The 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 orogeny by Kockel (2003).
Chapter 3 — Tectonic evolution

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 Early European Continents during the Caledonian and Variscan orogenic cycles (Ziegler, 1982a, 1989). High-precision radiometric techniques have revealed that such ‘orogens’ comprised several deformation phases, including the closure of orogenic basins, terrane docking or ‘soft collision’, post-collisional shortening and post-orogenic collapse (e.g. see Krawczyk et al., 2000b; Krijnen, 2006). Some of these phases overlapped, particularly during Palaeozoic times when rifting of Gondwana proceeded almost uninterrupted. Following the Variscan Orogeny, terranes to Cretaceous basin development was punctuated by phases of crustal extension and subsidence associated with breakup of the Panga Supercontinent. In Late Cretaceous to Tertiary times, pulses of intraplate compressive relation compared to the Alpine Orogeny resulted in widespread basin inversion. These tectonic phases are summarised in Figure 3.1.

1.2 Palaeozoic 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 margins of the Gondwana Palaeocontinent, which lay at low northern latitudes for much of the Palaeozoic (Torsvik, 1998). The Iapetus Ocean opened during Late Precambrian times and separated Gondwana from other large relics of the Rodinia-Pannotia Supercontinent (Dalziel, 1981, 1997) such as Laurentia and Baltica (Figure 3.2a). The Gondwana-derived terranes were transported northwards, pulled by subduction of the proto-Pacific Ocean by the now oceanic basement hosting them. Accretion to the margin of Baltica (Pharaoh, 1999; Pharaoh et al., 2006) was followed by strike-slip displacements along major crustal lineaments (Reimold & Poprawa, 2002). Further Gondwana-derived terranes were accreted to Laurussia during the Variscan Orogeny in Late Carboniferous to Early Permian times.

1.3 Permo-Carboniferous magmatic-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 stabilised at the end of the Variscan Orogeny by wrench-induced collapse and widespread alkaline and calc-alkaline magmatism (Ziegler, 1990a; Wilson et al., 2004), accompanied by profound thermal thinning of the lithosphere that started during the late Turonian and intensified during the Senonian and Paleocene (Ziegler, 1998). The inversion movements were remarkably heterogeneous in nature, with strain localisation in narrow zones separated by unrelated regions. Deformation was tectonically decoupled by the Zechstein salt, with overall compression in the salt-cored and localized basin deformation. The latter was constrained on the north-west-trending Bohemian Trough Zone of crustal weakness (along the northern margin of the Norwegian-Baltic Basin), on the Teisseyre-Tornquist Zone (underlying the axial part of the Polish Basin), and on the southern margin of the North-Saxon Basin (Krawczyk et al., 1999; Schenk, 2004). The north-westward trend of early inverted basins and transpressional overthrust faults indicate a tectonic setting controlled by north to north-westwards directed compressional stress (Kley & Voigt, 2008).

With the closure of the Alpine Tethys Ocean during the Palaeogene, the Alpine-Carpathian Orogen entered the continent-to-continent collision phase. The SPB was located in the foreland of this orogen and was affected by the related stress fields as well as by intermittent pulses of basin inversion (Boccaletti et al., 2004). The stress fields repeatedly changed in orientation and magnitude and controlled the evolution of the European Craton. The SPB belongs to the type of intracontinental or cratonic basins or to intracontinental or cratonic basins (Trench & Torsvik, 1992) and is characterised by pronounced coupled extension and magmatism 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 actively dipping subduction zones (Figure 3.3). Closure of the Tethys Ocean (Cocks & Fortey, 1982) involved a significant dextral-oblique component (Trench & Torelli, 1992) and eventually produced the North Sea area (Figure 3.3).

