth with doubling of the Dinantian thickness across the structure. It lies along-strike from the north-west-trending Glinton Thrust, a postulated Caledonian basement structure (Chapter 2; Figure 2.8) (Pharaoh, in Chadwick & Evans, 2005), and appears to be an extensional reactivation of this structure. The Eakring Fault displays a reverse throw of 250 m at the Namurian and Westphalian level as a consequence of Late Carboniferous to Early Permian inversion that concentrated on faults with a north-westerly trend (**Figure 3.6b**). Taking this into account, the base-Dinantian throw may have exceeded 750 m. The base of the Permo-Triassic is offset by a normal throw of about 60 m; a very clear example of the strong influence of inherited (in



a.



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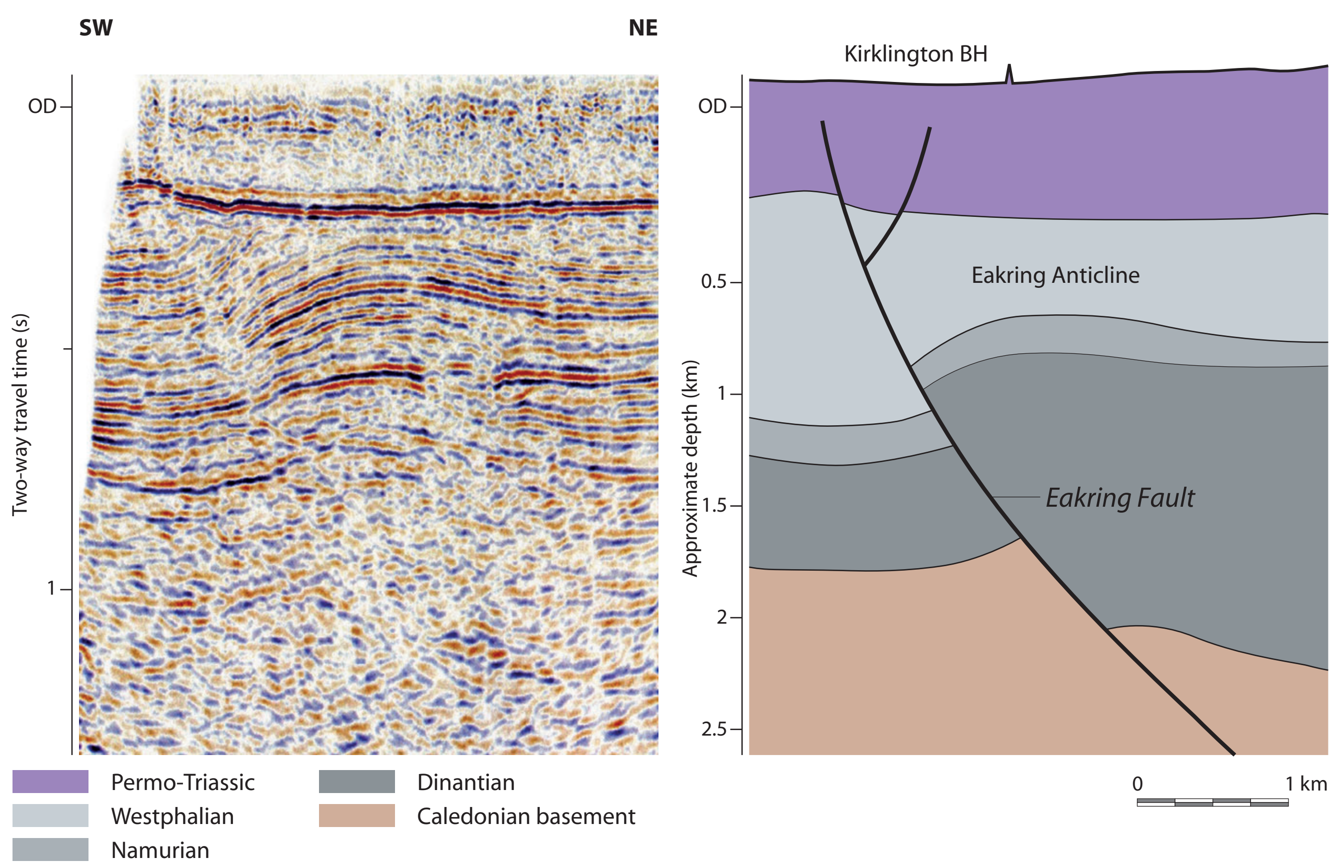

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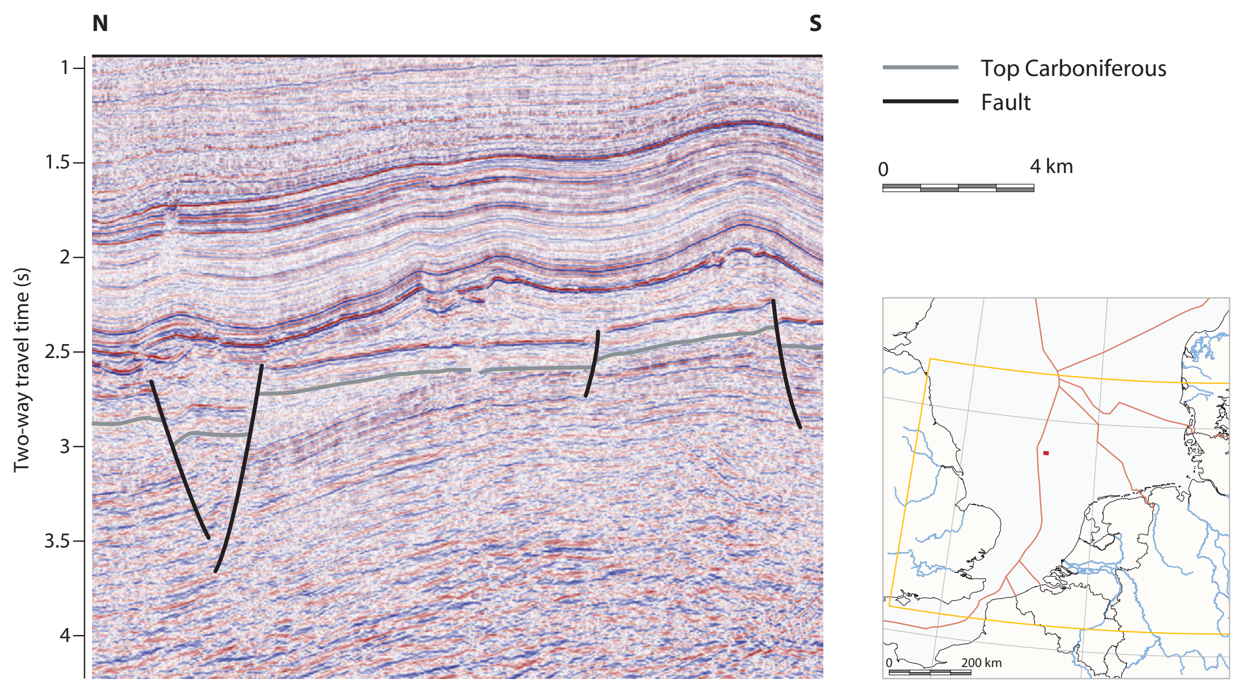

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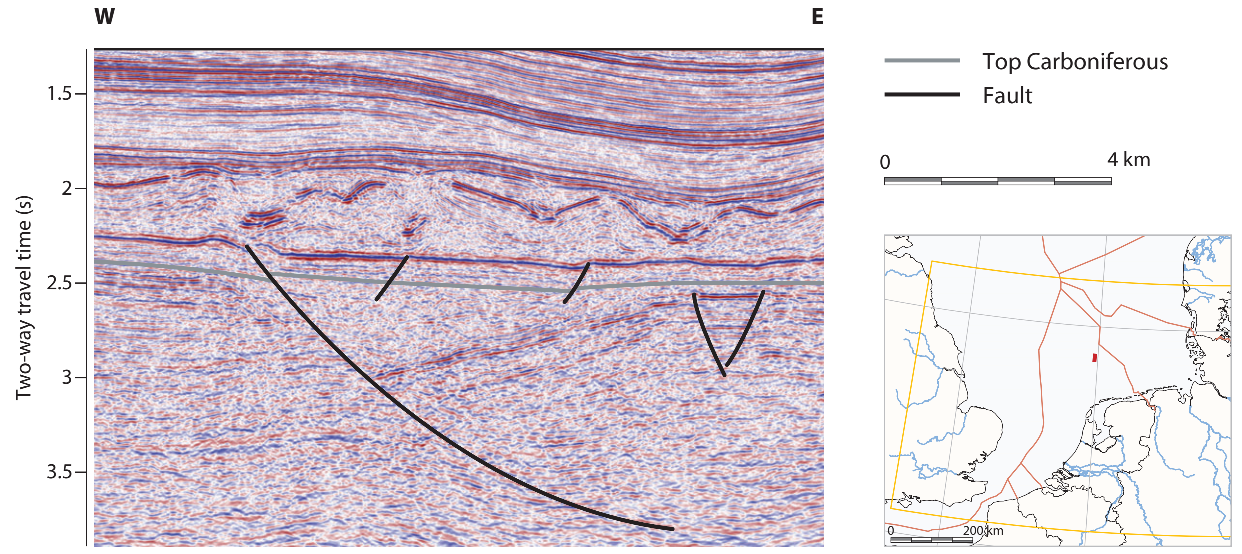


e.

Figure 3.6 *Seismic sections: control of Permian and younger cover by Variscan basement and Permo-Carboniferous magmatic structures. The diagrams show seismic examples of deformation of the Carboniferous foreland basin and one seismic example of the Variscan Front. The examples are mainly from the Dutch southern North Sea area (Cleaver Bank High), but are typical for much of the Carboniferous foreland basin (see for further reading Bartholomew et al., 1993; Hollywood & Whorlow 1993; Oudmayer & De Jager, 1993; Quirk, 1993; Ritchie & Pratsides, 1993; Corfield et al., 1996; Maynard et al., 1997; Quirk & Aitken, 1997; Besly, 1998; Maynard & Dunay, 1999; Smith, 1999; Geluk et al., 2002; Schroot & De Haan, 2003; De Jager, 2007).*

a. Seismic section across the SE margin of the Lublin Basin (Krzywiec, 2007a, 2009);

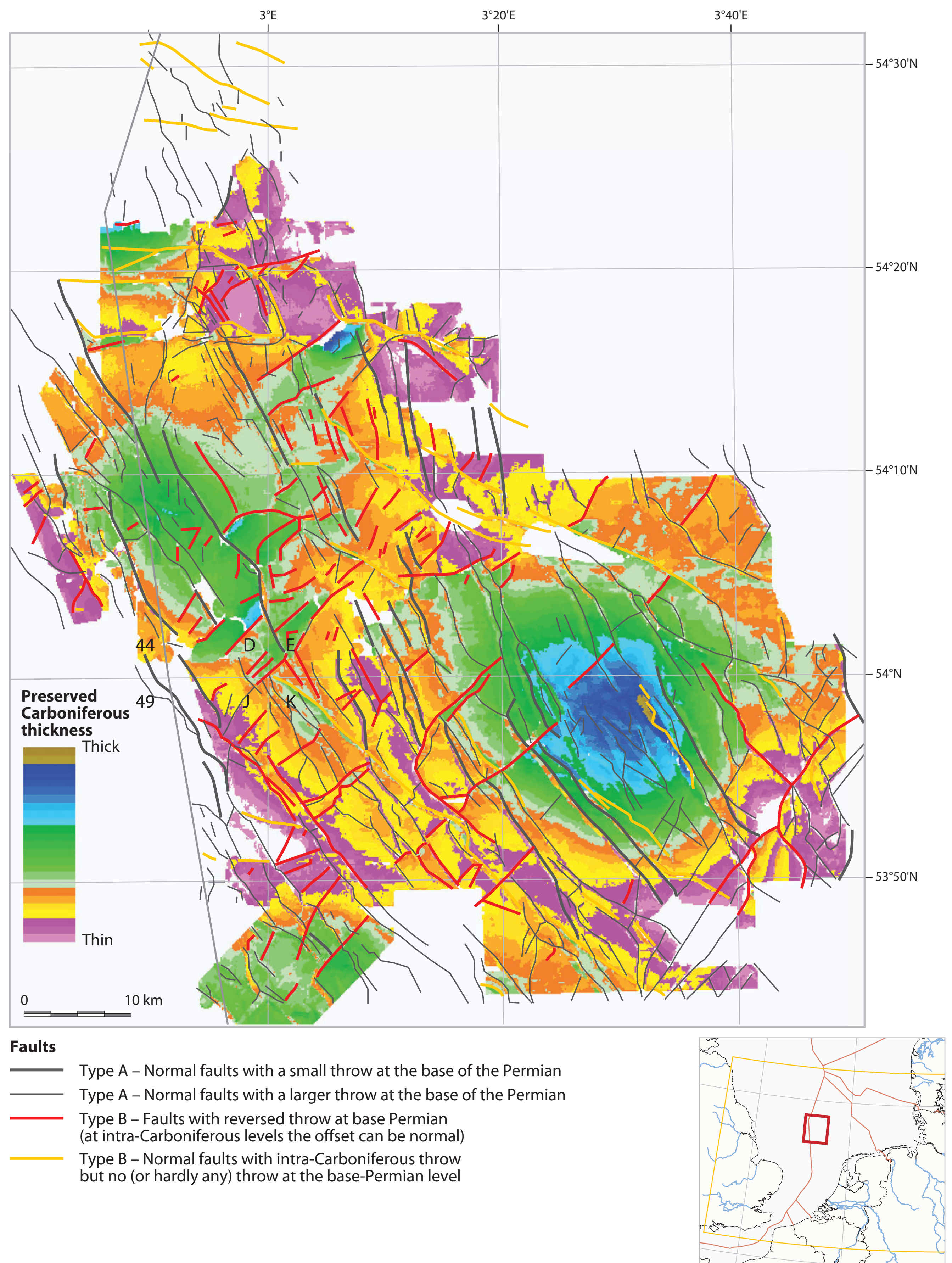
b. Location of the Pennine Basin (Chadwick & Evans, 2005);

c. Seismic section and interpretation for the Eakring Fault (from Chadwick & Evans, 2005);

d. Seismic section from block K5, Dutch offshore sector. The section shows a well-expressed angular unconformity at the base of the Permian. Faulted, westward-dipping Upper Carboniferous strata with distinct continuous high-amplitude seismic facies are truncated below the base of the upper Rotliegend Group. The Base Permian (Saalian) Unconformity here represents a time gap of 40 Ma. Deformation of the Carboniferous strata took place mainly during the latest Carboniferous and Early Permian phase of wrench tectonics (Ziegler, 1990a; Oudmayer & De Jager, 1993; Quirk, 1993; Geluk et al., 2002; Schroot & De Haan, 2003);

e. Seismic section of a small-scale Carboniferous half-graben from block E13 of the Dutch offshore sector. The graben is an example of Late Carboniferous / Early Permian extensional tectonics on the Cleaver Bank High. The listric fault clearly developed after deposition of the Coal Measures (Westphalian A-C). The infill of this graben, drilled by well E/13-1, consists of a succession of sand-prone Westphalian C-D sediments. The shoulders of the sub-graben have been significantly uplifted, with the upper Rotliegend almost directly overlying the Coal Measures. It is not possible to conclusively decide whether the graben was filled syntectonically due to the largely transparent seismic response of the succession. The fact that the sand-prone Westphalian C-D strata show a fair degree of lateral correlation (Besly, 1998; Geluk et al., 2002) suggests that the current geometry is mainly the result of post-depositional movements during the earliest Permian. Note that the NW fault has not been reactivated post-Permian;

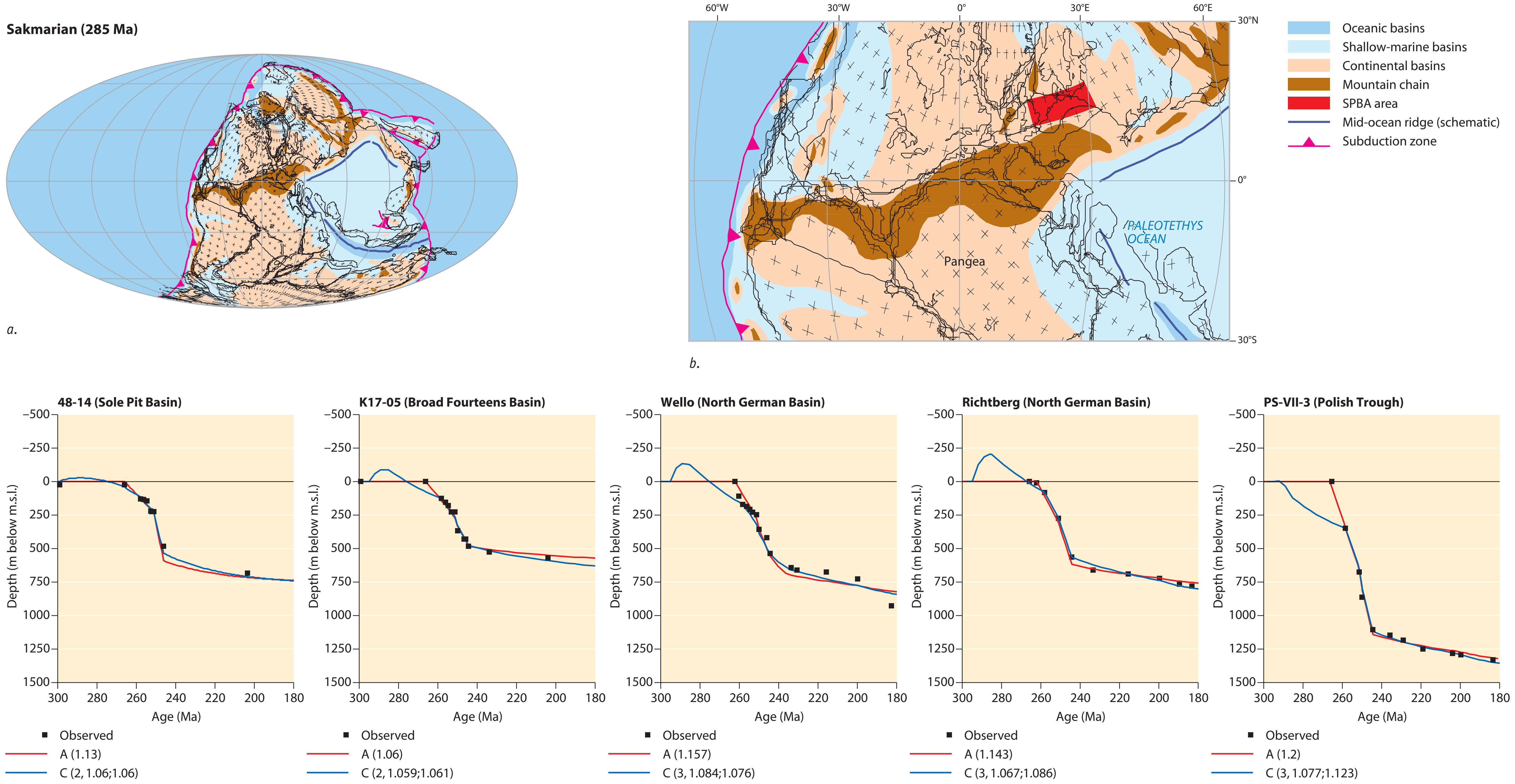
f. Preserved Carboniferous thickness (PCT) in part of the Cleaver Bank High. The map is based entirely on 3D-seismic data (Geluk et al., 2002; Schroot & De Haan, 2003), and shows the thickness of the interval between the top of the Coal Measures and the Base Permian Unconformity. Pink colours represent a low amount of PCT, green to blue colours a high amount of PCT. Active post-Westphalian B and pre-Base Permian Unconformity faults display rapid thickness changes over the fault traces. Later faults do not display these thickness variations. The large-scale structure of this area resembles an egg-box of three lows with a thick succession separated by highs with a strongly reduced succession. In addition to the large structures with a dimension of several tens of kilometres, there are also a number of small-scale structures such as the half-graben shown in the block above the letter E.

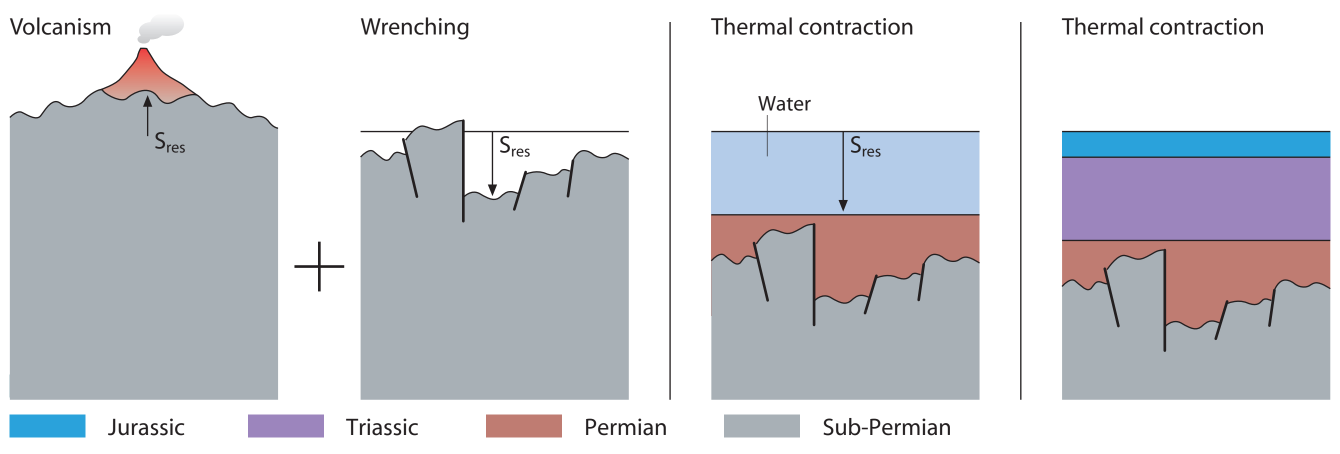


f.

this case, Caledonian) basement structure and the complex history of fault reactivation that may result. It is instructive to bear this in mind when considering the structural influence of the pre-Permian basement where it lies deeper in the SPB beneath thick Zechstein salt. The major Late Permian to Mesozoic basins in the southern North Sea are controlled by faults with a north-westerly trend, such as the Dowsing-South Hewett Fault System and Lower Rhine Lineament, which probably represent reactivations of the Caledonian basement grain (Pharaoh et al., 1995, 2006; Pharaoh, 1999).

The front of the Variscan thrust-and-fold belt runs approximately eastwards through Belgium, just south of the Netherlands, and then north-eastwards into Germany (Ziegler, 1990a) crossing the northern Münster Block. The front turns eastwards in the Bremen-Oldenburg area and has been encountered south of Lüneburg and east of the River Elbe (Pröttlin 1 well), it then crosses the River Odra to the south of Szczecin in Poland (Narkiewicz, 2007). The probable detachment horizon is Namurian black shales, which form a good magnetotelluric conductor that can be traced across the basin at depths between 6000 and 8500 m (Krawczyk et al., 2008b). The tectonic style in the sequence above this detachment is dominated by narrow, complex, thrust anticlines and wide synclines with flat bottoms (Krawczyk et al., 2002). The Devonian rocks are incorporated southwards into the nappe pile (Oncken et al., 2000). In general, the Variscan structure beneath the central SPB area is poorly known. 3D-seismic data show that folding is locally absent; wrench faults with flower structures partition the Variscan foreland into a number of blocks (Lohr et al, 2007; K. Schütz, pers. comm. 2009).





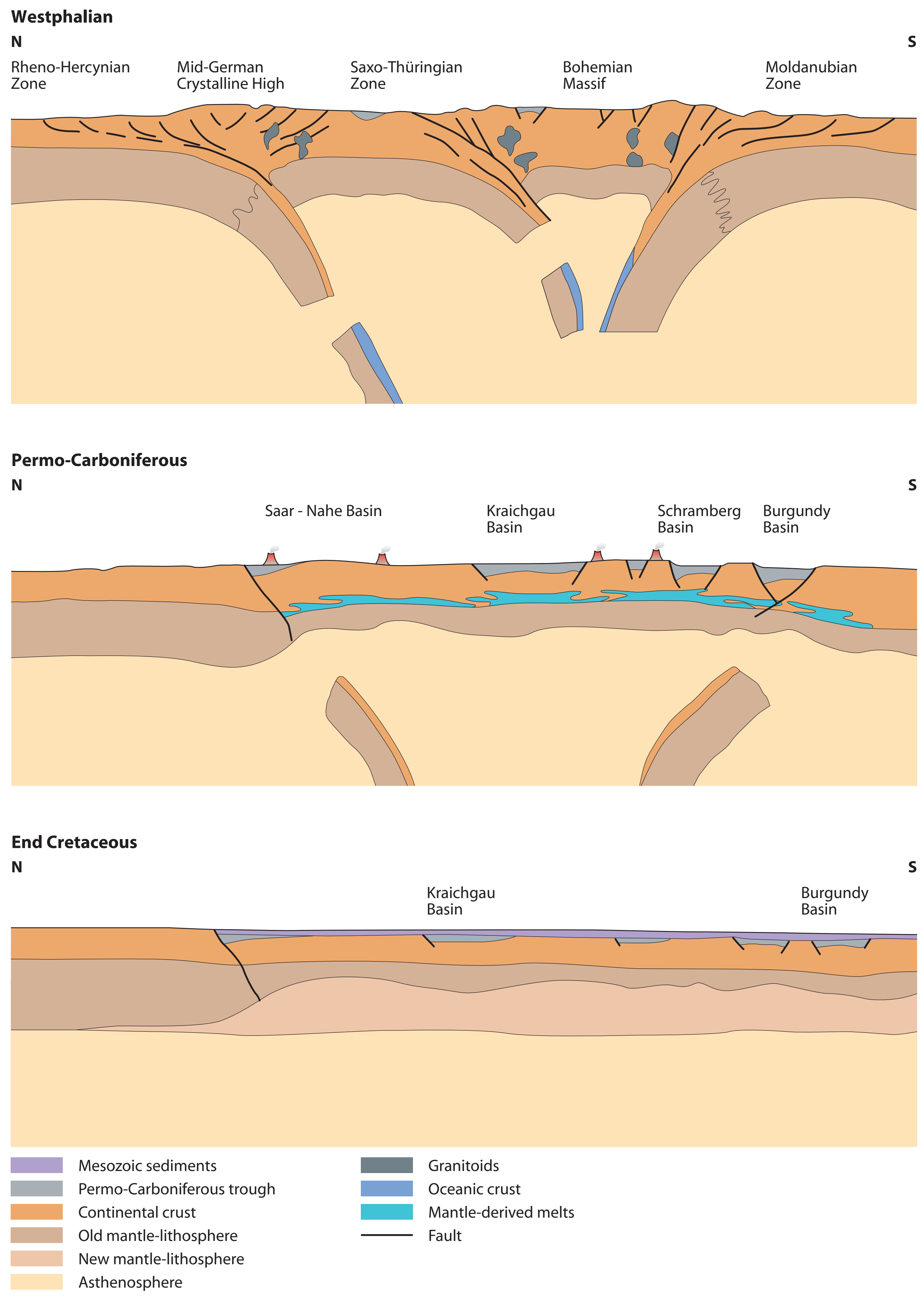


Figure 3.7 Orogenic collapse and Permo-Carboniferous magmatism:
a. Palaeogeographic map of the Sakmarian (285 Ma);
b. Detail of Sakmarian palaeogeographic map;
c. Observed and forward modelled tectonic subsidence curves at different locations in the SPB. Forward modelled tectonic subsidence curves for basin formation scenarios A (uniform stretching) and C (Early Permian mantle upwelling) (subcrustal stretching β > crustal stretching δ). For details, see Van Wees et al., 2000. Numbers correspond to two layered stretching (β and δ values separated by comma) or uniform stretching (β = δ single value). Individual phases are separated by semicolon. Please note that the apparent misfit for the early Permian evolution is not a real misfit but actually depicts the growing accommodation space as outlined by S_{res} in Fig. 3.7d;
d. Modelling philosophy cartoon of model scenario C (cf. Van Wees et al., 2000). Late Permian and Early Triassic subsidence is interpreted as delayed infill of pre-existing accommodation space following orogenic destabilisation followed by lithospheric contraction. S_{res} depicts the evolution of the palaeotopography below global sea level;
e. Conceptual model for the Late Carboniferous to end-Cretaceous evolution of the Variscan lithosphere along a transect from northern Germany to the Jura Mountains in the south (not to scale). Modified after Ziegler et al. (2006).

A gradational contact between Westphalian and Stephanian strata is only observed in the axial parts of the basin, the latter attaining a maximum thickness of about 600 m (Ziegler, 1990a). The Stephanian history of the SPB is poorly understood as this was a time of uplift and erosion that accompanied Variscan compression. Wrench-faulting was accompanied by widespread plutonic and volcanic activity (Heeremans et al., 2000) and may have post-dated it in part, as shown by erosion of the volcanics along the western bounding fault of the Schneverdingen Graben during the Rotliegend (late Havel Subgroup) (K.Schütz, pers. comm. 2009). Wrench-faulting along block margins in the Early Permian (e.g. the Lausitz Block) was contemporaneous with early Rotliegend magmatic activity (**Figure 3.9**). The north-west-trending Elbe, Rostock-Gramzower and Osning fault zones and faults bounding the Harz and Thüringian blocks, developed in response to dextral shearing (Betz et al., 1987). The north-east-trending Rhine Graben, Gifhorn, Emsland and Neuruppin lineaments were also initiated at this time (Ziegler, 1990a). In the Saale Trough, perhaps the most spectacular of the pull-apart basins, more than 2000 m of sediments and volcanics accumulated during the Early Permian, unconformably overlain (Saalian Unconformity) by upper Rotliegend clastic sediments. Similar relationships are observed in intramontane basins within the Bohemian Massif (Suk et al., 1984), for example in the Boskovice Trough.

Stephanian to Early Permian wrench-induced deformation caused deep disruption of the Variscan Foreland Basin, as indicated by the subcrop pattern of Carboniferous rocks at the Variscan Unconformity. The general absence of deep Early Permian transtensional basins in the southern North Sea and northern Germany suggest that transpressional deformation controlled the locally deep truncation of Westphalian and even Namurian strata. This is compatible with compressional features in the Carboniferous as imaged by seismic-reflection data in the north-western area of the southern North Sea (P. Ziegler, pers. comm., 2009). Parts of the SPB to the north of the Variscan Deformation Front were occupied by a Stephanian successor of the Westphalian foreland basin. This basin was open to cyclical marine transgressions from the Moscow Platform to the east. At the same time, wrench movements along the Teisseyre-Tornquist and Sorgenfrei-Tornquist zones activated tensional subsidence and magmatism in the Oslo Graben, starting during Westphalian D times.

3.5 *Pangea, the 'Supercontinent'*

Final suturing of Laurussia (the Old Red Continent) and Gondwana to produce Pangea (**Figure 3.5**) is recorded by the latest Carboniferous (Asturian) and Early Permian (Alleghenian) orogenic phases in Europe and America.

4 Late Paleozoic tectonic evolution: post-orogenic collapse and Permo-Carboniferous magmatism

4.1 *Collapse of the Variscan Orogen*

During latest Carboniferous and Early Permian times, a dextral mega-shear system, which connected ongoing crustal shortening in the Urals and the Appalachians (**Figure 3.4d**), transected the Variscan Orogen and controlled its collapse (Arthaud & Matte, 1977; Ziegler, 1982a, 1988, 1989). The Late Carboniferous palaeo-stress pattern was characterised by northward compression, which activated north-west-trending dextral strike-slip faults and subordinate north-north-east-striking sinistral faults (Mattern, 1996; Lamarche et al., 2002). Finite-element modelling suggests that gravitational collapse of the Variscan Orogen alone can not reproduce the observed timing and magnitude of Permo-Carboniferous crustal thinning, and that additional tensile plate-boundary forces are required (Henk, 1997). The Late Permian to Early Jurassic tectonic-subsidence curves for different parts of the SPB indicate a phase of Permian to Early Triassic rapid (extensional?) subsidence that was followed by asymptotically decreasing thermal subsidence (**Figure 3.7c**). However, a purely extensional model for the SPB is extremely problematic because active faulting during this time is 'minor' and only occurs in restricted areas of well-defined graben structures such as the Glückstadt and Horn grabens, and possibly the Mid-Polish Trough. Rift deformation related to these 'local' graben systems fails to explain the broad east–west saucer shape of the SPB, particularly during the Late Permian and Early Triassic. Recent studies have shown that this configuration can be related to thermal destabilisation of the lithosphere during the Permo-Carboniferous tectono-magmatic event that resulted in the collapse of the Variscan Orogen to the south (Van Wees et al., 2000; Ziegler et al., 2004, 2006).

Quantitative subsidence analysis and forward modelling of lithospheric processes can be used to assess processes governing the transformation of the orogenically destabilised Variscan lithosphere into Mesozoic stabilised cratonic lithosphere. Using inverse techniques to model the subsidence curves (**Figure 3.7c**), a significant component of the observed Late Permian and Triassic tectonic subsidence of the SPB can be explained by thermal relaxation of the lithosphere after its Early Permian thermal thinning, and by delayed in-filling of palaeotopographic depressions that developed during the late Early Permian (Van Wees et al., 2000; Ziegler et al., 2004, 2006; **Figure 3.7d**). In this interpretation, the orogen was characterised at the time of its end-Westphalian consolidation by 45 to 60 km-deep crustal roots that marked major sutures. Subducted lithospheric slabs were detached during the Stephanian to Early Permian wrench-induced collapse of the orogen (**Figure 3.7e**), causing upwelling of the asthenosphere, thermal thinning and/or partial delamination of the lithospheric mantle and regional uplift. By mid-Permian times, the crust had thinned to 28 to 35 km due to regional erosional unroofing, localised mechanical stretching, and the interaction of mantle-derived melts with the basal crust. Thermal subsidence of the lithosphere commenced once the temperature of the asthenosphere had returned to ambient levels (Ziegler et al., 2004). Similarly, Stephanian to Early Permian wrench-faulting in the area of the future SPB to the north of the Variscan Mountains controlled thermal destabilisation of the lithosphere, deep fracturing of the crust, disruption of its sedimentary cover, widespread magmatic activity, and deep erosion as a result of regional uplift. As volcanism and associated wrench tectonics ceased towards the end of Early Permian (Altmark) times, the SPB began to subside in response to thermal contraction of the lithosphere (Ziegler, 1990a; Bachmann & Hoffmann, 1997; Van Wees et al., 2000).

The Northern Permian Basin underwent a similar development with major crustal intrusions and subsequent thermally induced subsidence (Sørensen, 1986; Thybo & Schönharting, 1991; Thybo et al., 2006). During the late Early Permian, the evolving Northern and Southern Permian basins were isolated from the world oceans and apparently had subsided well below the level of the oceans during the deposition of the upper Rotliegend series. At the beginning of the Late Permian, these palaeotopographic depressions were catastrophically flooded by the Arctic seas (Ziegler, 1988). The accumulation of thick Zechstein sediments in the Northern and Southern Permian basins was controlled by continued thermal subsidence; cyclical eustatic sea-level fluctuations were essentially in balance with sedimentation and subsidence rates. Although the effects of Triassic rifting overprinted parts of the SPB area, its overall subsidence pattern persisted well into Jurassic times (Ziegler, 1990a).

4.2 Latest Carboniferous to Early Permian (early Rotliegend) volcanism and plutonism

The folded and imbricated Westphalian and older strata beneath the central SPB are discordantly overlain by Stephanian red beds (the Variscan Unconformity of northern Germany, eastern Netherlands and Dutch offshore sector), and locally by lower Rotliegend sediments (**Figure 3.9**). The latest Carboniferous was a time of intense volcanic and intrusive activity (Neumann et al., 2002; Wilson et al., 2004). Magmatic activity extended from the Oslo Graben to the north of the SPB to the Midland Valley of Scotland, northern England (Whin Sill), the central North Sea, on and around the Mid North Sea and Ringkøbing-Fyn highs, northern Germany, and southwards into the intramontane basins of the Variscan Mountains (Heeremans et al., 2004). Lava and ignimbrite flows more than 2500 m thick accumulated in eruptive centres such as Rügen, the Flechtingen-Altmark region, the Rheinsberg Lineament and the Mecklenburg-Vorpommern region. The flows are generally more than 500 m thick in eastern Lower Saxony, Holstein, Mecklenburg and Brandenburg, but less than 400 m thick to the west in the Bremen and Ems regions (Plein, 1995). Whereas alkaline magmatism is characteristic of the Oslo Rift (Neumann et al., 2002) in both the North German Basin and the intramontane transtensional basins (e.g. Saar-Nahe and Intra-Sudetic basins), it is predominantly intermediate-felsic, calc-alkaline with a strong anatectic component (Romer et al., 2001; Benek et al., 1996; Breitkreuz & Kennedy, 1999). Felsic granitoid intrusions (e.g. Brocken and Ramberg in the Harz Mountains, and the Magdeburg-Roxförde-Velpke-Asse intrusion) were also emplaced at about 295 Ma (Baumann et al., 1991). Deeper-seated intrusions include the South Rügen-Vorpommern granitic porphyry and the Middle Rügen basic intrusion. Other intrusions (e.g. at Erkelenz and Krefeld) are suspected from magnetic (Gabriel et al., 2008) and gravity signatures or coalification anomalies. All intrusions had a considerable impact on the maturity and hydrocarbon potential of Westphalian and older source rocks in their vicinity (>3% Ro at the top of the pre-Permian). Alkaline lavas more than 500 m thick were encountered in boreholes in the Central Graben, whereas lavas in the Horn Graben are mostly rhyolitic (Neumann et al., 2004). Thinner, more marginal volcanic sequences are known from the Ems and Dutch Central grabens and the Outer Rough Basin (Geluk, 2007a). The high vitrinite reflectances in these areas also appear to result from several local pulses of heat (Kettel, 1983).

4.3 The Base Permian (Variscan) Unconformity

In post-Carboniferous times, the area of western Europe became affected by late-Variscan post-orogenic tectonism. Wrench-faulting associated with intrusive and extrusive magmatism and thermal uplift caused widespread deep erosion (Ziegler, 1990a; Ziegler et al., 2004). Broad north-west–south-east-trending swells formed, which can be traced on the subcrop map of Westphalian units at the Base Permian Unconformity (BPU; **Figure 3.1**) (Geluk, 2005; De Jager, 2007; Van Buggenum & Den Hartog Jager, 2007). The north-west–south-east trend that was already established by mid-Paleozoic times was reactivated in response to Early Permian wrench deformation, while a conjugate set of north-east–south-west to north-north-east-trending faults also developed (Ziegler, 1990a). The subcrop pattern of the Westphalian against the BPU already clearly shows the shapes of the Mesozoic basins as areas of less uplift and erosion. The BPU varies in age throughout the SPB (Geluk, 2005), and is in fact an amalgamation of several unconformities (Glennie, 1998a) (See Section 2.1.5. in Chapter 7).

The Saalian Unconformity (Stille, 1924), which separates the volcanics and sediments of the Rotliegend Altmark Subgroup and the Müritz Subgroup sediments in eastern Germany is of Early Permian age. The Altmark I-III unconformities within the Rotliegend series (Hoffmann et al., 1989) are mid- to Late Permian (**Figure 3.1**).

5 Permian tectonic evolution: birth of the Southern Permian Basin system

5.1 Relationship of the Southern Permian Basin to Permian rifts in northern Europe

Definition of possible Early Permian extensional basins beneath the deeper parts of the SPB is hampered by generally poor seismic resolution at pre-Zechstein salt levels. In the western, shallower, parts of the basin, there is no evidence for Early Permian pull-apart basins or grabens. **Figure 3.8c** shows the main basins and plate boundaries during mid-Permian times (late Rotliegend), with a detail of the SPB area in **Figure 3.8d**. Seismic evidence suggests that neither the Central Graben (Ziegler, 1990a) nor the Horn Graben (Vejbæk, 1990, 1997) were active at this stage. Early Permian rifting is evident in East Greenland and on the Mid-Norway shelf (Ziegler, 1988; Brekke, 2000; Coward et al., 2003). The Norwegian-Greenland Sea rift propagated southwards during the Permian (**Figure 3.8c**). A northerly trending chain of Permian basins developed from the North Channel Basin and Mauchline Basin in Scotland, southwards via the East Irish Sea Basin, Cheshire Basin and Worcester Basin, all just west of the SPB area (Ziegler, 1990a; Coward et al., 2003). North-west–south-east-directed extension is inferred from Permian normal faulting in southern Britain (Chadwick & Evans, 1995). Deposition of the earliest Rotliegend sediments in small pull-apart basins close to the Teisseyre-Tornquist Zone (Kiersnowski & Buniak, 2006) and the 'Elbe Fault System' (Bachmann & Hoffmann, 1997; Gast, 1988; Scheck & Bayer, 1999) indicates that subsidence patterns were influenced by dextral shearing along the margins of the basin system. Inferred earliest Permian extensional

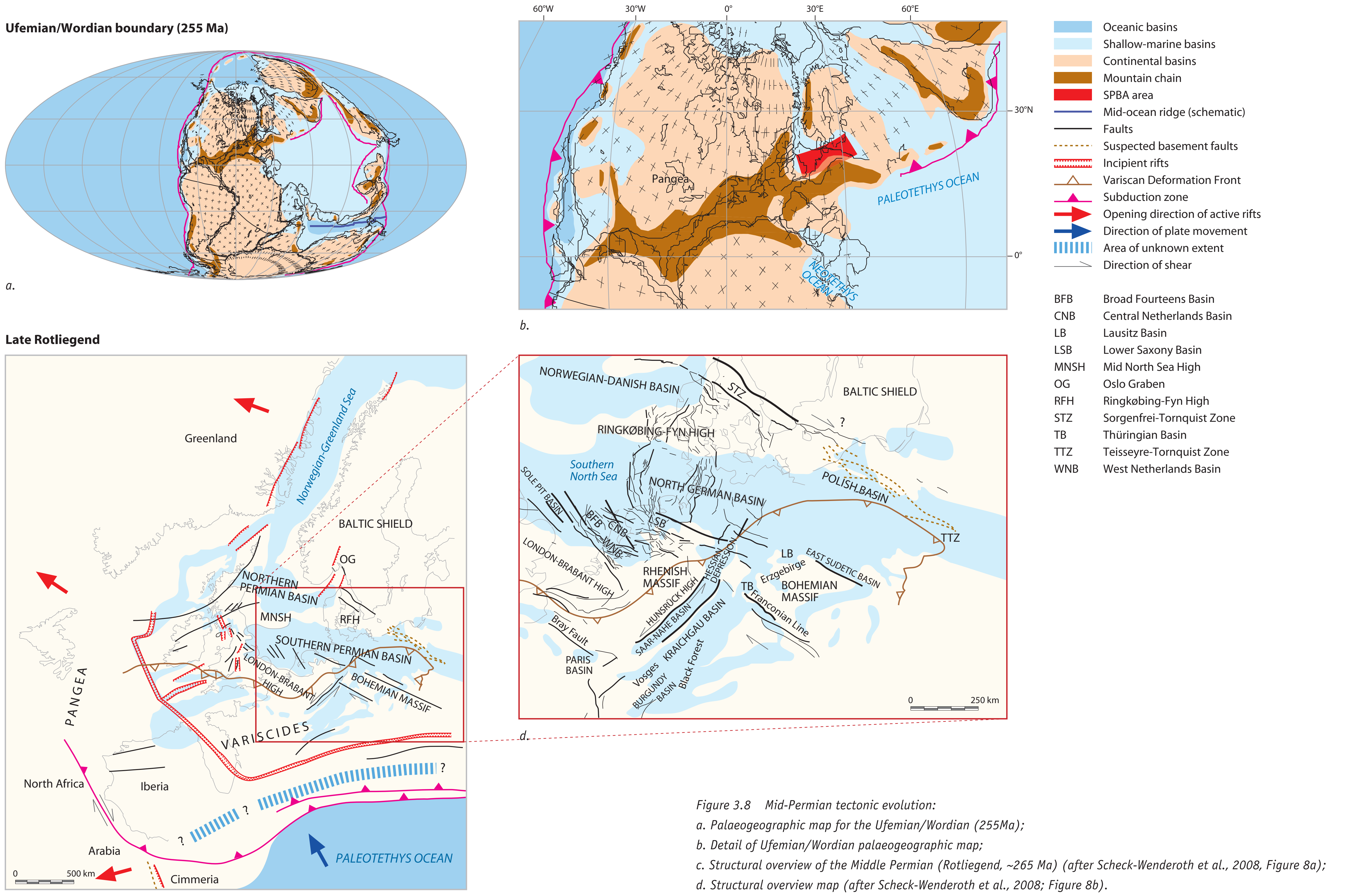


Figure 3.8 Mid-Permian tectonic evolution:
a. Palaeogeographic map for the Ufemian/Wordian (255Ma);
b. Detail of Ufemian/Wordian palaeogeographic map;
c. Structural overview of the Middle Permian (Rotliegend, ~265 Ma) (after Scheck-Wenderoth et al., 2008, Figure 8a);
d. Structural overview map (after Scheck-Wenderoth et al., 2008; Figure 8b).
Palaeogeographic reconstructions by C. Scotese, kindly supplied by Shell.

structures in northern Germany (Gast, 1988; Plein, 1995) and the Oslo Graben trend north-north-easterly and indicate a component of north-west-directed extension at the termination of the main wrench-faults. Following the Stephanian to Early Permian tectono-magmatic cycle, the SPB area started to subside in response to thermal re-equilibration of the lithosphere-asthenosphere system (Van Wees et al., 2000; Ziegler et al., 2004). During this process, upper Rotliegend sediments up to 2300 m thick accumulated in the axial parts of the North German Basin (Plein, 1995; Bachmann & Hoffmann, 1997; Scheck & Bayer, 1999) and about 1600 m in the Polish Basin (Dadlez et al., 1995; Kiersnowski et al., 1995; Dadlez, 1998b; Lokhorst et al., 1998; Scheck-Wenderoth & Lamarche, 2005; Kiersnowski & Buniak, 2006). The trend of the Mid North Sea, Ringkøbing-Fyn and Møn highs separates the SPB area from the Northern Permian Basin (Norwegian-Danish Basin) (**Figure 3.8c**) in which Rotliegend clastics were up to 600 m thick (Ziegler, 1990a; Lokhorst et al., 1998; Evans et al., 2003). In the internal parts of the Variscan Orogen of southern Germany and Poland, deposition of continental clastics during the Early Permian was restricted to often narrow, variably striking, transtensional intramontane basins developed on a Permo-Carboniferous wrench-fault template. The Saar-Nahe Basin (Stollhofen & Stanistreet, 1994; Schäfer & Korsch, 1998; Kukulus & Henk, 1999) Saale Basin (Rappsilber, 2003) and the Burgundy-Kraichgau Basin (Freudenberger, 1996a; Ziegler, 1990a) preserve a thickness of several hundred metres of Rotliegend sediments and volcanics. Similar structures developed within the Intra-Sudetic domain in eastern Germany and southern Poland (Dadlez, 1998b), where normal faulting took place along north-west-striking faults.

5.2 Late Early to Late Permian (late Rotliegend) basin evolution

For purely descriptive purposes, the saucer-shaped SPB was subdivided into three components by Ziegler (1990a): the Southern North Sea, North German and Polish sub-basins (see Figure 7.3), the latter two separated by a prominent saddle in the vicinity of the present-day German-Polish border. A similar approach is followed in this chapter, with the addition of a further element, the Fennoscandian Border Zone. Note that the boundary between the Southern North Sea and North German sub-basins is gradational and varies from epoch to epoch.