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 extensional and flooding of the SPB. Closure of the Alpine Tethys Ocean started at the beginning of the Late Cretaceous with counterclockwise rotations of Africa-Antarctica with Europe (Ribeauvillé et al., 2002), where compression switched to the build-up in the Alpine foreland (Ziegler, 1998, 1999; Boccaletti et al., 2004; Kley & Voigt, 2008). Early-Late Cretaceous subsidence of the SPB area was accompanied by broad extensional faulting and was followed by the build-up of intraplate compressional stresses as indicated by the inversion of Variscan tectonic basins and initiation of backbulge block faulting. The inversion movements were remarkably heterogeneous in nature, with strain localisation in narrow zones separated by unrelated regions. Deformation was tectonically decoupled by the Zechstein salt, with overall compression in the salt-cored and localized basin deformation. The latter was constrained on the north-west-trending Bohemian Trough Zone of crustal weakness (along the northern margin of the Norwegian-Baltic Basin), on the Teisseyre-Tornquist Zone (underlying the axial part of the Polish Basin), and on the southern margin of the North-Saxon Basin (Krawczyk et al., 1999; Schenk, 2004). The north-westward trend of early inverted basins and transpressional overthrust faults indicate a tectonic setting controlled by north to north-westwards directed compressional stress (Kley & Voigt, 2008).

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2 Early Palaeozoic 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 (Torsvik, 1998) took place during the Shelveian Phase in Ashgill times (Pharaoh et al., 1995; Pharaoh, 1999; Samrojan et al., 2002) and is associated with amphibolite-facies metamorphism in the mid-Northern Sea region (Frost et al., 1981; Pharaoh et al., 1995). Some of these phases of post-orogenic collapse (e.g. see Krawczyk et al., 2008a; Kröner et al., 2008). Some of these phases resulted in widespread basin inversion. These tectonic phases are summarised in Figure 3.1.

2.2 Avalonian (Ardennian) Orogenic Phase (closure of the Iapetus Sea)

Avalonia rifted away from Gondwana during the Early Ordovician (Figure 3.1b) (Trench & Torelli, 1992), driven northwards by opening of the Rheic Ocean (Cocks & Fortey, 1982) to the north of Avalonia, as well as by subduction of the Iapetus Ocean at a number of actively dipping subduction zones (Figure 3.3). Closure of the Tethys Ocean (Cocks & Fortey, 1982) involved a significant dextral-oblique component (Trench & Torelli, 1992) and eventually produced the North Sea area (Figure 3.3).

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

A flexural foreland basin developed on the margin of Baltica (Altmann & Tjøtta, 1998; B Bahlet, 1990) following its docking with Avalonia, in response to its leading by the extruding orogen wedge and its continued convergence with the foreland. Rapid exhumation (within 10 Ma of millions) of the colliding wall at the Thrust Surface in the central North Sea area led sediment reworking towards this flexural basin (Natusewski et al., 2002), which is filled with thick Silurian (middle Ordovician) strata that reflect rapid subsidence of the flexuring Iapetus (Sliapa, 2006). This also led to high (4+)% modification of the Middle to Upper Ordovician mixed rocks of Berentia, Scania, northern Denmark and the Skagerrak (middle Ordovician). Consequently, the Iapetus Ocean was closed to the south of Europe by the late Silurian or early Devonian (see Chapter 4). The Iapetus Ocean was closed to the south of Europe by the late Silurian or early Devonian (see Chapter 4).
Denniss, as shown by the development of intracratonic Old Red Sandstone basins and the widespread granite plutons commonly seen in northern England (Ziegler, 1988; Braithwaite et al., 2002). At the same time, a first compressional phase affected the Rheo-Hercynian back-arc basin, which was followed by renewed Late Devonian extension and volcanism (Ziegler, 1988, 1990a).

3 Late Paleozoic tectonic evolution: assembly of Pangea

3.1 The margins of Laurasia and the Rheo-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 Rheo-Hercynian Basin. Middle Devonian marine intercalations are found in the UK, in the vicinity of the Central Graken, in Schleswig-Holstein, the south-eastern North Sea and southern Baltic Sea. During Givetian and Late Devonian times, the marine facies persisted in the north-west beneath the central SFB area, whereas the southern, central and north-eastern parts as far north as Bremen were occupied by marine carbonate platforms and eustatic changes on the northern shelf of the Rheo-Hercynian Basin (Figure 3.4a). A mid-Famennian regression led to the development of widespread coastal sand bars of the Condino facies and evaporitic intras in north-western Poland. Carbonate platforms developed across much of the northern SFB area during Dinantian times. In northern England, grabens with more basinward strata that are up to 600 m thick (see Chapter 6) are flanked by carbonate platforms (Ziegler & Gaetjens, 1990). Similar features may occur in the southern North Sea (Kendrick, 2000) and beyond Hiddensee and Rügen Island.

Starved aeolian basins became widespread during Dinantian to earliest Namurian times; the Broadwell and Eddle Shale formations (UK), Gevern Formation (Netherlands) and Cheker Formation (Brabant Massif) include thick cherts that are important oil and gas source rocks in the UK and Netherlands (see Chapter 6) (Ziegler & Gaetjens, 1990).

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

Gondwanas began to converge with Laurussia in a clockwise-rotational mode during the Late Devonian. Initial contacts between Gondwana and the Laurussia-derived terrane of the Variscan Orogen were established in the Danish domain by the mid-Devonian (Ziegler, 1988). Northeast subduction of the Palaeotethys Ocean was accompanied by progressive closure of the oceanic basin that had separated the different Gondwana-derived terranes. Palaeomagnetic data (Tait et al., 1997; 2000) support independent motion between distinct Moldanubic and Perucian terranes within the Alpide Terrane Assemblage until at least the Late Devonian (~370 Ma).

After the Mid-Devonian Avalonian/Lapbian compressional phase and the subsequent extensional pulse, compressional stresses built up again during the latest Devonian. The "Bohemia" tectonic pulse spans the Devonian-Carboniferous boundary. Widespread high-pressure metamorphism in the Variscides reflects closure of the oceanic basins separating the Bohemian and Saxo-Thüringian terranes. Closure of the Rheo-Hercynian oceanic basin occurred (Franke, 1995a, 2000) and, during the Devonian, the entire Rheo-Hercynian Basin was invaded by south-eastern sourced clastics and flysch fans and subsequently overprinted by the Jurassic and east Rhenish nappe (Franke, 2000). The palaeoethos of Rheo-Hercynian magmas of south-west England and central Germany contains fragments of the southern margin of Baltica Arc (McKerrow et al., 1997; Ziegler et al., 2002). However, widespread Tournaisian to early Namurian faunal assemblage activity in the Rheo-Hercynian Basin testifies to the cessation of extension prior to its final closure during the Saarland Orogenic Phase, which started during the Visean (Ziegler, 1998, Franke, 2000; Ziegler et al., 2004).

Following collision of the Variscan magmatic wedge with the passive northern margin of the Rheo-Hercynian Basin, essentially thin-skinned thrusting during late Visean to Westphalian B times accounts for about 500 km of shortening in the Variscides (Huckenbeck et al., 2000); (Figure 3.4d). The Rheo-Hercynian Tuzurina can be traced along the phegony of the SFB from the south-eastern Hiddensee via Frankfurt-am-Main to the north-eastern tip of the Black Mountains-Wyregur Zone and the River Rhine (Figure 3.5). Collision-related compressional stresses exerted on the foreland at the Visean-Namurian boundary caused disruption of the Danubian carbonate platforms (Ziegler, 1994a) and late Westphalian inversion of Carboniferous rifts on the British Isles, such as the Montrose-Dundee-Seible-Northumberland trough and the Midland Valley Graben (Franke & Gaetjens, 1998; Ziegler, 1998a; Ziegler et al., 1995).

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

Following the end-Visean disruption of the Danubian carbonate platform and the development of a regional unconformity, the Rheo-Hercynian Shelf rapidly subsided during the Namurian in response to tectonic loading by the north-westward advancing Variscan orogen. 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. Cambrian (456 Ma)
d. Woodbark (435 Ma)

Reconstructions by C. Scotese, kindly provided by UWM.

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 McElhinny et al. (1987).

The Mid-Polish Trough is an inverted late Early Permian and Mesozoic tensional basin, which is superimposed on the north-west-trending foreland basin that was filled by oxygen-deprived deltaic and mud-laying deposits that contain sporadic marine intercalations. The axis of this foreland basin moved progressively north-westwards as its internal parts became involved in the advancing fold-and-thrust belt of the Variscan Extensive. Westphalian A and B units 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., 1986; Bedig, 1988).