5.2.1 North German Basin

The earliest known strata, the Müritz Subgroup, are only preserved in central Germany. The Altmark I (pre-Parchim) pulse led to formation of a northerly striking graben system (**Figures 3.1** & **3.9**) in eastern Lower Saxony (Gast, 1988), which was subsequently filled by volcanic fanglomerates and dune sands with good reservoir potential. Some of the grabens in the Flechtingen area and south-east Brandenburg are clearly older. In the northern Hessian Depression, between Göttingen in the south and Hamburg in the north, Lower Permian clastics and volcanics were deposited in a fan-shaped system of grabens (**Figure 3.9**) (Gast, 1988). The basin depocentre migrated westward into West Mecklenburg during deposition of the Mirow Formation; salt was deposited in the southern Schleswig-Holstein, northern Lower Saxony and eastern North Sea areas. The Parchim graben system was covered by nearshore, fluvial and sheet-flow sediments (Rieke et al., 2003). According to F. Kockel (pers. comm., 2009), north-north-east-trending swells such as the Hunte, Hunsrück-Oberharz and Eichsfeld-Altmark swells, which are partly devoid of Rotliegend clastics, and rifts such as those of the Ems-Horn, Glückstadt and Brunswick-Gifhorn areas, may have compartmentalised the west–east-trending SPB. However P. Ziegler (pers. comm., 2009) and K. Schütz (pers. comm., 2009) considered this very unlikely as the compartments are not reflected in the Zechstein facies patterns. The North German Basin gradually assumed an asymmetric saucer-shaped geometry with a steeper northern flank and a gentler southern flank (Ziegler, 1990a), the margins of which were progressively overstepped by younger parts of the sequence (Plein, 1978; Gast, 1988; Hoffmann et al., 1989). The North German Basin depocentre is slightly offset to the north of the area of greatest Early Permian magmatic activity and maximum crustal thinning (Plein, 1978; Bachmann & Grosse, 1989; Franke et al., 1989; Bachmann & Hoffmann, 1997), the area where the influence of the Baltic Shield ceases (DEKORP-BASIN Research Group, 1998, 1999).

5.2.2 Southern North Sea Basin

The Southern North Sea Basin province comprises several sub-basins with distinct subsidence and inversion histories throughout Permian and Mesozoic times (De Jager, 2007), although these elements were not strongly differentiated in mid-Permian times (**Figure 3.9**). Evolution of the Southern North Sea Basin area during the late Early Permian was dominated by thermal subsidence and its gradual incorporation into the westward-expanding Rotliegend Basin (Geluk, 2007a). In contrast, the Pennine High, Mid North Sea High and the London-Brabant Massif formed persistent highs that fringed the evolving Southern North Sea Basin to the west, north and south. Deposition of Rotliegend strata commenced later in this basin than in the North German Basin (Ziegler, 1990a; Geluk, 2007a) and was accompanied by minor faulting (Van Wijhe et al., 1980; Glennie & Boegner, 1981). In the Netherlands, the forerunners of the Central Graben, Broad Fourteens Basin and Lauwerszee Trough are subtly reflected in the pattern of Rotliegend isopachs (De Jager, 2007) (see Figure 7.3). According to H. Mijnlief (pers. comm., 2008), the Lauwerszee Trough has a Rotliegend succession several tens of metres thicker than that of the adjacent Friesland Platform and Groningen High (**Figure 3.9**). It is likely that the major faults delineating the Terschelling Basin and the Hantum and Rifgronden fault zones were already in existence in Early Permian times (De Jager, 2007); the Texel-IJsselmeer, Cleaver Bank and Elbow Spit highs were present but poorly defined (Van Hoorn, 1987; NITG-TNO, 2004). Thickness maps (see Chapter 7, Figure 7.3) show that Rotliegend strata are up to three times thicker in the Sole Pit Basin compared to surrounding areas. The Dowsing-South Hewett Fault Zone, which was active during the Carboniferous (Leeder & Hardman, 1990) and possibly earlier times (Pharaoh et al., 1995), may have exerted a syndepositional control. Rotliegend strata were not deposited on the Mid North Sea and Ringkøbing-Fyn highs, as indicated by their onlap geometry and that of the Zechstein series. They are also missing from the London-Brabant Massif, which apparently acted as a source area for the Rotliegend series in the North Sea.

5.2.3 Polish Basin

During the Permian stage of its evolution, the axial part of the Polish Basin, the Mid-Polish Trough, formed the easternmost part of the SPB (Kiersnowski et al., 1995; Kiersnowski & Van Wees, 2000; Kiersnowski & Buniak, 2006; McCann et al., 2006). With very few exceptions (e.g. Antonowicz et al., 1994), there is a lack of reliable seismic information on the basin's internal structure at sub-Zechstein levels. Only borehole data can therefore be used to infer tectonic activity during the late Early Permian initial subsidence phase of the Polish Basin. A recent model of sub-Zechstein basement tectonics shows a complex array of north-west-striking fault zones (cf. Krzywiec et al., 2006; Krzywiec, 2006a, 2006b), which were repeatedly reactivated during Mesozoic subsidence and subsequent inversion of the trough. Mild transtensional activity along fault systems of the Teisseyre-Tornquist Zone during the early phases of the early Rotliegend has been documented by Kiersnowski & Buniak (2006) who also showed that fault activity gradually diminished with time and had essentially ceased prior to the transgression of the Zechstein Sea. Anhydrite-shale sequences were deposited in ephemeral-lake or sabkha environments in the central Polish Rotliegend Basin (Pokorski & Wagner, 1975).

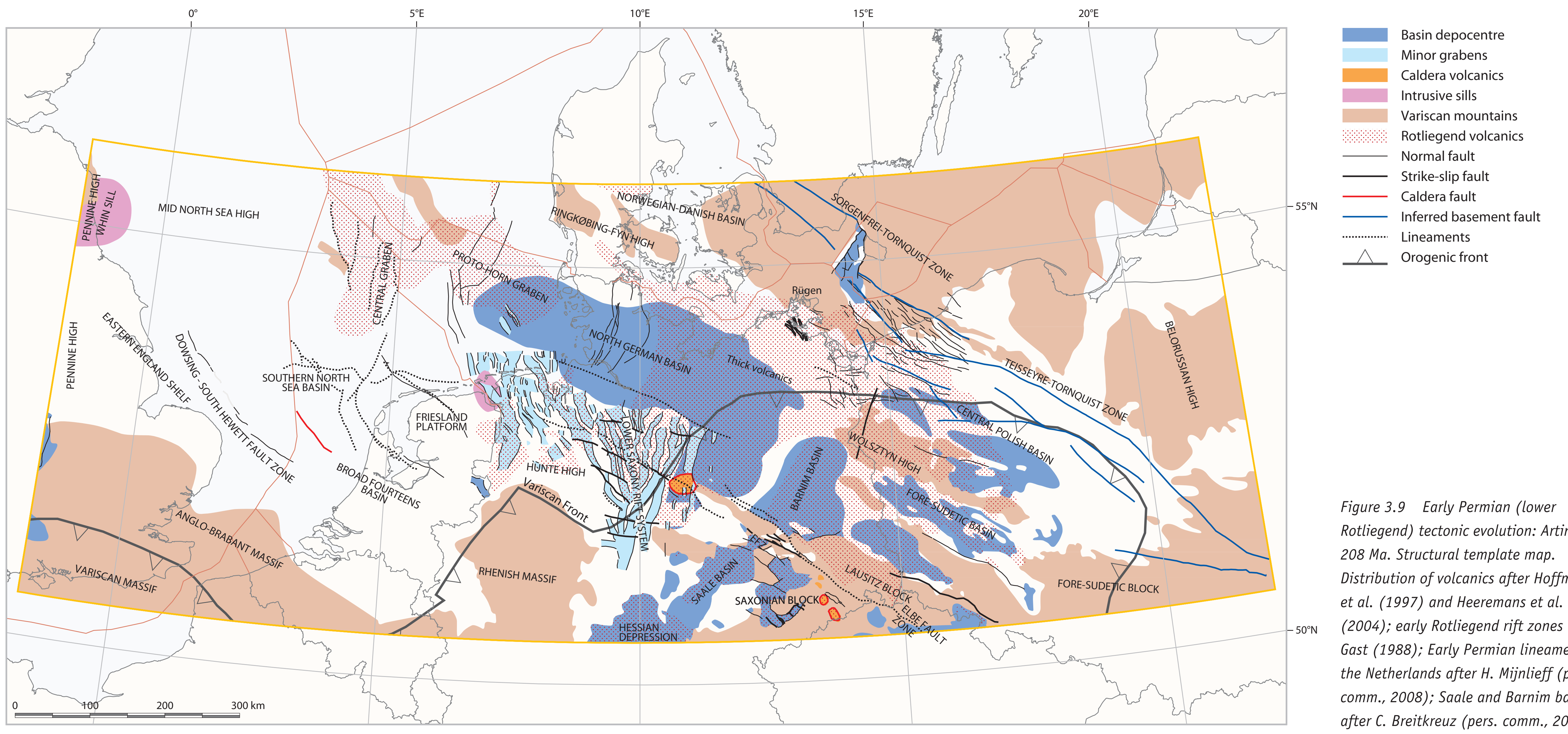


Figure 3.9 Early Permian (lower Rotliegend) tectonic evolution: Artinskian, 208 Ma. Structural template map. Distribution of volcanics after Hoffmann et al. (1997) and Heeremans et al. (2004); early Rotliegend rift zones after Gast (1988); Early Permian lineaments in the Netherlands after H. Mijnlieff (pers. comm., 2008); Saale and Barnim basins after C. Breitkreuz (pers. comm., 2008).

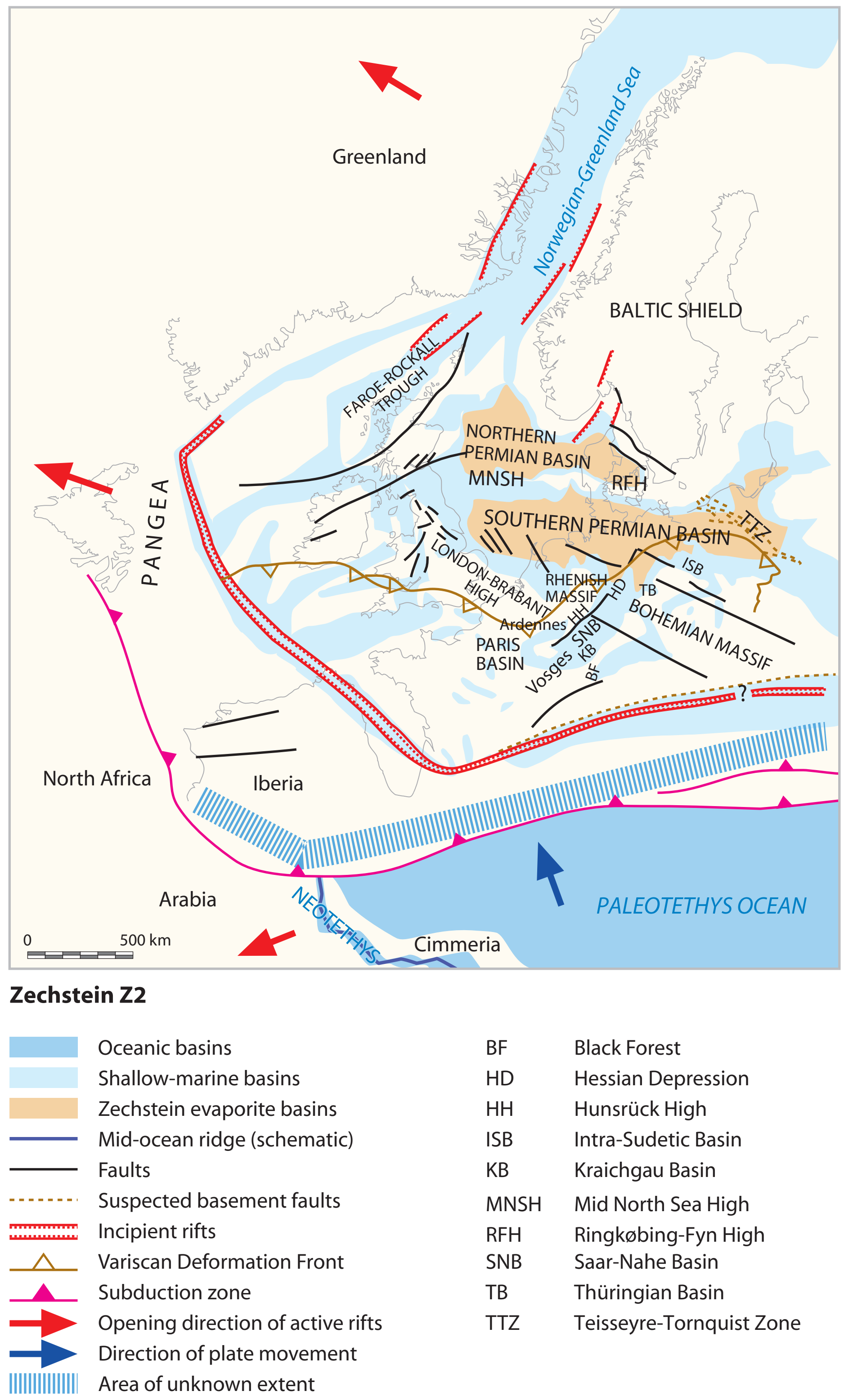


Figure 3.10 Structural overview map for the Late Permian (Zechstein, ~255 Ma) (after Scheck-Wenderoth et al., 2008; Figure 9a).

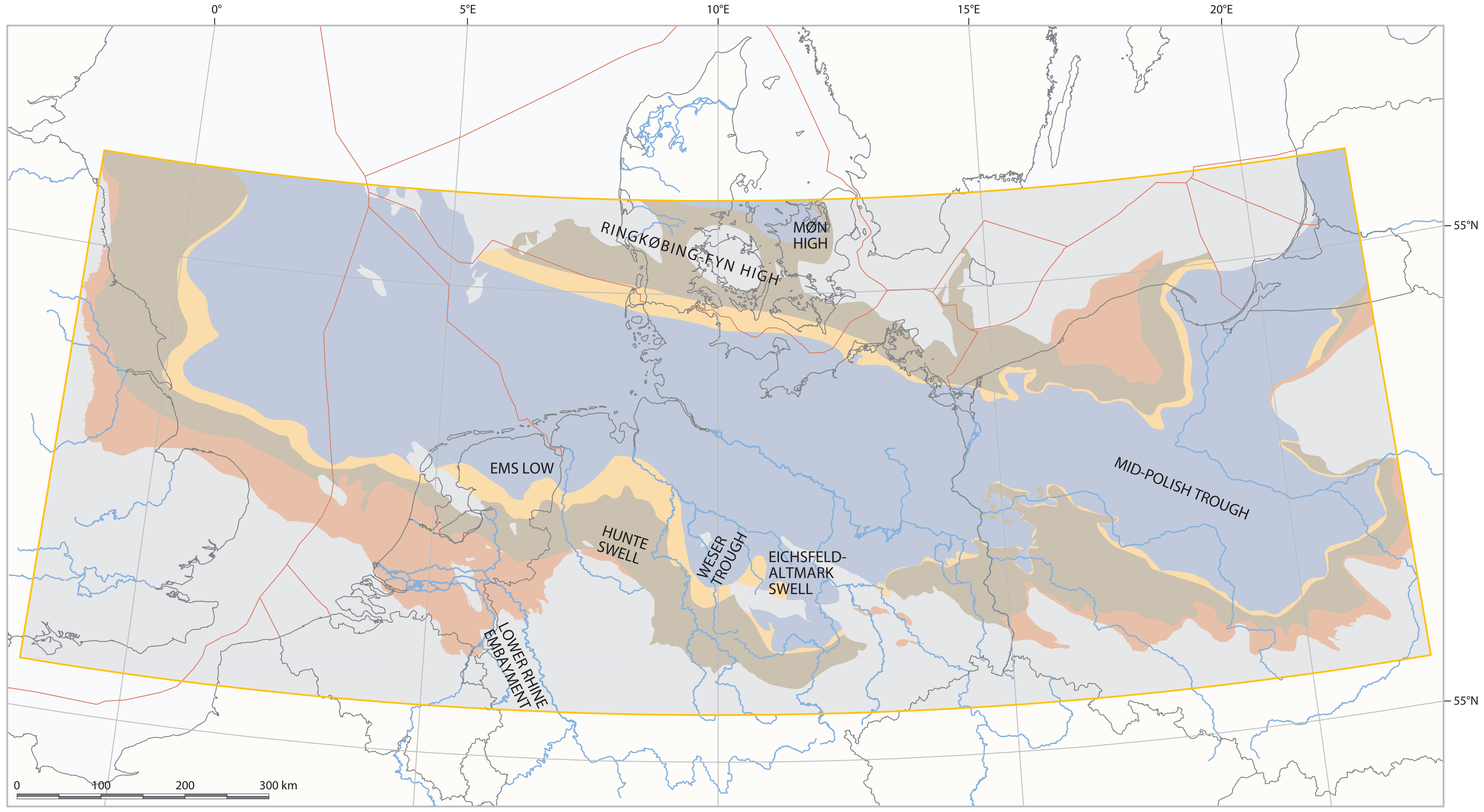
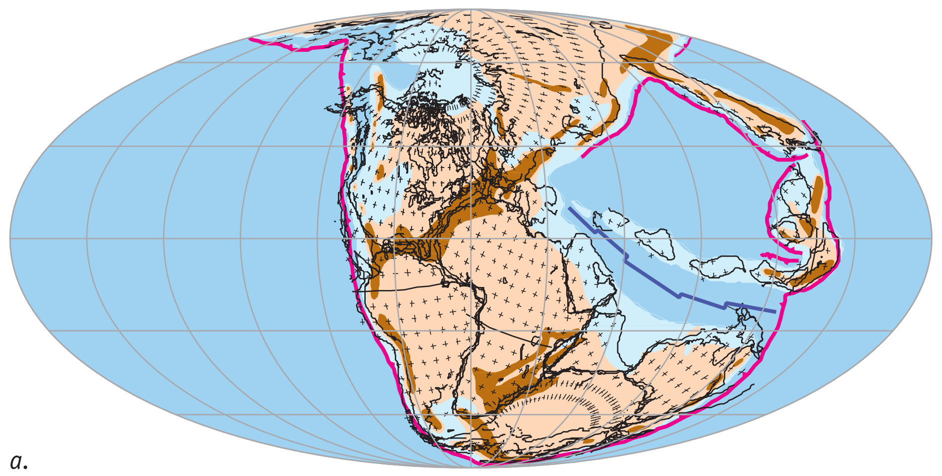
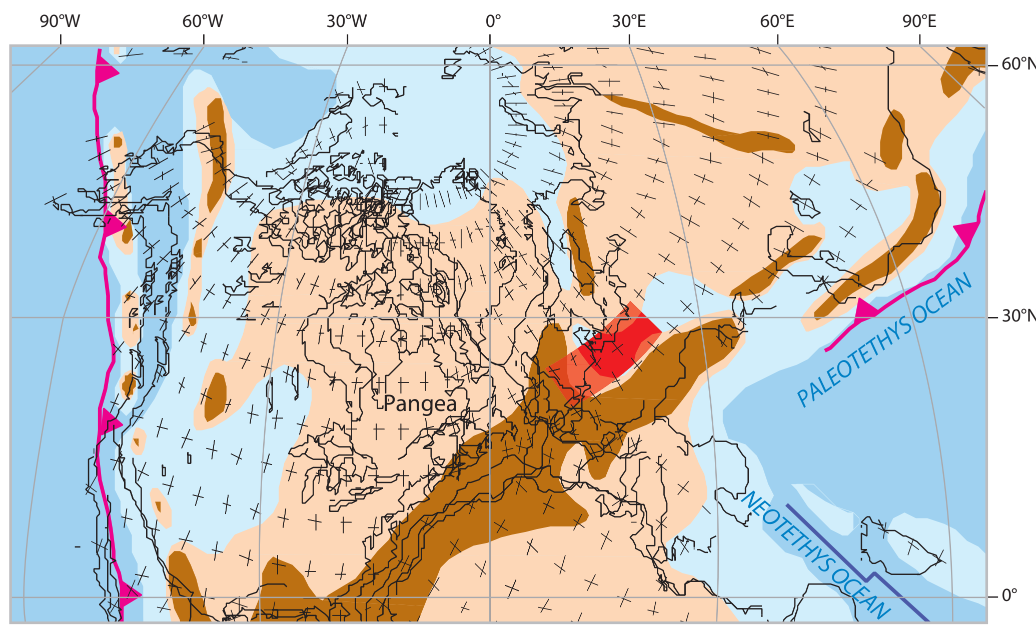


Figure 3.11 Late Permian (Zechstein Z2) tectonic evolution: Wuchiapingian, 255 Ma.

Ladinian (237 Ma)

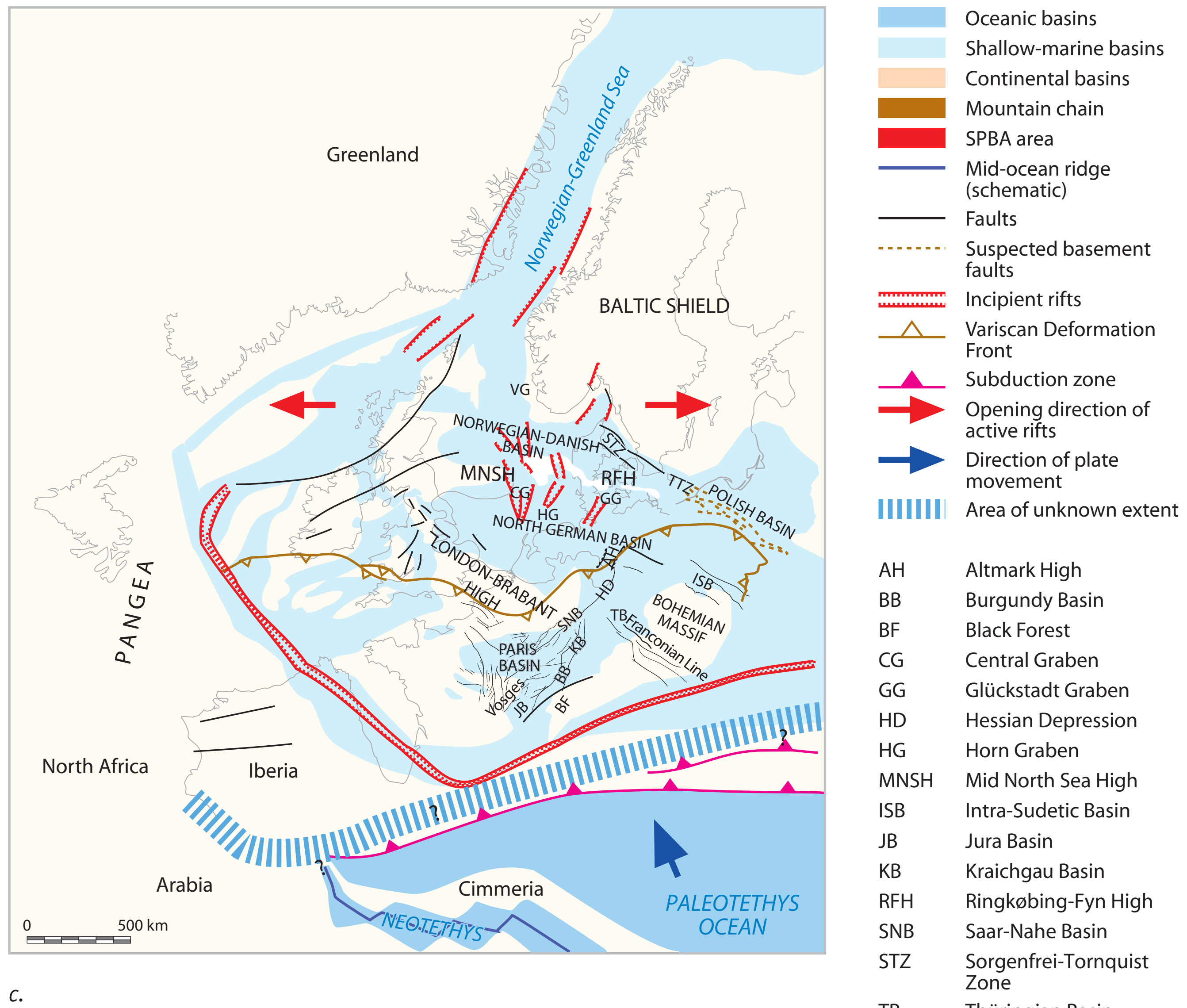


a.



b.

Early Triassic (~245 Ma)



c.

Figure 3.12 Early to Mid-Triassic tectonic evolution:
a. Palaeogeographic map for the Ladinian (237Ma);
b. Detail of Ladinian palaeogeographic map;
c. Structural overview map for the Early Triassic (Buntsandstein, ~245 Ma) (after Scheck-Wenderoth et al., 2008; Figure 9b).
Palaeogeographic reconstructions by C. Scotese, kindly supplied by Shell.

5.3 *Late Permian (Zechstein) tectonic evolution*

The Late Permian plate-tectonic setting of Europe (**Figure 3.10**) was dominated by the ongoing northward subduction of the Paleotethys Ocean beneath the southern margin of Europe (Stampfli & Borel, 2002). Persisting activity along the Arctic-North Atlantic rift system, combined with a glacioeustatic sea-level rise, facilitated the development of a seaway linking the Arctic Ocean with the Permian basins of western and central Europe (Ziegler, 1988, 1989, 1990a; Torsvik et al., 2002; Coward, et al., 2003). This transgression probably advanced to the margins of the Northern and Southern Permian basins through the Faroe-Rockall Trough, the fault-controlled basins of the Irish Sea and via the Solway-Vale of Eden Depression. The latter was connected during the Zechstein via the western flank of the Mid North Sea High, and later across it (Ziegler, 1990a; Taylor, 1998). There is no evidence for subsidence of the Central and Horn grabens as the Zechstein (Z2) Hauptdolomit banks trend east–west across the grabens (P. Ziegler, pers. comm., 2009).

By the beginning of the Late Permian, the Northern and Southern Permian basins had apparently subsided below global sea level. Although repeated temporary marine incursions into the SPB during late Rotliegend times have been recorded (Legler & Schneider, 2008), the deposition of the Kupferschiefer (Copper Shale) is evidence of catastrophic and permanent flooding of the Permian basins by the Zechstein transgression. This highly organic unit was deposited in basinal areas well below wave base under permanently stagnant (euxinic) bottom-water conditions in water depths of 200 to 300 m or more (Ziegler, 1990a). The Kupferschiefer has a relatively uniform thickness of about 0.5 m throughout the basin, marking a time of tectonic quiescence (Paul, 2006).

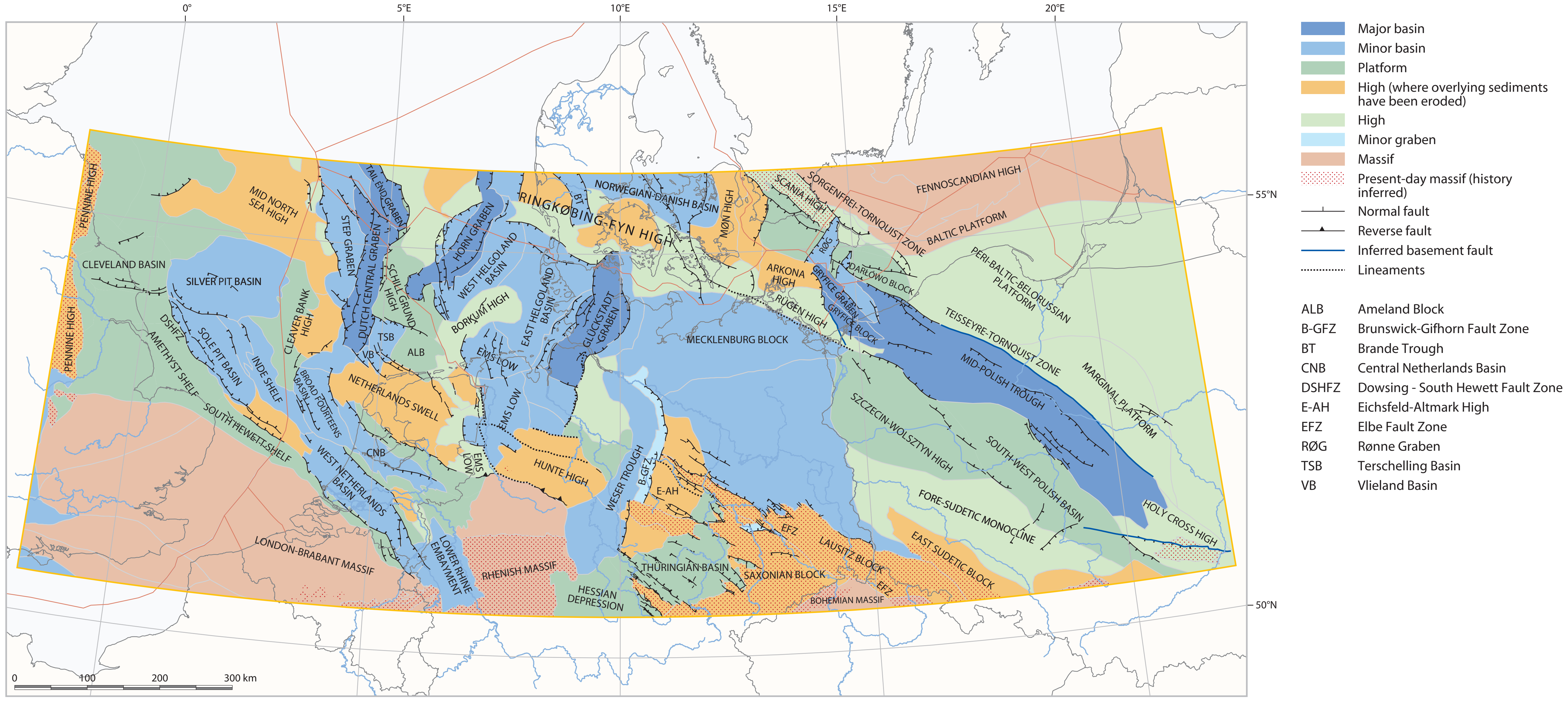
Late Permian evolution of the SPB continued to be dominated by thermal relaxation of the lithosphere (Ondrak et al, 1999; Van Wees et al., 2002; Scheck et al., 2003b; Scheck-Wenderoth et al., 2008) with crustal extension playing only a minor role. Cyclical glacioeustatic sea-level fluctuations controlled sedimentation patterns in both the Northern and Southern Permian basins (**Figure 3.10**). Carbonate and evaporite banks developed along the basin margins and on isolated highs during sea-level highstands, whereas basinal areas were starved of sediment; basinal areas were filled by thick halites during sea-level lowstands. Minor faulting reflecting east–west extension is reported from the southern margin of the SPB in the Netherlands and Germany (Ziegler, 1990a; Geluk & Röhling, 1997; Geluk, 2007a). Post-depositional salt mobilisation masks possible synsedimentary faulting in basinal areas (Scheck-Wenderoth et al., 2008). Facies development of the Zechstein and its onlap onto the Mid North Sea and Ringkøbing-Fyn highs indicate that these partly separated the Northern and Southern Permian basins (Vejbæk, 1997; Taylor, 1998; Clausen & Pedersen, 1999). In southern Germany and Poland, the Zechstein Sea intermittently inundated only the largest of the relatively slowly subsiding intramontane troughs (Ziegler, 1990a; Freudenberger, 1996a; Dadlez et al., 1998). Differences observed in Late Permian subsidence rates between basinal parts of the SPB and its flanking areas are attributed to differences in the magnitude of thermal destabilisation of the lithosphere during the Permo-Carboniferous tectono-magmatic cycle (Wilson et al., 2004; Ziegler et al., 2004).

Cyclic evaporation led to the deposition of the more than 1500 m-thick, halite-dominated Zechstein sequence in the axial part of the SPB area in northern Germany (Ziegler, 1988, 1990a). The northern margin of the basin was rather straight along the southern slope of the Ringkøbing-Fyn and Møn highs (**Figure 3.11**). In the south, the basin margin was differentiated into several north-north-east-trending swells and lows, such as the Ems Low, the Hunte Swell, the Weser Trough and the Eichsfeld-Altmark Swell. These structures were still evident during the first and second Zechstein depositional cycles and controlled the development of marginal carbonate and evaporite banks. With the progressive widening of the basins, these banks were overstepped and apparently ceased to influence facies developments. The Z2 (Stassfurt) cycle deposited the thickest sequence of evaporites; later cycles overstepped farther onto the margins of the basin, as recognised in Denmark, and as many as seven cycles have been recognised in the basin axis (Z1 to Z7; see Chapter 8). Due to later halokinetic movements, little is known about contemporaneous rifting. West-north-west-trending grabens are reported from the Z1 (Werra) cycle (Tubantian I) in the eastern Netherlands (Geluk, 2005) and the Lower Rhine Embayment, as well as the Z2 cycle on the Hunte Swell. A later (Tubantian II) phase of faulting postdates the Z4 (Aller) cycle in the Netherlands (Geluk, 2005) and Germany (F. Kockel, pers. comm., 2009). The fill of the Mid-Polish Trough includes a thick Zechstein salt layer within the north-western and central segments of the basin (Marek & Pajchlowa, 1997; Dadlez et al., 1998). As elsewhere in the SPB, the tectonic motif was one of thermal subsidence and broad overstepping of the Rotliegend Basin margins (P. Ziegler, pers. comm., 2009).

6 Triassic tectonic evolution: rifting of Pangea and juvenile development of the basin system

6.1 *Early Triassic rifting and subsidence*

Rifting intensified between Greenland and Scandinavia during the Early Triassic (**Figure 3.12c**) and propagated into the North Sea as well as the North Atlantic domain (Ziegler, 1988, 1990a; Roberts et al., 1995). The North Sea rift system, consisting of the Viking and Central grabens and the Horda half-graben, transected the Northern and Southern Permian basins (Ziegler, 1990a). Despite significant cooling, the lithosphere of these basins was presumably still considerably thinner and weaker than the lithosphere



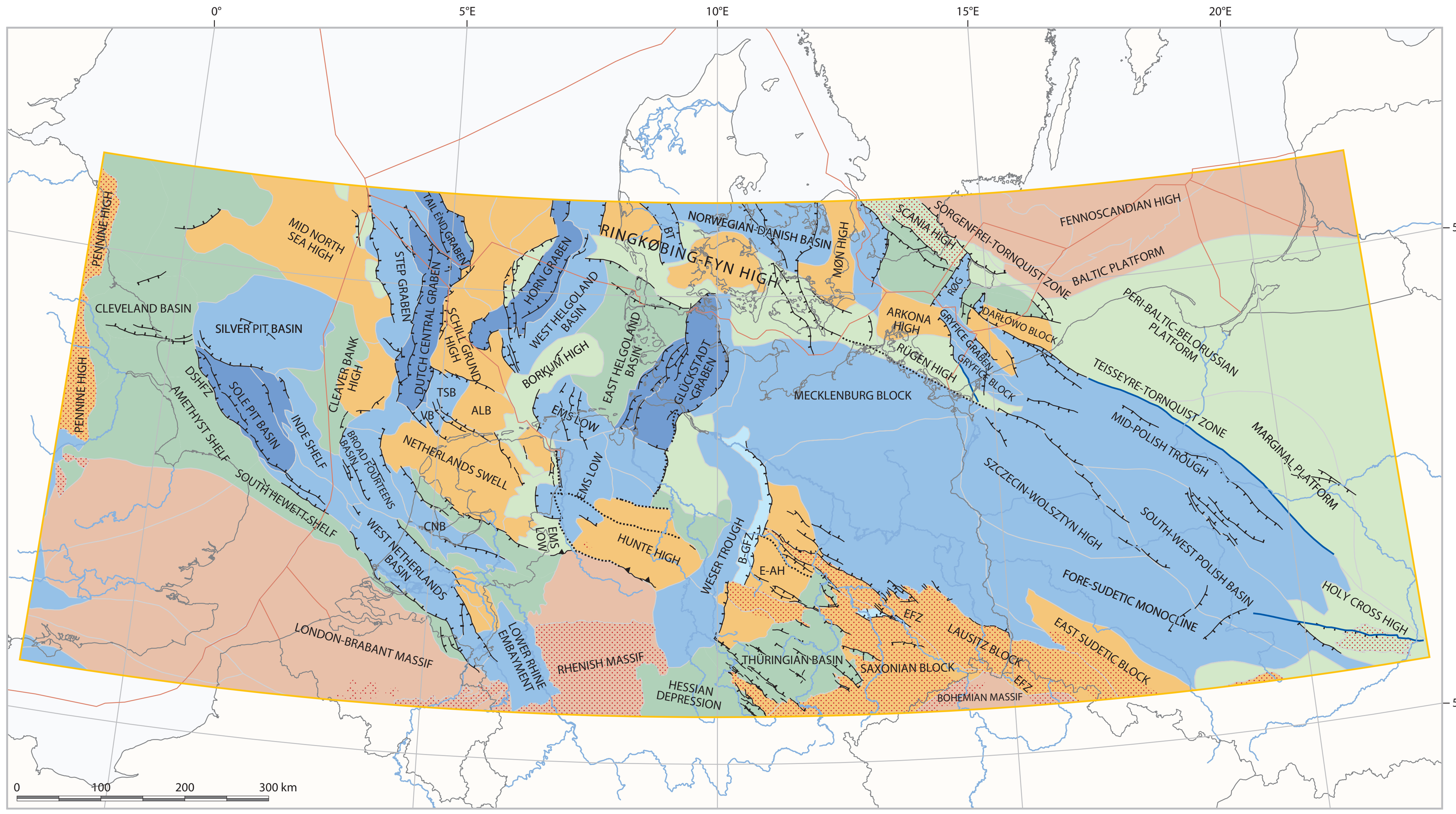


Figure 3.13 a. Early Triassic tectonic evolution for the Olenekian: 248 Ma; b. Mid-Triassic tectonic evolution: Ladinian 237 Ma. Compiled using information from the following published sources: Ziegler (1990a), Cameron et al. (1992), Vejbæk (1997), Dadlez et al. (1998), Lokhorst et al. (1998), Baldschuhn et al. (2001), De Jager (2007). Corroborated by SPBA thickness maps.

of Fennoscandia to the north and the Anglo-Brabant and Variscan massifs to south. The northward drift of the SPB to about 30°N by Early Triassic times (**Figure 3.12a** & **b**) was the result of Permo-Triassic 40° counter-clockwise rotation of Pangea. Continued northward subduction of the Paleotethys Ocean beneath Eurasia was accompanied by the northward drift of the Cimmerian Superterrane, which had rifted off the northern margin of Gondwana during late Early Permian times. Roll-back of the Paleotethys subduction zone resulted in back-arc rifting and the opening of the oceanic Küre and Meliata basins (Stampfli & Borel, 2002) (**Figure 3.14c**).

The facies development and thickness of the Bundtsandstein series (see Figure 9.3) are evidence that thermal subsidence of the SPB persisted during the Early Triassic (Ziegler, 1990a; Geluk, 2005). However, its broad saucer-shaped subsidence pattern was interrupted by the development of the northerly trending Central, Horn and Glückstadt grabens that were initiated at that time. Beyond these grabens, evidence of uniform subsidence is provided by seismic-reflection data throughout the North German Basin (Hoffmann & Stiewe, 1994; Kossow & Krawczyk, 2002; Maystrenko et al., 2005; Scheck et al., 2003a), on which Buntsandstein and Muschelkalk reflectors are apparently unaffected by syndepositional faulting (e.g. see **Figure 3.39**). Contemporary tectonic activity was restricted to north-west-trending structures parallel to the main basement faults.

The so-called 'Hardegsen Unconformity', which affects the margins of the North German Basin and for which a tectonic origin is often proposed (Beutler & Schüler, 1987; Röhling, 1991), may be the result of deformation of the crust in response to the build-up of intraplate stresses (Cloetingh, 1986; Ziegler, 1990a). The London-Brabant and Bohemian massifs remained important sources of clastic detritus, whereas the Rhenish Massif gradually subsided beneath the erosional base level, mainly in response to thermal re-equilibration of the asthenosphere-lithosphere system (Ziegler et al., 2004).

6.1.1 North German Basin

The Bundtsandstein series is up to 5000 m thick in the central North German Basin (see Figure 9.3) (Baldschuhn et al., 1996). Deposition of the Lower Buntsandstein Subgroup took place during a period of relative quiescence. Early Triassic movements of the north-north-east-trending swells and troughs along the southern margin of the SPB are reflected in the thickness variations of the Bundtsandstein series. These swells and troughs were reactivated during mid-Buntsandstein times, including the Netherlands Swell, the Borkum High, the Ems Low, the Hunte Swell, the Solling and Weser troughs, the Eichsfeld-Altmark Swell, the Thüringian and West Brandenburg basins and the East Brandenburg High (**Figure 3.13a**). Pulsed uplift of the swells caused erosion on their crests beneath the Early Triassic unconformities (Volpriehausen and Detfurth). During the more significant base-Hardegsen erosional hiatus (Wolburg, 1962; Schüler, 1980), erosion cut down to Lower Buntsandstein or even Zechstein levels (Wolburg, 1963; Baldschuhn et al., 1998; Röhling, 1991). This unconformity is not so well expressed in the intervening troughs. However, there is no clear erosion of graben-shoulders, suggesting that the causal mechanism is unrelated to extension on individual graben-systems. It is more likely that the build-up of regional tensional stresses gave rise to broadscale warping of the crust and lithosphere (Ziegler, 1990a). In the centre of the basin, north-north-east to west-north-west-trending grabens started to subside during deposition of the Buntsandstein series, namely the Ems-Horn and Glückstadt grabens, Brunswick-Gifhorn Trough, Rügen-Vorpommern and Adler-Kamien-Rønne graben systems (**Figure 3.13a**). Some of the west-north-west-oriented basement trends also became active, such as the Aller and Uelzen lineaments (**Figure 3.13a**). Subsequent (Late Triassic) stretching was so locally intense that the Buntsandstein cover sequence was torn apart into rafts and the Zechstein became tectonically eroded in areas such as the Ems-Horn and Brunswick-Gifhorn grabens and the Aller Lineament. The subcrop pattern beneath the Solling Formation (Hardegsen Unconformity) clearly illustrates the structural differentiation of the SPB (Geluk, 2005; see Chapter 9).

6.1.2 Southern North Sea Basin

During the Bundtsandstein depositional cycle, the Dutch Central Graben subsided faster than adjacent platforms such as the Terschelling and Vlieland basins, but not as rapidly as the Glückstadt and Horn grabens (**Figure 3.13a**). However, poor seismic-data quality at the edges of the Central Graben hampers the resolution of its structural geometry (NITG-TNO, 2004). The coarser-grained sediments of the Main Buntsandstein Formation overlie the fine-grained Lower Buntsandstein clastics. The predominantly sandy Solling Formation overlies earlier Triassic strata above the Hardegsen Unconformity. As no uplift of rift-shoulders can be demonstrated, this event is generally believed to reflect differential uplift across the region, with the greatest amount in the vicinity of the Netherlands Swell. The Broad Fourteens Basin started to subside during Late Permian / Early Triassic times; deposition of thick Zechstein salt in the northern half of the basin resulted in a strongly decoupled structural style in post-Permian strata. The Cleaver Bank High, Ameland Block and Schill Grund High formed intermediate platforms (**Figure 3.13a**) between the grabens and persisting highs such as the London-Brabant Massif. The latter was by-passed by rivers flowing northwards from Armorica in the Wessex, Campine and Central Netherlands basins (see Chapter 9).

6.1.3 Polish Basin

Rapid subsidence of the Polish Basin during Zechstein to Scythian times is attributed to a distinct extensional pulse overprinting its thermal subsidence (Dadlez et al. 1995; Stephenson et al., 2003). Salt movements in the Mid-Polish Trough were initiated during the Early Triassic, significantly modifying local subsidence patterns (cf. Krzywiec 2004a, 2009; 2006a). The Mid-Polish Trough, Norwegian-Danish Basin and the linking Rønne Graben all began to subside rapidly, implying linked reactivation of the Teisseyre and Sorgenfrei segments of the Tornquist Zone (Ziegler, 1990a). Buntsandstein strata are several hundred metres thick in the Norwegian-Danish Basin (see Figure 9.3) (Clausen & Korstgard, 1996; Clausen & Pedersen, 1999; Vejbæk, 1997) and the Polish Basin (Dadlez, 1998a; Krzywiec, 2002a). The Rønne Graben contains up to 1500 m-thick Triassic rocks resting on Lower Paleozoic sediments (Thomsen et al, 1987). The graben configuration of the Mid-Polish Trough, where Upper Permian sediments are up to 4000 m thick, was enhanced during the Triassic (Pożaryski & Brochwicz-Lewiński, 1987). Syndepositional fault movement was restricted to north-west trending structures (Dadlez, 1998b; Krzywiec, 2002a) parallel to the main inferred basement-fault systems (Krzywiec, 2006a). Slightly accelerated subsidence is seen during Zechstein to Scythian times (Dadlez et al., 1995; Dadlez, 1998b). In the Polish offshore sector, thickening of the Triassic succession towards a basement-fault zone indicates localised extension and subsidence (**Figure 3.42**) at the northern end of the Mid-Polish Trough. The Pomeranian segment of the trough shows a gentle increase in thickness of Lower Triassic strata towards the basin axis (Dadlez, 2003). This pattern may have resulted from basin-scale mechanical extensional decoupling of the basement from the post-salt series by Zechstein evaporites (Krzywiec, 2002a, 2002b, 2004b, 2006a; **Figure 3.41**). Within the central (Kuiavian) part of the trough, in the vicinity of the Kłodawa salt structure, basement faulting triggered early salt movements that resulted in the local thickness variations of the Buntsandstein strata (Krzywiec, 2004a, 2004b; see **Figures 3.29** & **3.42**).

6.2 Mid-Triassic development

The Muschelkalk sequences of the SPB were deposited in a shallow-marine environment during a global sea-level highstand (Aigner & Bachmann, 1992). In response to continued thermal subsidence of the lithosphere, the Muschelkalk strata were deposited over a wider area than the Buntsandstein series and onlap onto persisting highs such as the London-Brabant and Bohemian massifs. Differential subsidence of the Central, Horn and Glückstadt grabens (**Figure 3.13b**) is reflected in synsedimentary faulting and increased thicknesses of Muschelkalk strata compared to areas outside the grabens where they are uniformly a few hundred metres thick (Geluk, 2005, 2007a). This reflects continued thermal subsidence of the SPB as a whole, with tensional tectonics restricted to the Central, Horn and Glückstadt grabens and the Mid-Polish Trough (**Figure 3.13b**). Deposition of the Muschelkalk series was ended by a possibly eustatic regional regression (Ziegler, 1990a). Extension along major faults and uplift of the swells continued during Mid-Triassic times, although this is poorly documented. Differential subsidence continued in the principal rift structures, the Horn and Glückstadt grabens and the Emsland Trough (Westdorf Graben). East-north-east-trending subsidence axes in central and southern Germany, such as the Thüringian Basin, were also significant at this time (Walter, 1992).

Triassic sequences thicken into the nascent Dutch Central Graben and Broad Fourteens Basin (see Figure 9.5). Development of salt walls at the bounding faults of the Central Graben suggests that downfaulting was well underway by Mid-Triassic times (Remmelts, 1995, 1996). In the part of the northern Dutch offshore sector underlain by thick Zechstein salt, faults in the Triassic are decoupled from those below the thick salt layer. Thickness maps show that most of the salt swells formed at this time (see Figure 9.5); piercing salt domes and rim-synclines developed later (De Jager, 2007). Only minor thickness variation is visible across faults in the West Netherlands Basin to the south of the Zechstein Basin.

Connection with the Tethyan domain was established via the East Carpathian, Silesian-Moravian areas and the Burgundy and Alemannic gates (formerly referred to as the Western Gates by Szulc (2000); see Chapter 9). Progressive basement faulting in the central (Kuiavian) part of the Mid-Polish Trough resulted in further salt movements emphasised by local thickness variations of the Middle Triassic succession (Krzywiec, 2004a, 2004b). Within the south-eastern (Holy Cross Mountains) segment of the Mid-Polish Trough, basement faulting along the Nowe-Miasto-Iłża Fault Zone resulted in localised Triassic sedimentation within the axial part of the basin (Hakenberg & Świdrowska, 1997; Krzywiec, 2002a; see **Figures 3.22i** & **3.42f**).