Stephanian sediments unconformably overlie the Westphalian in the north Netherlands and Germany (Lower Saxony Basin) and testify to the termination of thrusting activity in the Variscan Extensive at the end of Westphalian D times (Joegler, 1996b); there are also remnants of Stephanian strata in the West Netherlands Basin and Cimmerian Block 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 tectonic fault-blocks are typical in the north (Franko & Hoffmann, 1997; Brinks et al., 2000). Parts of the Variscan Extensive, which are now buried beneath a thick pile of post-Carboniferous sediments in the eastern SPB, are characterised by thin-skinned deformation along gently southerly dipping fault planes (Franko & Hoffmann, 1997). Organic-rich shales in the Cambrian, Silurian, Devonian and Lowermost Carboniferous successions are the most likely detachment horizons (Franko et al., 1990; Hoffmann et al., 1996). In contrast, the Variscan Extensive at the southern SPB margin are characterised by thick-skinned, basement-cored 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 Tarnowskie-Góry-Troppau Zone (TTZ) (cf. Bukówka & Gładysz, 1972; Potyryński & Brytowicz-Lewicki, 1974; Badzian et al., 1989; Badzian, 1991a, 1991b; Bukówka & Ruszkewicz, 2005), one of the most fundamental lithostratigraphic boundaries in Europe (Thybo et al., 2002). The TTZ is characterized by a very complex internal structure, which has been imaged on deep-seismic-refraction data (Chapter 2) (BABEL Working Group, 1993; Gnad et al., 2002; Badzian et al., 2005; Gnad & Gutberch, 2000a, 2000b). The north-eastern boundary of the TTZ generally coincides with the north-western boundary of the East European Craton (cf. Gołębiewska et al., 1986; Grabowska & Bediak, 2001; Gnad et al., 2002a, 2002b, 2004a; Potyryński & Rumiński, 2000; Ełkowska & Pancerz, 1997, 2002) whereas its south-western boundary lies within the so-called Trans-European Suture Zone (see e.g. Kryn 1998). This is a wide zone encompassing the fronts of the Caledonian and Variscan orogens and various tectonic zones, some of which are only hypothetical, (for details see Thybo et al., 1984; Alenius & Thybo, 1999; 2000; Painzani, 1999; Gnad et al., 2002a; Muniz & Janowczik, 2001; Painzani & Papanicolaou, 2004; McCann et al., 2003; McCann, 2005b; Painzani et al. 2006) suggested that the TTZ may have originated at the rifted margin of Baltica.
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During the late Neoproterozoic breakup of the Rodinia-Pannotia Supercontinent (Dalziel, 1997), seafloor spreading indicated that the TTZ and its north-westward continuation, the Sorgenfrei-Tornquist Zone (TTZ) that transects the Precambrian crust, was certainly active by Stephanian to Early Permian times (Poulsen et al., 1990a; Ziegler, 1990a; Thybo, 2000). The geometry of these axes is consistent with down-to-southwest displacement of up to 20 km (Ziegler, 1990a). During these displacements, the Ranae Fault 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. Kunstmann, 2007). The Upright–Kamzisz-Liborz–Zamost Fault Zone (Figure 3.5a), the present-day north-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, Ringpoting–Kip and Men highs, which developed during Stephanian to Early Permian times (Ziegler, 1990a). Here, the distinctive characteristics of the crust are the result of emplacement of Carboniferous granites adjacent to the Thørvæ Bank. The trend of highs is interrupted by the Central and North grabens and Berde Trough. Lower Bathonian volcanic rocks have been encountered by wells both inside and outside these troughs, supporting the idea that they became significant passages through the highs only during the Triassic (Best et al., 1983; Tyler, 1990; Ziegler, 1990a).

The Permo-Triassic basin of eastern England lies within the Visean foreland and comprises a number of channel-shaped Carboniferous sub-basins, for example the Widmouth, Gainsborough and Stamford 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 accompanied by faulting on early north-south trends (Figure 3.6a). The Eakring Fault (Figure 3.6b) shows Dinantian syndepositional growth of faulting of the Dinantian thickness across the structure. It is a down-to-southwest faulting trend that developed during Late Carboniferous to Early Permian times that concentrated on faults with a north-western trend (Figure 3.6b). Taking this into account, the base-Dinantian thrust may have extended 750 m. The base of the Permian-Triassic is offset by a normal throw of about 80 m; a very clear example of the strong influence of inherited (in
Figure 3.4: Seismic sections: a. central frontier and purgatory cover by the Variscan basement and Variscan-Carboniferous magnetic structures; the diagram shows several examples of depocentres of the Carboniferous foreland basin and one scene example of the Variscan Front. The examples are mostly from the Kuirs and Western North Sea (Clausen Bank High), but are typical for much of the Carboniferous foreland basin (see for further reading Bartholomew et al., 1993; Adolfsen & Wrisley, 1981; B竹hinger & De Jong, 1983; De Jong, 1981; B竹hinger & Palevsky, 1983; Ceyneld et al., 1986; Maynard et al., 1987; De Jong & Karsten; 1987; B竹hinger, 1980; Maynard & Durieux, 1989; Smit, 1989; G鰀tch et al., 2002; Smit, & De Jong, 2005; in progress).  

This figure is interpreted for the analysis fault with large laterally: the abse of the Pennine (uppermost). A seismic section across the SE margin of the Lublin Basin (Krzywiec, 2007a, 2009).  

The section across the SE margin of the Lublin Basin (Krzywiec, 2007a, 2009) shows a number of small-scale structures, such as a. Seismic section across the SE margin of the Lublin Basin (Krzywiec, 2007a, 2009)  

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


d. Seismic section from block 53, Danish offshore sector. The section shows a well-segmented irregular unconformity at the base of the Permian. Faulted, westward-dipping Upper Carboniferous strata with distinct continuous high-amplitude seismic units are truncated below the base of the upper Rotliegend Group. The Base Permian (Upper) (Unconformity here represents a time gap of 40 Ma). Deposition of the Carboniferous strata took place mostly during the latest Carboniferous and Early Permian phase of wrench tectonics (Ziegler, 1990b, 1990c). Downdip & De Jong, 1983; De Jong, 1981; G鰀tch et al., 2002; Smit, & De Jong, 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 Clausen Bank High. The Early fault closely resembles after deconstruction of the Coal Measures (Krzywiec et al., 2007). The off-axis of this graben, drifted by wall (less).  

The Early fault closely resembles after deconstruction of the Coal Measures (Krzywiec et al., 2007). The off-axis of this graben, drifted by wall (less).  


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

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

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


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

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

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


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


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

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

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

This figure shows a number of small-scale structures, such as:

- a. Seismic section across the SE margin of the Lublin Basin (Krzywiec, 2007a, 2009)
- b. Location of the Pennine Basin (Duchamp & Evans, 2005)
- c. Seismic section and interpretation for the analysis fault with large laterally: the abse of the Pennine (uppermost)
- d. Seismic section from block 53, Danish offshore sector. The section shows a well-segmented irregular unconformity at the base of the Permian. Faulted, westward-dipping Upper Carboniferous strata with distinct continuous high-amplitude seismic units are truncated below the base of the upper Rotliegend Group. The Base Permian (Upper) (Unconformity here represents a time gap of 40 Ma). Deposition of the Carboniferous strata took place mostly during the latest Carboniferous and Early Permian phase of wrench tectonics (Ziegler, 1990b, 1990c). Downdip & De Jong, 1983; De Jong, 1981; G鰀tch et al., 2002; Smit, & De Jong, 2003).

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

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


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


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

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

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 Coal Measures. The major Late Permian to Mesozoic basins in the southern North Sea are controlled by faults with a north-westly trend, such as the Drenthe-South Hewett Fault System and Lower Rhine Lowstand, which probably represent reactivation of the Caledonian basement grain (Pharaoh et al., 1991, 1997, 1999).  

The front of the Variscan thrust-and-fold belt runs approximately eastwards through Belgium, past south of the Netherlands, and then north-eastwards into Germany (Ziegler, 1984a) missing the northern Moosblatt Block. The front turns eastwards in the Bromel-Golsberg area and has been encountered south of Luneburg and east of the River Elbe (Pottrell, J. well), then crosses the River Elbe to the south of Szczecin in Poland (Krawczyk et al., 2008). 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 interpreted seawards into the nappe pile (Brockes et al., 2003). In general, the transition from the Variscan to the Caledonian basement is poorly known. 3D seismic data show that folding is locally absent; wrench faults with linear structures partition the Variscan foreland into a number of blocks (Loh et al., 2007; E. Schütt, pers. commun).
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...Stephanian to Early Permian nows-induced deformation caused deep disruptions 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 transitional basins in the southern North Sea and northern Germany suggest that transtensional deformation controlled the locally deep truncation of Westphalian and upper Stephanian 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 SFB to the north of the Variscan Deformation Front were occupied by a Stephanian Foreland Basin, as indicated by the subcrop pattern of Carboniferous rocks at the Variscan Unconformity.