6.3 Late Triassic subsidence and rifting

The Late Triassic megatectonic setting of western and central Europe (**Figure 3.14a** & **b**) was dominated by:

a. accelerated activity along the Arctic-North Atlantic rift system that propagated southwards into the Central Atlantic domain during the Late Triassic and by contemporaneous uplift of the flanks of the Norwegian-Greenland rift as shown by the increase in clastic influx into the SPB from northern sources (Ziegler, 1988, 1990a);
b. continued northward subduction of the Paleotethys Ocean beneath Eurasia, contemporaneous opening of the Neotethys Ocean and collision of the Cimmerian Superterrane with the Paleotethys arc-trench system caused by closure of the Küre Basin during the early Cimmerian Orogeny (Stampfli & Borel, 2002; Nikishin et al, 2001). The resulting build-up of intraplate compressional stresses in the East European Platform controlled its uplift and the westward shedding of clastics into the SPB (Ziegler, 1990a);
c. accelerated rifting activity in the domain of the future Alpine Tethys Ocean (Ziegler 1988; Ziegler & Stampfli, 2001). In response to continued counter-clockwise rotation of Pangea, the area of the SPB moved to latitudes of 30 to 40°N by Late Triassic times (**Figure 3.14**).

The North Sea, Horn and Glückstadt grabens remained active during the Late Triassic. Similarly, crustal extension persisted along the Sorgenfrei-Tornquist Zone, Rønne Graben and the Mid-Polish Trough (Ziegler, 1990a; Scheck-Wenderoth et al., 2008). The absence of significant uplift of graben flanks is indicative of a shallow, extensional, lithospheric necking level (Van Wees & Cloetingh, 1996). Moreover, very minor volcanic activity and the absence of regional thermal doming of the North Sea rift suggests that its development was not associated with thermal destabilisation of the lithosphere (Ziegler, 1990a). Important base-Norian and base-Rhaetian unconformities (Early Cimmerian-I, II) are mostly seen along the SPB margins and are attributed to intraplate stresses operating on a lithospheric scale (Cloetingh, 1986; Ziegler, 1990a).

6.3.1 North German Basin

East–west-directed Mid- to Late Triassic extension is reflected in the continued subsidence of the northerly trending Horn and Glückstadt grabens (**Figure 3.15**). During the Late Triassic, discrete extensional pulses were evident during deposition of the Grabfeld and Weser formations (Lower and Upper Gipskeuper), which were separated by a period of tectonic quiescence during the deposition of the Stuttgart Formation (Schilfsandstein) (see Chapter 9). Both extensional pulses affected the Horn and Glückstadt grabens and several major west-north-west-oriented basement-fault systems such as the Aller Lineament. The Late Triassic grabens do not entirely coincide with the older grabens, but are wider due to footwall collapse. Diapiric ascent of Permian salts into the Mesozoic overburden started during Late Triassic times; salt diapirs generally straddle basement faults and so often mark the border faults of grabens. Rim-synclines, which subsided in response to salt migration into diapirs, can contain very thick Keuper strata making it difficult to assess the magnitude of tectonic graben subsidence. About 70% of the known salt domes in northern Germany entered the diapiric phase during mid-Keuper times. The first structural traps had formed in Rotliegend and basal Zechstein reservoirs and already had probably filled with natural gas generated from Paleozoic source rocks.

Seismic-reflection data show that widespread syndepositional normal faulting and salt mobilisation took place in the Central Graben (Michelsen et al., 1987; Sundsbø & Megson, 1993) and Horn Graben (Best et al., 1983; Clausen & Korstgard, 1996; Clausen & Pedersen, 1999) during the Late Triassic. The strongest Late Triassic extensional activity is recorded in the centre of the Glückstadt Graben where up to 5800 m-thick Triassic strata have been observed (Krawczyk et al., 2008a). Diapiric salt walls were initiated above normal faults in the sub-salt basement (Maystrenko et al., 2005) leading to the formation of salt walls several kilometres high and wide and up to 200 km long. At least some of this salt was extruded at the surface and re-deposited. Salt withdrawal played an important role in the creation of accommodation space in all of the grabens (Best et al., 1983; Baldschuhn et al., 1996; Kockel et al., 1996). Rifting activity was considerably slower following the basinwide, base-Norian Arnstadt Formation (Steinmergelkeuper) Unconformity (Early Cimmerian I), which is recognised in the Ringkøbing-Fyn High, southern Sweden, and across the Hunte Swell (Beutler, 1979; Schröder, 1982); crustal extension decreased further after the Late Triassic base-Rhaetian (Early Cimmerian II) Unconformity. The 180-km long Rheinsberg Trough, with up to 1500 m-thick Keuper strata, formed as a basin-scale rim-syncline (Scheck et al., 2003a). Seismic data and structural restoration indicate that salt withdrawal almost completely balanced extension in the salt cover (Scheck et al., 2003b).

The depositional area of the southern German basins widened, and their subsidence seems to have been largely thermally controlled (Ziegler et al., 2004, 2006), although some zonation following the Variscan structural grain is observed in the west-north-west-trending Burgundy-Kraichgau depocentres (Walter, 1992; Freudenberger, 1996b;).

6.3.2 Southern North Sea Basin

The Keuper series comprises mainly mudstones and fine-grained clastic sediments. Sequences thicken northwards into the nascent Dutch Central Graben and Broad Fourteens Basin, the only regions with active faulting. With the rapidly developing load of the sedimentary pile upon the thicker Zechstein salt in the northern Dutch offshore sector, piercing salt diapirs began to break out of the earlier salt swells to initiate the development of spectacular rim-synclines. The Dowsing and South Hewett fault zones at the western limit of the Sole Pit Basin continued to exhibit syndepositional displacement during the Late Triassic to Mid-Jurassic interval. The faults affecting the Upper Triassic were produced by dextral transtension and are anastomosing in plan with a negative flower structure in profile (Van Hoorn, 1987).

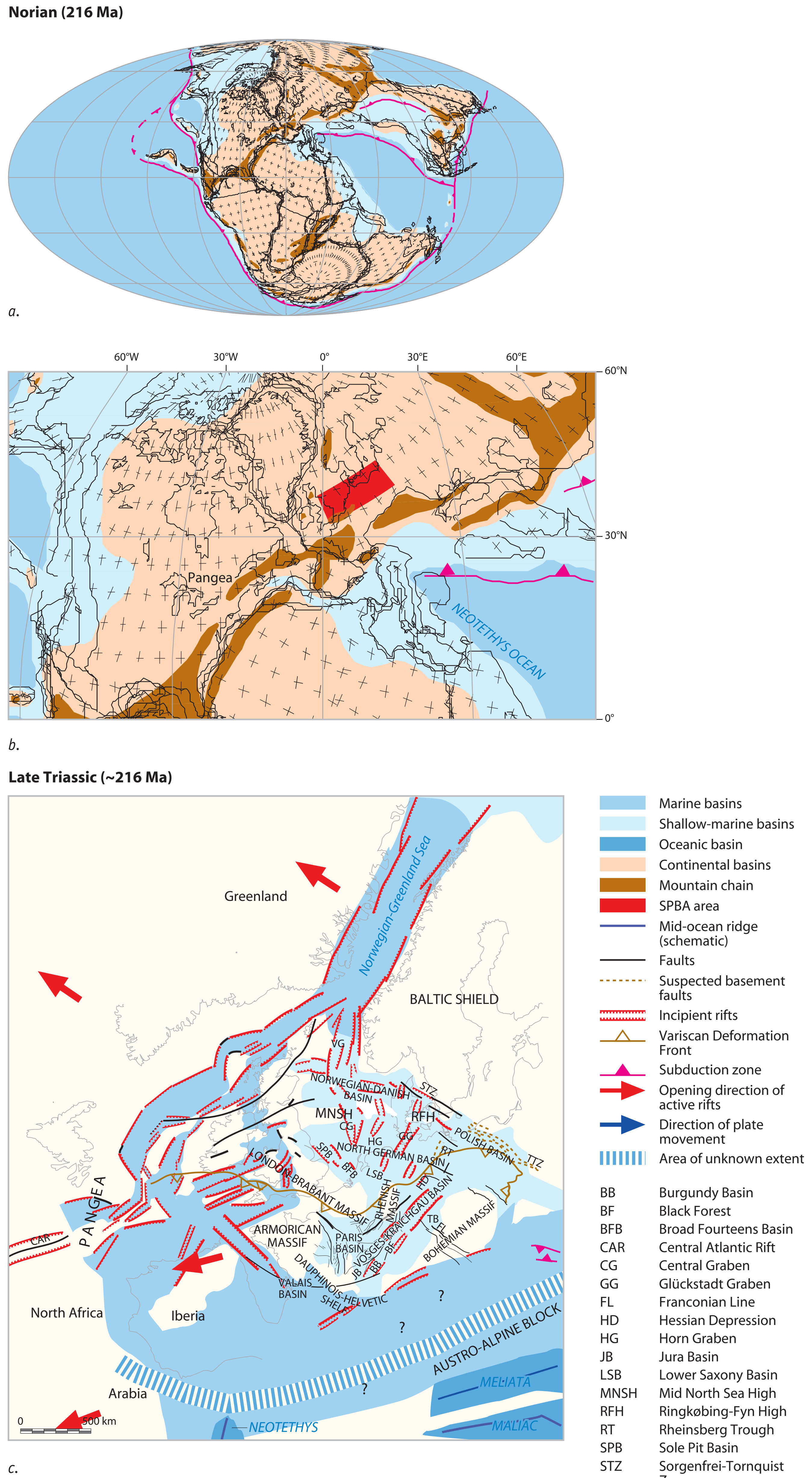


Figure 3.14 Late Triassic tectonic evolution:
a. Palaeogeographic map for the Norian (216 Ma);
b. Detail of Norian palaeogeographic map;
c. Structural overview map for the Mid- to Late Triassic (Keuper, ~216 Ma) (after Scheck-Wenderoth et al., 2008; Figure 10).
Paleogeographic reconstructions by C. Scotese, kindly supplied by Shell.

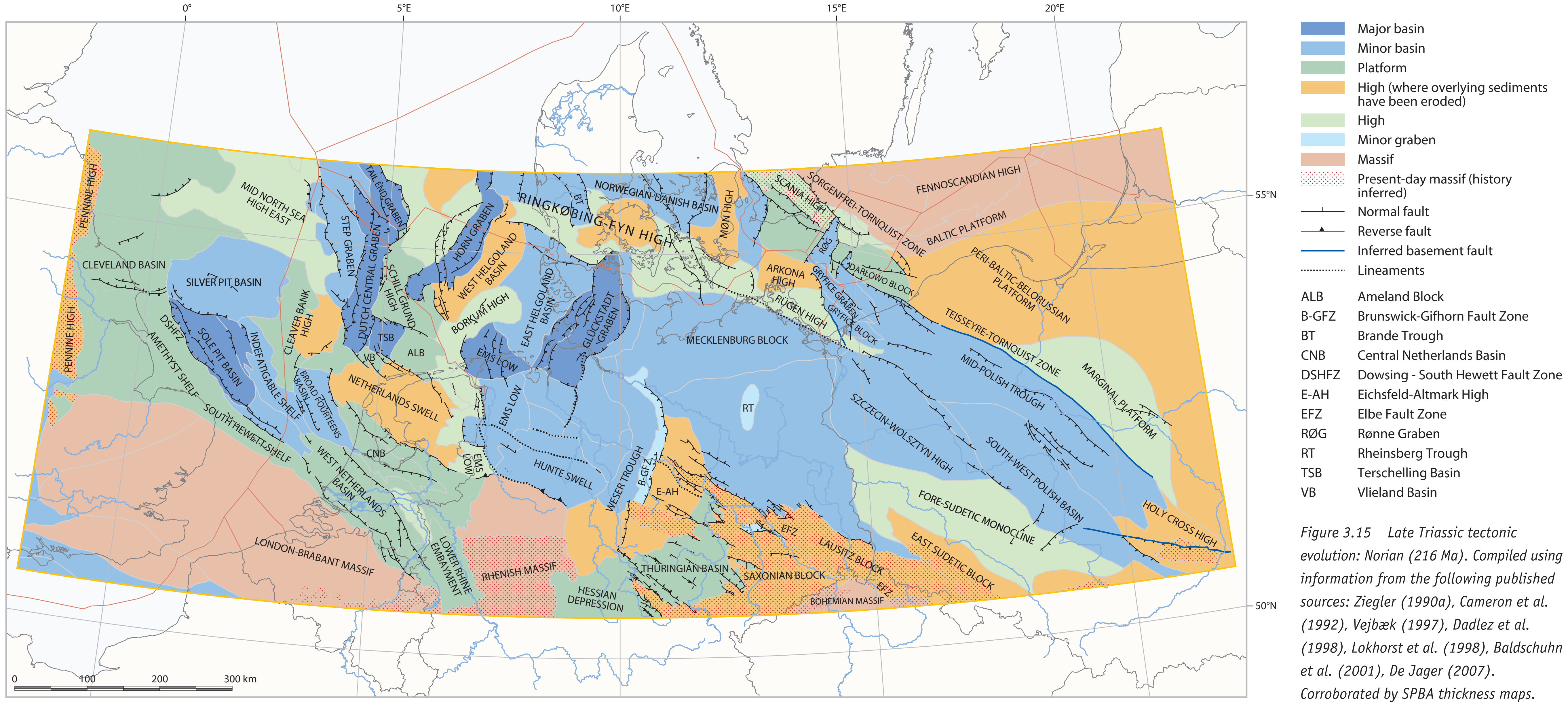
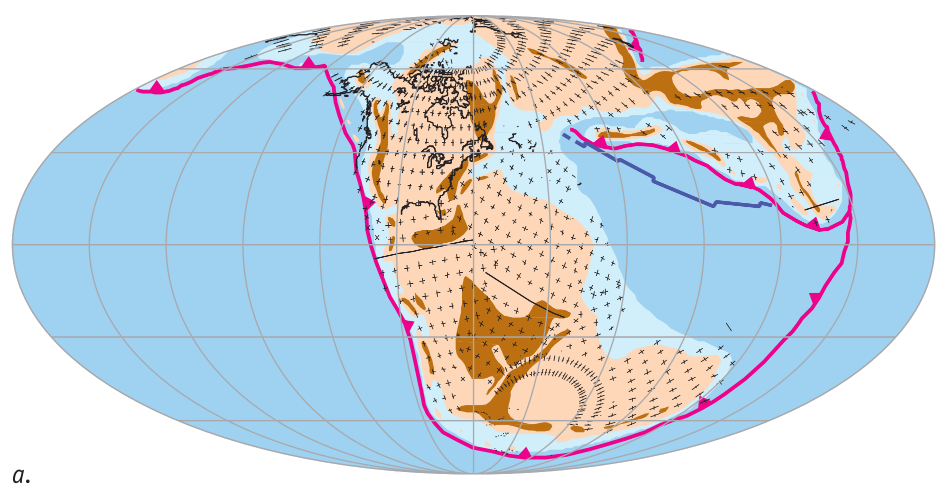
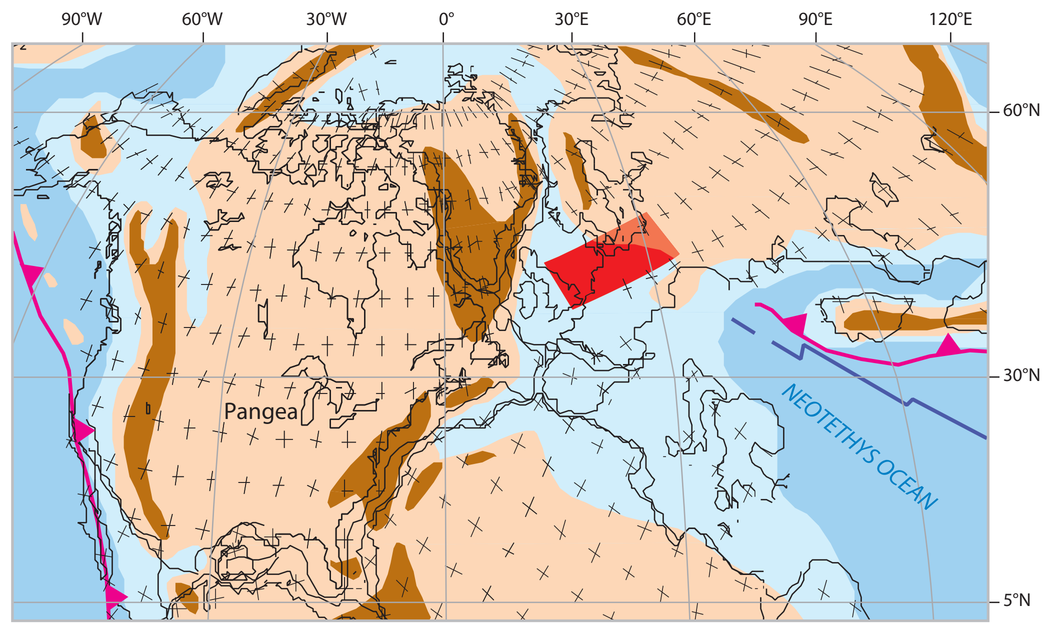


Figure 3.15 Late Triassic tectonic evolution: Norian (216 Ma). Compiled using information from the following published sources: Ziegler (1990a), Cameron et al. (1992), Vejbæk (1997), Dadlez et al. (1998), Lokhorst et al. (1998), Baldschuhn et al. (2001), De Jager (2007). Corroborated by SPBA thickness maps.

Sinemurian (195 Ma)

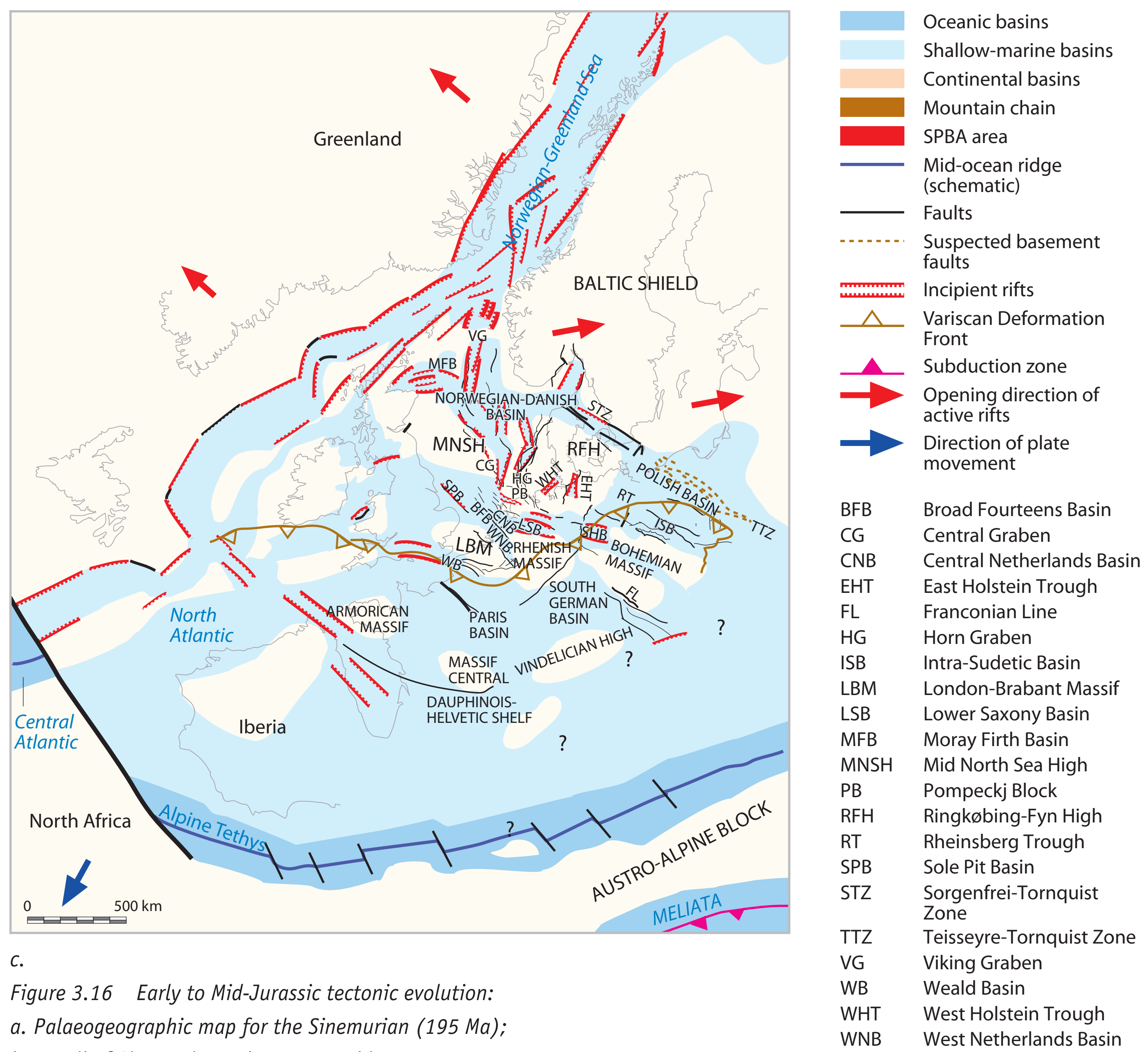


a.



b.

Mid-Jurassic (~160 Ma)



c.

Figure 3.16 Early to Mid-Jurassic tectonic evolution:
a. Palaeogeographic map for the Sinemurian (195 Ma);
b. Detail of Sinemurian palaeogeographic map;
c. Structural overview map for the Mid-Jurassic (~160 Ma) (after Scheck-Wenderoth et al., 2008; Figure 12).
Palaeogeographic reconstructions after C. Scotese, kindly supplied by Shell.

6.3.3 Polish Basin

Keuper sediments up to 2000 m thick were deposited in the north-west-trending Mid-Polish Trough (Czerminski & Pajchlova, 1975), where there was strong movement of Zechstein salt. This is documented in the Pomeranian segment by localised unconformities developed within the Upper Triassic succession (see **Figure 3.29f**). The Early Cimmerian I (base-Norian) Unconformity truncates the Keuper sequence down to the Muschelkalk level and is locally accompanied by normal faulting (Ziegler, 1990a). Basement faulting was most intense in the centre (Kuiavian) of the basin and eventually led to development of the spectacular Kłodawa salt diapir. This salt structure extruded onto the basin floor (see **Figures 3.29** & **3.42**) to form a large overhang onlapped by the uppermost Triassic deposits (Krzywiec, 2004a, 2004b); uppermost Triassic and Jurassic strata indicate only minor growth of this salt structure. In southern Poland and the western Carpathian domain, the Early Cimmerian II Unconformity caused extensive truncation of the Triassic sequence (Ziegler, 1990a) and cannot be a purely eustatic phenomenon as previously envisaged (Michalik, 1978).

7 Jurassic to Early Cretaceous (Aptian) tectonic evolution: evolution of the Tethyan margin and increasing differentiation of the basin system into sub-basins and intervening highs

7.1 Jurassic to Early Cretaceous changes in megatectonic setting of the Southern Permian Basin

Rifting in the Central Atlantic culminated in crustal separation and the onset of sea-floor spreading during Aalenian times. This was followed by northward propagation of rifting into the North Atlantic, Norwegian-Greenland Sea and Labrador-Baffin Bay domains (**Figure 3.16c**). Moreover, rifting accelerated in the Alpine Tethys domain in which sea-floor spreading started during Mid-Jurassic times (Ziegler, 1988; Stampfli & Borel, 2002, 2004). Eastwards, the Alpine Tethys spreading system linked up with that of the Neotethys Ocean, isolating the Carpathian and Tisza blocks and partly the Dacia Block (Schmid et al., 2008). Following collision of the Gondwana-derived Cimmerian Terrane with the southern margin of eastern Europe, which gave rise to the early Cimmerian Orogens, northward subduction of Neotethys was accompanied by intermittent back-arc extension and compression in the Black Sea domain. This gave rise to the Mid-Jurassic, mid-Cimmerian Orogeny and, at the Jurassic-Cretaceous transition, to the late Cimmerian Orogeny (Nikishin et al., 2001; Stampfli & Borel, 2002, 2004).

The Mid-Jurassic onset of sea-floor spreading in the Alpine Tethys Ocean underlies a fundamental reorganisation of the stress field that had dominated the evolution of western and central Europe during Triassic to Mid-Jurassic times. With the relaxation of tensional stresses in the Alpine Tethys domain, the stress field of western and central Europe became dominated by stresses controlling the evolution of the Arctic-North Atlantic rift system and the related clockwise, westward rotation of Laurentia-Greenland relative to Eurasia (Ziegler, 1988, 1990a; Torsvik et al. 2002).

During the Early Cretaceous, the Central Atlantic sea-floor spreading axis gradually propagated northwards into the North Atlantic domain (**Figure 3.18**). By end-Aptian times, crustal separation became effective between Iberia and the Grand Banks, and in the Bay of Biscay. Iberia, including the Alpine Briançonnais Terrane, became isolated from Europe, causing the Valais Ocean to open. At the same time, activity decreased in the North Sea rift system and concentrated on the Norwegian-Greenland Sea rift and the evolving Labrador Sea-Baffin Bay rift (Ziegler, 1988; Stampfli & Borel, 2002, 2004; Schmid et al. 2008).

7.2 Early Jurassic rifting

The North Sea rift system remained active during the Early Jurassic, as evident in the Viking and Dutch Central grabens. At the southern end of the North Sea rift system, crustal extension was compensated by activation of a system of west-north-west-trending transtensional basins along the southern margin of the SPB (**Figure 3.17**). Similarly, tectonic activity along the Sorgenfrei-Tornquist and Teisseyre-Tornquist crustal lineaments persisted during the Early Jurassic. Continued regional thermal subsidence of the Northern and Southern Permian basins during the Rhaetian and Hettangian, combined with a eustatic sea-level rise, controlled the development of a wide, shallow-marine basin. The basin was open to the Arctic seas via the English Midlands, the tensional basins of the Irish Sea and Atlantic seaboard, and to the Tethys Sea via the Thüringian and Kraichgau depressions and the Helvetic Shelf. The Viking Graben was also transgressed during the Sinemurian. Clastics were shed into this broad, regionally subsiding basin from the Fennoscandian Shield, East European Platform and Bohemian Massif (**Figure 3.17**). Stagnant-water stratification led to the deposition of the Posidonia Shale Formation during the Toarcian, the principal source rock for the oil provinces of the southern North Sea and northern Germany (Ziegler, 1990a).

7.2.1 North German Basin

The German part of the SPB area underwent major changes during Early Jurassic times. However, these changes are difficult to assess due to Mid-Jurassic to Early Cretaceous deep erosional truncation of Lower Jurassic strata, particularly in northern Germany. The Horn and Glückstadt grabens gradually became

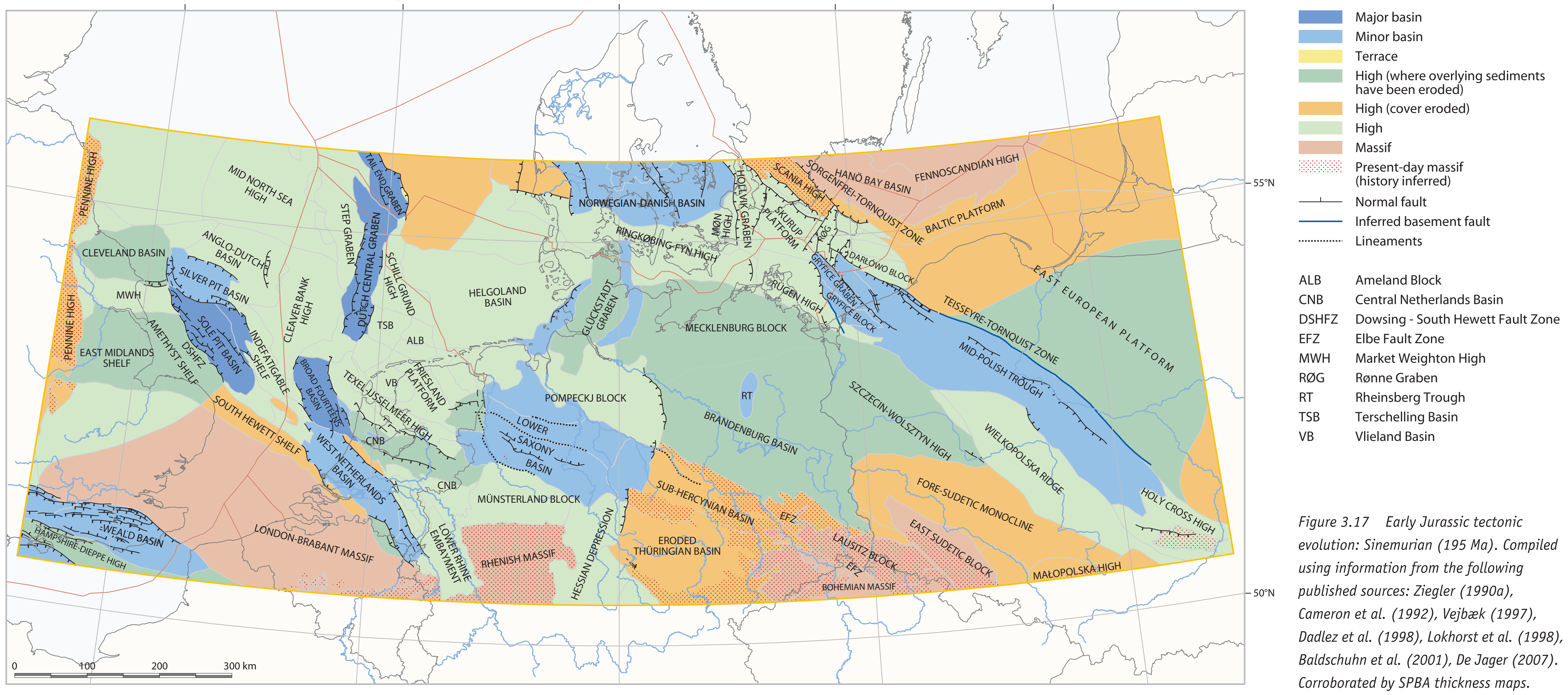


Figure 3.17 Early Jurassic tectonic evolution: Sinemurian (195 Ma). Compiled using information from the following published sources: Ziegler (1990a), Cameron et al. (1992), Vejbæk (1997), Dadlez et al. (1998), Lokhorst et al. (1998), Baldschuhn et al. (2001), De Jager (2007). Corroborated by SPBA thickness maps.

inactive during the Early Jurassic (**Figure 3.17**), although they are still expressed on thickness maps due to continuing salt withdrawal from rim-synclines into the diapirs straddling the graben margins and/or compaction-driven subsidence (see Figure 10.3). The Ems Low and Brunswick-Gifhorn Trough also became inactive at this time (Betz et al., 1987).

Conversely, new west-north-west-trending subsidence centres or sub-basins developed mainly along the southern margin of the SPB (**Figure 3.17**). These include the Lower Saxony Basin (Mazur & Scheck-Wenderoth, 2005), the Sub-Hercynian Basin (including parts of the later Harz Mountains), the Prignitz Basin (including the later Flechtingen Horst (Krawczyk et al., 1999)), the Grimmen Basin (Kossow & Krawczyk, 2002) and the South-east Brandenburg-Lausitz Basin (including the later Lausitz Block) (**Figure 3.17**). Local elongated and narrow subsidence centres also developed as grabens or half-grabens above major basement-fault zones. Until Tithonian times, these sub-basins did not represent individual and isolated palaeo-environmental units with local facies development, but were covered by shelf seas, as were the platform areas between. Sediment influx was not strongly influenced by subsidence of these sub-basins. Some, for example the Lower Saxony and South-east Brandenburg-Lausitz basins, are transtensional in origin; others, such as the Prignitz or Grimmen basins, are shallow depressions bordered by flexure zones with little or no marginal faulting (Kossow & Krawczyk, 2002). Transtensional subsidence and internal differentiation into elongated horst and graben features, especially in the Lower Saxony Basin, started during Hettangian times and continued until the end of the Aptian.

7.2.2 Southern North Sea Basin

The Lower Jurassic series was deeply truncated in the central North Sea during Mid- to Late Jurassic times, consequently hindering the assessment of its structural evolution. Nevertheless, it appears that the Early Jurassic was a period of relative tectonic quiescence, with faulting largely restricted to the Dutch Central Graben and locally to the Broad Fourteens Basin. Differential fault-controlled subsidence continued in the Dutch Central Graben. The Cleaver Bank High, Ameland Block and Schill Grund High remained platforms during much of the Early Jurassic and probably accumulated sediments hundreds of metres thick. Differential subsidence of the north-west-trending transtensional Sole Pit, Broad Fourteens and West Netherlands basins is indicated by the thickness of the Lower Jurassic series, which is about 1000, 500 and 750 m thick respectively. Subsidence of the Sole Pit Basin was controlled by movements along the Dowsing-South Hewett and Flamborough Head fault zones (Holloway, 1985), which also caused uplift of the Market Weighton High. Subsidence of the Weald Basin was controlled by westerly trending faults that represent extensional reactivation of Variscan compressional structures (Whittaker, 1985; Lake & Karner, 1987).

7.2.3 Polish and Norwegian-Danish basins

Differential subsidence of the Mid-Polish Trough during the Early Jurassic was accompanied by local uplift of the graben flanks (**Figure 3.17**); the trough had reached its greatest extent by Toarcian times (Ziegler, 1990a). Lower Jurassic clastic sediments up to 1100 m thick were derived from sources in the Baltic area, the Bohemian Massif and the East European Platform. Uplift of the latter (and inversion of the Donets Basin) was paralleled by the early Cimmerian Orogeny in the Black Sea domain (Ziegler, 1990a; Nikishin et al., 2001). A thickness of up to 750 m of Lower Jurassic shales and deltaic clastics is preserved in the Norwegian-Danish Basin to the west of the SPBA area. Tectonic activity along the Sorgenfrei-Tornquist zone accelerated during the Hettangian (Norling & Bergström, 1987).

7.3 Mid-Jurassic doming and uplift (mid-Cimmerian event)

Uplift of the central North Sea area started towards the end of the Aalenian, presumably in response to the impingement of a transient mantle plume on the lithosphere, which continued during the Bajocian and Bathonian (Ziegler, 1990a; Underhill & Partington, 1993; Surlyk & Ineson, 2003). Development of this large thermal dome (700 × 1000 km), which was transected by the Central Graben, caused deep truncation of Lower Jurassic and Triassic sediments and the development of the regional mid-Cimmerian unconformity in the central North Sea area. Uplift of the dome was accompanied by major volcanic activity at the triple junction of the Viking, Central and Moray Firth grabens, causing deltaic complexes to prograde into the Viking Graben, the grabens of the Atlantic seaboard, and the SPB area. This resulted in blocking of the connection between the Arctic seas and the freely connected Tethys and Central Atlantic oceans (Ziegler, 1988, 1990a). However, crustal extension across the North Sea rift system persisted during the uplift of this thermal dome as shown by continued fault-controlled subsidence of the Viking Graben, the subsidence of deep half-grabens containing continental series in the Central Graben, and continued tectonic activity in the array of transtensional basins along the southern margin of the SPB (Ziegler, 1990a). By late Mid-Jurassic times, the Central North Sea Dome had subsided sufficiently for open-marine conditions to be restored in the North Sea.

The northern SPB was uplifted and subjected to erosion during Mid-Jurassic times, whereas sedimentation in the south was essentially continuous. Sedimentation resumed variably during the Callovian or Late Jurassic in areas uplifted during Mid-Jurassic times (Ziegler 1990a; Vejbæk, 1997). The London-Brabant Massif was also uplifted during Mid-Jurassic times, its Triassic and Upper Paleozoic cover removed to expose the Lower Carboniferous core. Fission-track data suggest that a thickness of 3000 m of sediments was removed (Van den Haute & Vercoutere, 1990).

7.3.1 North German Basin

The eastern shoulder of the Central North Sea Dome started to rise at the end of the Aalenian, shedding large amounts of clastics from the North Sea and the Baltic into the North German Basin as far south as Lower Saxony. The Ringkøbing-Fyn High was re-established by early Bajocian times (Pieńkowski & Schudak, 2008). In the central North Sea a thickness of up to 2000 m of older sediments were removed in response to doming during the mid-Cimmerian event. The northern half of the SPB (Pompeckj Block) was uplifted slightly later than the central North Sea (Baldschuhn et al., 1996; Jaritz, 1987) where Jurassic strata are only preserved locally in small rim-synclines (Maystrenko et al., 2005) with typical thicknesses of about 400 m. Sediment thicknesses are up to 1000 m in the Gifhorn Trough and south of the Aller Lineament (Lohr et al., 2007). The uplift appears to be of Late Jurassic age in the north-eastern part of the North German Basin and possibly lasted into the Early Cretaceous (Kossow & Krawczyk, 2002).

The south-eastern part of the North German Basin reached its maximum extent at this time, although mid-Late Jurassic uplift took place in the north-west (Meyer & Schmidt-Kaler, 1996). The Hessian Depression was also uplifted and eroded with reactivation of north-west and north-north-east-trending faults (Meyer & Schmidt-Kaler, 1996). The long-standing Bohemian Massif was transected during the Callovian by the north-west-trending Saxonian Straight, which provided a new marine connection between the North German Basin and the Tethyan shelves of the Carpathian domain (Ziegler, 1990a).

7.3.2 Southern North Sea Basin

Much of the Dutch offshore area was uplifted during the mid-Cimmerian event (Ziegler, 1982a; Ziegler, 1990a). Erosion on the Cleaver Bank High (locally to the Carboniferous level) was most severe in the north adjacent to the Dutch Central Graben. A thickness of up to 2000 m of Lower Jurassic and Triassic strata were eroded from the high during Mid-Jurassic times (Glennie, 1986), and an unknown amount from the Mid North Sea, Ringkøbing-Fyn and Schill Grund highs. The Step Graben and Terschelling Basin were uplifted and eroded down to Triassic and Lower Jurassic strata. The Winterton High was up-arched, with subsequent pre-Cretaceous erosion cutting down to Upper Carboniferous strata in the core (Van Hoorn, 1987). Meanwhile, the Central Graben continued to subside (Ziegler, 1990a; Underhill & Partington, 1993) such that it contains a thick and complete Jurassic succession (Heybroek, 1975). The central North Sea thermal dome started to subside during Callovian times and from then on rifting accelerated in the North Sea rift system including the Central and Tail End grabens. The Horn Graben became inactive during the Mid-Jurassic (Ziegler, 1990a). In the Cleveland, Sole Pit and West Netherlands basins, Middle Jurassic strata comprise brackish to shallow-marine sediments of fluviodeltaic facies with a cyclicity that reflects strong eustatic control (Cameron et al., 1992). Clastic sediments were shed into these basins from the volcanic dome in the Central Graben, the Mid North Sea High, and more local areas of uplift such as the Cleaver Bank, Broad Fourteens, Winterton and Friesland highs. Sediments of oolitic-carbonate facies were deposited on the East Midlands Shelf.

7.3.3 Polish and Norwegian-Danish basins

Sedimentation became restricted to the differentially subsiding axial trough (Kutno Depression) during Aalenian to Bajocian times and the Mid-Polish Trough became isolated from other basins within the SPB (Dadlez, 1998a). A connection to the Tethys Ocean opened via the East Carpathian Gate. A eustatic sea-level rise during the late Bathonian resulted in wide overstep onto the basin margins (Pieńkowski & Schudak, 2008) such that there is a strong contrast between the thickness of Bajocian to Bathonian strata in the axial part of the basin (600 m) and its flanking platforms (50-100 m). Intraformational conglomerates accompanied the continuing differential subsidence of the Mid-Polish Trough.

The Norwegian-Danish Basin was significantly affected by Mid-Jurassic doming of the central North Sea (Nielsen, 2003). However, tectonic activity accelerated along the fault systems of the Sorgenfrei-Tornquist Zone and there is evidence for Aalenian to Bajocian volcanic activity in Scania (Norling & Bergström, 1987).

7.4 Late Jurassic to Early Cretaceous rifting

At the southern end of the North Sea rift system, a wrench-dominated regime intensified during the Late Jurassic and Early Cretaceous and controlled transtensional subsidence of north-west-oriented basins and transpressional uplift of narrow highs along the southern SPB margin (**Figures 3.18d** & **3.19a**). This stress system relaxed during the transition to Late Cretaceous times with the deactivation of the North Sea rift system (Ziegler, 1990a; Scheck-Wenderoth et al., 2008).

A eustatic sea-level lowstand at the Jurassic-Cretaceous transition, combined with stress-induced deflection of the lithosphere, led to earliest Cretaceous emergence and erosion of large parts of western and central Europe (Ziegler, 1990a). Crustal extension across the North Sea graben system gradually decreased during the Early Cretaceous and essentially ended during the Aptian to Albian (Ziegler, 1990a; Torsvik et al., 2002;

Coward et al., 2003). Nevertheless, there is evidence of late-stage deformation as for instance in strata underlying the late Hauterivian unconformity in the Danish Central Graben (Vejbæk & Andersen, 1987, 2002). The tectonic evolution of the North German Basin during this time is very similar to that of the German and Dutch areas of the southern North Sea.

7.4.1 North German Basin

North-west-oriented sub-basins along the southern SPB margin have an *en-echelon* relationship (**Figure 3.18d**). Rapid differential subsidence of the Lower Saxony, Sub-Hercynian and Altmark-Brandenburg sub-basins is linked to that of the sub-basins farther west in the Dutch sector.

Previously exposed areas of the SPB were gradually transgressed during the Late Jurassic, whereas dextral-wrench deformation intensified in areas flanking its southern margin (**Figure 3.18d**). The early salt domes and pillows entered the diapiric phase during the Oxfordian, with erosion along their crests. The Lower Saxony and Brandenburg basins became the principal depositional areas. The north-west–south-east trending Saxonian Straight connecting the North German Basin with the Carpathian Tethys shelves persisted during the Late Jurassic. Accelerated subsidence of the transtensional Lower Saxony, Sub-Hercynian and Altmark-Brandenburg sub-basins was accompanied by uplift of the Rhenish Massif and progressive closure of the Trier Embayment and the Thüringian Basin.

During the transition to the Early Cretaceous, rapid crustal extension across the North Sea rift system was accompanied by accelerated wrench-induced subsidence of the Lower Saxony, Sub-Hercynian and Altmark-Brandenburg sub-basins, uplift of their flanking platforms, closure of the Saxonian Straight (**Figure 3.18c**), and reactivation of Permo-Carboniferous fault systems flanking the Bohemian Massif to the south-west. By Early Cretaceous times, the London-Brabant, Rhenish and Bohemian massifs formed a coherent barrier between the Tethys shelves and the transtensional basins of Germany and the Netherlands (Ziegler, 1990a; Walter, 1992; Schwerd, 1996). These late Cimmerian lithosphere-scale deformations, combined with an earliest Cretaceous eustatic sea-level lowstand, led to temporary isolation of the Lower Saxony, Sub-Hercynian and Altmark-Brandenburg sub-basins, which became highly differentiated into local swells and troughs. In contrast, the platforms between these basins became areas of erosion during the Tithonian and Berriasian, for example, the Pompeckj and Münsterland blocks (**Figure 3.19b**). Although apparently isolated by the late Cimmerian tectonic movements, the structural development of these wrench-induced basins continued across the Jurassic-Cretaceous boundary without a break in sedimentation. Latest Jurassic to earliest Cretaceous dextral transtension was the controlling mechanism for their formation (Jaritz, 1980, 1987; Betz et al., 1987; Ziegler, 1990a; Nalpas et al., 1995; Scheck et al., 2002a). The intervening blocks and flanking highs were uplifted during transpressional deformation, whereas deep-crustal fracturing triggered repeated igneous activity (Ziegler, 1990a). Metallic hydrothermal vein deposits formed in areas such as the Harz Mountains and southern Lower Saxony Basin, contemporaneous with crustal dilation and magmatism (Dohr et al., 1989). A new pulse of salt mobilisation started in late Oxfordian times accompanied by rapid subsidence of the Lower Saxony and Sub-Hercynian sub-basins (Jaritz, 1980, 1987; Betz et al., 1987; Nalpas et al., 1995; Scheck et al., 2002a). A seismic transect across the Lower Saxony Basin (Mazur & Scheck-Wenderoth, 2005; **Figure 3.35**) clearly shows strong stratigraphic thickening during the Jurassic to Early Cretaceous, for example across the Allertal Fault. The Pompeckj Swell to the north was uplifted during the Kimmeridgian, subjected to erosion, and gradually transgressed again during the later Early Cretaceous. Correspondingly, the Late Cimmerian Unconformity is evident in large areas of northern Germany (Baldschuhn et al., 1996). Whereas Lower Cretaceous sediments are 50 to 100 m thick in platform areas, they are up to 2000 m thick in the Lower Saxony Basin. The magnitude of Late Jurassic and Early Cretaceous subsidence varied considerably between the different wrench-induced basins and their sub-basins; facies development was similarly very variable. The Kimmeridgian and Tithonian series are up to 700 m thick in the Lower Saxony Basin and developed in carbonate and evaporite facies. These are overlain by Berriasian lacustrine and deltaic shales and sands that are up to 500 m thick. Valanginian to Albian marine shales and deltaic sands are up to 1500 m thick; the sands are important hydrocarbon reservoirs (Betz et al., 1987). Upper Jurassic and Lower Cretaceous rocks in the Dutch sector of the Lower Saxony Basin thin westwards to the Friesland Platform (**Figure 3.19b**), where there are no extensional half-grabens.

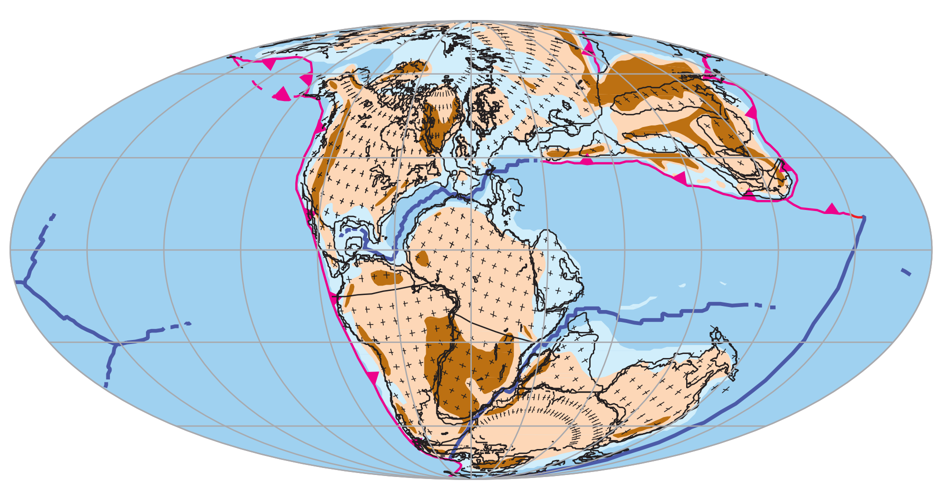
7.4.2 Southern North Sea Basin

The main tectonic elements of the Dutch sub-surface developed during the Late Jurassic and Early Cretaceous, comprising the late Cimmerian rift pulses. Extensional faulting and subsidence accelerated in the northerly trending Dutch Central Graben (**Figures 3.19a** & **b**), progressing in time from north to south (Heybroek, 1975; Schroot, 1991). In Mid-Jurassic times, the graben straddled the uplifted Central North Sea Dome (NITG-TNO, 2004). Thick fluviolacustrine to shallow-marine sequences accumulated in the graben during Late Jurassic and Early Cretaceous times. Volgian to Ryazanian shales are kerogenous in the northern Central Graben (Herngreen & Wong, 1989). In the southern graben, the provenance of clastic sediments was the Cleaver Bank-Broad Fourteens High, which was uplifted during Callovian times. Adjacent highs such as the Friesland High were uplifted and eroded at the same time. The Schill Grund High formed a stable platform area on the eastern flank of the Central Graben. The Step Graben and Terschelling Basin subsided more slowly than the Central Graben during the Late Jurassic and accumulated thinner sequences. Salt walls developed along the main bounding faults of the Central Graben and in the Outer Rough Basin (**Figure 3.19a**), which had been little affected by mid-Cimmerian uplift. Late Jurassic uplift of the Friesland Platform resulted in erosion down to Lower Triassic and, locally, to Zechstein levels (NITG-TNO, 2004).

Farther south, the Sole Pit, West Netherlands, Central Netherlands, Roer Valley and Vlieland basins (like the Lower Saxony Basin) all trend north-west–south-east and probably developed by transtensional reactivation of pre-existing basement structures (**Figure 3.18d**). These basin-controlling faults accommodated the east–west extension evident in the Central Graben. However, due to the complex reactivation history, unambiguous evidence of dextral transtensional displacement is only available locally, for example, in the Rifgronden Fault Zone between the Terschelling Basin and the Schill Grund High (De Jager, 2007). During Callovian to Oxfordian times, the uplift of intervening highs such as the Broad Fourteens, Winterton and Friesland highs (**Figure 3.19a**) shed clastics into the adjacent rapidly subsiding basins. A similar tectonic regime has been suggested for the Vlieland Basin (Herngreen et al., 1991), which linked the Central Graben and Lower Saxony Basin during the Kimmeridgian to Tithonian. The Zuidwal alkaline volcanic complex (Kimmeridgian) developed during the late Cimmerian rifting phase; the associated thermal anomaly slowed the Late Jurassic subsidence of the Vlieland Basin. In the Terschelling Basin, tectonic events were slightly delayed relative to the Dutch Central Graben; uplift occurred before the end of the Mid-Jurassic and a thin, younger, Upper Jurassic sequence rests on the Triassic, whereas the Lower Cretaceous sequence is thicker than in the Central Graben.

The Broad Fourteens Basin, and the Central Netherlands Basin (**Figures 3.19a** & **b**) with which it merges, started to subside rapidly during the Kimmeridgian and developed into pronounced fault-bounded transtensional basins during the Late Jurassic and Early Cretaceous. The margins of these basins were overstepped by post-rift deposits during the Valanginian-Barremian transgression. Lower Cretaceous marine clastics were deposited in the Central Netherlands Basin, while at the same time continental sedimentation took place in the Roer Valley Graben. The Zandvoort Ridge remained an important wrench-induced feature separating the West and Central Netherlands basins, which were filled with up to 2500 m-thick Upper Jurassic to Lower Cretaceous sands and clays. At the same time, the adjacent platforms were uplifted and

Kimmeridgian (152 Ma)



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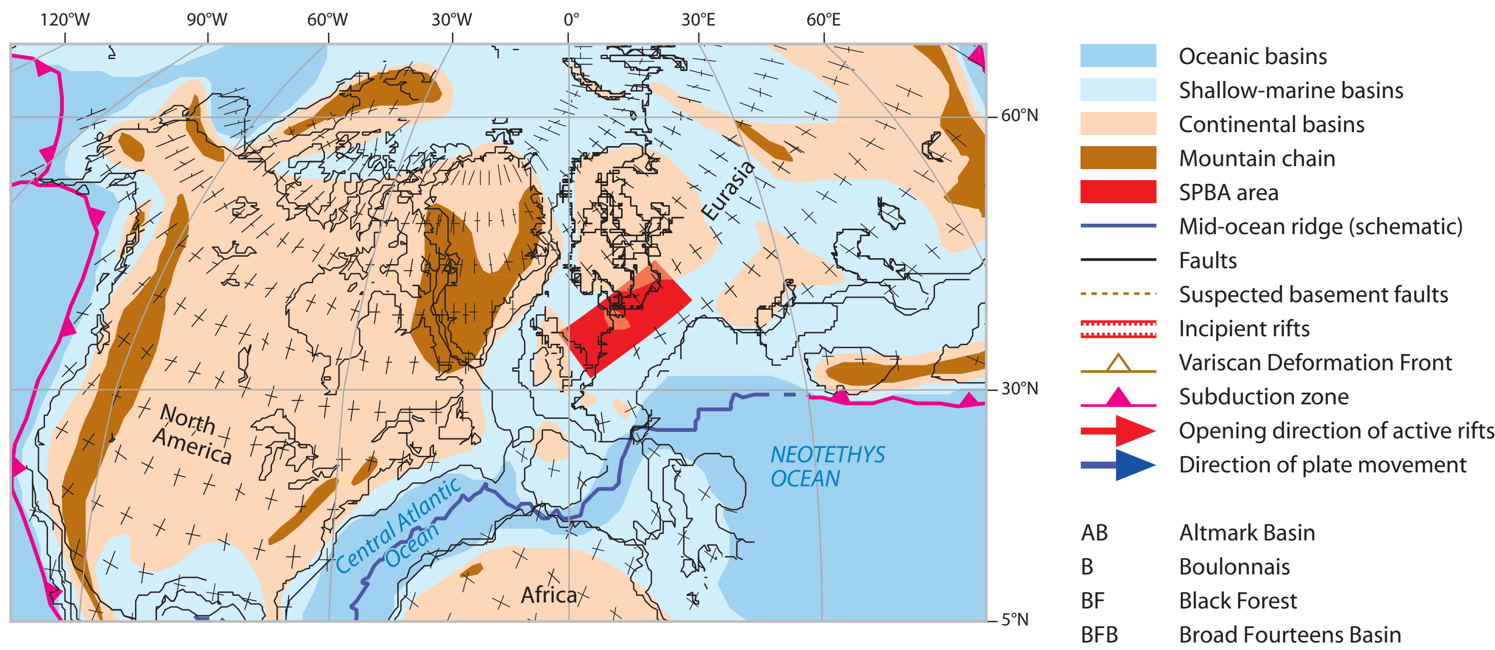
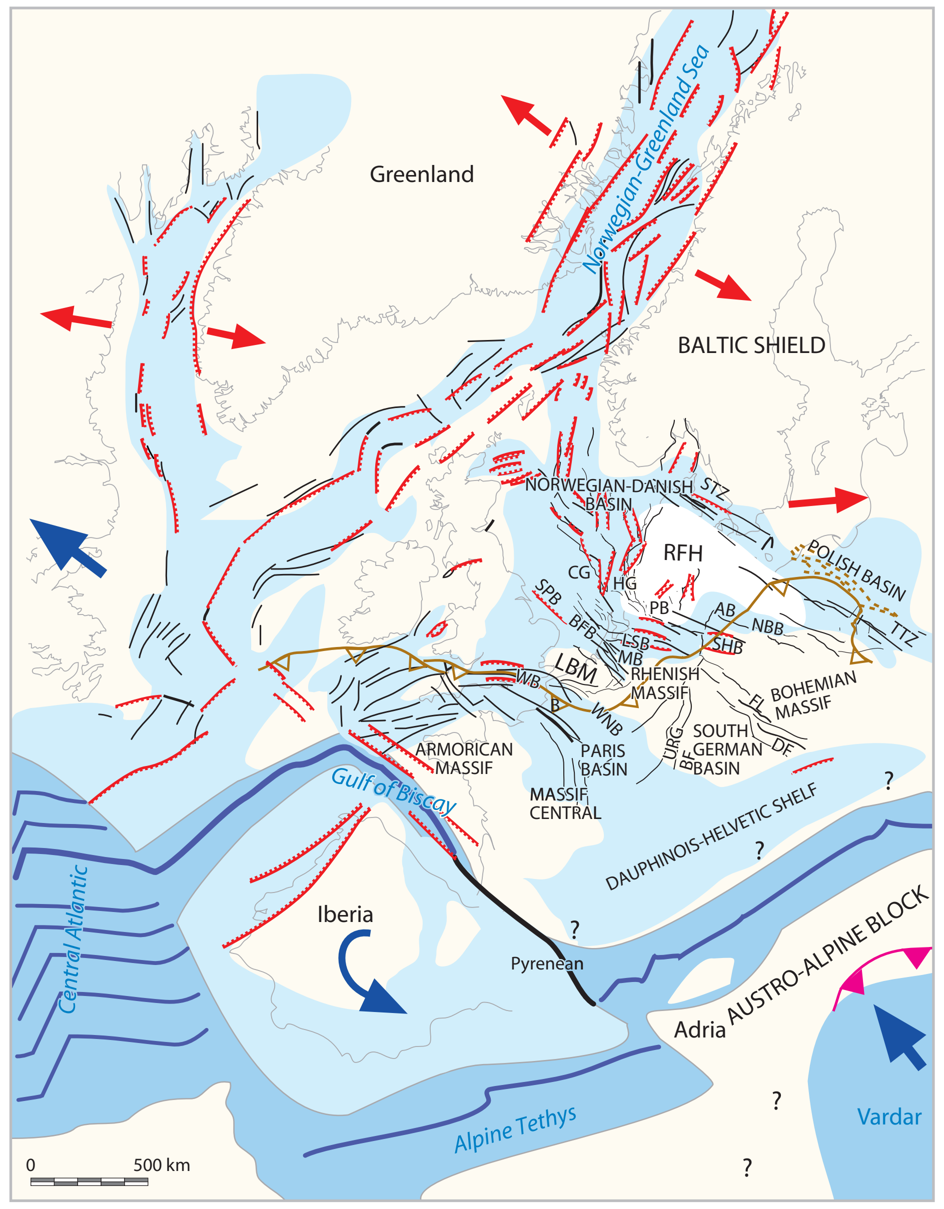


b.

Early Cretaceous (~120 Ma)



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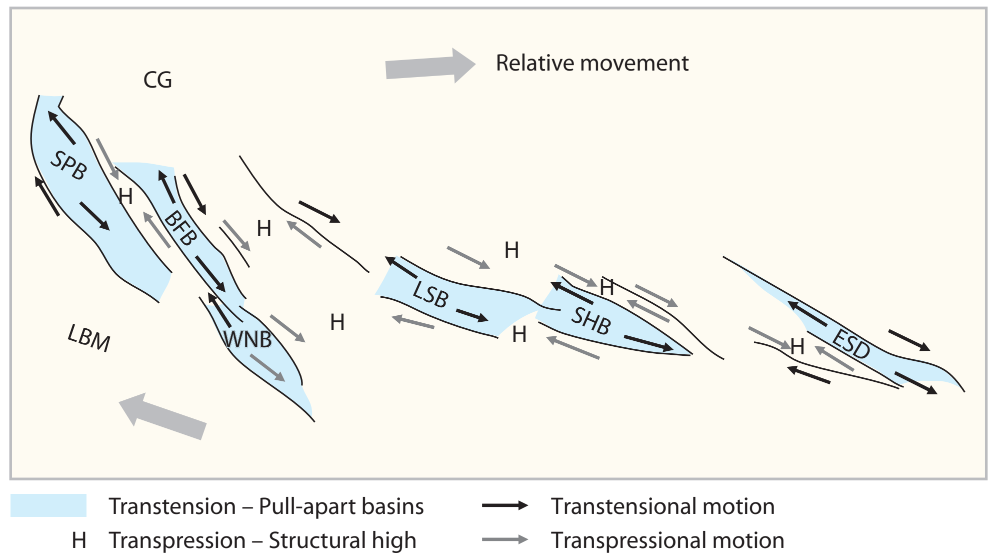


d.

Figure 3.18 Late Jurassic – Early Cretaceous tectonic evolution:
a. Palaeogeographic map for the Kimmeridgian (152 Ma);
b. Detail of Kimmeridgian palaeogeographic map;
c. Structural overview map for the Early Cretaceous (~120 Ma) (after Scheck-Wenderoth et al., 2008; Figure 13);
d. Effect of transpression and transtension on the 'marginal basins' (after Scheck-Wenderoth et al., 2008; Figure 13b).
Palaeogeographic reconstructions after C. Scotese, kindly supplied by Shell.

strongly eroded (Late Cimmerian Unconformity), the effect of which was enhanced by a regional sea-level lowstand. In the West Netherlands Basin, where Zechstein salt is absent, extension resulted in a series of half-grabens filled with Upper Jurassic to Lower Cretaceous clastics and associated basalts. The Cleaver Bank High, the Ameland Block and Schill Grund High (**Figure 3.19a** & **b**), which were platforms during much of Triassic to Early Jurassic times, were uplifted and eroded during the mid- to late Cimmerian rifting phases. Upper Jurassic and Lower Cretaceous syn-rift strata are consequently missing from these highs, where Triassic and Permian strata are unconformably overlain by thin post-rift Lower Cretaceous and thicker Upper Cretaceous rocks (De Jager, 2007). Hundreds of metres of Triassic to Middle Jurassic sediments were probably removed from these highs. Late Cimmerian uplift of the Maasbommel High led to erosion of most of its Middle Jurassic to Permian cover. The Texel-IJsselmeer and Winterton highs (**Figure 3.19b**) were eroded even more severely, down to the Westphalian. The thick Rijnland Group (latest Ryazanian to Albian) sucession, comprising mainly fine-grained clastics, was subsequently deposited across a large open-marine basin.

In the Sole Pit Basin, subsidence was not interrupted by mid-Cimmerian (Mid-Jurassic) uplift, although it strongly affected the Cleaver Bank and Winterton highs (**Figure 3.19b**). A minor basal Bajocian unconformity and a stronger Late Cimmerian Unconformity are evident in the Cleaver Bank High (Ziegler, 1990a). The hanging wall of the Dowsing-South Hewett Fault Zone was a locus for deposition of relatively thick Lower Cretaceous strata, locally up to 1000 m thick (Cameron et al., 1992). The basal Cretaceous Spilsby Sandstone Formation is apparently in continuity with the underlying Upper Jurassic clays (Glennie & Boegner, 1981). The Lower Cretaceous is not preserved in the axial zone of the Sole Pit Inversion, but there is evidence of substantial amounts of missing overburden in this region (Marie, 1975; Glennie & Boegner, 1981), compatible with the deposition of a thick Lower Cretaceous sequence across the basin, which was subsequently removed by erosion following inversion. A progressive sea-level rise led to starvation of the basin from late Hauterivian times. The Swarte Bank Hinge Zone is a rather complex braided-fault zone, which delimits the eastern side of the Sole Pit Basin (**Figure 3.19**). Development of the hinge zone started significantly later than the Dowsing Fault Zone, where the initial *en-echelon* asymmetric grabens are Late Jurassic in age. The zone is defined by listric faults that sole in the Zechstein salt (Cameron et al., 1992), which appears to have played a passive role in the deformation (Gibbs, 1986). The master faults dip away from the Sole Pit Inversion axis (Walker & Cooper, 1987). Whereas strike-slip movements may have taken place along the Dowsing Fault Zone over a long period, and may be related to movements on deeper basement fractures (Glennie & Boegner, 1981), the Swarte Bank Hinge Zone has been interpreted as an altogether shallower and more superficial structure. The hinge zone may have been the result of gravitational creep of post-Permian sedimentary rocks from the rising north-east flank of the Sole Pit Basin, mainly during the Early Cretaceous (Walker & Cooper, 1987).

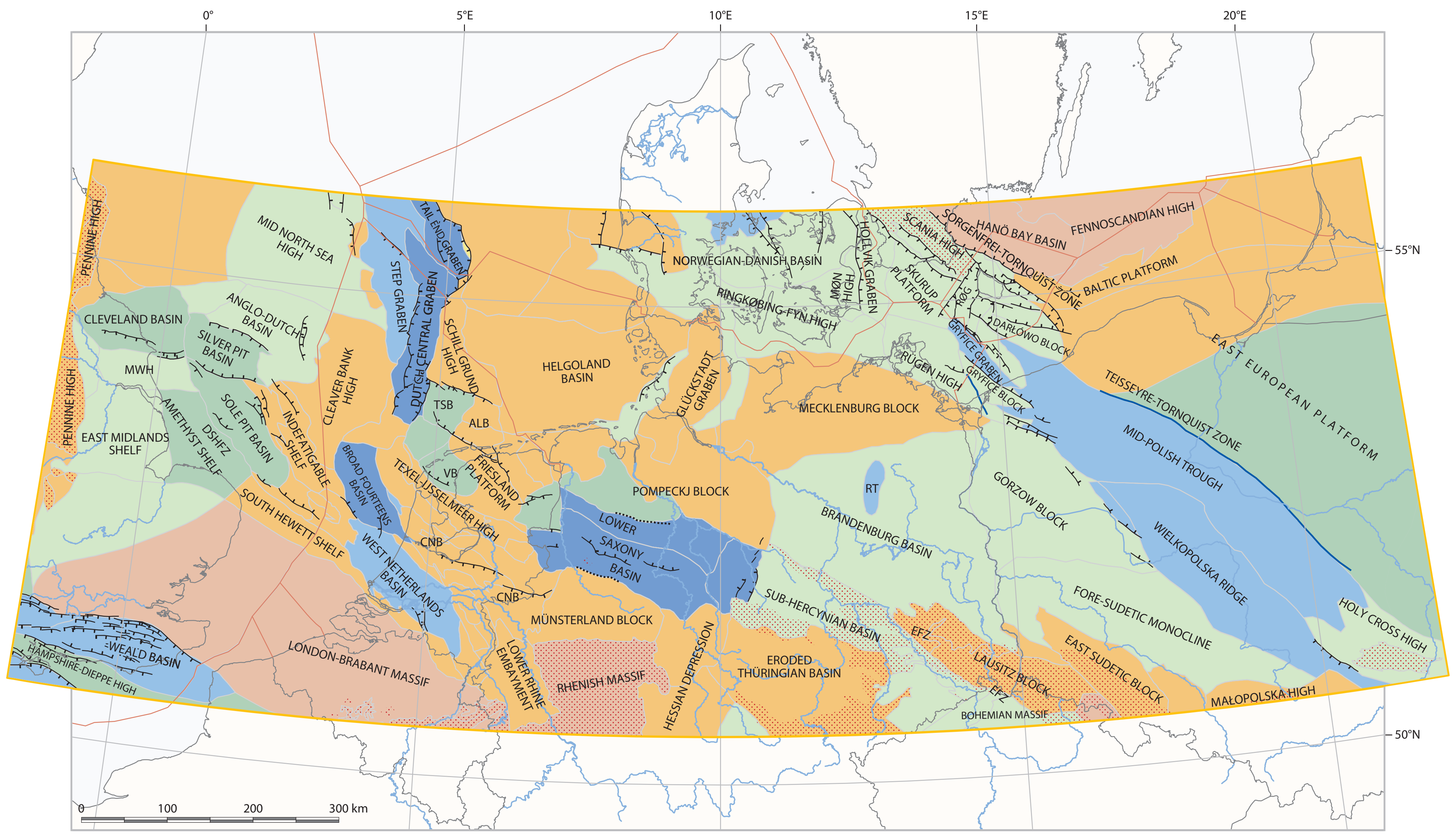
The Weald sub-basin (Wessex Basin) was initiated in Early Jurassic times then rapidly subsided. It was intermittently connected to the Paris Basin. Normal faulting along an east–west trend resulted from extensional reactivation of Variscan basement thrusts (Whittaker, 1985; Hansen et al., 2002). The basin lies south of the London-Brabant Massif, a clastic provenance for most of Mid- to Late Jurassic times (Cope et al., 1980). Similarly, the Mons Basin is framed by east–west-trending faults. It is still debated if its collapse structure, which was initiated during Hauterivian-Barremian times, is attributable to pull-apart and/or evaporite dissolution in the underlying Devonian-Carboniferous basin (Vandycke, 2002 vs Delmer, 2004). The Cleveland Basin was also subsiding at this time due to fault-controlled crustal extension, and can be seen as a north-westward extension of the Sole Pit Basin. It lies between blocks (e.g. Market Weighton) underpinned by Caledonian granites. In contrast, subsidence of the intervening, little-faulted East Midlands Shelf was much slower, except during the Late Jurassic (Cameron et al., 1992), and the thin succession was greatly affected by eustatic influences. The Late Cimmerian Unconformity is most distinctive in the basin-marginal and shelf areas.

7.4.3 Polish and Norwegian-Danish basins

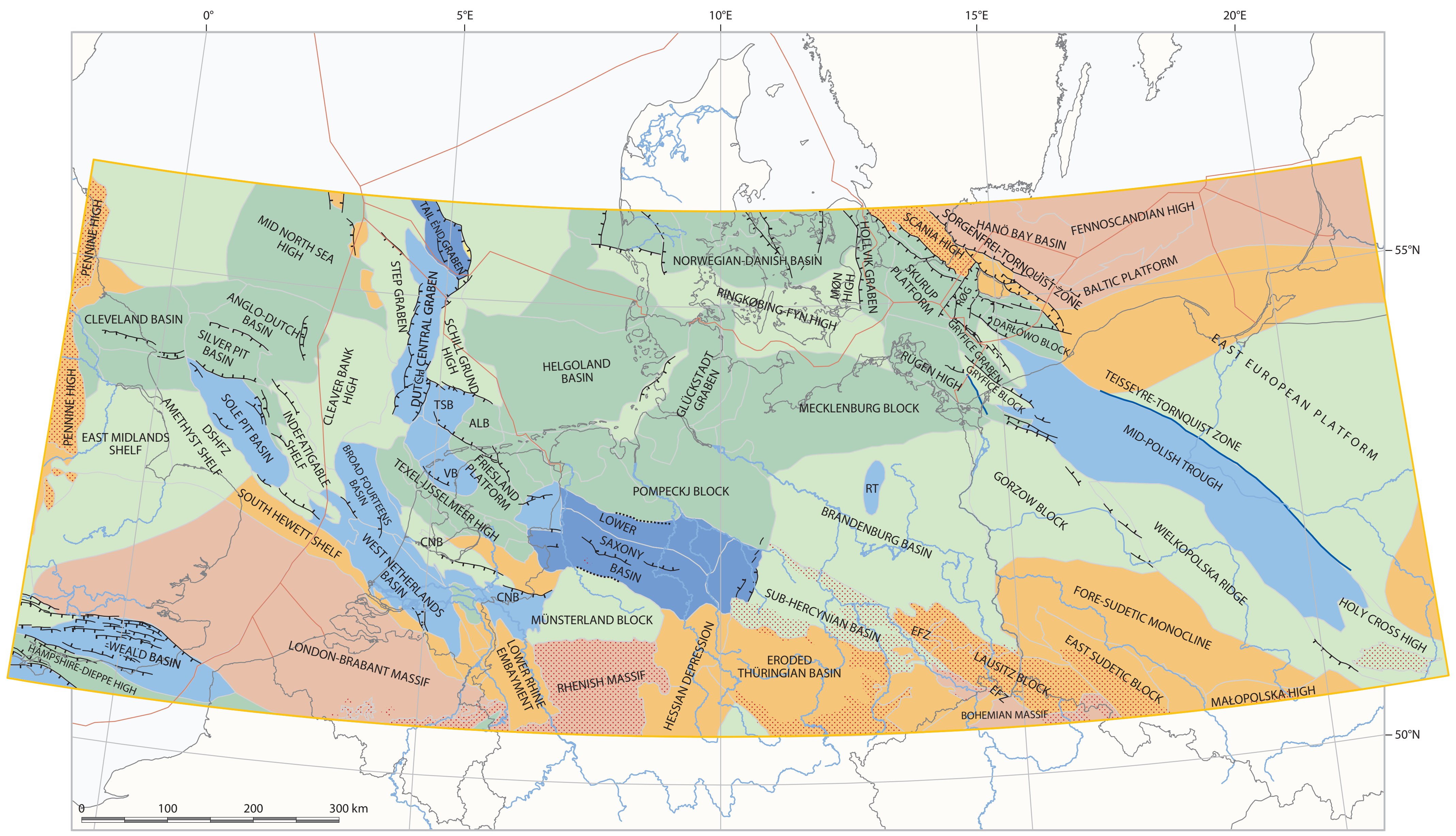
Sedimentation was restricted to axial parts of the Mid-Polish Trough during the early Callovian sea-level lowstand. This arm of the Tethys Ocean was separated from the North German Basin by a land barrier joining the Lausitz Block to the Pompeckj Swell and the eastern Ringkøbing-Fyn High. The barrier was overstepped during the late Callovian transgression, linking the Arctic and Tethys faunal provinces until latest Tithonian times (Ziegler, 1990a). A pulse of extension-related accelerated tectonic subsidence affected the Polish Basin in Oxfordian to Kimmeridgian times (Dadlez et al., 1995; Stephenson et al, 2003). Extension was focussed on the Teisseyre-Tornquist Zone, with maximum subsidence within the Mid-Polish Trough, at least partly related to basement faulting (Dadlez et al., 1995; 1998). Mechanical decoupling of the pre- and post-Zechstein salt series caused peripheral extension on the flanks of the Mid-Polish Trough, whereas its axial parts subsided partly in response to salt withdrawal (Krzywiec, 2002a, 2004b). The south-eastern trough, located in the transitional area between the peri-Tethyan and Tethyan domains, exhibits a complex pattern of Late Jurassic to Early Cretaceous subsidence (Pożaryski & Żytko, 1981; Kutek, 1994, 2001; Hakenberg & Świdrowska, 1997; Gutowski et al., 2005; Krzywiec et al., 2009). Seismic data from the Holy Cross Mountains segment show a much thicker Jurassic series in the inverted and uplifted axial parts of this basin than in adjacent areas (Krzywiec, 2002a, 2009; Krzywiec et al., 2009; **Figure 3.42**). Within the Nida Trough between Kraków and the Holy Cross Mountains, a progressive increase in thickness of the Jurassic succession is seen towards the north-east (**Figure 3.42**), confirming the presence of the depocentre in the Holy Cross domain (Kutek & Głazek, 1972; Gutowski & Koyi, 2007).

The Late Cimmerian Unconformity is only evident along the margins of the Mid-Polish Trough, whereas in its axial parts sedimentation continued across the Jurassic-Cretaceous boundary. During the Early Cretaceous, sedimentation was virtually restricted to the axial, differentially subsiding, parts of the Mid-Polish Trough (**Figure 3.19b**). Large areas of the East European Platform were overstepped by sedimentation during the late Albian transgression, when the Polish Lowlands were incorporated into the evolving North-west European Basin (Ziegler, 1990a).

The Sorgenfrei-Tornquist Zone was repeatedly reactivated during the Late Jurassic and Early Cretaceous. Evidence for this is seen in Scania by the development of angular unconformities in Callovian to early Oxfordian, early Berriasian and early Aptian times, and by scattered magmatic activity (Norling and Bergström, 1987; Klingspor, 1976). During the Late Jurassic and Early Cretaceous, the depocentre of the elongated north-west-trending Norwegian-Danish Basin was closely associated with the fault systems of the Sorgenfrei-Tornquist Zone (Surlyk, 2003) and provided repeated connections between the Arctic and Tethys seas via the Mid-Polish Trough (Ziegler, 1990a).



a.



b.

Figure 3.19 a. Late Jurassic tectonic evolution: Kimmeridgian (152 Ma); b. Early Cretaceous tectonic evolution: Hauterivian (132 Ma). Compiled using information from the following published sources: Ziegler (1990a), Cameron et al. (1992), Vejbæk (1997), Dadlez et al. (1998), Lokhorst et al. (1998), Baldschuhn et al. (2001), De Jager (2007). Corroborated by SPBA thickness maps.

8 Late Cretaceous tectonic evolution

Following Aptian crustal separation in the North Atlantic and Bay of Biscay (**Figure 3.20c**), sea-floor spreading continued in the latter area until the early Campanian when the Pyrenean subduction system was activated. Crustal separation was achieved at the same time in the Labrador Sea (**Figure 3.20c**). During the Late Cretaceous, crustal extension between Greenland and Europe focused on the Rockall-Faroe Trough, the Hatton-Greenland rift, and on the Norwegian-Greenland Sea rift. Crustal extension across the North Sea rift system had abated by Albian times and post-rift thermal subsidence commenced. At the same time, the transtensional basins flanking the SPB ceased to subside differentially (Ziegler, 1988, 1990a; Torsvik et al., 2002; Stampfli & Borel, 2004).

Africa-Arabia converged with Eurasia in a counter-clockwise rotational mode related to Cenomanian crustal separation between Africa and South America in the Equatorial South Atlantic, (Dewey et al., 1989; Ziegler & Stampfli, 2001; Rosenbaum et al., 2002). The ensuing Late Cretaceous plate reorganisation in the Alpine-Mediterranean domain reflects the build-up of regional compressional stresses and the deformation of the weakest elements of its continental and oceanic lithosphere. In the Alpine Tethys domain, this involved Cenomanian-Turonian activation of the Piemont-Penninic-Vahic subduction system along the northern margin of the Austro-Alpine-ALCAPA block (Schmid et al., 1996, 2004, 2008; Stampfli & Borel, 2004), as well as Campanian activation of the Pyrenean and the Alboran subduction systems along the northern and south-eastern margin of Iberia respectively (Dèzes et al., 2004; Ziegler & Dèzes, 2007). Moreover, compressional stresses built up during late Turonian and Senonian times in the west and central European platform, as indicated by the 'Sub-Hercynian' inversion of Mesozoic tensional basins (Ziegler, 1990a, 1998; Kley & Voigt, 2008). Build-up of these north-east to northerly directed intraplate compressional stresses in the foreland of the evolving Alpine Orogen during subduction of the Piemont-Penninic-Vahic Ocean suggests that Africa-Europe convergence was not fully compensated by this subduction system and that the European lower plate was not entirely decoupled from the overriding Austro-Alpine upper plate. The Pyrenean and Alboran subduction systems in the Iberian domain largely compensated for the Africa-Europe convergence. Upon Paleocene closure of the Piemont-Penninic-Vahic Ocean, subduction resistance of the west and central Alpine Briançonnais Terrane, and the European passive margin in the Austro-Alpine and Carpathian domains, accounted for the build-up of collision-related intraplate compressional stresses, which underlay the 'Laramide' pulse of basin inversion and upthrusting of basement blocks (Ziegler 1990a; Ziegler et al, 1995, 1998; Dèzes et al., 2004; Ziegler & Dèzes, 2007).

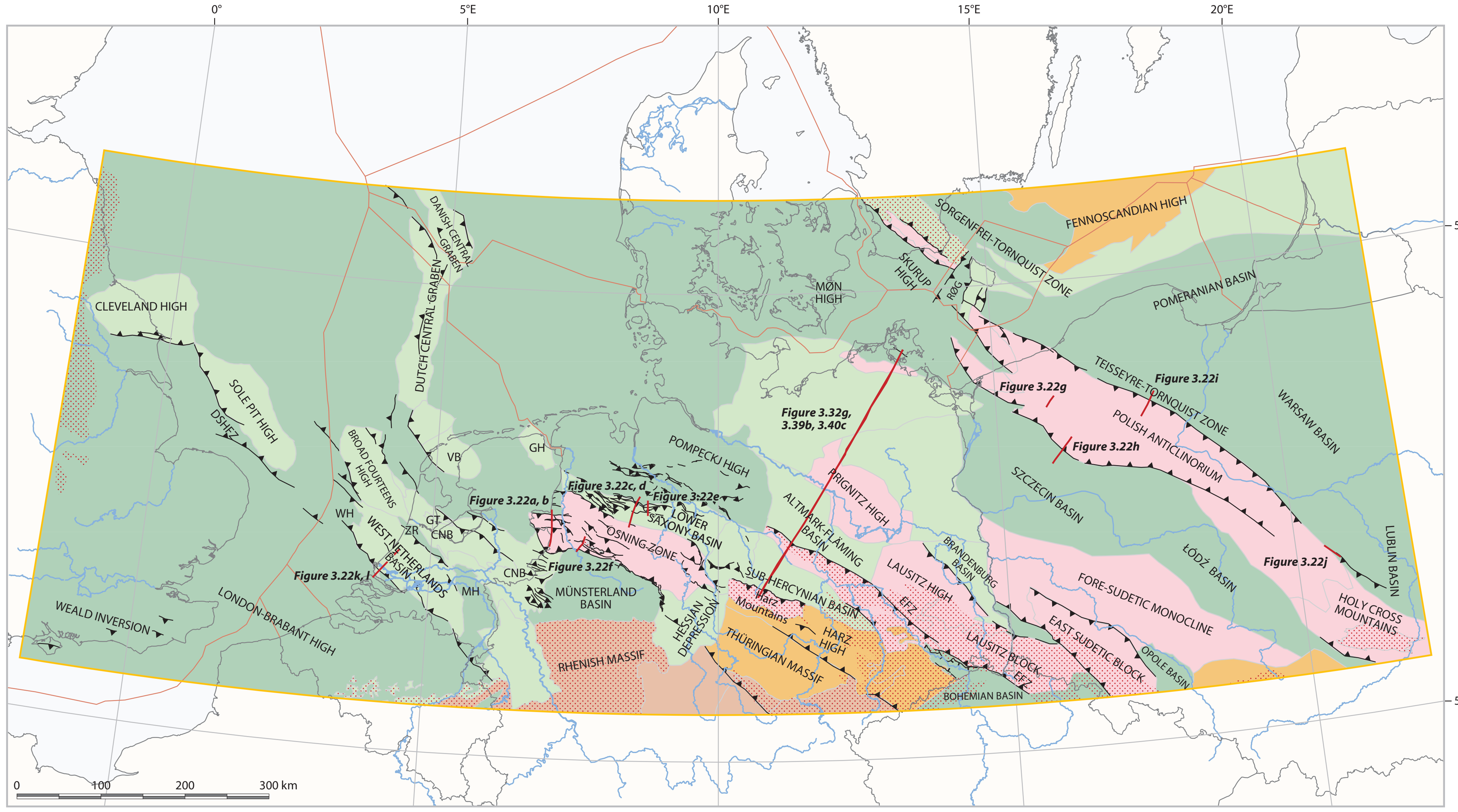
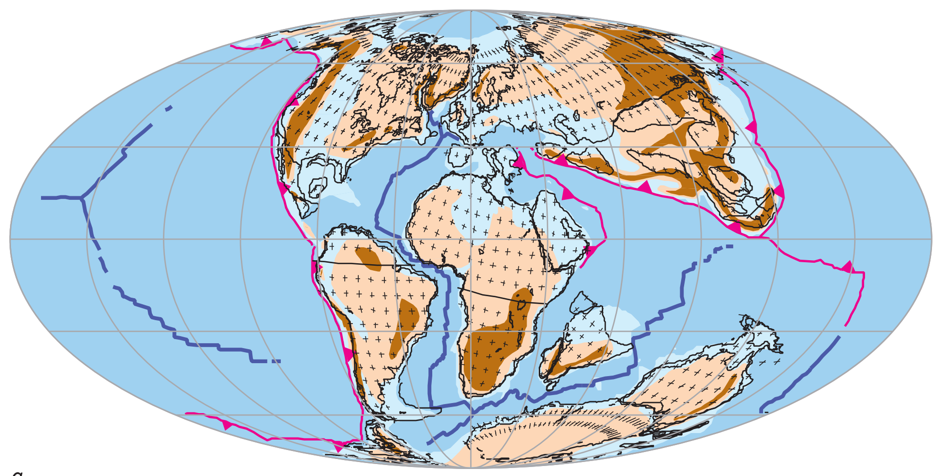
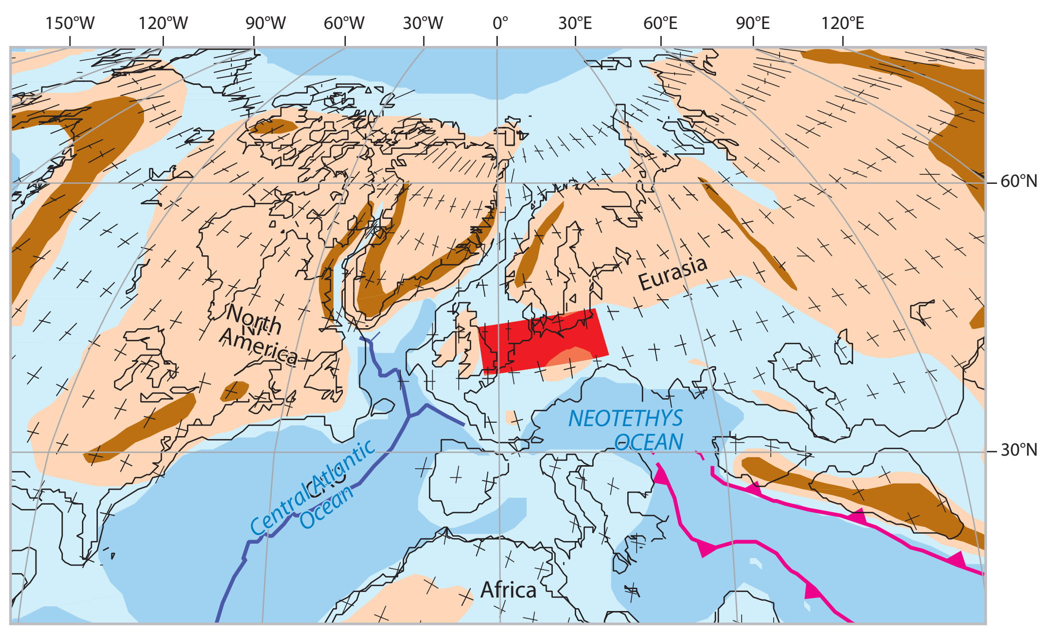


Figure 3.21 Late Cretaceous tectonic evolution: Santonian (85 Ma). Inversion elements after Ziegler (1990a), Baldschuhn et al. (2001), Vejbæk & Andersen (2002), Mazur et al. (2005) and Scheck-Wenderoth & Lamarche (2005), as summarised in Kley & Voigt (2008). Compiled using information from the following published sources: Ziegler (1990a), Cameron et al. (1992), Vejbæk (1997), Dadlez et al. (1998), Lokhorst et al. (1998), Baldschuhn et al. (2001), De Jager (2007). Corroborated by SPBA thickness maps.

Cenomanian-Turonian (94 Ma)

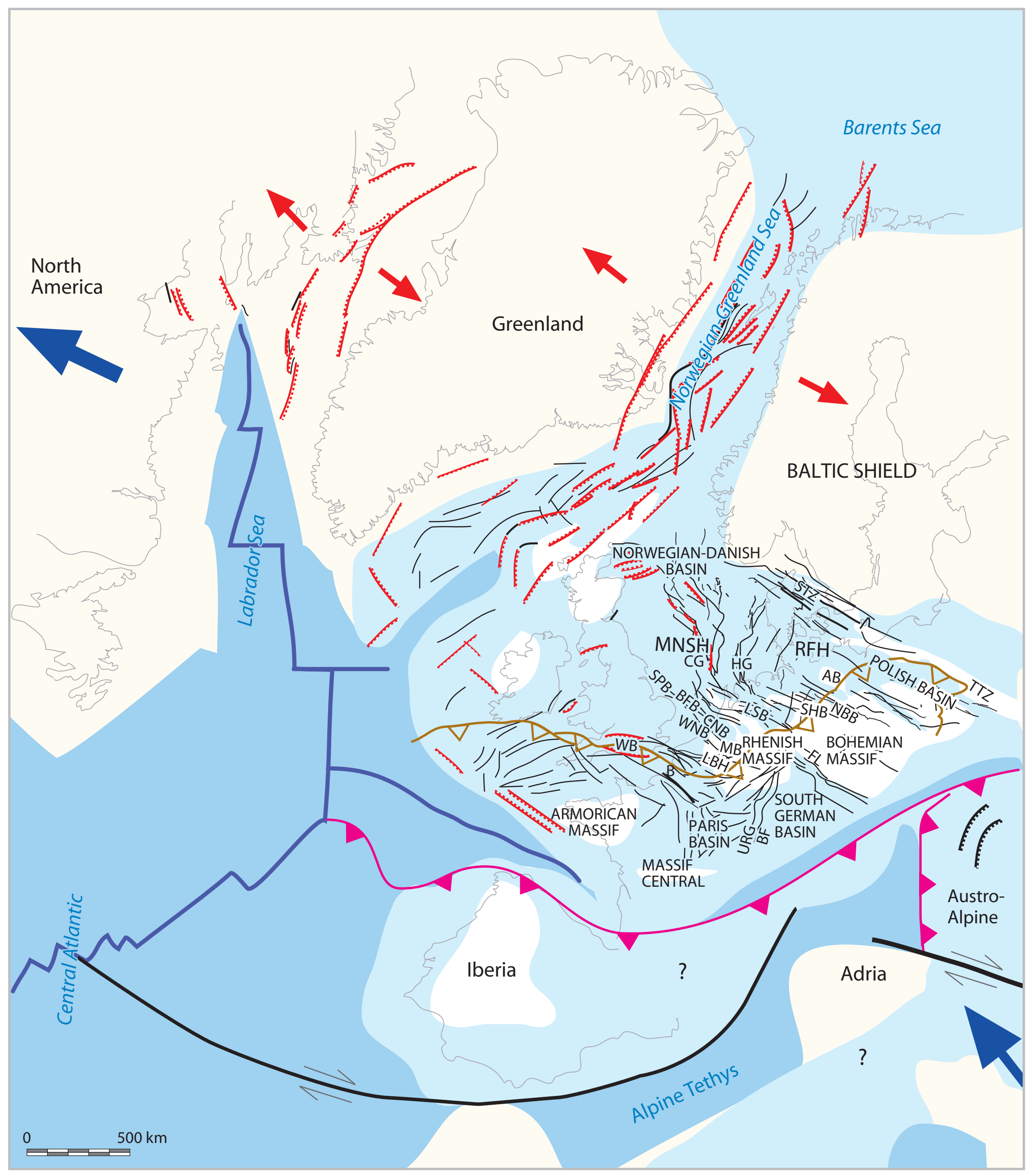


a.



b.

Late Cretaceous (Cenomanian; ~70 Ma)



c.

Figure 3.20 Late Cretaceous tectonic evolution:
a. Palaeogeographic map for the Cenomanian-Turonian (94 Ma);
b. Detail of Cenomanian-Turonian palaeogeographic map;
c. Structural overview map for the Late Cretaceous (~70 Ma) (after Scheck-Wenderoth et al., 2008, Figure 14). Palaeogeographic reconstructions after C. Scotese, kindly supplied by Shell.

8.1 *Transgression and thermal subsidence (Albian to Turonian)*

Regional thermal subsidence of the North Sea Basin started during the Hauterivian and Barremian in combination with gradually rising sea levels, such that the SPB was incorporated by Aptian-Albian times into a vast shallow-marine basin. During the Late Cretaceous, this basin further expanded to reach its maximum extent in response to thermal subsidence and sea-level rise to about 100-200 m above the present-day level. The Upper Cretaceous Chalk series is up to 2000 m thick in the basin (Ziegler, 1990a).

Transtensional activity in the North Sea along the Central Graben and the Sorgenfrei-Tornquist Zone gradually abated during the Early Cretaceous, although Albian to early Cenomanian basaltic dykes testify to persisting activity along these crustal-scale fault systems (Norling & Bergström, 1987; Van Bergen & Sissingh, 2007). Similarly, the deep-seated intrusions inferred beneath the Lower Saxony Basin may have been emplaced during the Aptian (Betz, et al., 1987; Ziegler, 1990a).

Thermal subsidence of the North Sea rift system continued during the Cenomanian, Turonian and later Cretaceous times. Compressional stresses related to convergence of Africa-Arabia with Europe (**Figure 3.20**) started to build-up during the late Turonian, leading to reactivation of pre-existing crustal discontinuities in the SPB area.

8.2 *The Sub-Hercynian inversion phase (Coniacian to Maastrichtian)*

The lithosphere of the SPB area was subjected to compressional intraplate stresses starting in the late Turonian to Coniacian, when the tensional and transtensional Mesozoic basins of the SPB began to experience pulsating phases of inversion that continued well into the Cenozoic. The first of these pulses, the Sub-Hercynian Phase, peaked during Campanian times. Deformation was mainly localised on north-west-striking fault zones of inherited crustal weakness (**Figures 3.21** & **3.22**), and to a lesser extent on north-trending grabens such as the Dutch Central Graben (**Figure 3.36**) (Ziegler, 1990a, 1998; Kley & Voigt, 2008). The geometry of inversion structures and transpressively reactivated (north-west–south-east) faults indicates north–south to north-east–south-west-directed compression (Scheck-Wenderoth et al., 2008; Kley and Voigt, 2008) throughout the SPB area. This has been deduced from seismic data in the North German Basin (Kossow et al., 2000; Otto, 2003; Franzke et al., 2004; Voigt et al., 2004) and is consistent with independent studies that investigated the palaeo-stress field evolution of the Polish Basin, for example in the Holy Cross and Sudetic mountains (Lamarche et al., 1999, 2002). Ziegler (1990a) has advocated a strong transpressional component during inversion-related reactivation of fault zones. Development of major anticlinoria, such as the Osning Zone of the Lower Saxony Basin and the Polish Anticlinorium (**Figures 3.21a** to **d**; **Figures 3.22g** & **h**), involved shortening of the thick sedimentary fill in the inverting basins. Inversion of tensional basins was paralleled by upthrusting of the basement blocks forming the Bohemian Massif, as well as the Harz Mountains and Lausitz Block (Malkovsky, 1987). Similarly, upthrusting of the Romeleås basement block commenced along the Sorgenfrei-Tornquist Zone and was accompanied by basaltic dyke intrusions (Norling & Bergström, 1987).

8.2.1 North German Basin

The North German Basin was strongly affected by intraplate compressional stresses from Coniacian to end-Campanian times (Boigk, 1981; Baldschuhn et al., 1985, 1991, 2001; Betz et al., 1987; Kockel, 2003; Kley & Voigt, 2008). Seismic sections across the Lower Saxony and Altmark-Brandenburg basins (**Figures 3.39** & **3.40**) demonstrate that their inversion axes coincide with their north-west trending Late Jurassic to Early Cretaceous depocentres (Lohr et al., 2007; Scheck-Wenderoth et al., 2008). The section crossing the Lower Saxony Basin clearly shows that the area with thickest Lower Cretaceous strata underwent maximum uplift during the Late Cretaceous. This section also images normal faults of Early Cretaceous age, which were reactivated as reverse faults during the Late Cretaceous inversion phase (Betz et al., 1987; Mazur & Scheck-Wenderoth, 2005; Lohr et al., 2007). Similar phenomena are seen in the Sub-Hercynian Basin (Kossow et al., 2000; Otto, 2003; Franzke et al., 2004; Voigt et al., 2004). Inversion was accompanied by a new phase of salt movement during which north-west-trending salt diapirs formed parallel to the uplifted blocks. The sections shown in **Figure 3.29** illustrate the degree of tectonic decoupling caused by the salt. Basement-involved deformation is restricted to the southern North German Basin margin, where Zechstein salt is thin or missing. Along the Gardelegen Fault, for example, the pre-Zechstein basement south of the fault is uplifted by about 5000 m relative to the basin (**Figure 3.39**) (DEKORP-BASIN Research Group & Krawczyk, 1999; Kossow & Krawczyk, 2002). In contrast, compressional deformation in the central North German Basin is mostly concentrated above the Zechstein salt and is expressed in folding of the cover in salt-cored anticlines, and in salt-withdrawal synclines. Inversion movements in the Lower Saxony Basin (**Figure 3.22**) are restricted to the tensional sub-basins and grabens associated with major basement faults and did not affect the intervening platforms.