4 Late Palaeozoic tectonic evolution: post-orogenic collapse and Permo-Carboniferous magnetism

4.1 Collapse of the Variscan Orogen

During late Carboniferous and Early Permian times, a dextral mega-shear system, which connected ongoing crustal shortening in the Urals and the Appalachians (Katsanevakis et al., 2001) and transected the Variscan Orogen and controlled its collapse (Arthaud & Matte, 1977; Ziegler, 1982a, 1988, 1989). The Late Carboniferous palaeo-stress pattern was characterized by north-south compression, which activated north-west-trending dextral strike-slip faults and subordinate north-south east-striking normal faults (Katsanevakis, 1998; Lamanthe et al., 2002). Plate-motion modelling suggests that post-orogenic collapse of the Variscan Orogen alone can not reproduce the observed timing and magnitude of Permo-Carboniferous crustal thinning, and that additional tectonic plate-boundary forces are required (Brink, 1997). The Late Permian to Early Jurassic tectonic subsidence curves for different parts of the SFB indicate a phase of Permo-Carboniferous tectonic subsidence that was followed by asymmetrically decreasing tectonic subsidence (Ziegler, 1998). However, a purely extensional model for the SFB is extremely problematic because active faulting during this time is minor and only occurs in restricted areas of well-defined
In post-Carboniferous times, the area of western Europe became affected by late-Viennese post-orogenic tectonics. Witness-faulting associated with intrusives and extensional magmatism and thermal uplift caused widespread erosion, particularly in the north-west–south-east trending Axial Ranges (Ziegler, 1990b; Ziegler et al., 1994). Broad north-west–south-east trending swells formed, which can be traced on the subcrop map of Metahyaline units at the base of the Basin (Figure 3.1). The subcrop pattern of the Metahyaline against the BPF already closely shows the shape of the Mesozoic basin as a series of lens uplift and cones. The SPF varies in age throughout the SPF (Galak et al., 1990), and is in fact an amalgamation of several unconformities (Glenia, 1986). (See Section 2.1.5. in Chapter 7).

The崤nian Overdeepening (Stilla, 1982), which separates the volcanics and sediments of the Rifted Altmann Subgroup and the Mittel Rücken sediments in eastern Germany is of Early Permian age. The Altmann-10 exploration well within the Rifted Altmann series (Hoffmann et al., 1989) is an mid- to Late Permian (Figure 3.1).

In the Lower Rotliegend times, the area of western Europe became affected by late-Viennese post-orogenic tectonics. Witness-faulting associated with intrusives and extensional magmatism and thermal uplift caused widespread erosion, particularly in the north-west–south-east trending Axial Ranges (Ziegler, 1990b; Ziegler et al., 1994). Broad north-west–south-east trending swells formed, which can be traced on the subcrop map of Metahyaline units at the base of the Basin (Figure 3.1). The subcrop pattern of the Metahyaline against the BPF already closely shows the shape of the Mesozoic basin as a series of lens uplift and cones. The SPF varies in age throughout the SPF (Galak et al., 1990), and is in fact an amalgamation of several unconformities (Glenia, 1986). (See Section 2.1.5. in Chapter 7).