Two phases of inversion can be observed: 1) dislocational deformation (folding and thrusting?) during the Coniacian to early Campanian and 2) uplift of the inversion structures in late Campanian to Maastrichtian times. Phase 1 was preceded during the Turonian to Coniacian by slumped chalks, turbidites and olistostromes indicating increasing tectonic instability at the southern basin margin (Voigt, 1977). During the dislocational phase, the bounding normal faults of the former grabens became reverse faults or undercompensated reverse ('phaeno-normal') faults. The sedimentary fill within the basins and grabens was thrusted over the former graben shoulders (**Figure 3.22f**) and onto the adjacent stable uninverted Pompeckj (north) and Münsterland (south) platforms (**Figure 3.21**). Inversion is only expressed at the major fault zones transecting these blocks. Sub-horizontal thrusts extending over 4000 m occur within the Röt Salt along the northern margin of the Lower Saxony Basin (**Figure 3.22e**). Extensional faults developed in the crestal regions of the inversion structures. The rising antiforms were immediately eroded and debris was shed into contemporaneous 'marginal sinks' on the former graben shoulders. Their floors became intensely fractured, thrusted and uplifted during inversion of the Lower Saxony, Sub-Hercynian and Altmark-Brandenburg basins. In the Devonian-Carboniferous 'basement' of the Lower Saxony Basin, up to 8000 m of shortening occurred along listric thrust planes that can be traced into the lower crust. Vertical uplift of the sub-Zechstein basement and erosion of the Upper Jurassic to Lower Cretaceous strata locally amounts to 7000 m; up to 2500 m of strata accumulated in the adjacent syntectonic basin. Total vertical displacement is therefore of the order of 9 to 10 km (Kley & Voigt, 2008). Salt plays a significant role in the architecture of inversion structures. For instance, the Triassic Röt Salt provides the detachment horizon for thin-skinned thrust sheets as shown in **Figure 3.22e**. Furthermore, adjacent to pre-existing salt diapirs, Zechstein salt can be injected into Röt Salt levels to form salt 'wedges' that can be several kilometres wide and tens of kilometres long (**Figure 3.22e**). The salt of some pre-existing diapirs was squeezed out and formed large salt overhangs, probably extruding as salt glaciers onto the Cretaceous sea bed (**Figures 3.29b** & **c**). During this process, the width of diapir stems was reduced or completely squeezed off from the basal Zechstein salt layer. Open-folds, wrench-induced positive flower structures, and small-scale thrust faults are all seen at shallow levels (**Figure 3.22b**) (Ziegler, 1990a).

Inversion structures in the North German Basin strike mainly west-north-west as seen in the Lower Saxony, Sub-Hercynian and South-east Brandenburg-Lausitz basins. However, inversion structures have also been observed along north-north-east to north-north-west-striking elements, such as the eastern margin of the Brunswick-Gifhorn, Egge and Adler-Kamien-Rønne Graben, the Marienburg-Eicklingen zones and the southern end of the Glückstadt Graben (Bremen Graben).

The central Lower Saxony Basin was deeply eroded prior to the late Campanian transgression. During inversion, Santonian to Campanian depocentres migrated away from the Lower Saxony Basin inversion axis northwards onto the Pompeckj Block and southwards onto the Münsterland Block. The Dutch sector of the Lower Saxony Basin was only mildly inverted. According to Ziegler (1990a), the Lower Saxony Basin was significantly affected by the Laramide Phase of inversion although according to F. Kockel (pers. comm. 2008) there is little evidence for this; however, Betz et al. (1987) record inversion movements as late as the Paleocene.

The Harz Mountain basement block was thrusted up along a south-west-dipping reverse fault starting in mid-Senonian times and continuing into mid-Paleocene times (Boigk, 1968; Führer, 1988; Ziegler, 1990a). The vertical throw at the frontal Harz Mountain thrust amounts to 5000 m at the base-Permian level (**Figure 3.39b**). A similar inversion history is likely for the Flechtingen High, formerly the Prignitz Basin (Kossow & Krawczyk, 2002). The Bohemian Massif and overlying Cretaceous basin are transected by a number of north-west-trending thrusts and reverse faults including the Lužice (Lausitz) and Main Intra-Sudetic Faults, and the western boundary zone of the massif, the Franconian Fault (Reicherter et al., 2008). The KTB borehole demonstrated imbrication of the crust along a zone of strong dextral transpression (O'Brien et al., 1997; Wagner et al., 1997).

8.2.2 Southern North Sea Basin

In the southern North Sea area, early Late Cretaceous thermal subsidence was followed during the Late Cretaceous by strong inversion of the north-west-trending Sole Pit, Broad Fourteens, West and Central Netherlands basins (**Figures 3.22k** & **l**) and the northerly trending Central Graben (**Figures 3.35a** & **3.36b**). Inversion of these basins started variably during the late Turonian or early Senonian and initially peaked during the Campanian Sub-Hercynian pulse. Chalk sequences thin towards the inversion axes with Maastrichtian and Danian chalk onlapping unconformities (**Figure 3.35**) in the Broad Fourteens and Central and West Netherlands basins, the Dutch Central Graben and the Roer Valley Graben, which variably cut down into Lower Cretaceous and Jurassic rocks (Gras & Geluk, 1999; De Jager, 2007). However, the last inversion phase took place during the Paleocene-Eocene Laramide pulse of intraplate compression, as indicated by the strong truncation of the Mesozoic series (**Figure 3.22k**) at the unconformity at the base of the Paleogene Lower North Sea Group (De Jager, 2007). During basin inversion, the thick Zechstein salts have a distinct decoupling effect as indicated by upwarping of the Mesozoic fill of the Central Graben into a broad anticline (**Figures 3.35a** & **3.36b**).

The Danish Central Graben (Vejbæk & Andersen, 1987, 2002; Cartwright, 1993) was inverted during the Senonian and Paleocene, with three particularly strong phases during the latest Santonian, mid-Campanian, late Maastrichtian and two phases in the Early Cenozoic, as evidenced by synkinematic sedimentation (Vejbæk, 1997). Early inversion was focussed in narrow zones associated with the reverse reactivation of pre-existing normal faults; later phases of inversion are dominated by gentler flexuring and folding attributed to 'stress relaxation' by Nielsen et al. (2005). In the Danish Central Graben, as elsewhere, inversion is focussed in the area with thickest Upper Jurassic to Lower Cretaceous strata. This is also seen in the Sole Pit Basin (**Figure 3.33a**) (Badley et al., 1989; Nalpas et al., 1995; Buchanan et al., 1996) and the Broad Fourteens Basin (**Figure 3.35b**) (Nalpas et al., 1995; De Lugt et al., 2003). In the latter, severe inversion and the presence of Zechstein salt has led to the development of spectacular low-angle reverse faults (De Roos & Smits, 1983; NITG-TNO, 2004), particularly at the north-eastern margin (**Figure 3.35b**). Inversion of the Broad Fourteens Basin resulted in the development of broad north-west-trending anticlinal arches (Dronkers & Mrozek, 1991; Nalpas et al., 1995). A thin chalk conglomerate (block P9 wells), presumably Maastrichtian to Danian age, suggests that inversion of the Broad Fourteens Basin was completed mainly during the Sub-Hercynian Phase, and that the Laramide and Pyrenean phases mainly affected the southern part of the basin. Details on basin inversion are provided by Van Wijhe (1987a, 1987b), Hooper et al. (1995), Huyghe & Mugnier (1995) and Nalpas et al. (1995). The Cleaver Bank High to the north of the Broad Fourteens Basin was affected by compression, generating localised pop-up structures at Rotliegend levels (**Figure 3.35a**). These follow north-west–south-east, north-east–south-west and west-north-west trends, which were reactivated during Late Cretaceous and Early Cenozoic compression (De Jager, 2007).

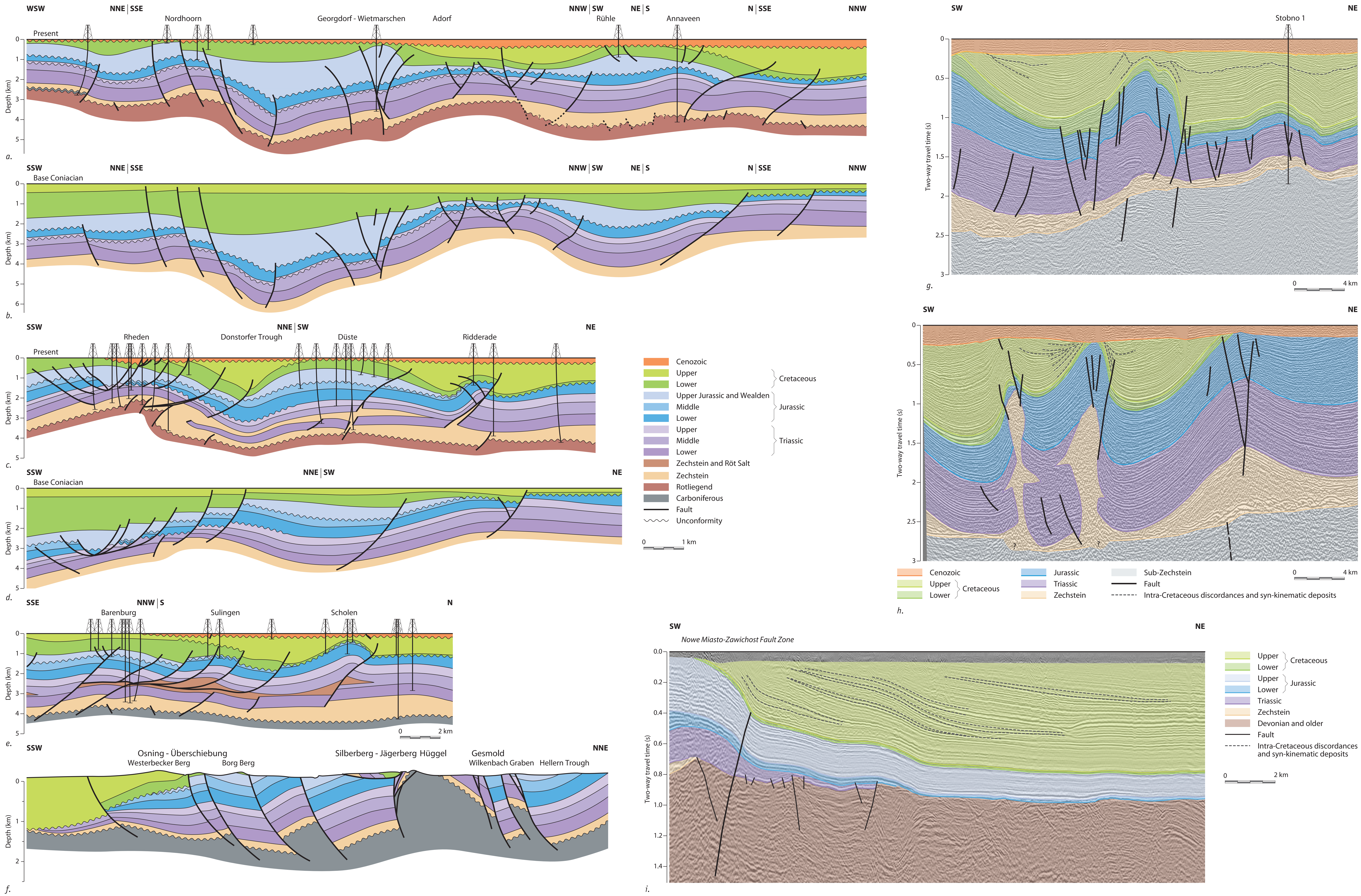
The inverted West and Central Netherlands basins (**Figure 3.35c**) are located onshore on the same trend as the Broad Fourteen Basin. The chalk has been eroded across the crests of their anticlinoria, which began growing in late Santonian times, whereas it is up to 1500 m thick along their flanks (Ziegler, 1990a). The structural relief generated by the combined Sub-Hercynian and Laramide inversions is 2000 m at the base-Zechstein level and up to 3500 m at shallow Mesozoic levels (Heybroek, 1974; Ziegler, 1983). Pre-existing faults were reactivated in reverse during inversion of the West Netherlands Basin, with transpressional movements giving rise to a series of prominent north-west-trending flower structures (**Figures 3.22k** & **l**). Many of these transpressionally reactivated faults still display normal offsets at deeper levels. Thinning of the Chalk Group towards the inverted basin in the north-west, and several unconformities imaged by seismic data, show that these flower structures were formed during the Sub-Hercynian inversion phase. Strong inversion in the Central Netherlands Basin led to erosion of most of the Upper Jurassic to Lower Cretaceous 'syn-rift' sequence, except in the Gouwzee Trough depocentre (**Figure 3.35c**) where thick Jurassic sediments are preserved between two inversion axes. The Sub-Hercynian Phase is not well documented here as Upper Cretaceous strata are not preserved; however, the Danian Chalk Group overlies pre-Cretaceous successions in several wells (NITG-TNO, 2004). The Zandvoort-Maasbommel High (**Figure 3.35c**) separating the West and Central Netherlands basins was only slightly inverted (Van Wijhe, 1987a; NITG-TNO, 2004). Sub-Hercynian inversion of these transtensional basins resulted in uplift of their sedimentary fill, erosion of most of the Chalk Group and locally older series, and contemporaneous rapid subsidence of the adjacent highs (**Figures 3.35c** & **3.36d**; North Holland Platform, Zandvoort High and Voorne Trough). The Roer Valley Graben (**Figure 3.36e**) was also affected by the Sub-Hercynian inversion, most strongly in the north-west, whereas the adjacent Maasbommel High subsided (Geluk et al., 1994; Gras & Geluk, 1999). The inversion axes were overstepped by the Maastrichtian to Danian Chalk Group, reflecting a brief quiescent interlude prior to the Laramide inversion pulse that strongly affected all Late Jurassic to Early Cretaceous basins in the southern North Sea area (Ziegler, 1982a, 1990a, 1998; De Jager, 2007).

The entire London-Brabant Massif was overstepped by Upper Cretaceous sediments, but was partly denuded during subsequent upwarping of the Weald-Artois Axis (Dusar & Lagrou, 2007; P. Ziegler, pers. comm. 2009). Inversion of the Sole Pit Basin started during the late Turonian and peaked during the early Campanian, as indicated by the onlap of Campanian Chalk Group sediments against the inversion axis (Van Hoorn, 1987; Badley et al., 1989; **Figure 3.24**). Late Campanian to Maastrichtian chalks resting unconformably on Jurassic rocks show that Sub-Hercynian inversion movements in the Sole Pit Basin ended during mid-Campanian times (Ziegler, 1990a); the Laramide inversion movements were very mild in contrast.

8.2.3 Polish Basin

Inversion of the Mid-Polish Trough was accompanied by Late Cretaceous and younger collisional events within the Carpathians (cf. Krzywiec, 2006b; Scheck-Wenderoth et al., 2008). In the Inner Carpathians, these include thrusting of the Fatric Basin (mid-Albian to late Turonian), emplacement of the Fatric-Hronic nappe system (late Turonian), uplift of the Veporic core and further nappe thrusting (late Turonian to early Campanian), final stacking along the outer Tatric edge (mid-Campanian to early Maastrichtian), and closure of the Penninic-Vahic oceanic trough (mid-Maastrichtian to Danian) (Plasienka, 1997). The Pieniny Klippen Belt underwent early compression during the Aptian to Turonian or Senonian and main subduction and compression during Campanian, Maastrichtian and Paleocene times (Birkenmajer, 1986). The Outer Carpathian compressional phases reflect the collision between the European continent and intra-oceanic arcs and include: Turonian inversion of the Silesian, Sub-Silesian and Skole basins; Campanian and younger inversion of the Magura Basin; Late Eocene to Early Miocene final closure of the Outer Carpathian basins; and finally Miocene folding and thrusting accompanied by formation of the Carpathian foredeep basin (Oszczypko, 2004). The Late Cretaceous to Paleocene compressional stresses generated within the continental collision zone were presumably transferred into the Carpathian foreland where they led to reactivation and eventual inversion of the Mid-Polish Trough.

Late Cretaceous reactivation of older normal faults is seen along the Teisseyre-Tornquist Zone in the axial part of the inverted Polish Basin (**Figure 3.42**) (Erlström et al., 1997; Krzywiec, 2002a; Krzywiec et al., 2003). The Nowe Miasto-Zawichost Fault Zone, a strand of the Teisseyre-Tornquist Zone, was active as a normal fault zone in Permian to Jurassic times and was compressionally reactivated during the Sub-Hercynian and Laramide inversion (**Figure 3.21**) of the south-eastern Mid-Polish Trough (Krzywiec, 2007a, 2009; Krzywiec et al., 2009). Inversion of the Mid-Polish Trough during the Late Cretaceous and Paleocene resulted in uplift and erosion of its axial part (Mid-Polish Swell or Anticlinorium). Strike-slip tectonic activity took place during the inversion (Krzywiec, 2002a; Krzywiec et al., 2003, 2009; Gutowski & Koyi, 2007). The amount of uplift and erosion makes it difficult to determine the exact timing of onset, peak and end of inversion, but it may have started as early as the late Turonian and continued until Maastrichtian to Paleogene times (e.g. Leszczyński, 2000, 2002a; Krzywiec, 2002a, 2002b, 2004a, 2006a, 2007a, 2009; Krzywiec et al., 2003, 2009; Mazur et al., 2005; Resak et al., 2008) and ended during the Early Cenozoic. Ziegler (1990a) has suggested that most of the growth of this structure was a Laramide and not a Sub-Hercynian phenomenon. In parts of the basin such as the Nida Trough (**Figure 3.42f**), basement-fault zones were still in extension during the Late Cretaceous and were reactivated as reverse faults during the Early Cenozoic (Scheck-Wenderoth et al., 2008). Despite inversion and uplift, the sub-Zechstein basement still forms a regional syncline in the central (Kuiavian) Mid-Polish Trough (**Figures 3.41b** & **3.42d**), descending to a depth of about 8000 m. In contrast, it lies at much shallower depths (3000-4000 m) in the Pomeranian and offshore areas of the Mid-Polish Trough where it is flanked by two minor depressions (cf. profiles 1 and 2 from **Figure 3.42**). A similar pattern is also seen in the Holy Cross Mountains, where the sub-Zechstein basement is uplifted and exposed along the axis of the trough. The configuration of the top sub-Zechstein level contrasts with that of the top-base Upper Cretaceous level. The base Upper Cretaceous level along the axis of the inverted Polish Trough describes a 750 km-long anticlinal axis, the amplitude of which increases from about 1500 m in the north-west to over 3000 m in the Holy Cross Mountains (Pożaryski, 1977a; Dadlez, 1980b) where the pre-Permian floor of the trough is exposed. The Base Cenozoic Unconformity consequently truncates the lowermost Jurassic /


WSW
NNE | SSE
Nordhoorn
Georgdorf - Wietmarschen
Adorf
NNW | SW
Rühle
NE | S
Annaveen
N | SSE
NNW
Present
Depth (km)
a.
SSW
Base Coniacian
NNE | SSE
NNW | SW
NE | S
N | SSE
NNW
b.
SSW
Rheden
Donstorfer Trough
NNE | SW
Düste
Ridderade
NE
Present
c.
SSW
Base Coniacian
NNE | SW
NE
d.
SSE
Barenburg
NNW | S
Sulingen
Scholen
N
0
2 km
e.
SSW
Osning - Überschiebung
Westerbecker Berg
Borg Berg
Silberberg - Jägerberg Hüggel
Gesmold
Wilkenbach Graben
Hellern Trough
NNE
f.
Cenozoic
Upper
Lower
Cretaceous
Upper Jurassic and Wealden
Middle
Lower
Jurassic
Upper
Middle
Lower
Triassic
Zechstein and Röt Salt
Zechstein
Rotliegend
Carboniferous
Fault
Unconformity
0
1 km
SW
NE
Stobno 1
Two-way travel time (s)
0
4 km
g.
SW
NE
Cenozoic
Upper
Lower
Cretaceous
Jurassic
Triassic
Zechstein
Sub-Zechstein
Fault
Intra-Cretaceous discordances and syn-kinematic deposits
0
4 km
h.
SW
NE
Nowe Miasto-Zawichost Fault Zone
Two-way travel time (s)
Upper
Lower
Cretaceous
Upper
Lower
Jurassic
Triassic
Zechstein
Devonian and older
Fault
Intra-Cretaceous discordances and syn-kinematic deposits
0
2 km
i.

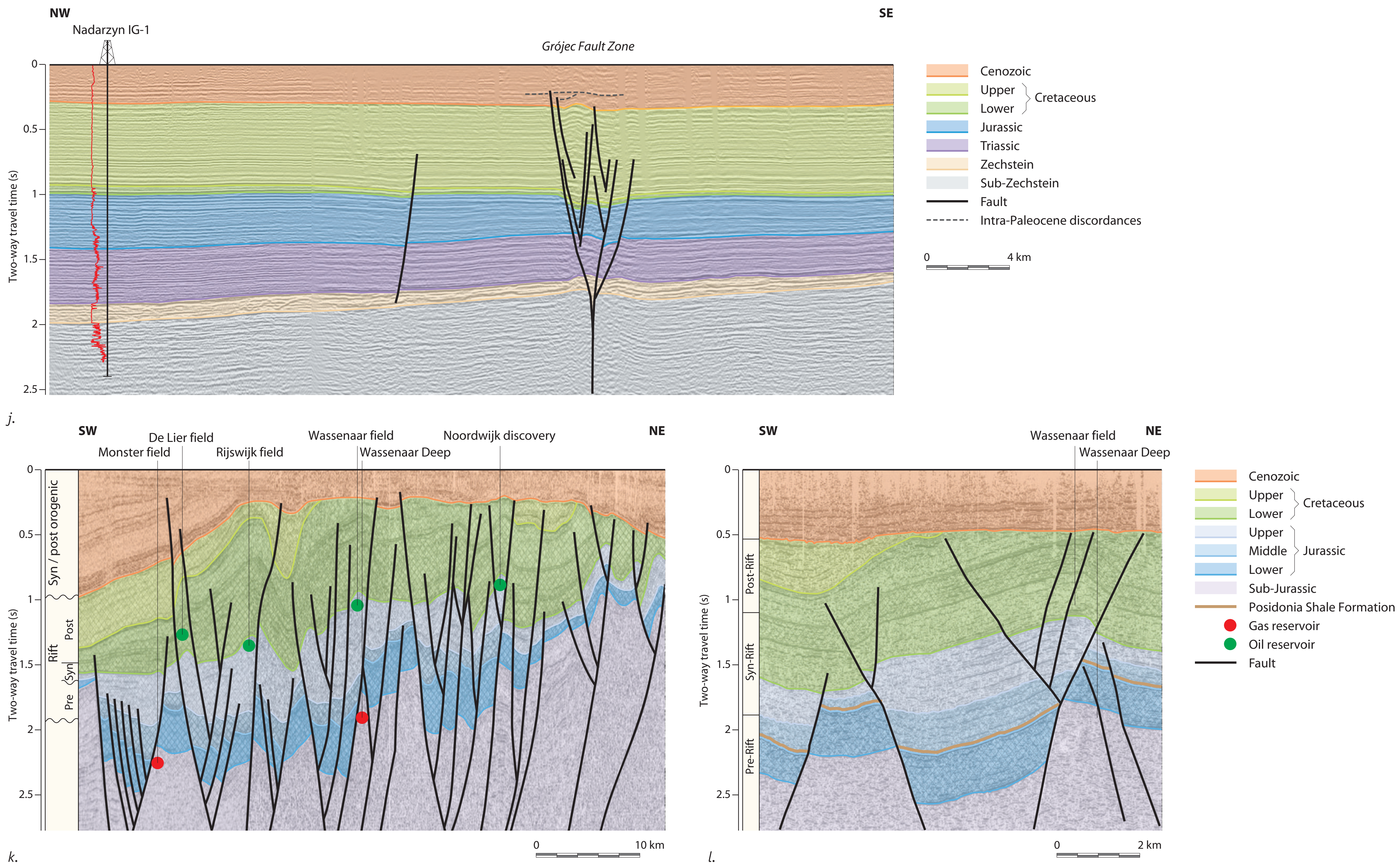


Figure 3.22 Seismic lines and cross-sections showing examples of Sub-Hercynian and Laramide inversion structures:
a. Cross-section across the Nordhoorn, Georgdorf-Wietmarschen and Rühle-Annaveen inversion structures, Emsland, western Lower Saxony Basin (Kockel, 2003). The fill of the Rühle Trough was upwarped during inversion in Coniacian-Santonian times. A hydrocarbon trap formed in the Bentheim Sandstone Member (Valanginian) and the northern boundary fault transformed into a partly under-compensated reverse fault. The adjacent anticline (Adorf Swell) was also mildly inverted. The Georgdorf-Wietmarschen Graben was converted into a pop-up structure, modified by the mobilisation of Upper Jurassic salt. These structures have undergone no further deformation since the Paleogene (Baldschuhn et al., 1991). As no Maastrichtian and Danian sediments are preserved, a component of Laramide inversion cannot be excluded;
b. Restoration of cross-section 3.22a to base-Coniacian level (Kockel, 2003). From the Late Jurassic onwards, the Rühle-Annaveen structure was a subsiding trough bordered by a synsedimentary fault in the north and by the Adorf Anticline (with crestal graben) in the south. The Georgsdorf-Wietmarschen structure was a synsedimentary graben;
c. Cross-section across the Ridderade-Düste inversion structures at the northern margin of the Lower Saxony Basin (Kockel, 2003). The Düste Basin became an anticline during Coniacian-Santonian times while the Ridderade Graben became a bivergent pop-up structure flanked by deep marginal troughs. Zechstein salt intruded the Triassic Röt Salt layer. The Rheden structure was upwarped and the former normal boundary faults reversed. The Donstorfer Trough is another Coniacian-Santonian marginal trough;
d. Restoration of cross-section 3.21c to base-Coniacian level (Kockel, 2003). The Düste Basin formed as a platform on the northern step-faulted margin of the central Lower Saxony Basin during the late Early Cretaceous;
e. Cross-section across the Barenburg, Sulingen, Scholen inversion structures at the northern margin of the Lower Saxony Basin (Kockel, 2003). The Zechstein salt is thick and there is a decollement at this level and in the Upper Buntsandstein. Gently dipping thrusts are detached in both Permian and Triassic salt;
f. Cross-section across the Osning Lineament-Nordwestfalen-Lippe Swell at the southern margin of the Lower Saxony Basin. The Zechstein salt is thin and not involved in the structures. Steep listric faults root directly in the pre-Permian basement and are under-compensated. A marginal trough south of the southernmost thrust is filled with 1000 m of Coniacian-Santonian strata (Baldschuhn & Kockel, 1998);
g. Seismic section across the inverted Mid-Polish Trough and the Drawno salt structure. Example from the SW flank of the Pomeranian segment of the inverted Mid-Polish Trough showing the compressionally reactivated Drawno salt structure (based on Krzywiec, 2006b).The Turonian-Campanian succession above this salt structure thins towards the Drawno salt structure and is characterised by intraformational local unconformities, which reflects synkinematic sedimentation above and adjacent to the diapir during the Sub-Hercynian Phase. The entire Upper Cretaceous succession is folded and truncated across the Drawno salt structure as well as across the entire inverted Mid-Polish Trough, and is unconformably overlain by Cenozoic deposits, suggesting a post-Maastrichtian final inversion pulse. Paleogene palaeogeographic maps (Piwocki, 2004) suggest that there are Eocene sediments in the entire Pomeranian segment of the Mid-Polish Trough, apart from its axial parts. This suggests that the final inversion pulse took place during the Paleocene (Laramide Phase), with Eocene deposits forming the basal part of the post-tectonic succession;
h. Seismic section across the inverted Mid-Polish Trough and the Chojnice structure. Example from the NE flank of the Pomeranian segment of the inverted Mid-Polish Trough showing the compressionally reactivated Koszalin-Chojnice structure, which in this case is almost completely decoupled from its sub-Zechstein basement (based on Krzywiec, 2006b). The Triassic-Jurassic cover is characterised by localised thickness increase across this structure, suggesting that it was active during the Mesozoic extension and subsidence of the Polish Basin. Numerous intra-Upper Cretaceous thickness variations and local unconformities document several stages of inversion tectonics; the first was late Turonian in age. A final pulse of inversion took place during late/post-Maastrichtian times, resulting in the development of a regional erosional unconformity that was overstepped by a transgressive Paleogene series (Paleocene – Eocene?, Piwocki, 2004);
i. Seismic section across the Nowe Miasto-Zawichost Fault Zone, one of the components of the Teisseyre-Tornquist Zone (after Krzywiec, 2009; Krzywiec et al., 2009). The variations in thickness of the Upper Cretaceous sequence are related to basin inversion of the Holy Cross segment of the Mid-Polish Trough. This is a detail from the profile illustrated in Figure 3.42f;
j. Seismic section showing strike-slip activity (positive flower structure) along the NE-trending Grójec Fault Zone oriented perpendicular to the axis of the Mid-Polish Trough. The profile is one of few that clearly demonstrates that the inversion-related wrenching was active until mid-Paleocene times (Laramide Phase);
k. Seismic section across the strongly inverted West Netherlands Basin showing the location of the Wassenaar and other fields;
l. Seismic section showing the detailed structure in the vicinity of the Wassenaar field. See Figure 3.21 for locations.

topmost Triassic strata in the Pomeranian segment of the Mid-Polish Swell, the Paleozoic/Precambrian in the Holy Cross Mountains, and the Upper Jurassic in the Kuiavian part (Dadlez, 2001a). Inversion of the Mid-Polish Trough could not compensate for the total Permo-Mesozoic subsidence of the pre-Zechstein level in the central Kuiavian segment of the trough (**Figure 3.42d**), where the pre-Zechstein level is still clearly deeper than in flanking areas. This contrasts with the Baltic (offshore), Pomeranian and the Holy Cross Mountain areas, which were strongly inverted (**Figures 3.42a** & **b**).

Mechanical decoupling of the pre- and post-Zechstein salt sequences strongly influenced the architecture of structures resulting from inversion of the Mid-Polish Trough. Basement-controlled, extensional-fault zones in the south were reactivated as reverse faults and caused uplift of particular basement blocks. However, in the central Kuiavian and north-western Pomeranian areas, this uplift was buffered by the thick Zechstein evaporites. As a result, the Mesozoic cover was mostly deformed by fault-related folding (**Figure 3.42b** to **d**), locally accompanied by reverse reactivation of peripheral structures that developed along the flanks of the trough and are detached within Zechstein evaporites.

8.2.4 Sorgenfrei-Tornquist Zone

Transpressional reactivation of the Sorgenfrei-Tornquist Zone started during the Coniacian and intensified in Santonian to Campanian times, as indicated by the partial inversion of the Norwegian-Danish Basin (**Figures 3.38d** & **e**). Inversion movements slowed during the Maastrichtian to Danian and became pronounced again during the Late Paleocene (Liboriussen et al., 1987; Vejbæk & Andersen, 2002). Inversion movements focussed on the north-west-trending Late Jurassic to Early Cretaceous depocentre of the Norwegian-Danish Basin (Vejbæk 1997), whereas in Scania they involved the transpressional reactivation of fault systems dating back to the Permo-Carboniferous and the injection of basalt dykes. Upthrusting of the Romeleås basement block was accompanied by subsidence of the adjacent Alnarp Trough in which Upper Cretaceous rocks are up to 1800 m thick (Norling & Bergström, 1987). Similarly, inversion of the deep Triassic to Early Cretaceous Rønne Graben took place during the Senonian and Paleocene (Ziegler, 1983; Petersen et al., 2003a).

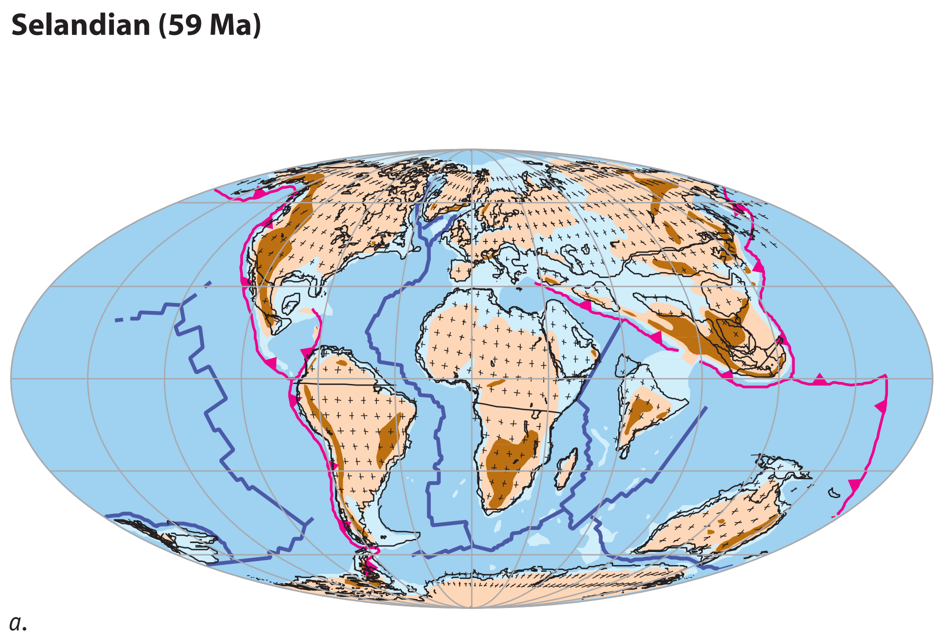
9 Cenozoic tectonic evolution

The carbonate-dominated depositional regime that had prevailed in the SPB area during Late Cretaceous and Danian times gave way to a clastic-dominated depositional regime in response to crustal deformation at the transition to the Selandian, particularly along the southern SPB margin during the Laramide pulse of intraplate compression (Ziegler, 1990a, 1998). Significantly, this change in depositional regime did not occur during the Sub-Hercynian Phase of intraplate compression and basin inversion, which apparently was not as intense as the Laramide Phase. Related basinwide regressions led to the erosional planation of Sub-Hercynian and Laramide inversion structures and the development of the Late Paleocene (Landen, Thanetian) and Early Eocene (London A) unconformities.

During the Cenozoic, the western SPB area was incorporated into the continuously subsiding North Sea thermal sag basin, as indicated by the thicknesses of the North Sea Group deposits (Ziegler, 1990a; Wong et al., 2007b). The Cenozoic depositional history of the North Sea Basin has allowed the establishment of a detailed sequence-stratigraphic framework (Vandenberghe et al., 1998; 2004). The Cenozoic evolution of the SPB area reflects repeated fluctuations in the magnitude and orientation of the intraplate stress field and related subtle reactivations of pre-Cenozoic structural elements (Reicherter et al., 2008). These changes in stress field were controlled by the interaction of the Alpine-Carpathian and Pyrenean orogens with their northern foreland and, following earliest Eocene crustal separation between Greenland and Europe, by increasing ridge-push forces exerted by the Arctic-North Atlantic sea-floor spreading axes (Ziegler, 1988; Gölke & Coblentz, 1996; Torsvik et al., 2002; Dèzes et al., 2004; Jarosiński et al., 2009). Cenozoic sedimentation patterns in the SPB area were strongly influenced by the progressive uplift of the Variscan massifs along its southern flank and by the Late Miocene and Pliocene upwarping of the Fennoscandian Shield and the Plio-Pleistocene subsidence acceleration of the North Sea and North German basins (Scheck-Wenderoth & Lamarche, 2005; Cloetingh et al., 2006, 2008; Ziegler & Dèzes, 2007). Development of the European Cenozoic rift system had only an indirect and marginal affect on the SPB (Ziegler, 1990a; Geluk et al, 1994; Dèzes et al, 2004; Jarosiński et al., 2009).

9.1 The Laramide, Pyrenean and Savian phases of intraplate compression

In most sub-basins of the Netherlands, the strongest inversion took place during the Laramide (mid-Paleocene) pulse, which marked the end of deposition of the Chalk Group and the start of deposition of the siliciclastic North Sea Group (De Jager, 2007). It has also been argued that most growth of the Polish Anticliniorum and inversion of the Danish Basin took place at this time (Ziegler, 1990a). Hardly any Maastrichtian and Danian sediments are preserved in the Lower Saxony Basin, which led Baldschuhn et al. (1991) and F. Kockel (pers. comm., 2007) to infer that inversion movement had ceased during



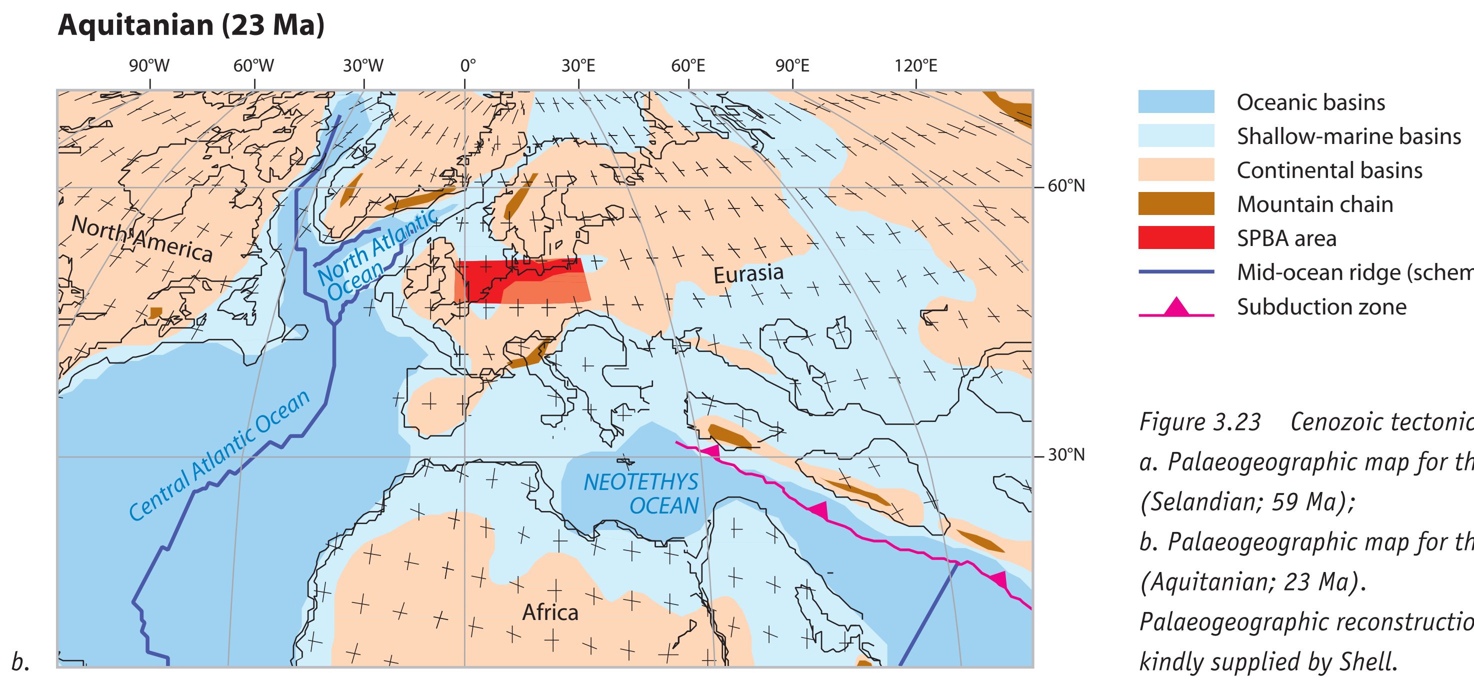


Figure 3.23 Cenozoic tectonic evolution:
a. Palaeogeographic map for the Paleocene (Selandian; 59 Ma);
b. Palaeogeographic map for the Early Miocene (Aquitanian; 23 Ma).
Palaeogeographic reconstructions after C. Scotese, kindly supplied by Shell.

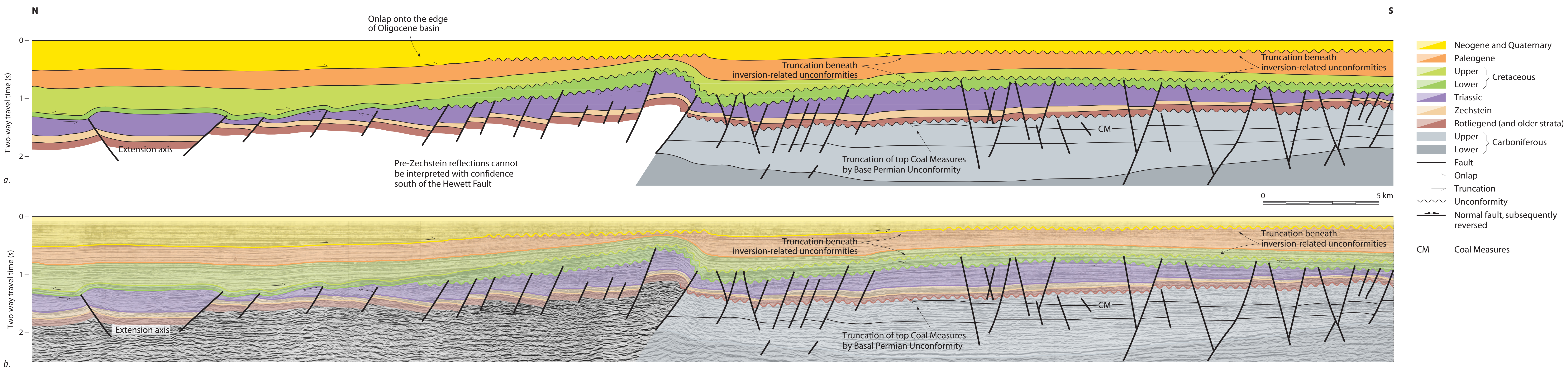
the Campanian, the youngest preserved Upper Cretaceous series. In contrast, Betz et al. (1987) made a case for strong Laramide deformation. The Pyrenean (end-Eocene) pulse caused broad uplift of the Mesozoic West and Central Netherlands basins, the amplitude of which decreases into the Broad Fourteens Basin. The final Savian (end-Oligocene / Early Miocene) pulse led to further significant uplift in the West Netherlands Basin (De Jager, 2007), on the Ringkøbing-Fyn High and Central Graben in the Danish sector (Rasmussen, 2009), and in the Mesozoic Sole Pit, Weald and Cleveland basins (Whittaker, 1985) of the UK sector, but apparently little in the Netherlands or Germany. Simpson et al. (1989) estimated that the Weald Basin was uplifted by up to 1500 m during its Late Eocene to Miocene inversion. The observed progressive westward shift of basin inversion after the Laramide pulse reflects the latest Eocene change in convergence direction of the Adriatic indentor with the European foreland, from a northward direction during the Paleocene and Eocene, to a north-westerly direction during the Oligocene and Miocene (Ziegler, 1987, 1990a; Ziegler et al., 1995; Schmid & Kissling, 2000; Dèzes et al., 2004; Ziegler & Dèzes, 2007).

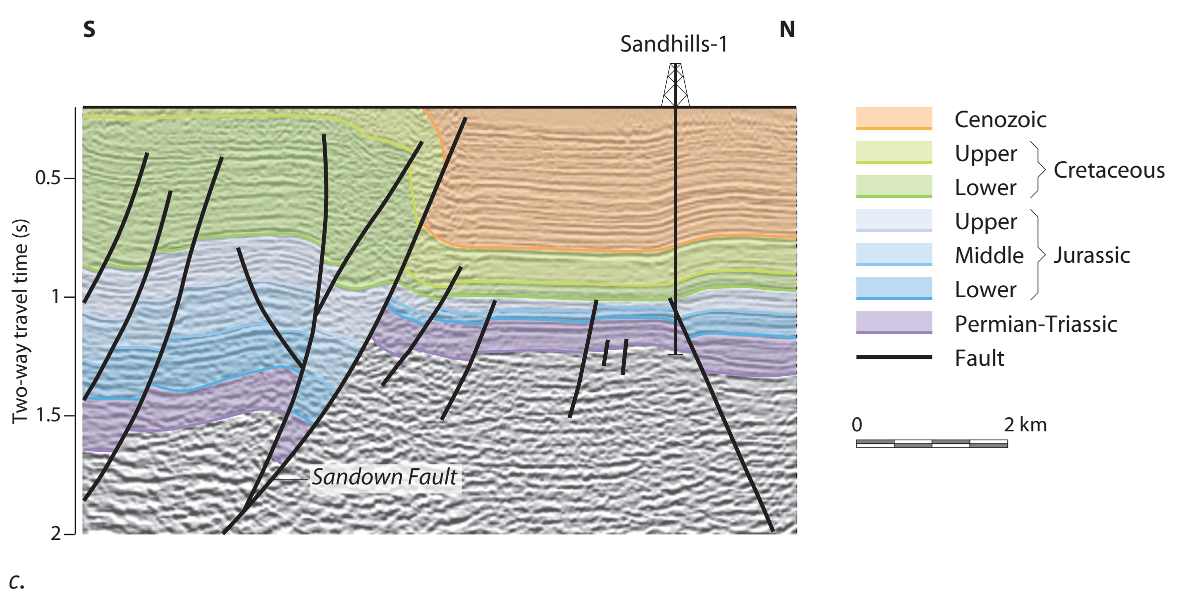
The presence or absence of thick Zechstein salt plays an important role in the architecture of inverted basins (De Jager, 2007). Reverse reactivated faults dominate the structural style of Jurassic and Cretaceous series in basins with little salt, whereas in basins with thick salt (e.g. the Dutch Central Graben) these are involved in broad uplifts. Vitrinite-reflectance, fission-track and fluid-inclusion studies suggest that the total uplift during basin inversion can reach 2500 to 3000 m, as seen in the Mid-Polish Swell (Dadlez, 1980b) and along the Sorgenfrei-Tornquist Zone (Petersen et al., 2003a), but is generally in the order of 1000 to 1500 m (De Jager, 2007). In contrast, upthrusting of basement blocks such as the Harz Mountains and Bohemian Massif along the southern SPB margin can involve crustal imbrication with vertical and horizontal displacements of 3000 to 4000 m along reverse faults (Reicherter et al., 2008).

9.1.1 Southern North Sea Basin

During the Early Cenozoic, the Central Graben and its flanking highs subsided regionally as part of the North Sea thermal sag basin, the axis of which is offset to the west of the Dutch Central Graben. Cenozoic uplift of up to 600 m is recognised in the Danish Central Graben and the Ringkøbing-Fyn High, most of which is of Early Miocene age (Rasmussen, 2009). This uplift may be related in part to regional uplift of the Fennoscandian Shield (Rohrmann et al., 1995) associated with opening of the North Atlantic Ocean.

The southern part of the former Broad Fourteens Basin was affected by both Laramide and Pyrenean inversion (Oele et al., 1981; Van Wijhe, 1987a). A Riedel fault pattern at base-Cenozoic level suggests dextral shearing during the Pyrenean, controlled by reactivation of deeper faults (Oudmayer & De Jager, 1993). The Savian inversion, which is very pronounced in the West Netherlands Basin (De Jager, 2007), barely affected the Broad Fourteens Basin (**Figure 3.35b**). Uplift attributable to the Pyrenean and Savian phases is of the order of 200 to 300 m (Van Wijhe, 1987a). In the West Netherlands Basin, and to a lesser extent the Roer Valley Graben and the graben shoulder of the Campine Basin where Zechstein salt is absent (**Figures 3.36d** & **e**), Sub-Hercynian and Laramide inversion resulted in reverse reactivation of pre-existing faults forming prominent ridges of flower structures on dominantly west-north-west and north-north-west trends. The resulting fault pattern is complex and anastomosing, and generally appears to support the regional model of dextral displacement compatible with both east–west late Cimmerian





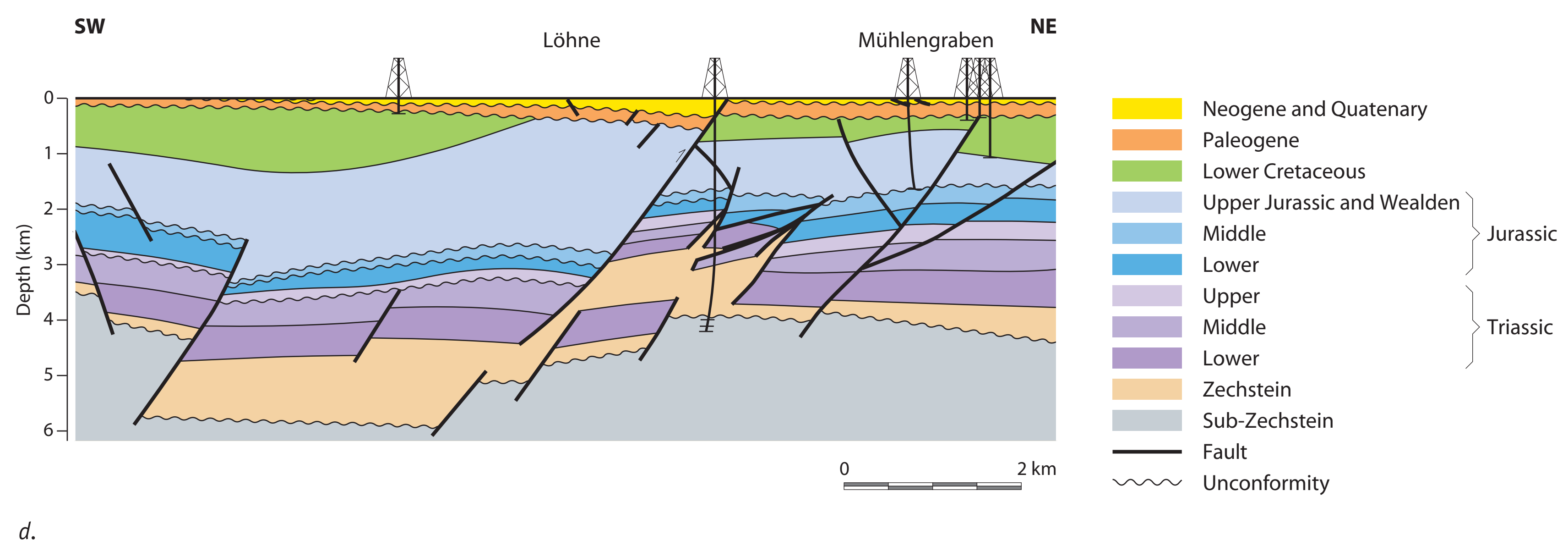


Figure 3.24 Seismic lines showing typical Pyrenean and Savian inversion structures in the UK and Netherlands at the the Sole Pit margin, Broad Fourteens Basin and Isle of Wight, and young normal faulting in Germany:
a. Seismic section through UKCS Quadrant 53 showing the main structural features of the South Hewett Shelf, South Hewett Fault and southern limit of the Sole Pit inversion (after Badley et al., 1989);
b. Section in Figure 3.24a showing the timing of inversion events (modified after Badley et al., 1989). Two generations of structures result from compression in the Sub-Hercynian and Savian phases;
c. Seismic section across the Sandown structure, Isle of Wight, UK (Chadwick & Evans, 2005). Migrated stack with interpretation. Note the dramatic thickening of the Lower Cretaceous syn-rift sequence to south of the Sandown Fault, and reverse displacement of the top of the Chalk Group pick beneath the Sandown (Isle of Wight) Monocline;
d. Cross-section of the Lohne structure. The reverse fault active during Coniacian-Campanian times was reactivated as an extensional fault during the Miocene (F. Kockel, pers.comm. 2008).
See Figure 3.26 for locations.

extension and north–south Alpine compression (Rossa, 1986; De Jager, 2007). In contrast, the Pyrenean inversion caused broad basin uplift without fault reactivation. Relaxation of the stress regime resulted in localised normal reactivation of faults at the end of the Pyrenean inversion movement. Most of the inverted parts of the West and Central Netherlands basins were subjected to erosion during the Oligocene.

The latter is a deeply eroded and broken-up basin (**Figure 3.36e**) in which Cenozoic deposits rest on Permian to Triassic sediments; Cretaceous sediments have been removed completely. Lower Jurassic sediments are locally preserved in lows. Relatively thin Zechstein salt forms minor salt pillows and local detachment zones.

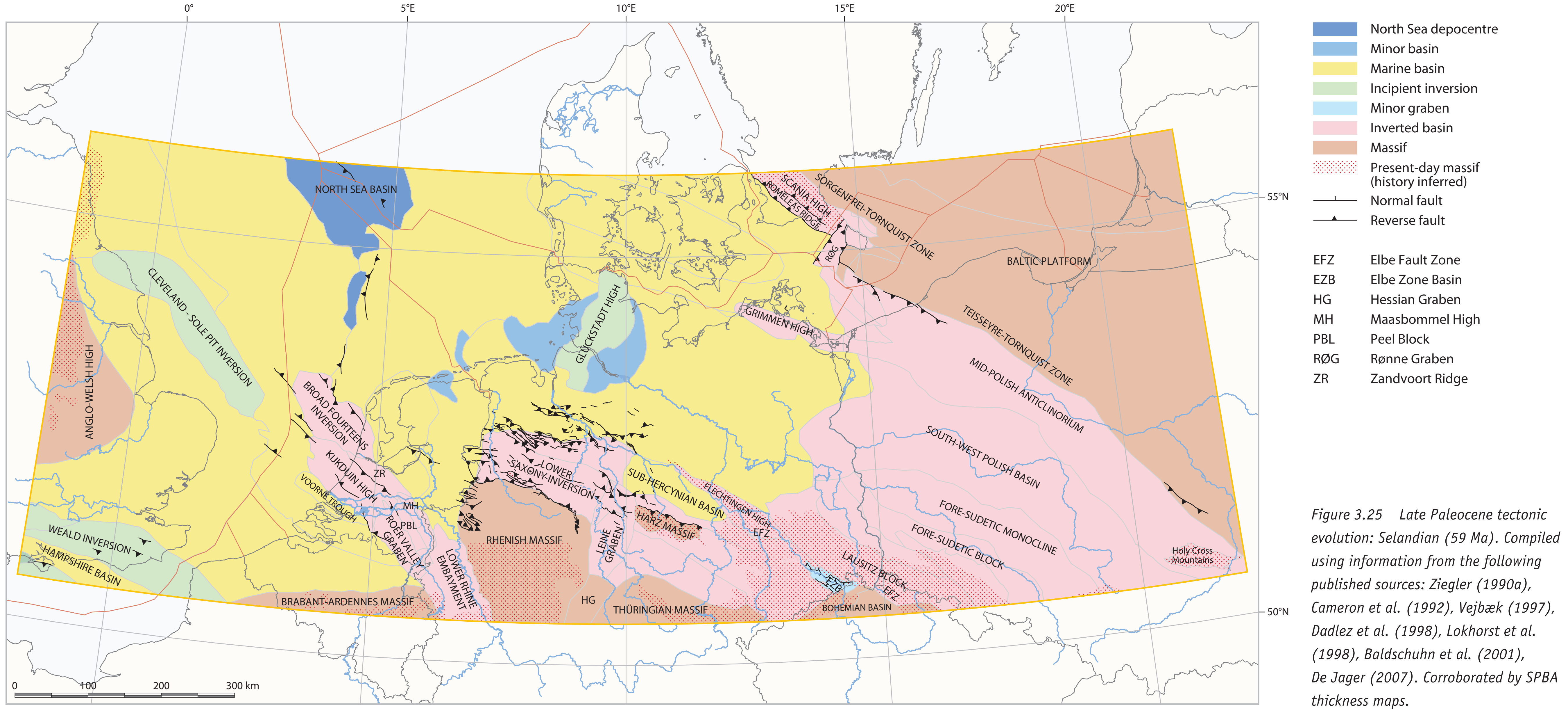


Figure 3.25 Late Paleocene tectonic evolution: Selandian (59 Ma). Compiled using information from the following published sources: Ziegler (1990a), Cameron et al. (1992), Vejbæk (1997), Dadlez et al. (1998), Lokhorst et al. (1998), Baldschuhn et al. (2001), De Jager (2007). Corroborated by SPBA thickness maps.

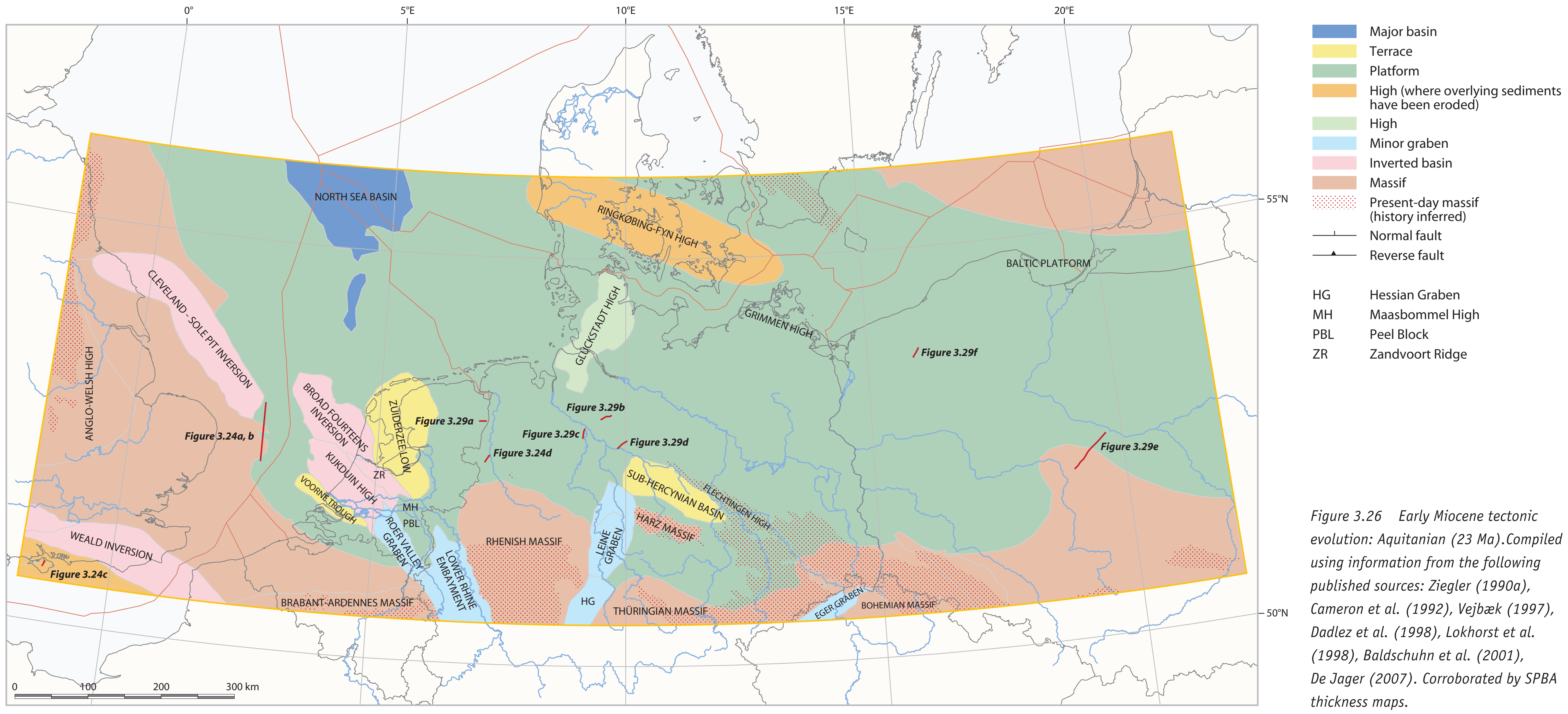


Figure 3.26 Early Miocene tectonic evolution: Aquitanian (23 Ma).Compiled using information from the following published sources: Ziegler (1990a), Cameron et al. (1992), Vejbæk (1997), Dadlez et al. (1998), Lokhorst et al. (1998), Baldschuhn et al. (2001), De Jager (2007). Corroborated by SPBA thickness maps.

As a consequence of the strong inversion of the Mesozoic West and Central Netherlands basins, their Cenozoic cover is at most only 500 m thick compared to 1000 m in the northern part of the inverted Broad Fourteens Basin. During the Laramide inversion, the West and Central Netherlands basins were uplifted and deeply eroded, locally down to the Triassic. Broad uplift of the West Netherlands Basin during the Pyrenean pulse focussed on the Mid-Netherlands Fault Zone (**Figure 3.35**), forming the Kijkduin High (**Figure 3.25**) in Late Eocene times. The Voorne Trough is an asymmetric basin to the south-west of this high, which was filled with Paleocene and Eocene deposits (NITG-TNO, 2004). In the Lauwerszee Trough, fault maps at the Cretaceous and Cenozoic levels indicate that both north-north-west and east-north-east-trending faults were reactivated during the Cenozoic (De Jager, 2007). Paleogene strata in this basin are up to 1000 m thick; the whole Cenozoic sequence is more than 1750 m (NITG-TNO, 2004). Subsidence along the eastern and western (Hantum Fault Zone) boundary faults of the trough was accommodated by Zechstein salt flowing towards the basin margin (NITG-TNO, 2004). Reverse faults and low-angle thrusts have resulted in much more shortening than in the Roer Valley Graben, where Cenozoic deposits are up to 2000 m thick (Geluk et al., 1994; Van den Berg et al., 1994). Significant post-Oligocene subsidence of this graben was associated with propagation of the Lower Rhine rift system into the Netherlands.

The Sole Pit Basin, which was hardly affected by the Laramide inversion phase, was strongly inverted during the Savian Phase, involving dextral strike-slip on the Dowsing-South Hewett Fault Zone (**Figures 3.34a** to **c**) (Glennie & Boegner, 1981; Van Hoorn, 1987). This was followed by erosion of 200 to 400 m of Paleogene

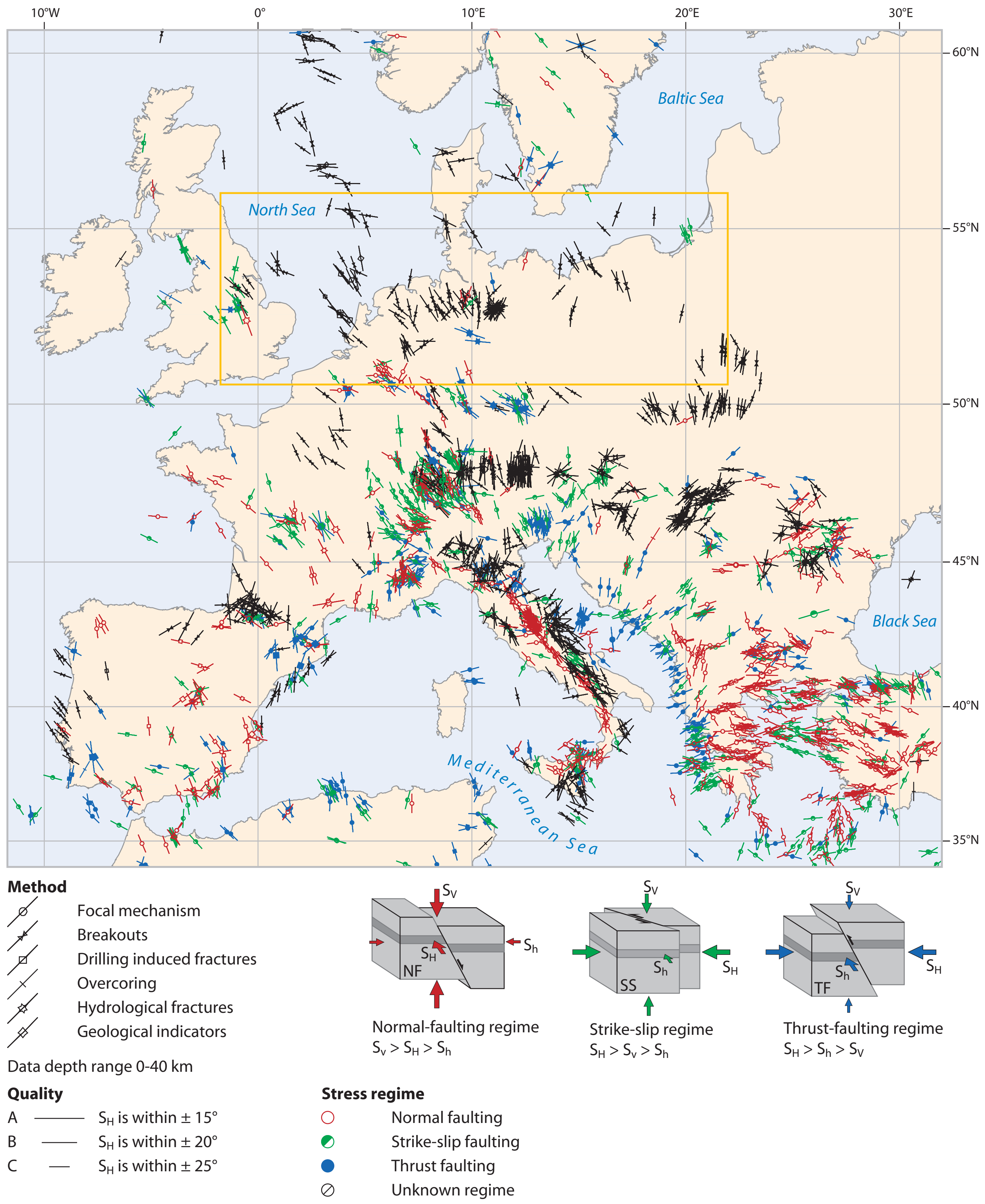


Figure 3.27 Present-day stress field for the SPB area, displaying the present-day orientation of maximum horizontal stress (S_{hmax}). Symbols stand for the stress indicators, and the length of the lines represents the data quality with A being the highest quality. Extracted from the World Stress Map Database (Reinecker et al., 2005). Available online at www.world-stress-map.org.

strata (Cameron et al., 1992). The same phase affected the conjoined Cleveland High, where according to Ziegler (1990a) this phase had a far greater effect than the preceding Laramide pulse. Estimates for the total uplift of the basal Permian strata achieved by the two phases of inversion range from between 1200 and 1800 m (Marie, 1975) to as much as 2500 m (Bulat & Stoker, 1987). The structure of the Silver Pit Basin is dominated by numerous north-west-trending salt pillows and walls. Widespread diapirism of Zechstein salts was triggered by the Pyrenean reactivation of basement faults under a dextral transpressional regime. North–south to north-east–south-west directed compression during Laramide and/or Pyrenean inversion has been deduced from seismic data in the southern North Sea (Nalpas et al., 1995), the Weald Basin (Whittaker, 1985; Hansen et al., 2002), and also from palaeo-stress studies in the Chalk Group of south-east England and north-east France (Vandycke & Bergerat, 2001). The structural effects of the Savian Phase were significant in the Weald Basin, for example, in the Isle of Wight (**Figure 3.24f**) and Hog's Back monoclines.

The different inversion events probably caused significant remigration of gas in the Rotliegend reservoirs of the southern North Sea. The southern bounding fault of the North Dogger Shelf (**Figure 3.19**) shows evidence for minor Mid-Cenozoic reverse reactivation (Jenyon, 1985; Cameron et al., 1992). By analogy with the Mesozoic development of the Vale of Pickering-Flamborough Head Fault Zone, this structure may have developed by sinistral transcurrent Mesozoic movement on a deep-basement lineament, followed by dextral compression during the Cenozoic.

9.1.2 North German Basin

Following the Sub-Hercynian and Laramide pulses of basin inversion, all faults active during the Cenozoic have normal throws in the North German Basin. Many of these faults were already active in Triassic times and were reactivated during the Cenozoic. Many of the faults bounding the Jurassic to Early Cretaceous transtensional basins, which during their inversion were transformed into thrusts, reverse faults or under-compensated reverse faults, were also reactivated as normal faults during Cenozoic times (**Figure 3.24d**). Diapirism is geographically limited, for example some salt domes to the west and east of the Glückstadt Graben entered their diapiric stage. The North German Basin gently subsided during the Cenozoic and was tilted westwards in conjunction with thermal subsidence of the North Sea Basin. In northern Germany, local depocentres developed along the flanks of actively growing major salt walls and diapirs. Subsidence of the North German Basin and the North Sea Basin accelerated during Plio-Pleistocene times (Scheck-Wenderoth & Lamarche, 2005; Ziegler & Dèzes, 2007).

9.1.3 Polish Basin and Fennoscandian Border Zone

During the Cenozoic, the south-eastern part of the inverted Mid-Polish Basin was overridden by the Outer Carpathian orogenic wedge (Oszczypko, 2004; Oszczypko et al., 2005). This was accompanied by the development of the flexural Carpathian foredeep basin in response to thrust- and slab-loaded deflection of the foreland lithosphere (Krzywiec, 2001; Oszczypko, 2006). The Mid-Polish inversion anticlinorium (**Figures 3.42a** & **b**) extends north-westwards into the Baltic Sea (Pożaryski & Brochowicz-Lewinski, 1979; Bergström et al., 1990) where it links up via the inverted Rønne Graben with the Laramide structures of Scania and the Norwegian-Danish Basin (Liboriussen et al., 1987; Norling & Bergström, 1987; Thomsen et al., 1987; Petersen et al., 2003a). The Bornholm Block and Christiansø Horst (see **Figure 3.38c**) were also uplifted at this time (Kumpas, 1979; BABEL Working Group, 1993). The Romeleås Ridge was upthrusted by at least 2500 m on a steep reverse fault (Norling & Bergström, 1987) under a dextral wrench regime, whereas Upper Cretaceous strata in the adjacent basin are 2000 m thick (Ziegler, 1990a).

In the Polish Lowlands, as seen in northern Germany, the Cenozoic sequence is thin (typically about 250 m) and shows no sign of differential subsidence (**Figure 3.42b**), even in the vicinity of the former Mid-Polish Trough. The remnant topographic relief of the Mid-Polish Anticlinorium was reduced by erosion during the Eocene and was completely overstepped by mid-Oligocene times (Ziegler, 1990a), with no evidence for Pyrenean or Savian inversion (Jarosiński et al., 2009).

9.2 Cenozoic rift systems

The SPB area was only marginally and indirectly affected by the development of the European Cenozoic rift system (ECRIS), which extends about 1100 km from the Dutch coast to the western Mediterranean.

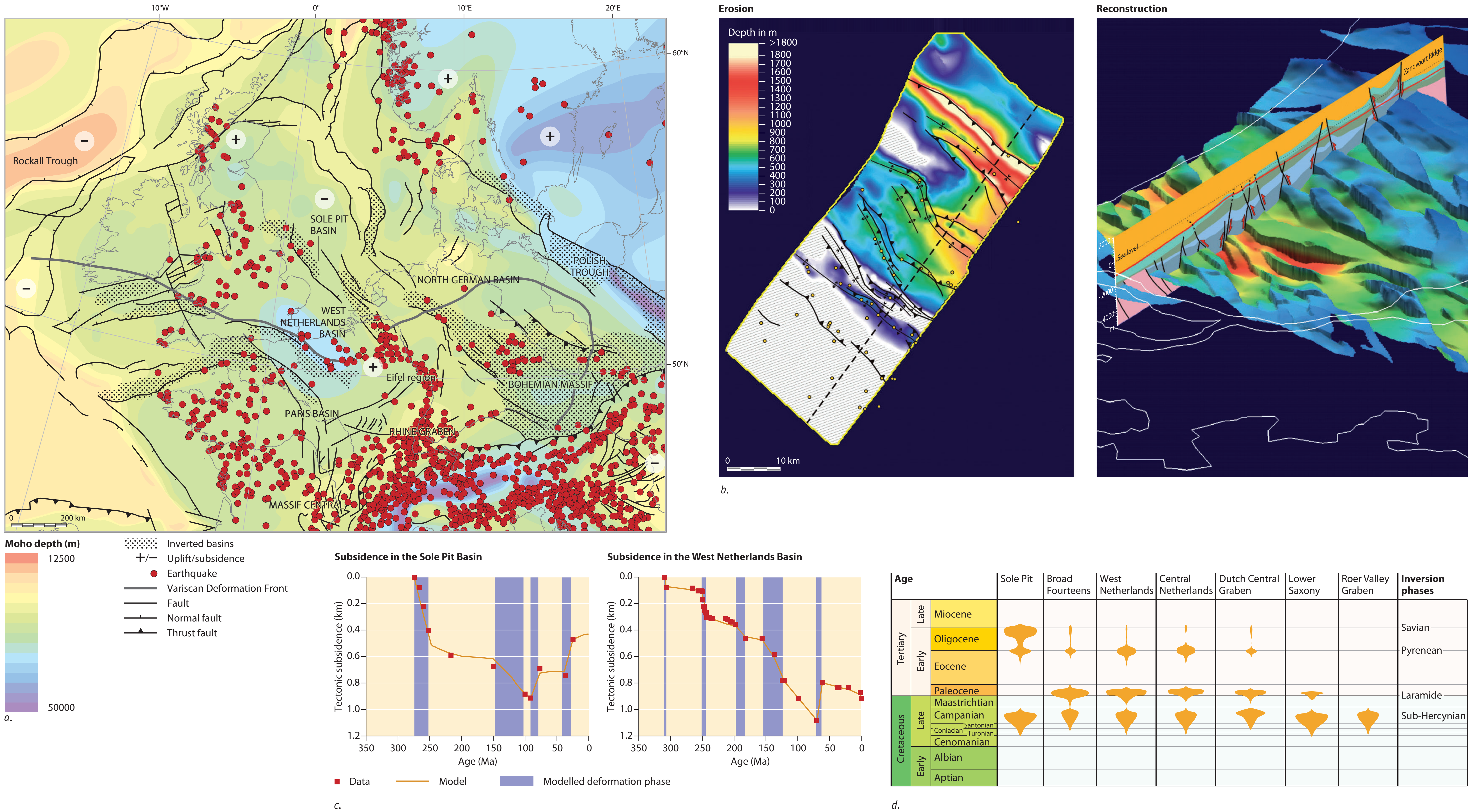


Figure 3.28 Mesozoic-Cenozoic subsidence and inversion:
a. Seismicity superimposed on map showing the depth to the Moho in north-west Europe (Cloetingh, 1986). Neogene uplift and subsidence anomalies are indicated by circled plus and minus symbols respectively (cf. Japsen & Chalmers, 2000). Locations of inverted basin structures are also shown (modified after Brun & Nalpas, 1996). Note the concentrations of intraplate seismicity in areas of crustal thickness, the contrast between the basinal and Paleozoic massif areas, as well as in areas of crustal contrast along the NE Atlantic rifted margins. A notable exception is the seismicity along the neotectonically active Rhine Graben. The West Netherlands Basin is modified from Cloetingh & Van Wees (2005);
b. West Netherlands Basin inversion, left: estimated erosion for the Base Cenozoic Unconformity derived from over-compaction and structural analysis, right: reconstructed geological cross-section at Late Cretaceous times, eroded down to the Base Cenozoic Uniformity. The colour-coded surface is the base of the Jurassic derived from 3D-seismic interpretation in agreement with the reconstructed geological cross-section (modified after Worum, 2004);
c. Subsidence curves and modelled subsidence for the Sole Pit Basin (modified after Van Wees & Beekman, 2000) and the West Netherlands Basin (modified after VanWees et al., 2009). See Figure 3.28a for locations. The West Netherlands Basin is mainly marked by Late Cretaceous to Early Cenozoic inversion (Sub-Hercynian to Laramide phase), constrained by the erosion estimates in Figure 3.28b;
d. SPBA inversion history. Modified after De Jager (2007).

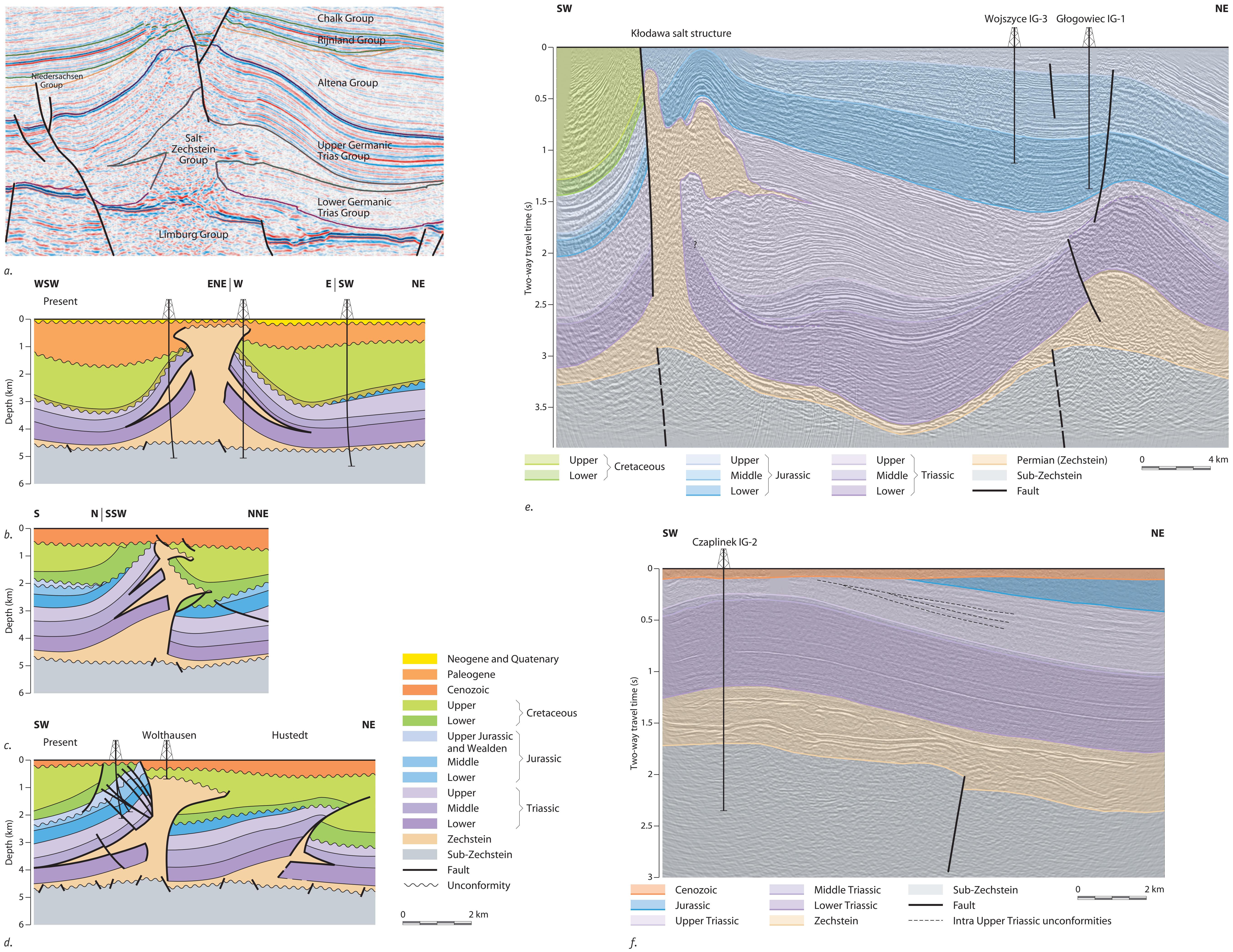


Figure 3.29 Seismic sections illustrating the style of salt tectonics:
a. Seismic section across salt intrusion kbVI. Zechstein salt with wedges into the Triassic;
b. Cross-section through the Söhlingen salt structure. Wedges of Zechstein salt intruded into the Upper Buntsandstein Röt salt level during the Sub-Hercynian (Coniacian-Campanian) inversion phase. The wedges are tilted by post-Cretaceous rise of the diapir (Baldschuhn et al., 1998);
c. Cross-section across the Eitzendorf salt structure straddling the inverted Aller Lineament (eastern Lower Saxony Basin). The diapiric phase began in Early Cretaceous times, with a secondary rim-syncline in the SW. The Zechstein salt wedges intruded into the Upper Buntsandstein Röt and middle Muschelkalk salt levels during the Sub-Hercynian inversion phase at the same time as the overhang formed in the NE (Baldschuhn et al., 1998);
d. Cross-section across the Wolthausen salt structure (eastern Lower Saxony Basin) straddling the Aller Lineament (Kockel, 2003). The diapiric phase was reached in Late Jurassic-Early Cretaceous times with a secondary rim-syncline forming SW of the structure). Salt was expressed from the diapir during inversion to form a salt wedge into the Röt Salt level and a large overhang, which may have been extruded onto the Santonian sea floor. Further rise of the structure took place during Campanian and Maastrichtian times (Baldschuhn et al., 1998);
e. Seismic section from the central (Kuiavian) segment of the inverted Mid-Polish Trough (cf. Krzywiec, 2004a). The Kłodawa salt structure has a very large salt overhang located within the Upper Triassic (Keuper) sedimentary cover. The internal architecture of the entire Triassic sedimentary cover reflects the influence of the sub-salt basement faulting and salt movements on a gross depositional pattern – changing from salt-pillow growth (thinning towards the salt structure; Early to Mid-Triassic), salt diapirism (thickening towards the salt structure; Late Triassic) that followed extrusion of the salt overhang onto the basin floor, to post-diapiric stage (latest Triassic) (cf. Krzywiec, 2004b). The salt overhang was folded together with the Triassic-Jurassic (presently partly eroded) sedimentary cover most probably during the Late Cretaceous-Paleogene inversion of the Polish Basin, including the Mid-Polish Trough;
f. Seismic section from the NE flank of the Pomeranian segment of the inverted Mid-Polish Trough, showing intra-Upper Triassic unconformities and thickness changes (based on Krzywiec, 2006b; cf. Krzywiec, 2006a; Krzywiec et al., 2006). Such depositional architecture reflects Late Triassic salt movements along the NE flank of the Mid-Polish Trough. These are age-equivalent to the salt movements documented for example in the vicinity of the Kłodawa salt structure.
See Figure 3.26 for locations.

The ECRIS was activated during the Mid- to Late Eocene and propagated northwards during the Oligocene from the Upper Rhine Graben into the Lower Rhine-Roer Valley (**Figure 3.36e**) and Hessian grabens (Ziegler, 1990a, 1994; Geluk et al., 1994; Dèzes et al., 2004). Evolution of the ECRIS was punctuated by mantle-plume related volcanic activity, which intensified during the Oligocene and Miocene and contributed, by thermal thinning of the lithosphere, to the uplift of the Ardenno-Rhenish Massif and partly the Bohemian Massif (Ziegler & Dèzes, 2006, 2007). The Peel Boundary Fault (**Figure 3.36e**) has a post-inversion throw of up to 1000 m (De Jager, 2007) and separates the Roer Valley Graben from the Peel Block to the north-east. On the south-western graben margin, bounding faults step-up more gradually to the London-Brabant Massif (Geluk et al., 1994; NITG-TNO, 2004). Recent earthquakes along the bounding faults of this system (Roermond event) indicate that tectonic activity still occurs (Camelbeeck et al., 2007; Van Balen & Houtgast, 2007). Evidence for recent faulting is very rare elsewhere, except in the coastal area of the Netherlands where faults dissect early Pleistocene deposits. Minor earthquakes have been recorded in northern Germany with hypocentres at known basement faults at depths of about 8 to 17 km (Leydecker, 1986; Kaiser et al., 2005). A system of narrow and relatively shallow grabens developed on the Fore-Sudetic Monocline along the south-eastern SPB margin during the latest Eocene and Oligocene. This graben system propagated northwards into the Polish Lowlands at the end of the Oligocene and remained active until Mid-Miocene times (Reicherter et al., 2008; Jarosiński et al., 2009).

The Bohemian Massif was peneplaned following its strong inversion during the Sub-Hercynian and Laramide phases (**Figure 3.25**). Widespread Oligocene volcanic activity preceded the subsidence of the shallow Eger Graben, which ended with the mid-Burdigalian upwarping of the Bohemian Massif. The fault systems of the Bohemian Massif started to reactivate during the Late Miocene (**Figure 3.26**) in response to the build-up of the present-day compressional stress field. This resulted in accentuation of its marginal highs, such as the Thüringian-Bohemian and Bavarian Forest and the Erzgebirge, Lausitz, Sudetic and Moravo-Silesian blocks, and a reorganisation of its drainage system (Malkovsky, 1980, 1987; Suk et al., 1984; Ziegler & Dèzes, 2006). Remnants of Lower Oligocene marine sediments on the central Harz Mountains (König & Blumenstengel, 2005, quoted by Reicherter et al., 2008) testify to their uplift by some 300 m during Late Miocene and Pliocene re-structuring of the Bohemian Massif.

9.3 Formation of the present-day North Sea Basin (Miocene-Pleistocene)

The North Sea Basin has undergone regional thermal subsidence since Aptian to Albian times, and the Mid North Sea and Ringkøbing-Fyn highs were overstepped (Ziegler, 1990a; Wong et al, 2007a; Chapters 11 & 12). Regional post-rift thermal subsidence of the North Sea Basin continued during the Paleogene and Neogene (Kooi et al, 1989; Ziegler, 1990a). Progressive uplift of the Ardennes and the Rhenish and Bohemian massifs resulted in increased clastic influx into the North Sea Basin during the Miocene. Major delta systems have prograded westwards into the deeper-water North Sea Basin since Mid-Miocene times (Cameron et al., 1993). The most important delta, the Eridanos delta (Overeeem et al, 2001), developed due to gradual uplift of the Fennoscandian Shield. Shallow-water conditions were established throughout the southern North Sea area by mid-Pleistocene times (~1.7 Ma) (Gibbard & Lewin, 2003; Wong et al., 2007b). Quaternary deposits are up to 1000 m thick in the northern part of the Dutch offshore sector. Burial curves show a sharp increase in the rate of subsidence during the last few million years (De Jager, 2007), considered to be the flexural response of the lithosphere to the build-up of the present-day, north-north-west-directed, compressional stress field (Kooi et al., 1989; Van Wees & Cloetingh, 1996). Subsidence in the south-eastern North Sea was rather rapid, especially during the early Pleistocene, contemporaneous with rapid uplift of the Variscan massifs on the southern flank of the North German Basin, such as the Harz, Rhenish Massif and the Ardennes. Similarly, the connection between the southern North Sea via the Channel to the Atlantic shelves of France and Ireland was opened intermittently by flooding or breaching of the Melantois-Artois High during the Cenozoic (Van Vliet-Lanoë et al, 1998; Gibbs & Lewin, 2003; Ziegler & Dèzes, 2007).

9.4 Present-day stress field

The present-day maximum horizontal stress pattern is based largely on studies of earthquake focal mechanisms, analyses of borehole break-outs and hydraulic fracturing experiments at levels below the Zechstein salt (**Figure 3.27**). The stress pattern shows a predominant north-north-west-oriented direction for the maximum horizontal compressional stress trajectories, reflecting a combination of North Atlantic ridge push and collisional coupling of the foreland lithosphere with the Alpine-Carpathian Orogen (Gölke & Coblentz, 1996; Gölke et al., 1996). At about the longitude of Hamburg and Hannover, the trajectories of SH_{max} stress axes are generally oriented north–south, whereas a general north-easterly trend prevails east of the River Elbe; in the North Sea area, stress trajectories are predominantly to the north-west (Marotta et al., 2002; Reinecker et al., 2005; Jarosiński, 2006). No correlation between the present-day stress pattern and the direction of the major basement faults has been recognised (Kaiser et al., 2005). Major deviations of the stress patterns occur along strong contrasts in the lithospheric structure and so mimic the inherited structure from underlying terranes and orogens (Krawczyk et al., 2008a). At supra-salt levels, the SH_{max} stress trajectories strongly deviate from those at sub-salt levels, indicating strong decoupling by the Zechstein salt (Lempp & Lerche, 2006) and both regional and local influences (Kley et al., 2008).

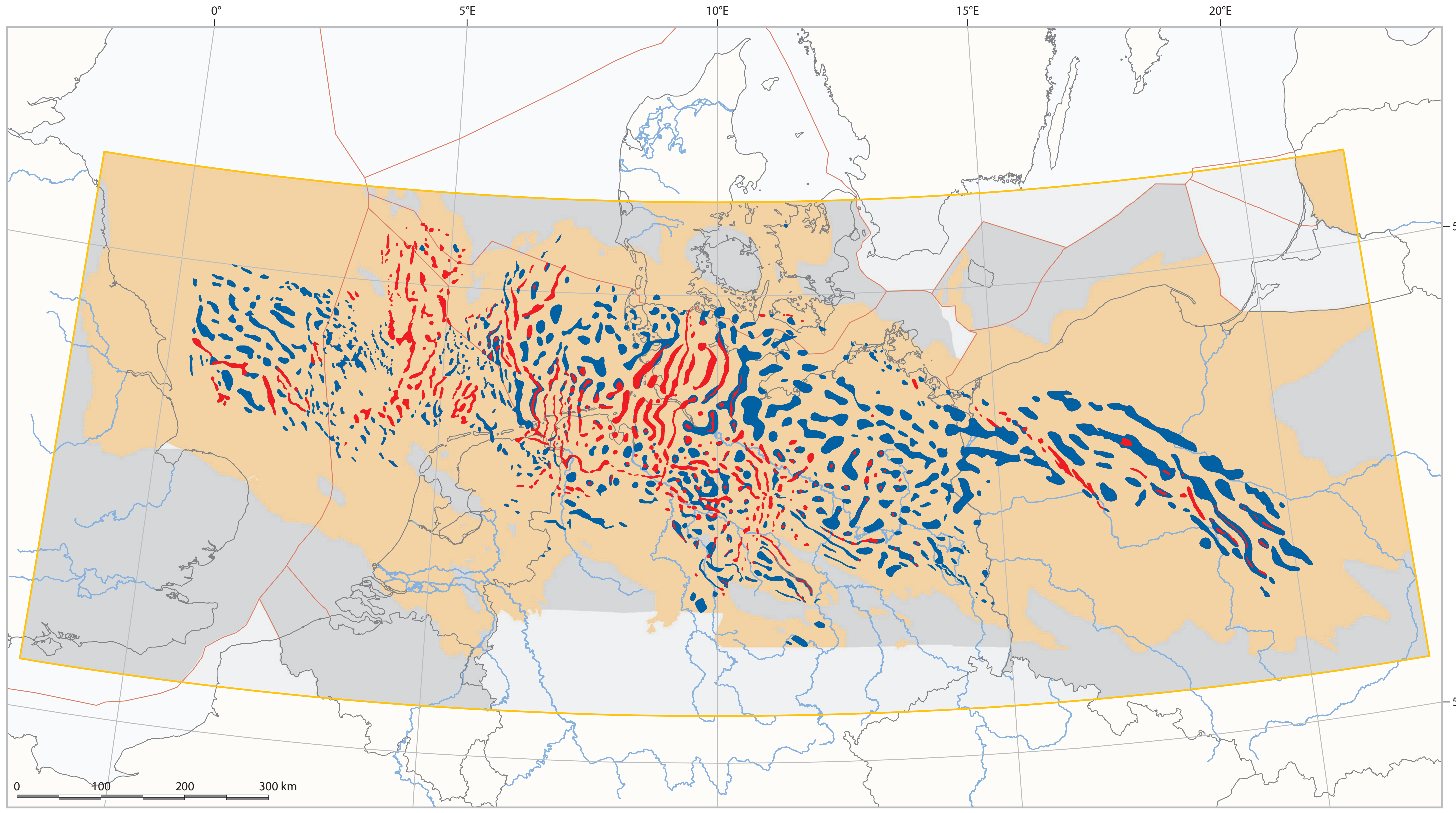


Figure 3.30 Salt tectonics (Lokhorst et al., 1998, with contributions from Mazur et al., 2005; Y. Maystrenko, pers. comm. 2008).

The present-day stress field evolved from the Mid-Miocene stress field, but intensified during the Pliocene (Dèzes et al., 2004; Ziegler & Dèzes, 2007). This is compatible with the accelerated Pliocene subsidence of the North Sea Basin and the North German Basin (Van Wees & Cloetingh, 1996; Scheck-Wenderoth & Lamarche, 2005; Cloetingh et al., 2008).

10 Models of Mesozoic-Cenozoic basin evolution

10.1 Basin inversion and tectonic subsidence

Multiple phases of Late Cretaceous to Cenozoic basin inversion have occurred in various parts of the SPB (**Figure 3.28a**), generally subdivided into Sub-Hercynian (Senonian), Laramide (Paleocene), Pyrenean (Late Eocene to Early Oligocene) and Savian (Late Oligocene to Early Miocene) pulses (Ziegler, 1990a; Worum & Michon, 2005). Until recently, most authors (e.g. Ziegler et al., 1995; Marotta et al., 2001; Krzywiec, 2005) have interpreted the Sub-Hercynian event as a consequence of convergence or early collision of the Alpine-Carpathian Orogen with Europe's southern margin in a classic orogenic foreland collision model. Kley & Voigt (2008) present an alternative hypothesis, arguing that this interpretation is unlikely because the Late Cretaceous kinematic history of the two regions is incompatible, whereas the location of the developing Alpine chain on the Adria Plate in recent plate reconstructions (e.g. Stampfli & Borel, 2004) means that it lay far to the south of central Europe across an ocean basin. They therefore propose that the Sub-Hercynian event reflects the onset of Africa-Iberia-Europe convergence; Alpine collision with southern Europe did not commence until Paleocene or Eocene times. Kley & Voigt (2008) also suggest that the Late Cretaceous event was caused by the pinching of west central Europe's thin lithosphere between the East European Platform and Africa.

Inversion has compressionally/transpressionally reactivated Mesozoic transtensional basins, resulting locally in deep erosion of their sedimentary fill. Quantitative dating, structural style and magnitude of inversion have been established through reconstruction of erosion by over-compaction and maturation modelling (e.g. Hillis, 1995; Petersen et al., 2003a), geochronological techniques including apatite fission-track analysis (e.g. Brun & Nalpas, 1996), and geological interpretation (e.g. Worum & Michon, 2005; **Figure 3.28b**). Incorporating the estimated erosion into subsidence analysis and forward modelling of lithosphere deformation (**Figure 3.28c**) clearly demonstrates pervasive lithosphere deformation during basin inversion, accounting for crustal shortening in the order of 10-15% in strongly inverted areas. Quantitative basin analysis shows shifting patterns of basin inversion through time. East of the Netherlands, basin inversion largely ceased after the Late Cretaceous and earliest Cenozoic (Polish Trough, North German and Lower Saxony basins), whereas inversion continued in the west (Sole Pit, Broad Fourteens and West Netherlands basins) until the Late Oligocene to Early Miocene (**Figure 3.28d**) (Van Hoorn, 1987; Ziegler, 1990a, 1998; De Lugt et al., 2003; De Jager, 2007). This may relate to a delay in the mechanical coupling of the collisional system towards the west, i.e. between the Pyrenean Orogen and central Europe.

10.2 Thermo-mechanical controls in Late Cretaceous - Early Cenozoic basin inversion and Neogene lithosphere folding

The compressional intraplate stress field controlling basin inversion is generally assumed to originate from the interaction of the evolving Alpine Orogen with its foreland, which in turn is controlled by the convergence of Africa with Europe. Space was constrained in the Alpine Tethys domain during the Late Cretaceous and Paleocene due to convergence of Africa with Europe. Evidence for this is provided by the activation of subduction processes and the build-up of intraplate compressional stresses controlling the Sub-Hercynian and Laramide pulses of inversion tectonics (Ziegler et al., 1995, 1998, 2001; Stampfli & Borel, 2002; Dèzes et al., 2004; Kley & Voigt, 2008). Distinct pulses of inversion are well-correlated with Alpine compression phases (Ziegler, 1990a; Ziegler et al., 1998) clearly demonstrating the key role of far-field compressional intraplate stress. Most of the Mesozoic tensional basins and crustal-scale faults in the SPB area were inverted during the Late Cretaceous and Paleogene (Scheck-Wenderoth & Lamarche, 2005; **Figure 3.28a**). The evolution of these basins has been characterised by repeated reactivation of their fault fabric. Thermo-mechanical modelling of the lithospheric strength of the inverted basins in the SPB shows a rheologically-strong lithosphere for a long period after rifting (Van Wees & Beekman, 2000). This prediction does not agree with observations of long-term repeated basin reactivation during multiple phases of basin extension and inversion. However, crustal-scale weakness zones, such as faults, cause a permanent strong reduction of integrated strength values (Ziegler et al., 1995; Van Wees & Beekman, 2000), and so play an important role in basin reactivation at lithospheric scales. Such weak zones at shallow levels correspond to relatively weak basement faults, which have been intrinsically adopted in analogue and numerical models (e.g. Brun & Nalpas, 1996). The inversion of the North German Basin has been studied using numerical modelling by Marotta et al. (2002). The north-west–south-east-orientation of Alpine compressive stress was confirmed, and the basin, which is not in isostatic equilibrium, remains in a state of horizontal compression. A less deformable area in the northern part of the basin model corresponds to an area of low observed seismicity and fan-like SH_{max} stress pattern. Marotta et al. (2002) infer that this region in the transition to the Baltic Shield has a stronger lithosphere that acts as a barrier to compressional stress.

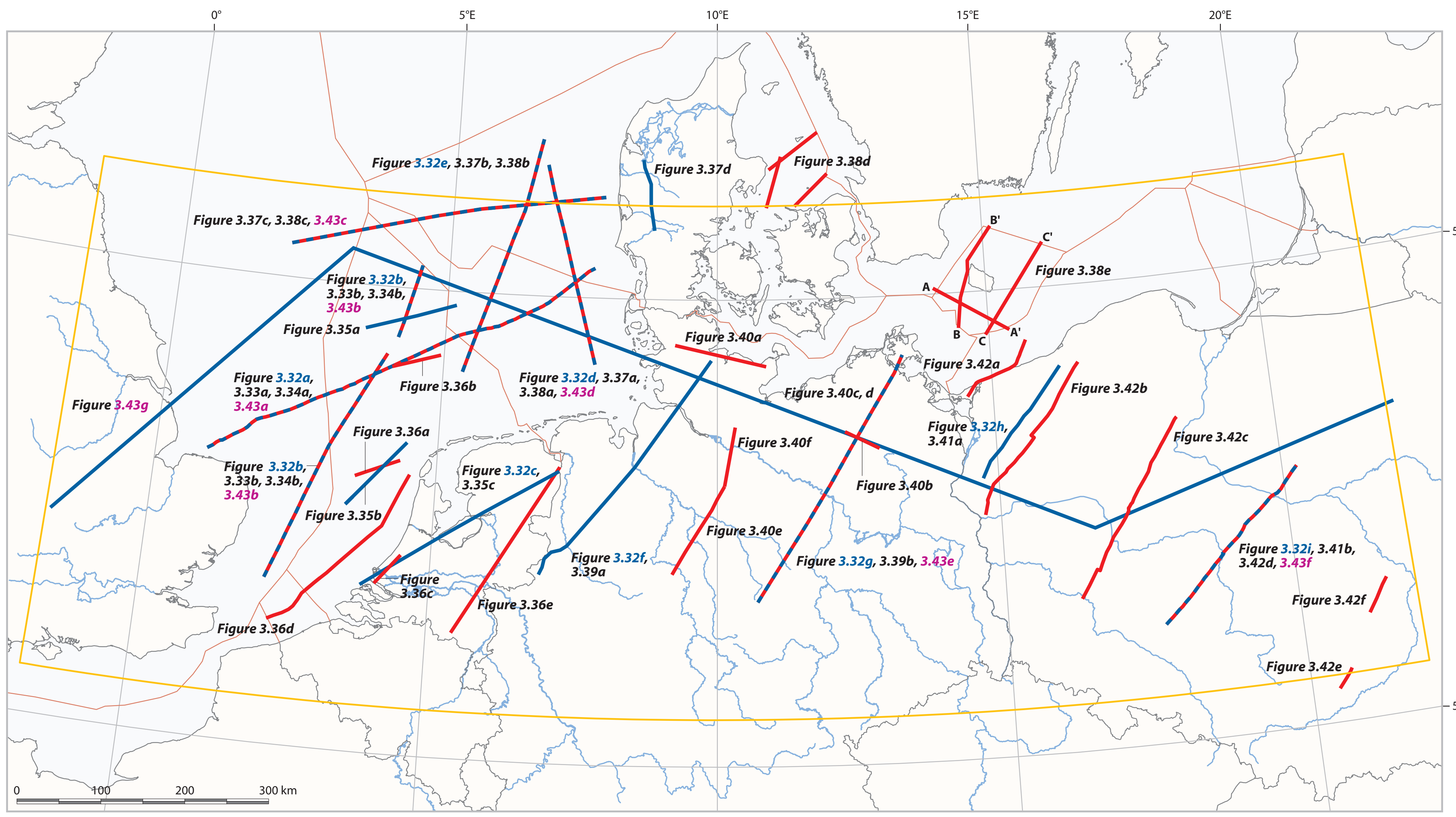


Figure 3.31 Locations of regional seismic sections shown in Figures 3.32 to 3.43 and referred to in the text.

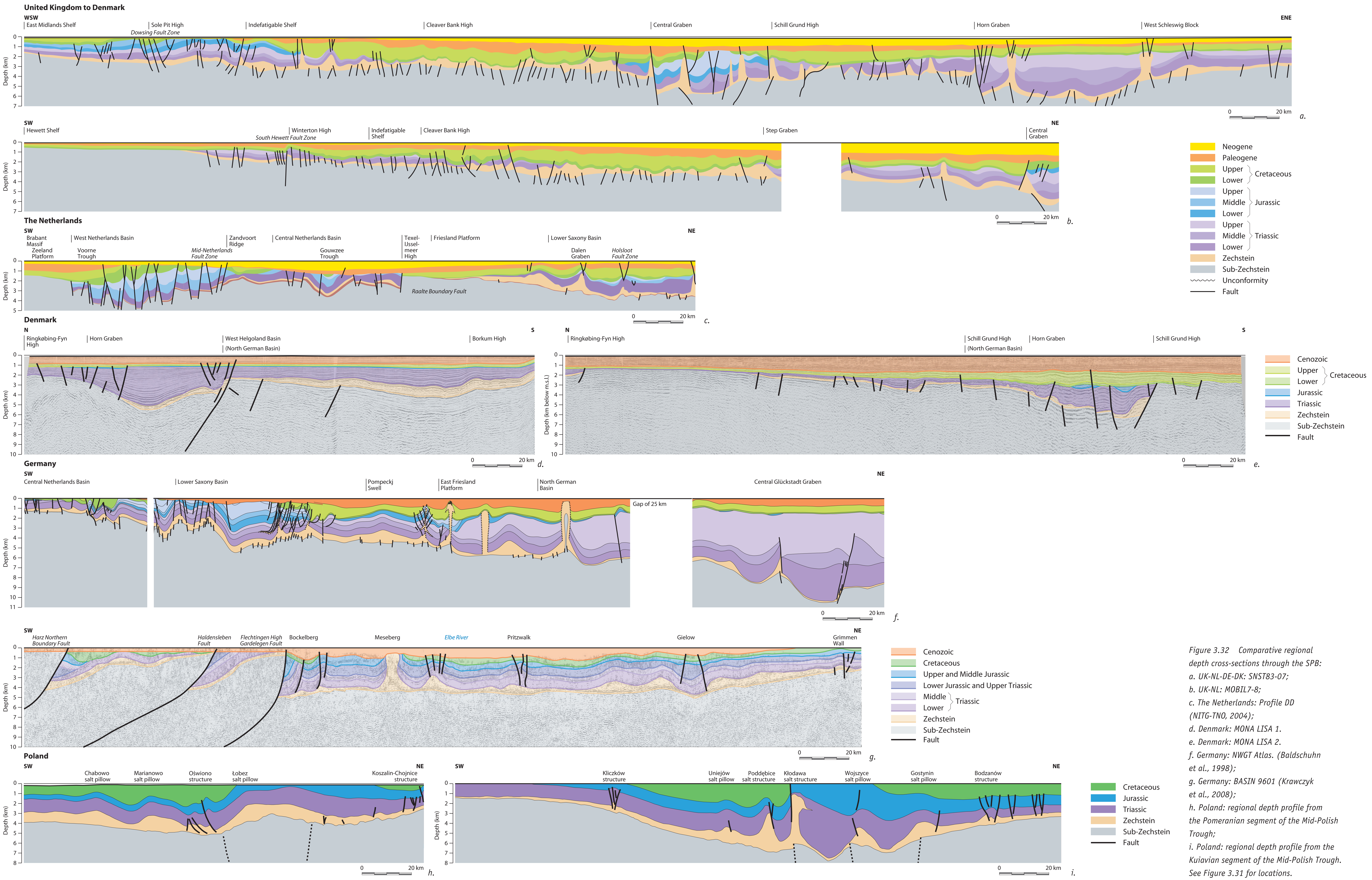


Figure 3.32 Comparative regional depth cross-sections through the SPB:
a. UK-NL-DE-DK: SNST83-07;
b. UK-NL: MOBIL7-8;
c. The Netherlands: Profile DD (NITG-TNO, 2004);
d. Denmark: MONA LISA 1.
e. Denmark: MONA LISA 2.
f. Germany: NWGT Atlas. (Baldschuhn et al., 1998);
g. Germany: BASIN 9601 (Krawczyk et al., 2008);
h. Poland: regional depth profile from the Pomeranian segment of the Mid-Polish Trough;
i. Poland: regional depth profile from the Kuiavian segment of the Mid-Polish Trough.
See Figure 3.31 for locations.

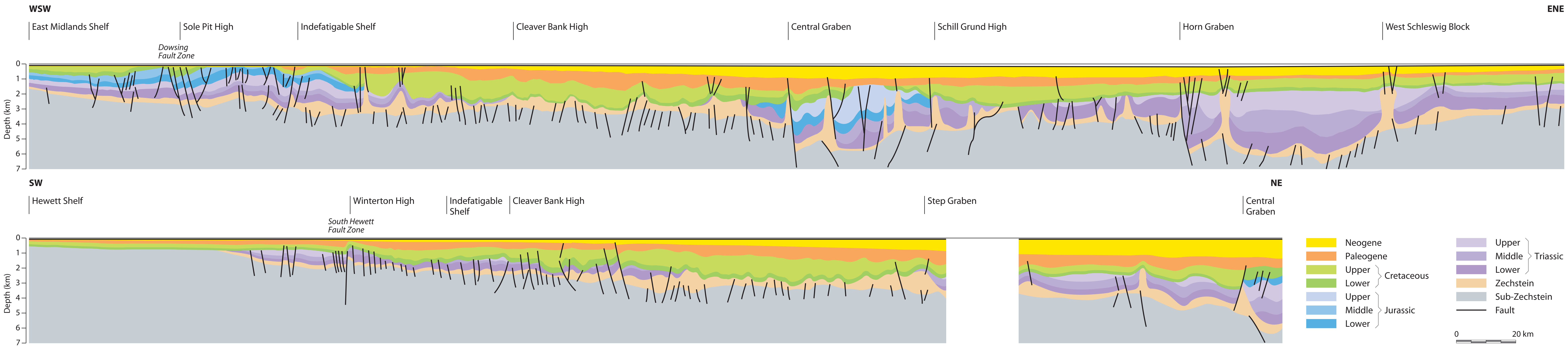


Figure 3.33 Regional depth cross-sections: UK-NL-DE-DK; SNST83-07, MOBIL7-8.

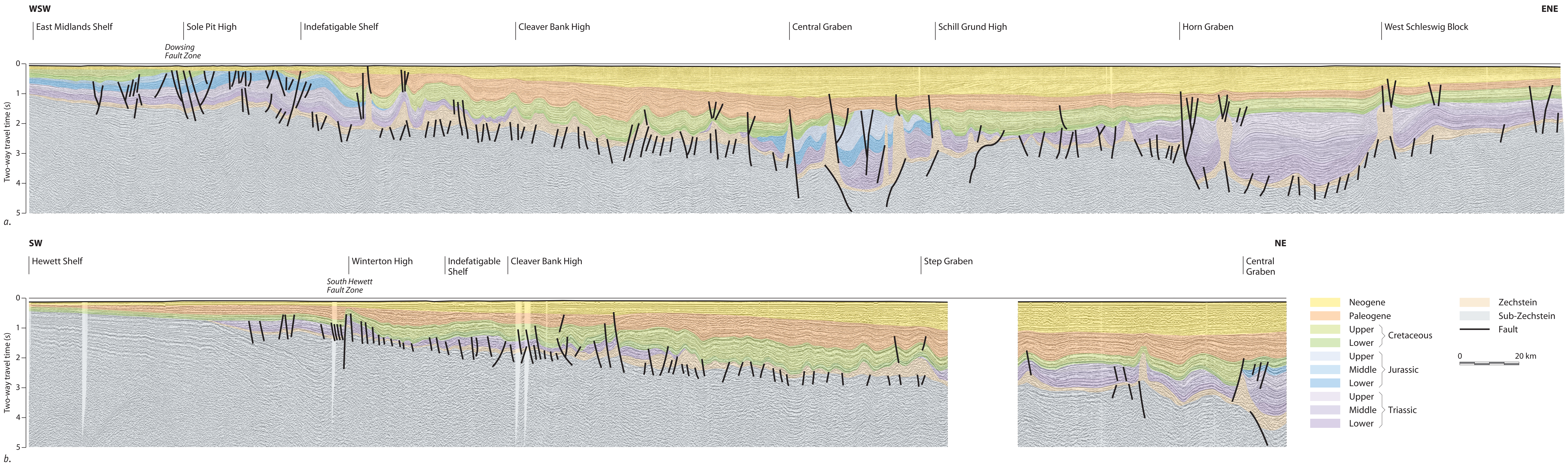


Figure 3.34 Regional seismic profiles in time: UK-NL-DE-DK; SNST83-07, MOBIL 7-8 plus detail panels (e.g. after Badley et al., 1989):

a. PCS-NOPEC SNST 83-07. Southern North Sea Tie (SNST) Line 83-07 was acquired by NOPEC A/S as part of a regional speculative survey and included within the SPBA dataset by kind permission of PCS NOPEC A/S and GECO. An interpretation of a large segment of this profile was published by Van Hoorn (1987). The profile runs NE from The Wash, across the East Midlands (Amethyst) Shelf and Sole Pit Inversion, onto the Indefatigable Shelf and Cleaver Bank High. The west end of the profile crosses the East Midlands Shelf, the Mesozoic strata of which are little affected by faulting and thin towards The Wash. The profile crosses the Dowsing Fault Zone at the western margin of the Sole Pit Basin, and the limit of inversion. On this section, the near-surface displacement is normal, with relatively thick Lower Cretaceous strata downthrown against Jurassic in the footwall. Evidence for inversion is therefore not as good as that seen on MOBIL 7. The profile cuts across the middle of the Sole Pit inversion axis. Van Hoorn (1987) published an interpretation of part of the line to demonstrate the complex evolution of the Sole Pit Basin. Across the inversion, the Rotliegend reflector rises from 2.0 s TWT to 1.25 s before dipping to 1.6 s TWT just NE of the Dowsing Fault Zone. A thick sequence of Triassic to Upper Jurassic strata was deposited in the Sole Pit Basin and truncated by an early phase of uplift in Late Jurassic-Early Cretaceous (late Cimmerian) times. The Upper Cretaceous sequence is thin and attenuated across the Sole Pit High, but Paleogene strata were deposited with apparent conformity on latest Cretaceous sediments (Van Hoorn, 1987). The final inversion of the basin took place during the Late Oligocene, when the area was affected by the Alpine orogenic movements. Van Hoorn (1987) estimates that up to 1500 m of uplift was achieved by these two phases of inversion combined. Pre-Pliocene erosion resulted in the truncation of the uplifted Lower Cenozoic strata along the eastern flank of the Sole Pit High in the vicinity of the Swarte Bank Hinge Zone. The western margin of the Cleaver Bank High is complex, has preserved Jurassic strata, and exhibits evidence for multiple phases of halokinesis. The Rotliegend reflector lies at a relatively uniform depth of about 2.5 s TWT across the Cleaver Bank High. Most of the Jurassic and Triassic section was removed prior to deposition of Cretaceous and Paleogene strata, which make up most of the section. The main growth of salt pillows followed this phase of burial and affected strata as young as the Paleogene. The transect crosses the Dutch Central Graben, where the Jurassic sequence is significantly thicker. Strata younger than Triassic are absent in the Horn Graben. The profile ends on the West Schleswig Block in Danish waters;

b. BIRPS MOBIL 7/8. This profile is featured in the BIRPS Atlas (Klemperer & Hobbs, 1991). It extends SSW across the Cleaver Bank High at the western margin of the Broad Fourteens Basin in the Dutch sector, then across the adjacent Winterton Sub-basin and Winterton Arch of the UK sector. The western margin of the Broad Fourteens Basin is interpreted as an overthrust (Van Wijhe, 1987a) produced by Laramide (Late Cretaceous – Early Cenozoic) inversion. The effect of the overthrust is less obvious on this profile, largely as a consequence of extensive halokinesis. Permian-Jurassic strata were eroded across the Winterton Arch prior to deposition of Lower Cretaceous strata. Permo-Triassic strata are downthrown to north, although the displacement occurs over a broader zone and involves much synthetic domino-style faulting. There is good evidence for syndepositional displacement in the Permo-Triassic sequence. Cretaceous and Cenozoic strata are buckled into a monocline indicating reactivation of a controlling basement fault with southward downthrow during the Alpine inversion events. This may reflect the fact that, south of the Sole Pit, most of the effect of the inversion is taken up within the South Hewett Fault Zone alone. Klemperer & Hobbs (1991) have suggested that the fault zone may have undergone pre-Permian displacement. On the South Hewett Shelf, Cretaceous strata rest unconformably upon the Upper Triassic. Beneath the Permian sequence, Carboniferous strata exhibit good reflectivity, thinning from about 1.5 s TWT to the basin margin. The extension of the profile (MOBIL 8) extends from the shoulder of the Central Graben towards the NE. See Figure 3.31 for locations.

In the North Sea area, the mode of intraplate compressional deformation changed during Neogene times as Atlantic ridge-push forces became increasingly important. Basin inversion ended during the Miocene and gave way to broad-scale lithospheric folding and intracontinental seismicity, which particularly affects the non-basinal areas (Ziegler et al., 1998; Cloetingh et al., 1999, 2008; **Figure 3.28a**). This change from localised deformation to distributed seismicity and lithospheric folding reflects changes in lithosphere strength due to the interplay between Neogene thermal perturbations by mantle plumes and stress-induced intraplate deformation. As a result, inversion tectonics in the North Sea area ceased and gave way to lithospheric folding, which is most pronounced in onshore Variscan massif areas that were strongly weakened by thermal perturbations (Cloetingh & Van Wees, 2005).

10.3 Salt movement

Salt diapirism in the southern North Sea and North German Basin accompanied the Triassic development of large north-trending grabens, such as the Central, Horn and Glückstadt grabens and Rheinsberg Trough (Clausen & Pedersen, 1999). Halokinetic structures are progressively younger towards the margins of these grabens and distinct pulses of halokinesis are seen in the Jurassic and Neogene (Jaritz, 1987; Maystrenko et al., 2005). This may relate to footwall collapse of the grabens with time. In the north-westerly oriented basins at the southern SPB margin (Sole Pit, Broad Fourteens, Münsterland, West and Central Netherlands, Lower Saxony, Altmark, Sub-Hercynian and North Bohemian basins), phases of salt movement along north-westerly trends accompanied their rapid Late Jurassic to Early Cretaceous subsidence and their Late Cretaceous and Paleogene inversion. The axes of major salt structures in the Polish Basin generally trend north-west–south-east. Salt movement started during the Triassic (Dadlez & Marek, 1974; Krzywiec, 2002b; 2004a), particularly the Late Triassic when intense basement faulting and lateral salt withdrawal (**Figure 3.29e**) led to extrusion of salt on the floor of the central part of the basin (Krzywiec, 2004a, 2004b). Salt movement continued throughout the Jurassic (Sokolowski, 1966) with further reactivation during the Sub-Hercynian and Laramide inversion of the Mid-Polish Trough (Krzywiec, 2002a, 2002b, 2004a, 2004b, 2006b). Salt movement further continued into the Cenozoic, reflected by the base-Tertiary depth map (see Chapter 12; Figure 12.1). Moreover, the Pliocene build-up of the present-day stress field was associated with a reactivation of diapirism in the North Sea and possibly in other parts of the SPB, suggesting that increasing confining pressure enhanced salt flowage (Warren, 2006, 2008; Kukla et al., 2008).

11 Regional cross-sections

A series of regional cross-sections are shown in **Figures 3.32** to **3.42**. The extended captions that accompany the figures describe the geology of the SPB.

Acknowledgements

The contribution of T.C. Pharaoh appears with the approval of the Executive Director, British Geological Survey (Natural Environment Research Council). TCP gratefully acknowledges the support of his family during the difficult gestation of this project. Paul Lappage is thanked for drafting many of the original diagrams. The assistance of Manfred Stiller and Stig B. Lyngsie with seismic processing are gratefully acknowledged. The enthusiastic support of Jan de Jager of Shell, as Netherlands representative during the early stages of this project are also acknowledged; many of the diagrams are based on originals supplied by him or by Shell. Peter Ziegler made significant contributions to the text; he and other reviewers (Karsten Schütz and John Karlo) are gratefully acknowledged.

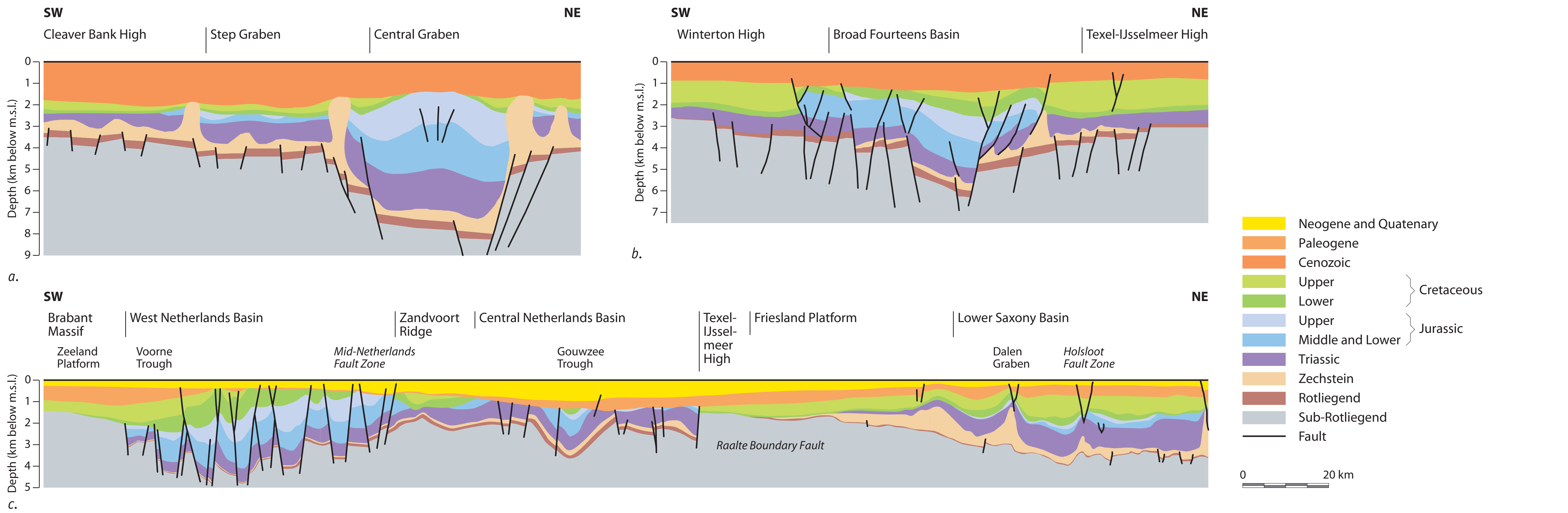


Figure 3.35 Regional depth cross-sections from the Netherlands:
a. Dutch Central Graben (De Jager, 2007);
b. Broad Fourteens Basin (De Jager, 2007);
c. West Netherlands and Central Netherlands basins (NITG-TNO, 2004). The three depth sections are adapted from the Geological Atlas of the Subsurface of the Netherlands and publications (NITG-TNO, 2004; De Jager, 2007) to give an impression of the Mesozoic and Cenozoic structuration of the Netherlands. Although these basins are quite different in detail, a general two-fold subdivision in deformation style can be made between basins with and without (thick) Zechstein salt. The Central and West Netherlands basins are clear examples of those without Zechstein salt. The same applies to the southern part of the Broad Fourteens Basin; in contrast, the northern part of the Broad Fourteens Basin, the Central Graben and the Lower Saxony Basin are examples of basins where the Zechstein salt forms a clear detachment horizon. For more details see Van Wijhe (1987a, 1987b), Dronkers & Mrozek (1991), De Jager (2003, 2007).
See Figure 3.31 for locations.

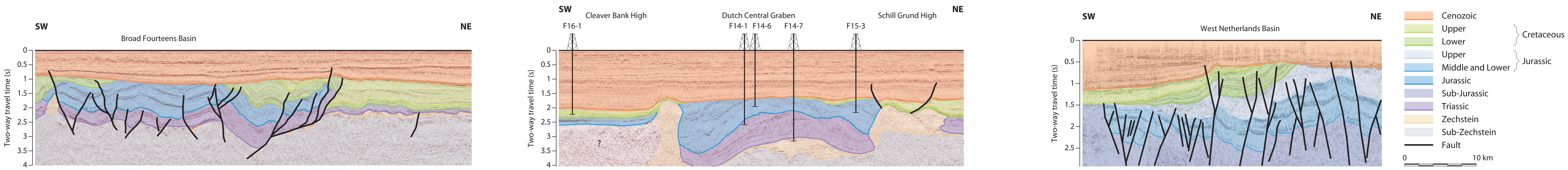
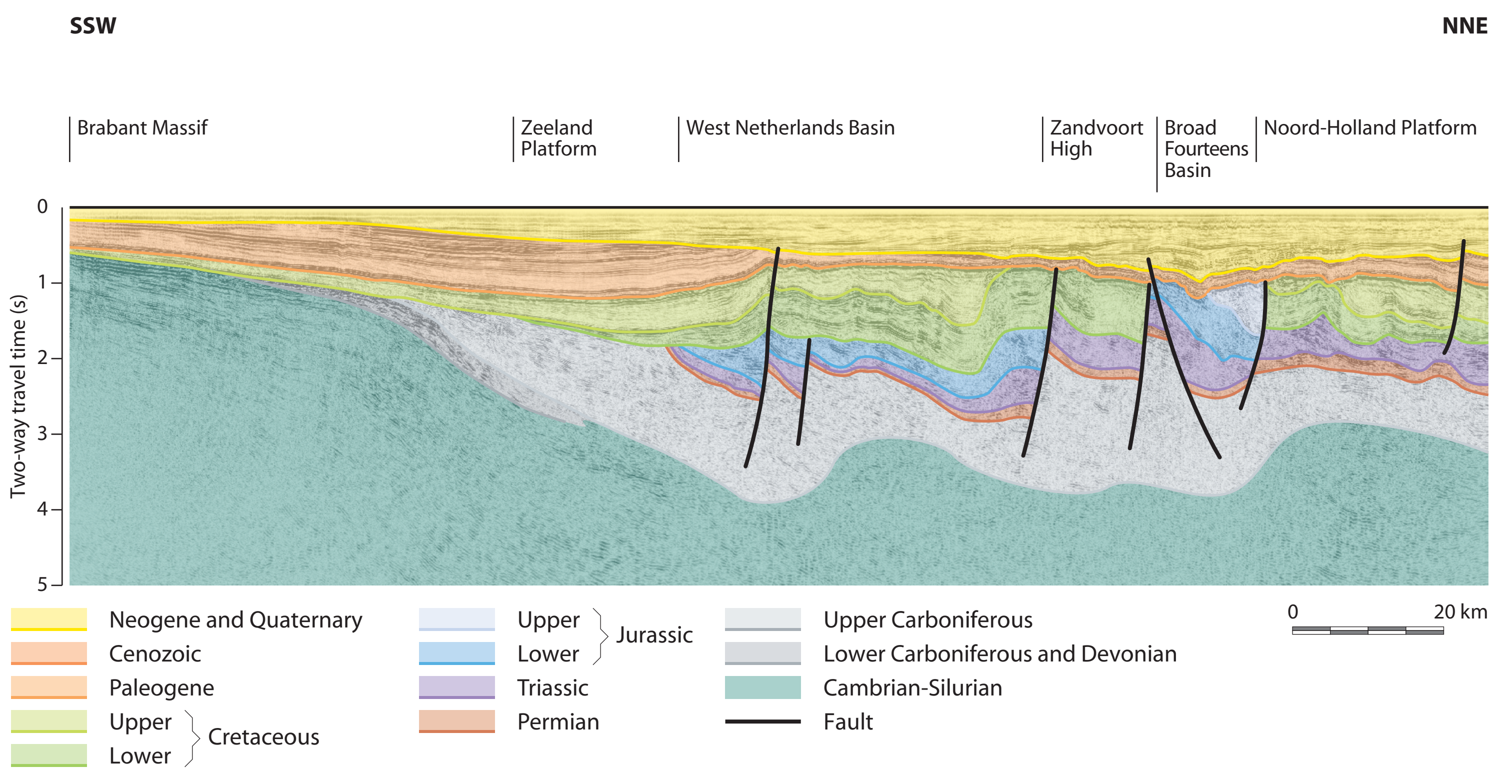


Figure 3.36 Regional geoseismic profiles in time: from the Netherlands:
a. Seismic section across the Broad Fourteens Basin (BFB) (De Jager, 2007). There is a thick Jurassic and Lower Cretaceous succession in this inverted basin, whereas the platforms adjacent to the basins are characterised by a thin Lower Cretaceous succession. The inversion of the BFB resulted in an impressive thrust fault along the NE flank of the basin, where Jurassic and Triassic rocks are thrusted on pre-inversion Lower Cretaceous and Chalk Group sediments. The BFB was already active as a basin in Late Permian and Triassic times (see Chapters 7 to 9); however, its main structuration took place during the Jurassic. The BFB was affected by the same structural events as the Central and West Netherlands basins, but displays a different structural style (Van Wijhe, 1987a, 1987b). The main difference in style is caused by the Zechstein salt in the northern half of the basin. Structuration resulted in a series of tilted fault blocks below this salt at the Rotliegend level. Inversion of the half-grabens filled with Upper Jurassic sediments, characteristic of the West Netherlands Basin, is less clearly developed because structuration above the salt was decoupled from sub-salt faulting. The Sub-Hercynian and Laramide inversion resulted in the development of broad NW-SE-trending anticlinal arches (Dronkers & Mrozek, 1991; Nalpas et al., 1995). A thin, presumably Maastrichtian-Danian chalk conglomerate in some wells in block P9 suggests that inversion of this basin was mainly completed during the Sub-Hercynian Phase. The Eocene Pyrenean inversion mainly affected the southern sector of the basin. Oligocene inversion, which is very pronounced in the West Netherlands Basin, did not affect the BFB (De Jager, 2007). From a structural point of view this basin represents the most complex inverted basin in the Netherlands, especially the NE margin. Details on the inversion are provided by Van Wijhe (1987a, 1987b), Hooper et al. (1995), Huyghe & Mugnier (1995) and Nalpas et al. (1995);
b. Dutch Central Graben (De Jager, 2007). Section showing the Cleaver Bank High, Central Graben and the Schill Grund High. Both the Schill Grund and the Cleaver Bank highs are characterised by a block-faulted pre-Zechstein substrate and the presence of several salt pillows and diapirs. The Central Graben, best known for its Mesozoic history, is an old structure that already existed during Mid-Devonian times (De Jager, 2007; see Chapter 4). Subsequent activity of the graben is known from Carboniferous and Permian times (see Chapters 6-8). The Zechstein succession in the graben is thin due to large amounts of salt withdrawal associated with growing salt diapirs bordering the graben. During the Triassic and Early Jurassic, the Dutch Central Graben had already subsided more than the adjacent platforms, but not as much as the Glückstadt and Horn Grabens in Germany. The Central Graben continued to subside during regional uplift of the Central North Sea Dome, (Ziegler, 1990a; Underhill & Partington, 1993) and contains a thick and complete Jurassic succession. The main rifting of the Central Graben took place during Late Jurassic times, progressing in time from N to S (Heybroek, 1975; Schroot, 1991). Rifting ceased during Early Cretaceous times. The Lower Cretaceous sediments deposited in the north of the Central Dutch Graben are thin, and thicken to the south. The Central Graben was inverted in Late Cretaceous times, as is evident from the thinning of the Chalk Group deposits on the crest of the graben; most of the Chalk Group represents post-inversion deposits of Maastrichtian to Danian age. The inversion removed all Lower Cretaceous deposits from the graben. In Cenozoic times, both the graben and the adjacent highs subsided regionally as part of the North Sea Basin;
c. West Netherlands Basin (De Jager, 2007). In the West Netherlands Basin, which is devoid of Zechstein salt, late Cimmerian extension resulted in a series of half-grabens filled with Upper Jurassic and Lower Cretaceous syn-rift clastics of the Delfland Subgroup. Younger Cretaceous post-rift deposits onlap and overstep the margins of these rift basins. Pre-existing faults were reactivated reversely during the Late Cretaceous and Early Cenozoic inversion, with transpressional movements forming a series of prominent NW-SE-trending flower structures. A reversed sense of movement can still be deduced clearly along many individual faults: normal offsets occur at deeper stratigraphic levels and reverse offsets at shallower levels. Thinning of the Upper Cretaceous chalk from the SE towards the inverted basin, with several internal unconformities visible on seismic data, clearly shows that the flower structures were formed during the Late Cretaceous Sub-Hercynian inversion. The West and Central Netherlands basins were both further uplifted and deeply eroded during the Paleocene Laramide inversion, locally down to the Triassic. The Sub-Hercynian inversion in the Central Netherlands Basin is less well documented, as no Upper Cretaceous deposits are preserved due to deep erosion during later phases, although Danian Chalk overlies pre-Cretaceous successions in several wells (NITG-TNO, 2004). Inversion of the basin continued during the Early Cenozoic. The Lower Cretaceous strata thin towards the inverted basin at the SW margin of the West Netherlands Basin. Erosion took place during the Oligocene in the most inverted sector of the basin, as well as in the Central Netherlands Basin. It is noteworthy that, although the Sub-Hercynian and Laramide inversions resulted in reverse reactivation of pre-existing faults, the Pyrenean inversion caused broad basin uplift without fault reactivation. Relaxation of the stress regime at the end of the inversion movement resulted locally in normal reactivation of faults.
See Figure 3.31 for locations.



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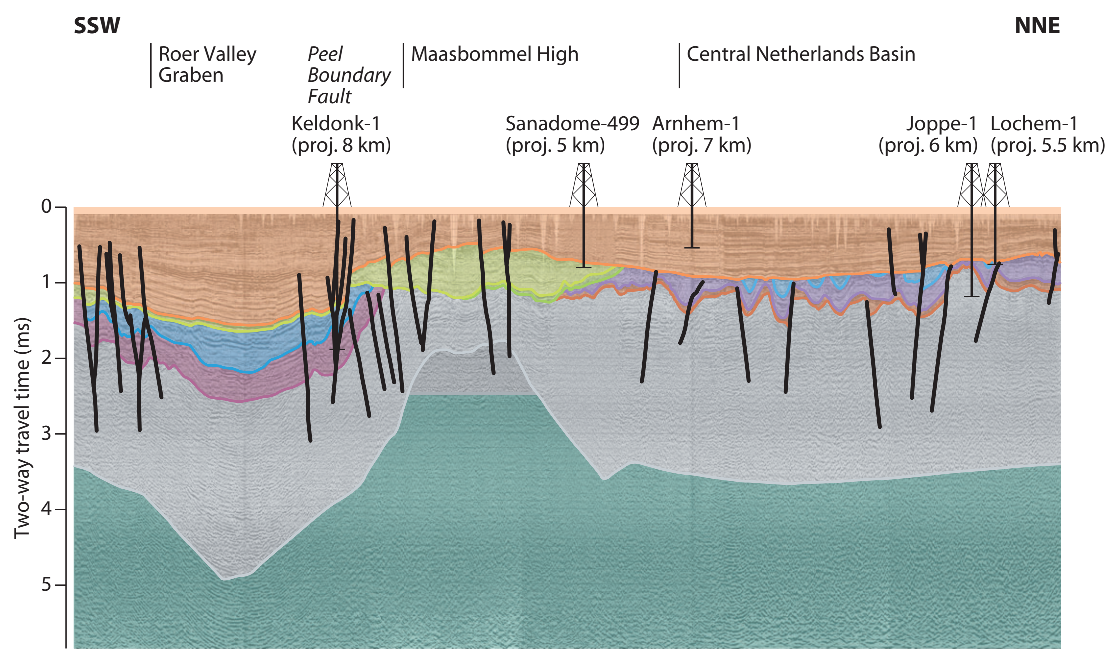
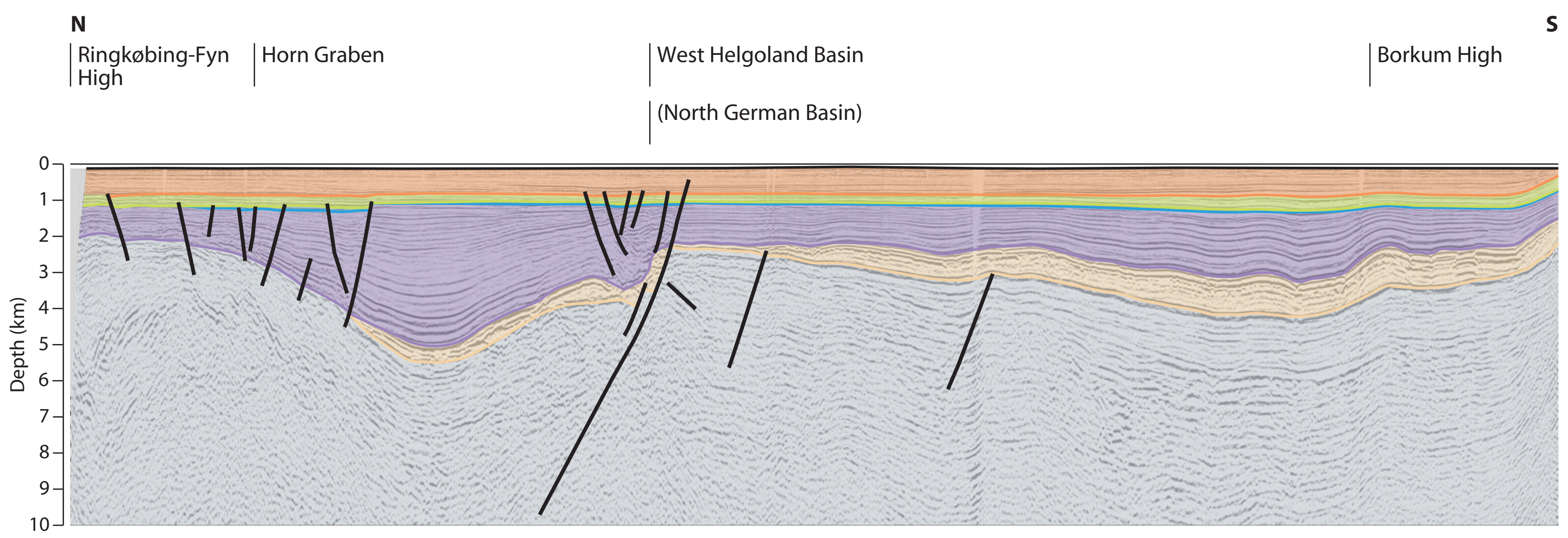


e.

Figure 3.36 d. MPNI-9101. This deep seismic section provides a good overview of the main structures of the southern North Sea: the Brabant Massif, the Zeeland Platform (with a northward-thickening Carboniferous succession), the inverted West Netherlands and Broad Fourteens basins and the bordering Zandvoort High and Noord-Holland Platform. The section illustrates the relatively shallow position of Carboniferous and pre-Carboniferous rocks in the southern Netherlands and their deep burial below the Mesozoic basins. In the south-westernmost part of the section, only a thin Chalk Group and Cenozoic succession rests unconformably on Caledonian deformed Cambro-Silurian deposits (as encountered in the Belgian wells Knokke, Knokke-Heist and Eeklo (Legrand, 1968; De Vos et al., 1993)). Northwards, a rapidly expanding wedge of Devonian and Carboniferous deposits marking the Variscan Foreland Basin infill north of the Variscan Mountains, are situated below the Chalk Group and Cenozoic on the Zeeland Platform. Thick Carboniferous deposits, often in excess of 5000 m, are present below the West Netherlands and Broad Fourteens basins. Permian to Middle Jurassic deposits were deposited in almost the entire area. A gradual overstepping of the Zeeland Platform and the Brabant Massif took place during Late Permian and Triassic times. Only a thin cover of sediments has been deposited on the top of the Brabant Massif, which was removed during Late Jurassic times (Van den Haute & Vercoutere, 1990). The West and Central Netherlands basins are two inverted late Cimmerian rift basins. Although the onset of differential subsidence of these basins was as early as during the Permian, the main rifting took place during the Late Jurassic. Differential movements ceased in Early Cretaceous times, and from the Valanginian onwards the margins of the basins were gradually overstepped. This process continued during most of the Cretaceous as can be seen by the onlap of Lower and Upper Cretaceous sediments onto the Zeeland Platform and the Brabant Massif. Inversion of the rift basins took place during Coniacian-Santonian and Paleocene times (see above). This inversion resulted in uplift of the basins, removing most of the Chalk Group and locally older deposits, and a contemporaneous rapid subsidence of the adjacent highs (thick Chalk Group on the Noord-Holland Platform, Zandvoort High and Voorne Trough). The entire area was situated in the southern part of the North Sea Basin during Cenozoic times as can be seen by the northward thickening of deposits of the Lower, Middle and Upper North Sea Groups. Note that the pre-Cenozoic structures were mostly no longer active;

e. Section DG 8601 (parts D-H). This NNE-SSW-oriented deep seismic section displays the tectonic structure of the central and southern Netherlands onshore. It shows two Late Paleozoic-Mesozoic basins, the Roer Valley Graben and the Central Netherlands Basin, characterised by a thick succession of Permian to Jurassic sediments and a general absence of Cretaceous sediments due to Late Cretaceous inversion tectonics. Four main tectonic events are responsible for the current structure: Early Permian uplift and wrenching, Late Jurassic – Early Cretaceous extension, Late Cretaceous compression and Neogene rifting. Pronounced Early Permian uplift of the Maasbommel High resulted in erosion of the Namurian-Westphalian succession, as shown by well data and an angular unconformity at the base of the Permian in the NW of the Roer Valley Graben (NITG, 2001). Permian to Middle Jurassic sediments were deposited over the entire area and are thicker in the Roer Valley Graben and Central Netherlands Basin and thinner on the Maasbommel High. Strong uplift of the Maasbommel High took place during the Late Jurassic – Early Cretaceous, removing most of the Permian to Middle Jurassic sediments. Lower Cretaceous marine clastics were deposited in the Central Netherlands Basin while continental sedimentation took place in the Roer Valley Graben at the same time. Differential subsidence of the basins continued until Santonian times, after which compressional tectonics resulted in the inversion of the Roer Valley Graben and Central Netherlands Basin. At the same time, the Maasbommel High started to subside rapidly and erosional products from the inverted basins were deposited on the high (Gras & Geluk, 1999). In contrast to the MPNI-9101 line, Cenozoic differential subsidence plays an important role here. The Roer Valley Graben, the NW branch of the Rhine Graben rift system underwent strong Late Oligocene and Neogene subsidence (Geluk et al., 1994). The Peel Boundary Fault Zone separates the Roer Valley Graben from the Peel Horst. The Roer Valley Graben and Central Netherlands Basin display quite different structures. The Roer Valley Graben is essentially a relatively simple faulted synclinal structure with mainly normal faults, and only some indications of small reversal movements. The Carboniferous depocentre underlying the Mesozoic/Cenozoic basin is slightly offset, with its axis on the SW flank of the Cenozoic graben. A thin cover of Maastrichtian-Danian post-inversion Chalk Group sediments overlies the Jurassic (Lower-Middle Jurassic). In contrast, the Central Netherlands Basin is a deeply eroded and broken-up basin where Permian-Triassic sediments occur below the Cenozoic. Cretaceous sediments have been removed completely. Lower Jurassic sediments are preserved locally in lows. Relatively thin Zechstein salt has created minor salt pillows and local detachment zones. The Permian shows local thickness variations across faults, probably reflecting mainly early salt flow in salt-filled half-grabens. Reverse faults and low-angle thrusts have resulted in much more shortening than in the Roer Valley Graben.
See Figure 3.31 for locations.



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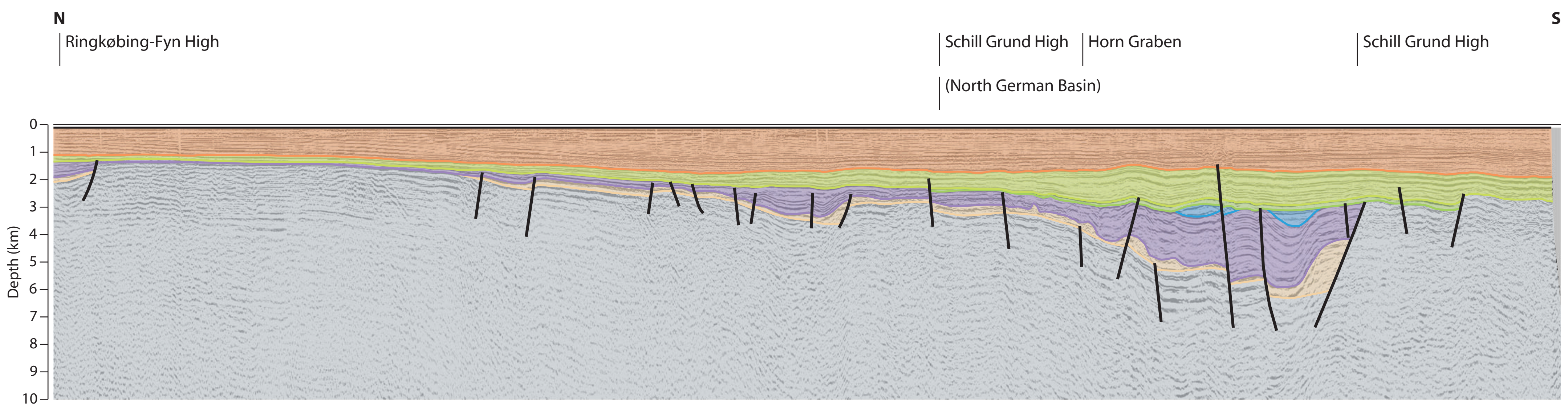

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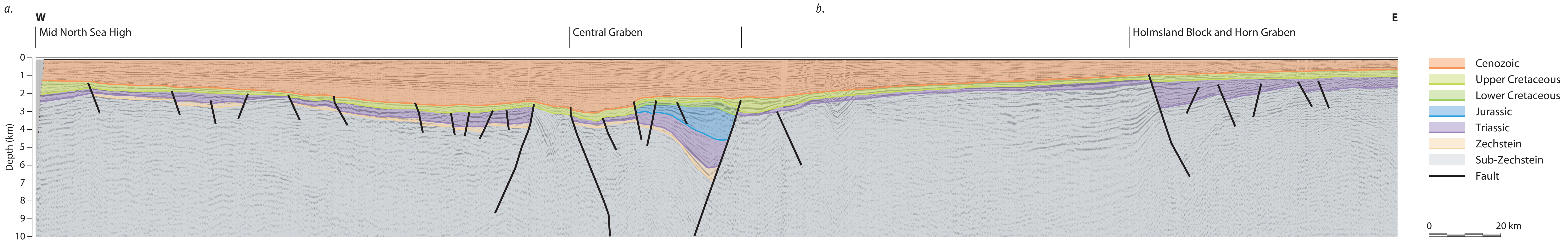

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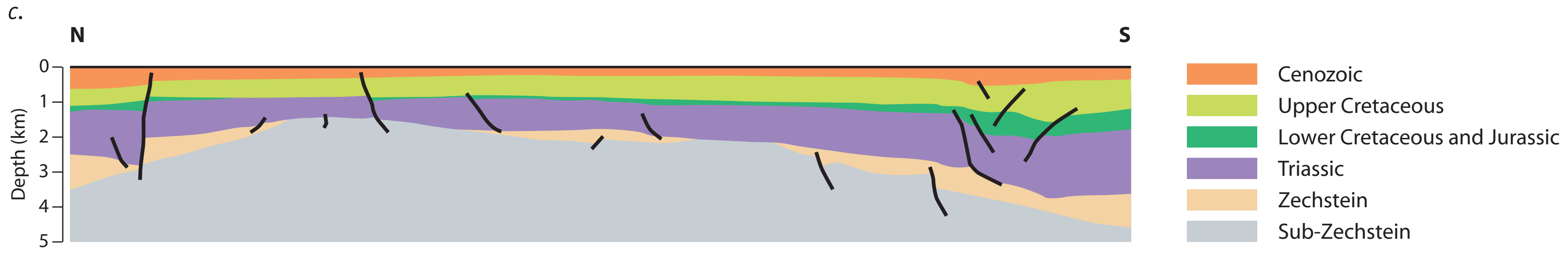


d.

Figure 3.37 Regional depth cross-sections in Denmark:
a. MONA LISA 1. Depth-migrated section across the Triassic-filled Horn Graben and Ringkøbing-Fyn High with a thin cover of Zechstein and Triassic strata;
b. MONA LISA 2. Depth-migrated section across the Ringkøbing-Fyn High to the southern limit of the Horn Graben;
c. MONA LISA 3. Depth-migrated section across the Mid North Sea High, Elbow Spit, Outer Rough Basin, Tail End Graben and the Holmsland and Grindsted blocks at the southern edge of the Norwegian-Danish Basin;
d. N-S profile onshore Jylland. See Figure 3.31 for locations.

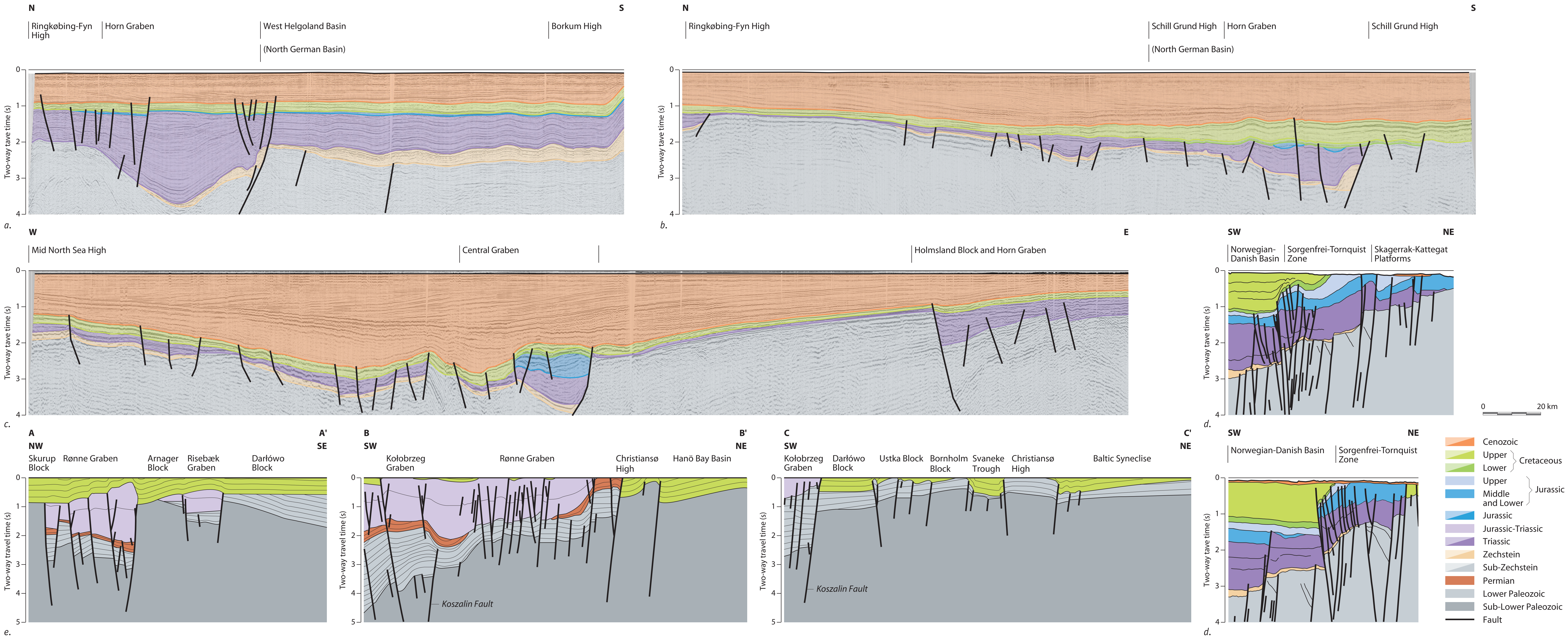


Figure 3.38 Regional seismic profiles in time: from Denmark:
a. MONA LISA 1. Seismic section across the Triassic-filled Horn Graben and Ringkøbing-Fyn High with a thin cover of Zechstein and Triassic strata;
b. MONA LISA 2. Seismic section across the Ringkøbing-Fyn High to the southern limit of the Horn Graben;
c. MONA LISA 3. Seismic section across the Mid North Sea High, Elbow Spit, Outer Rough Basin, Tail End Graben and the Holmsland and Grindsted blocks at the southern edge of the Norwegian-Danish Basin.
d. Seismic sections across the Sorgenfrei-Tornquist Zone close to the northern edge of the SPB (Vejbæk, 1997);
e. Geosections around Bornholm (Vejbæk et al., 1994).
See Figure 3.31 for locations.

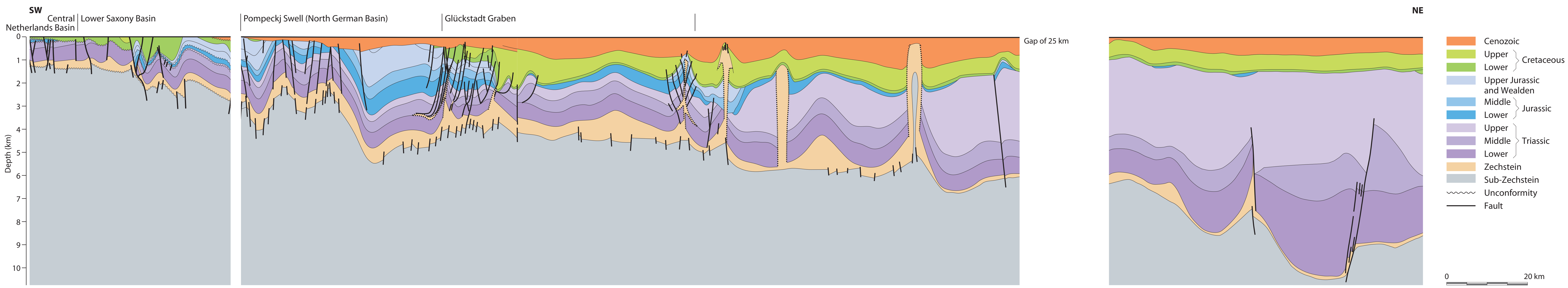
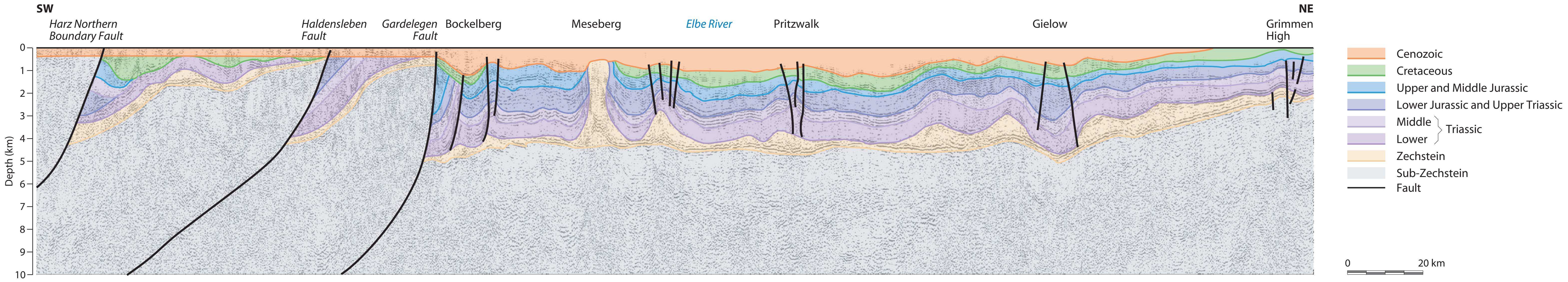
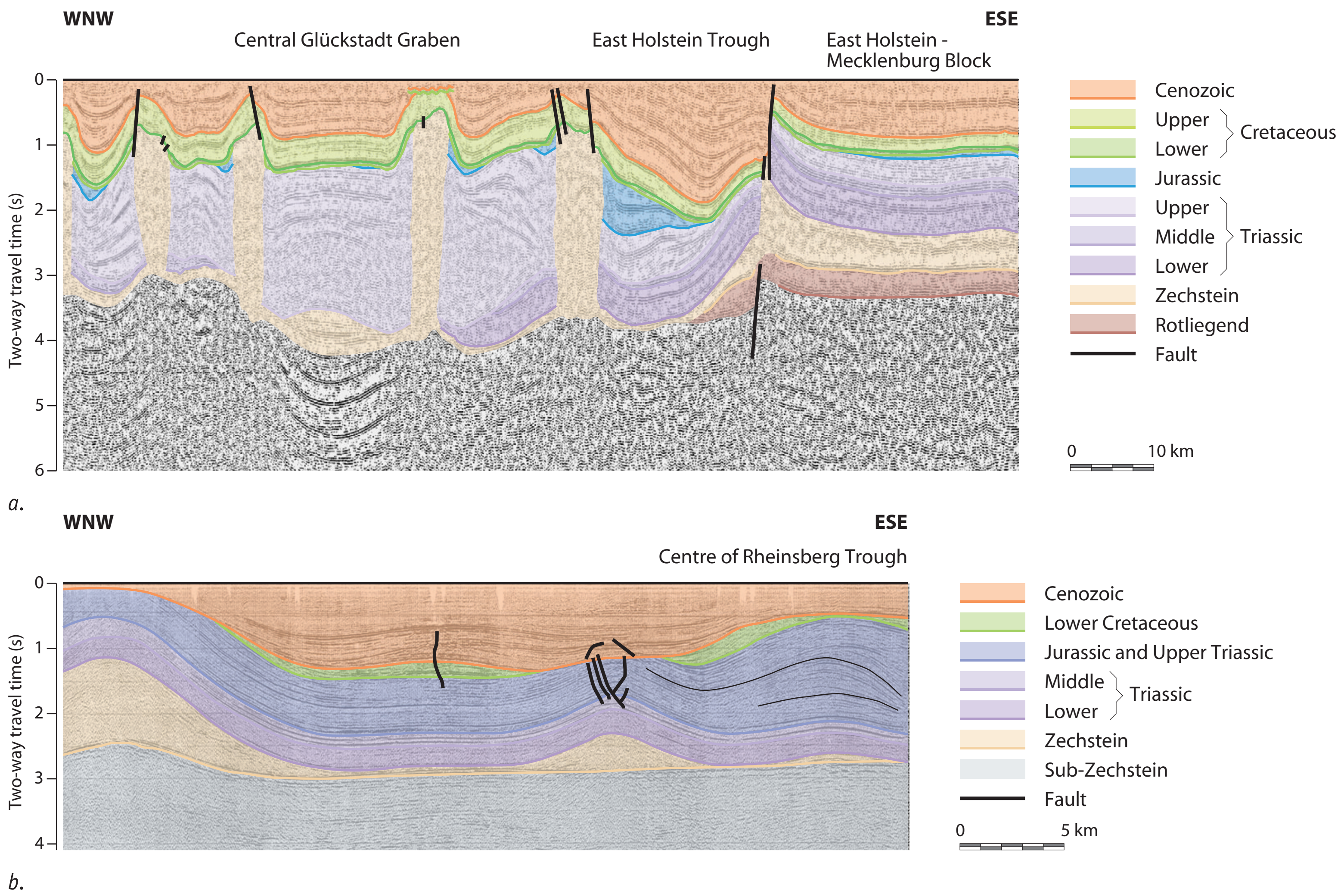


Figure 3.39a Regional cross-section across the Lower Saxony Basin-Pompeckj Swell-Glückstadt Graben. Montaged from profiles 80, 94, 95 and 18 in Baldschuhn et al. (2004). The section starts in the eastern extremity of the Central Netherlands Basin, part of the Ems Low during Triassic times. The section crosses the Gronau-Waldhugel Fault Zone and the SW boundary of the Lower Saxony Basin. Most of the Cretaceous cover was removed during strong Sub-Hercynian inversion. The complex NW boundary of the Lower Saxony Basin (Rheder Moor Lineament with significant salt movement) is shown. The Jurassic sequence is much thinner on the Pompeckj Block (Oldenburg Swell) than in the Lower Saxony Basin, and the smaller amount of inversion results in preserved Upper Cretaceous strata. The section crosses the Bremen Low (Triassic) marked by increasing thickness of Triassic strata and continues into the Glückstadt Graben. Note the evidence for early (pre-Jurassic) salt movement (Neuenhuntorf), with further movement during Alpine (Late Cretaceous-Paleogene) inversion. See Figure 3.31 for location.

Figure 3.39b Depth-migrated seismic-reflection profile (BASIN9601) crossing the entire North German Basin with interpretation (modified after DEKORP-BASIN Research Group (1999) and Krawczyk et al. (2008b)). The basal and top Zechstein salt layers are clearly imaged by two distinctive reflectors across much of the region at 2.5 to 5 km depth. Salt-pillow structures with amplitudes increasing from the NE to the SW are overlain by the Triassic, but are partially truncated by Jurassic to Cenozoic reflectors, indicating increasing halokinesis during the late Mesozoic and Cenozoic. Significant reflectivity is seen near the depocentre beneath the basal Zechstein reflector, decreasing in intensity both north- and southwards. These are interpreted as Namurian-Rotliegend sedimentary and volcanic rocks. At the southern basin margin, the basal Zechstein reflector shows ~5000 m vertical offset at the Gardelegen Fault, indicating Late Cretaceous inversion. Offsets of 2500 m are also seen at the Haldensleben and Harz Boundary faults. See Figure 3.31 for location.





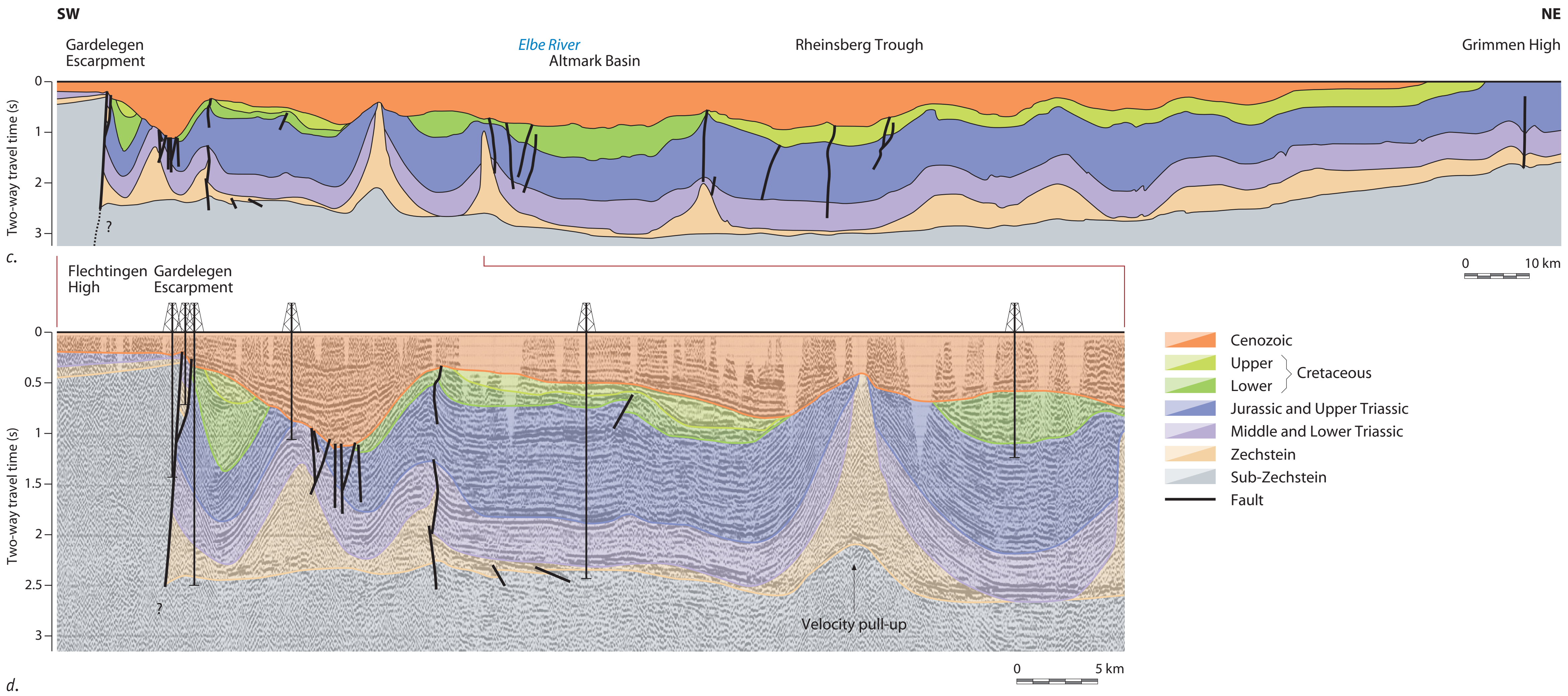


Figure 3.40 Regional seismic profiles in time: from Germany:

a. Migrated seismic section across the central Glückstadt Graben (after Scheck-Wenderoth et al., 2008) showing that the interpretation of the deepest part of the graben is still a subject of debate. Note that different interpretations can be made on the basis of deep-seismic profiles such as this, concerning the thickness of the Lower Triassic Buntsandstein and the location of the base Triassic. Major salt movements took place at the beginning of the Keuper, when the Glückstadt Graben was affected by extension. The internal seismic pattern of the Keuper, lithostratigraphic data and palynological investigations (Trusheim, 1960), indicate that Permian salt extruded onto the palaeosurface and was dissolved and redeposited within the Keuper strata. The Early Jurassic extension and related normal faulting documented in the Lower Saxony Basin and within the Pompeckj Block (Kockel, 2002) may have also affected the Glückstadt Graben. Thick Jurassic sediments are only observed around salt structures and thin with increasing distance from salt walls or salt stocks. Parts of the Jurassic were eroded in Late Jurassic-Early Cretaceous times. The Upper Cretaceous strata have an almost constant thickness and their parallel reflection pattern indicates a quiet tectonic setting with very minor salt movements during the Late Cretaceous;

b. Migrated seismic section with interpretation (below) across the western shoulder of the Rheinsberg Trough (after Scheck-Wenderoth et al., 2008). The onset of salt mobilisation is observed to be synchronous with the development of the NNE-SSW striking Rheinsberg Trough in the Late Triassic Keuper. This is inferred from stratigraphic thickening in the reflections interpreted as Late Triassic (Keuper) to Jurassic toward the trough centre and showing indications for synsedimentary salt movements in the reflectivity pattern. The salt is almost completely removed below the trough. While normal faults are present in the Mesozoic cover, the base Zechstein appears as a strong continuous signal below the trough;

c. Interpreted seismic section across the North German Basin perpendicular to the strike of the Permo-Triassic basin axis and to the WNW-ESE striking inversion structures from south of Rügen to the southern basin margin parallel to the NNE-SSW striking Rheinsberg Trough (line DEKORP BASIN9601) (after Scheck-Wenderoth et al. 2008). The line indicates continuous subsidence during the Triassic-Jurassic, but inversion in the Late Cretaceous. Deformation is tectonically decoupled by the Zechstein salt. Basement deformation is localised and most intense at the Elbe Fault System, where the base Zechstein is displaced by 2 s TWT along the Gardelegen Fault. Some smaller basement faults are visible to about 50 km north of the Gardelegen Fault and at the northern margin. The base Zechstein appears as a strong continuous signal below salt-depleted areas as well as below salt pillows and diapirs, but the amount of deformation in the salt basement is negligible compared to the amount in the cover or at the southern margin;

d. Detail of migrated part of line shown in Figure 3.40c.

e. Interpreted seismic section across the central North German Basin to the southern margin (across the Lower Saxony Basin) (after Scheck-Wenderoth et al. 2008). The section illustrates continuous subsidence during Early to Mid-Triassic times (Buntsandstein and Muschelkalk) represented by parallel, continuous reflections. The overlying reflections are interpreted as Upper Jurassic-Lower Cretaceous and show local stratigraphic thickening indicating accelerated subsidence in the Lower Saxony Basin at the southern margin of the SPB compared to the Pompeckj Block farther north. The tectonic decoupling by the Zechstein salt across large parts of the basin is obvious, as well as localised, basement-involved uplift in the Lower Saxony Basin where salt is thin or absent. The salt cover is affected by folding and faulting in the basinal part, whereas the salt basement shows almost no faulting (modified after Mazur & Scheck-Wenderoth, 2005);

f. Detail of migrated and coherency filtered seismic line from Figure 3.40e;

See Figure 3.31 for locations.

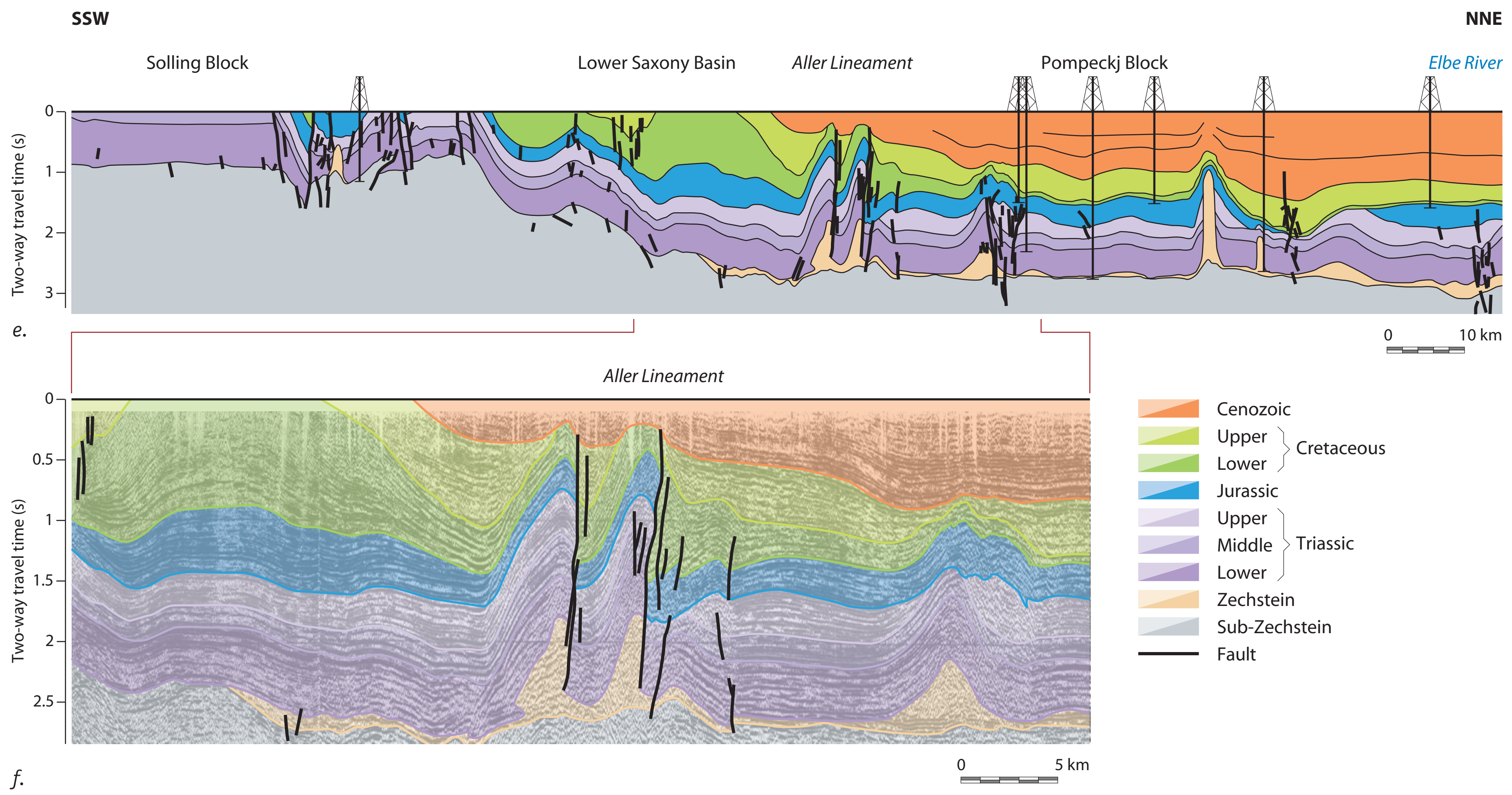
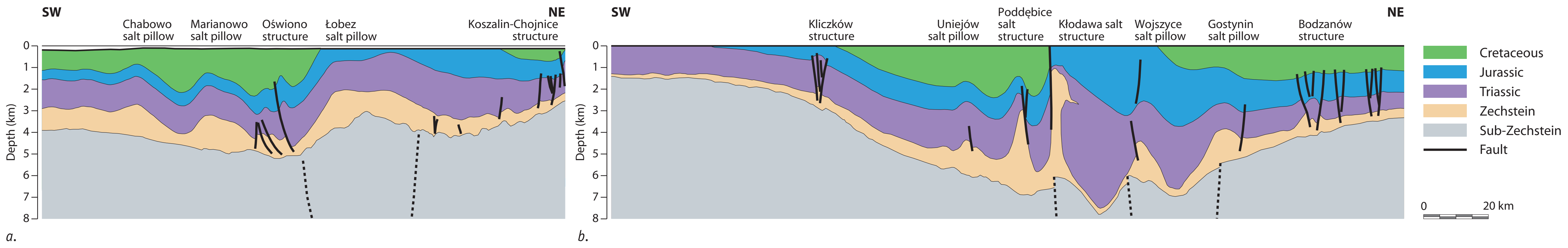


Figure 3.41 Regional depth cross-sections in Poland: a. Geological cross-section across the Pomeranian segment of the Mid-Polish Trough based on regional seismic profiles (cf. Krzywiec, 2006a, 2006b; Krzywiec et al., 2006). Sub-Zechstein faults are inferred fault zones responsible for subsidence and subsequent inversion of the Mid-Polish Trough. Thickness distribution of the Mesozoic sedimentary cover is much more symmetrical in comparison to the Kuiavian example, suggesting more uniform tectonic subsidence along both flanks of the Mid-Polish Trough;
b. Geological cross-section across the Kuiavian segment of the Mid-Polish Trough based on regional seismic profiles (cf. Krzywiec, 2006a, 2006b; Krzywiec et al., 2006). Sub-Zechstein faults are inferred fault zones responsible for subsidence and subsequent inversion of the Mid-Polish Trough. Note significant thickness variations of the Triassic and Jurassic deposits in the vicinity of the Kłodawa salt structure, in particular the very asymmetric thickness distribution of the Triassic succession, suggesting localised normal faulting beneath the Kłodawa salt structure. See Figure 3.31 for locations.



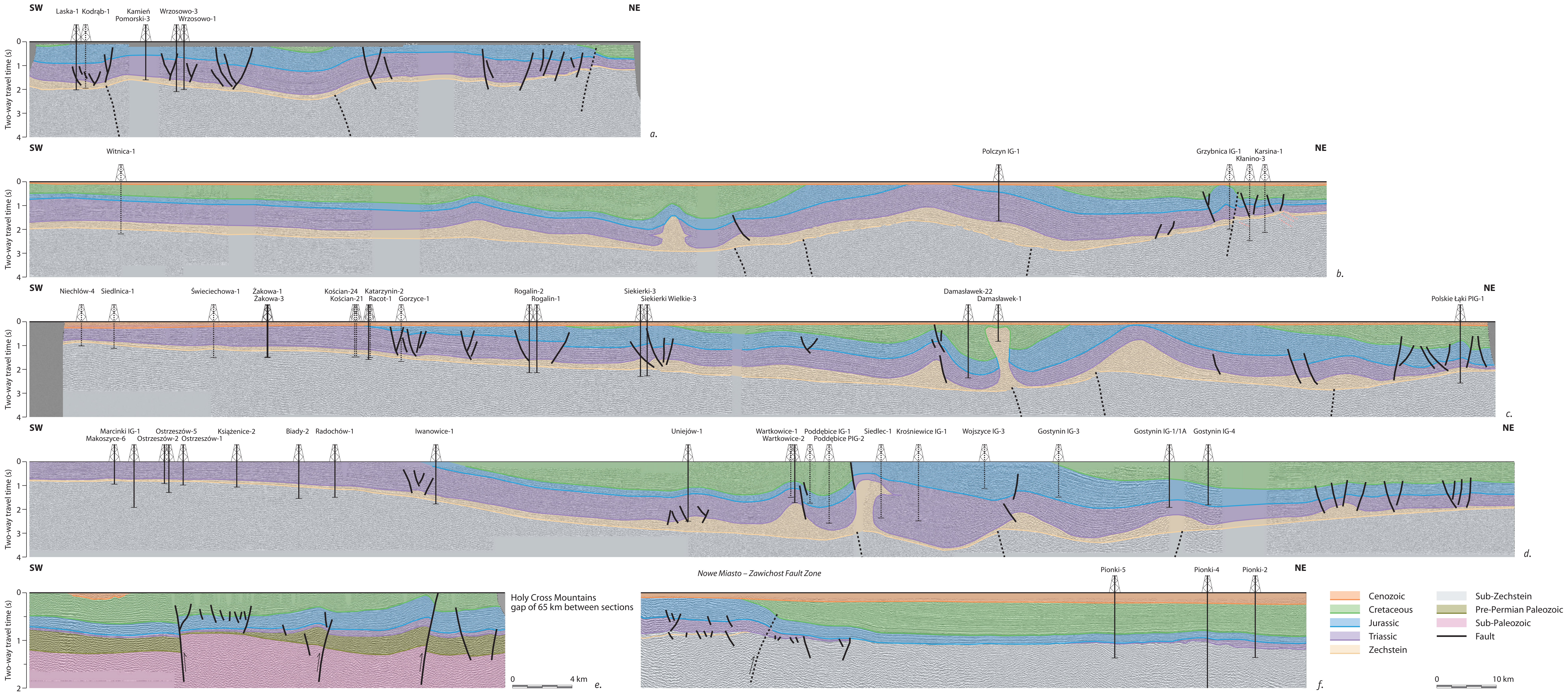


Figure 3.42 Regional seismic profiles in time: from Poland:
a. Seismic section across the NW (Baltic) segment of the Mid-Polish Trough, showing major inversion structures such as NW-SE trending Kamień-Pomorski-Adler, Trzebiatów and Koszalin-Chojnice fault zones (based on Krzywiec, 2006a; Krzywiec et al., 2006; cf. also Vejbæk et al., 1994; Schlüter et al., 1997). Sub-Zechstein faults are inferred fault zones responsible for subsidence and subsequent inversion of the Mid-Polish Trough;
b. Seismic section from the NW (Pomeranian) segment of the Mid-Polish Trough, showing the structure of the inverted Polish Basin (based on Krzywiec, 2006a; Krzywiec et al., 2006). The axial part of the basin, the Mid-Polish Trough, underwent strongest subsidence and was characterised by thickest (although presently partly eroded) Mesozoic sedimentary cover. The Mid-Polish Trough also focussed inversion of the Polish Basin, and presently forms the so-called Mid-Polish Swell, a regional antiformal structure clearly visible on this seismic profile. Sub-Zechstein faults are inferred fault zones responsible for subsidence and subsequent inversion of the Mid-Polish Trough. Some hard-linked basement faulting is observed in the NE part of the profile, which propagated into the Mesozoic cover;
c. Seismic profile from the NW (Pomeranian) segment of the Mid-Polish Trough, showing the structure of the inverted Polish Basin (based on Krzywiec, 2006a; Krzywiec et al., 2006). The Mid-Polish Trough, underwent strongest subsidence and was characterised by the thickest (although presently partly eroded) Mesozoic sedimentary cover. Zechstein faults are inferred fault zones responsible for subsidence and subsequent inversion of the Mid-Polish Trough. The salt structures that developed above basement fault zones (Szubin salt pillow and the fully pierced Damaslawek salt diapir) are seen in the central part of the profile. Along the SW flank of the basement, a system of cover deformations is observed detached above Zechstein evaporites. Their formation may have been partly triggered by sub-Zechstein strike-slip movements (Kwolek, 2000);
d. Seismic profile from the central (Kuiavian) segment of the Mid-Polish Trough, showing the structure of the inverted Polish Basin (based on Krzywiec, 2004a; Krzywiec, 2006a; Krzywiec et al., 2006). Sub-Zechstein faults are inferred fault zones responsible for subsidence and subsequent inversion of the Mid-Polish Trough. The inferred basement fault zone located beneath the Kłodawa salt diapir focused both extension and normal faulting during basin subsidence, as well as reverse faulting and uplift during basin inversion. Note that Figure 3.29f is a detail from this profile;
e. Profile from the SE segment of the Mid-Polish Trough to the SW of the Holy Cross Mountains (based on Krzywiec, 2002a and Scheck-Wenderoth et al., 2008). Note the thickness changes, local angular unconformities and progradational pattern within the Upper Turonian(?) to Maastrichtian deposits, which suggest they are related to inversion and indicate uplift of the axial part of the Mid-Polish Trough. The top of the Triassic and Zechstein successions within the hanging wall are based on long-range correlations across major fault zones due to the lack of deep wells and should be regarded as approximate;
f. Profile from the Nida Trough (based on Scheck-Wenderoth et al., 2008). On the NE flank of the Holy Cross Mountains, the uppermost Cretaceous deposits are characterised by increased thicknesses towards the axial parts of the inversion anticlines suggesting that these areas were still sites of subsidence during the Late Cretaceous, and that inversion took place later in Paleogene times. The Miocene sediments that partly cover the Mesozoic succession are part of the infill of the Carpathian foredeep basin. A progressive thickness increase of the Jurassic succession is observed towards the NE within the Nida Trough. Note that Figure 3.22i is a detail from this profile.
See Figure 3.31 for locations.

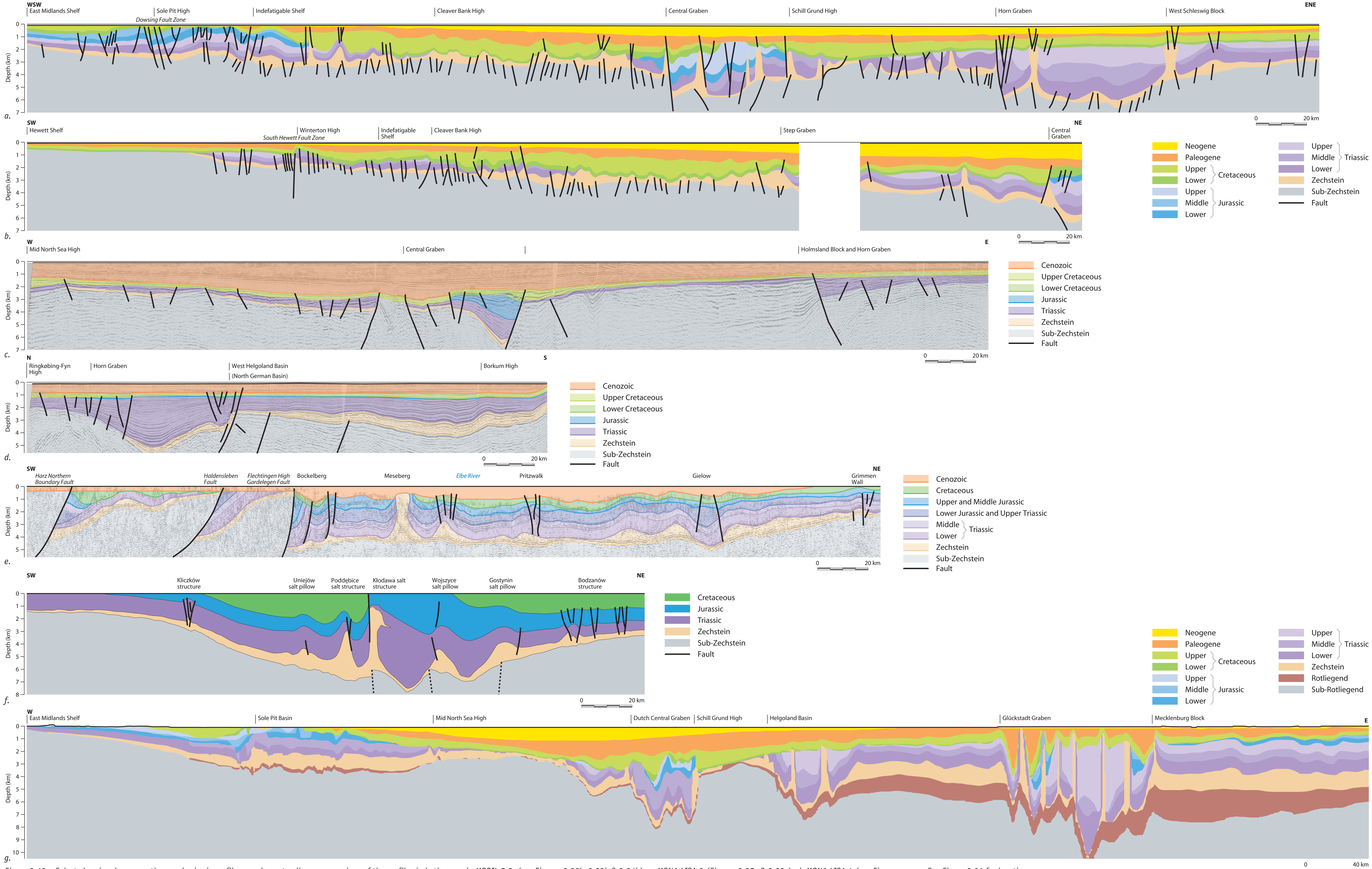
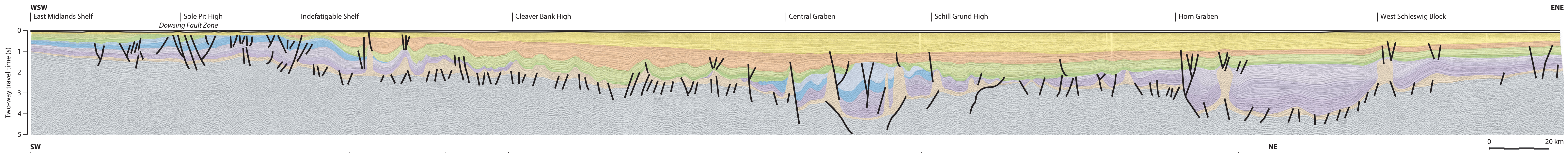
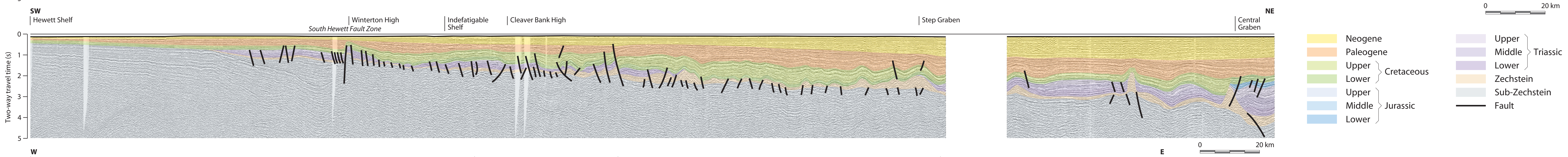
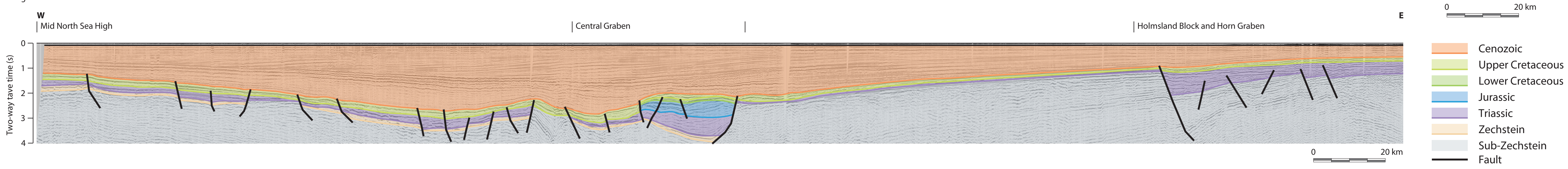


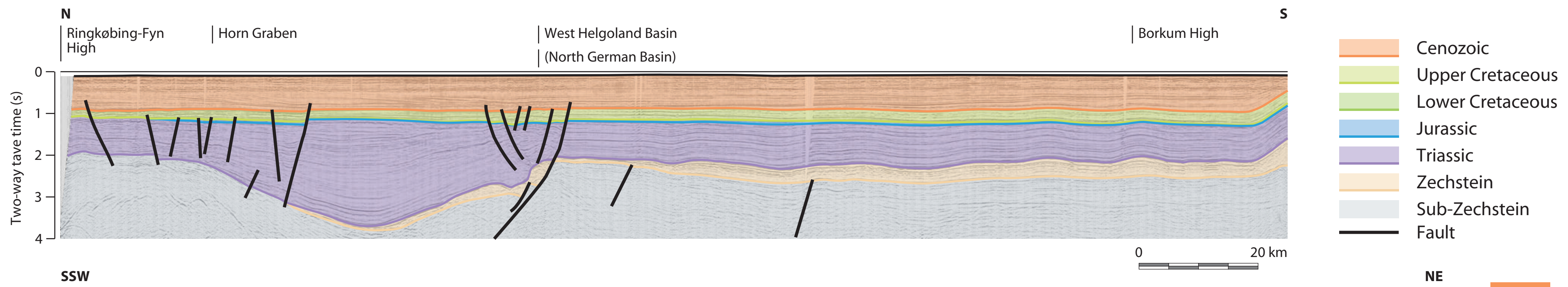
Figure 3.43 Selected regional cross-sections and seismic profiles are shown to allow a comparison of the profiles in both depth (left page) and time (right page). The profiles are presented at similar vertical and horizontal scales. See the captions to the figures given in brackets for an explanation of the geology. a. Profile SNST83-07 (see Figures 3.32a, 3.33a & 3.34a); b. MOBIL 7-8 (see Figures 3.32b, 3.33b & 3.34b); c. MONA LISA 3 (Figures 3.37c & 3.38c); d. MONA LISA 1 (see Figures 3.32d, 3.37a & 3.38a); e. BASIN 9601 (see Figures 3.32g & 3.39b); f. Profiles from the central (Kuiavian) segment of the Mid-Polish Trough (see Figures 3.32i, 3.41b & 3.42d); g. Regional line crossing the entire SPBA area from the UK to Poland.

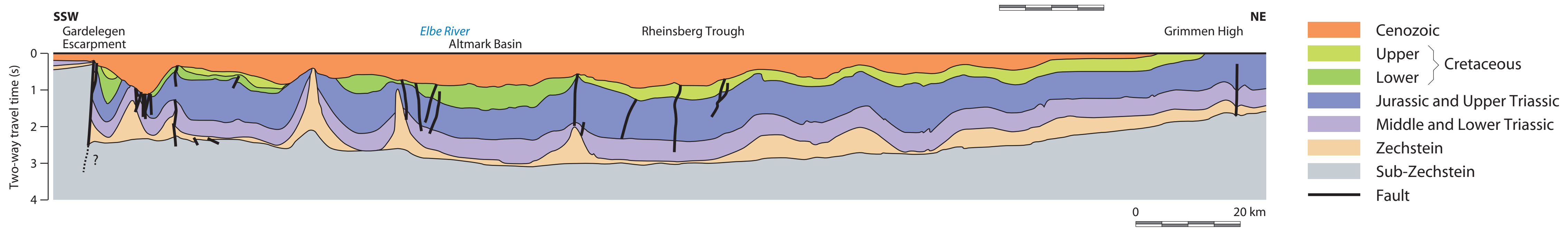
See Figure 3.31 for locations.

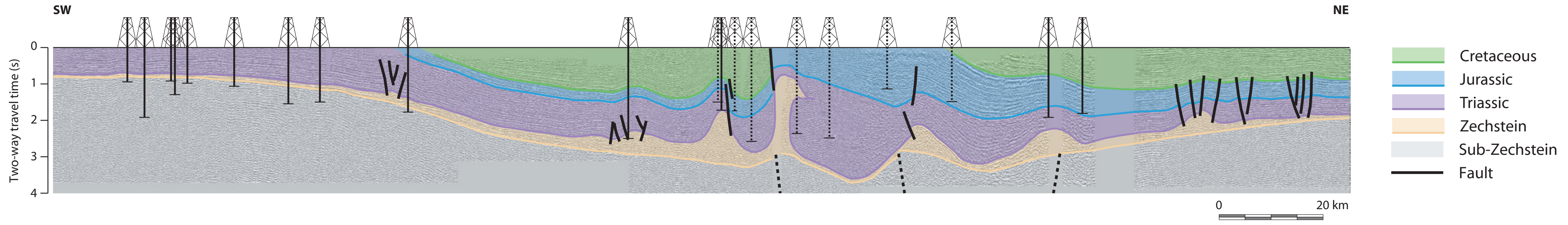

WSW
ENE
East Midlands Shelf
Sole Pit High
Dowsing Fault Zone
Indefatigable Shelf
Cleaver Bank High
Central Graben
Schill Grund High
Horn Graben
West Schleswig Block
Two-way travel time (s)
0
1
2
3
4
5
0
20 km

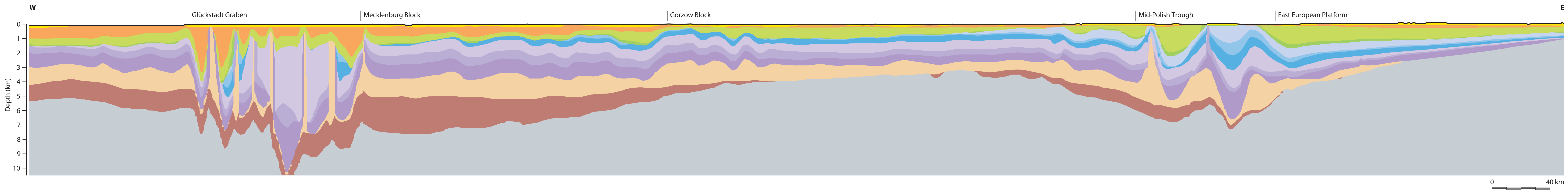

SW
NE
Hewett Shelf
Winterton High
South Hewett Fault Zone
Indefatigable Shelf
Cleaver Bank High
Step Graben
Central Graben
Two-way travel time (s)
0
1
2
3
4
5
0
20 km
Neogene
Paleogene
Upper
Lower
Cretaceous
Upper
Middle
Lower
Jurassic
Upper
Middle
Lower
Triassic
Zechstein
Sub-Zechstein
Fault


W
E
Mid North Sea High
Central Graben
Holmsland Block and Horn Graben
Two-way tave time (s)
0
1
2
3
4
0
20 km
Cenozoic
Upper Cretaceous
Lower Cretaceous
Jurassic
Triassic
Zechstein
Sub-Zechstein
Fault


N
S
Ringkøbing-Fyn High
Horn Graben
West Helgoland Basin
(North German Basin)
Borkum High
Two-way tave time (s)
0
1
2
3
4
0
20 km
Cenozoic
Upper Cretaceous
Lower Cretaceous
Jurassic
Triassic
Zechstein
Sub-Zechstein
Fault


SSW
NE
Gardelegen Escarpment
Elbe River
Altmark Basin
Rheinsberg Trough
Grimmen High
Two-way travel time (s)
0
1
2
3
4
?
0
20 km
Cenozoic
Upper
Lower
Cretaceous
Jurassic and Upper Triassic
Middle and Lower Triassic
Zechstein
Sub-Zechstein
Fault


SW
NE
Two-way travel time (s)
0
1
2
3
4
0
20 km
Cretaceous
Jurassic
Triassic
Zechstein
Sub-Zechstein
Fault


W
E
Glückstadt Graben
Mecklenburg Block
Gorzow Block
Mid-Polish Trough
East European Platform
Depth (km)
0
1
2
3
4
5
6
7
8
9
10
0
40 km