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structures in northern Germany (Gast, 1988, Pleus, 1995) and the Olistograben 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-magnetic cycle, the SPB area now started to subside in response to thermal re-equilibration of the lithosphere-asthenosphere system (Yao Wei et al., 2005; Ziegler et al., 2006). During this process, upper Rotliegend sediments up to 2000 m thick accumulated in the axial parts of the North German Basin (Pins, 1995; Bachmann & Hoffmann, 1997; Schlich & Ries, 1991) and about 1600 m in the Polish Basin (Bjellin et al., 1995; Kiersnowski et al., 1995; Scheck-Wenderoth, 1998; Scheck-Wenderoth & Lamarche, 2005; Kiersnowski & Buniak, 2006). The trend of the Miel North Sea Basin, Romanghai Fan and Nien hugs separation the SPB area from the Northern Permian Basin (Norwegian-Danish Basin) (Figure 3.8) in which Rotliegend clastics were up to 600 m thick (Ziegler, 1994a; Lokhorst et al., 1998). In the northern Hessian Depression, between Göttingen in the south and Hamburg in the north, Late Permian clastics and sediments were deposited in a fan-shaped system of grabens (Figure 3.8) (Gast, 1988). The basin depocentre migrated southwest into West Mecklenburg during deposition of the Marine Formation, salt was deposited in the northern Schleswig-Holstein, northern Lüneburg and eastern North Sea areas. The Parchim graben system was covered by nearshore, fluvial and sheet-flow sediments (Rabe et al., 2001). According to F. Koelsch (pers. comm., 2003), north-west-trending sediments such as the Mahr, Rosenburg-Oberrhein and Schwerin-Altvulkan Fault systems which are partly derived of Rotliegend clastics, and rifts such as the Aurich-Borglück and Braunschweig-Gifhorn areas, may have compartmentalized the north-west-trending SPB. However P. Ziegler (pers. comm., 2005) and K. Schulte (pers. comm., 2003) consider this very unlikely as the compartments are not reflected in the Zechariah facies pattern. The North German Basin gradually assumed an asymmetric saucer-shaped geometry with a deeper northern flank and a gentler southern flank (Ziegler, 1994d), the margins of which were progressively overstepped by younger parts of the sequence (Pins, 1978; Hoffmann et al., 1995). The North German Basin depocentre is slightly offset to the north of the area of greatest Early Permian magnetic activity and maximum crustal thinning (Pleus, 1978; Bachmann & Gros, 1989; Frankel et al., 1990; Bachmann & Hoffmann, 1997), the area where the influence of the Baltier Shield occurs (ZIEGELD-B ASI Research Group, 1998, 1999).

5.1.2 Southern North Sea Basin

The Southern North Sea Basin province comprises several sub-basins with distinct sub-basins and inversion histories throughout Permian and Mesozoic times (De Jager, 2007), although these elements were not strongly differentiated in mid-Permian times (Figure 5.3). 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 Permo-Triassic relationships in the South German Sea and the Lüderitz-Rekken Plume formed persistent highs that inhibited spreading 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, 1994a, 1997a) and was accompanied by marine faulting (Van Mijlert et al., 1990; Grönvold & Borgen, 1983). In the Netherlands, the development of the Central Graben, Buskamp Basin and Lower Trough are reflected in the pattern of Rotliegend compaction (De Jager, 2007) (see Figure 3.7). According to H. Wijffels (pers. comm., 2000), the Lower Trough has a Rotliegend succession several tens of meters thicker than that of the adjacent Flevoland Platform and Groningen High (De Jager, 2007). It is likely that the major faults defining the Onslow Trough and the Harum and IJsselmonde fault zones were already in existence in Early Permian times (De Jager, 2007); the Teal-Dunkirk, Coalball Bank and Elbow Spur highs were present but poorly defined (Van Hoorn, 1987; NITG-TNO, 2004). Thick marine (see Chapter 7, Figure 7.1) show that Rotliegend strata are up to three times thicker in the Sole Pit Basin compared to surrounding areas. The Dressing-Brooke Hewett Fault Zone, which was active during the Carboniferous (Geevor & Blackman, 1990) and possibly earlier times (Piehlo et al., 1995), may have exerted a syndepositional control. Rotliegend strata were not deposited on the Mid North Sea and Ringhals-Fyn highs, as indicated by their relay geometry and that the Zechariah Series. They are also missing from the Llandovery-Mississippian, which apparently acted as a source area for the Zechariah series in the North Sea.

5.2 Polish Basin

During the Pennsylvanian stage of its evolution, the eastern part of the Polish Basin, the Mid-Polish Trough, formed the easternmost part of the SPB (Kiersnowski et al., 1995; Grzybowski & Van Man, 2000; Kiersnowski & Buniak, 2006; McCann et al., 2006). With very few exceptions (e.g. Autonomin et al., 1996), there is a lack of reliable seismic information on the basin’s internal structure at sub-Zechariah levels. Only chronostratigraphic 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-Zechariah basement tectonics shows a complex array of north-west-striking fault zones (e.g. Kiersnowski et al., 2000; Kiersnowski, 2003), which were repeatedly reactivated during Jurassic subsidence and subsequent inversion of the trough. Mid transpressional activity along fault systems of the Tarnobrzeg-Torzym 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 Zechariah Sea. Asteroid-chaîne sequences were deposited in ephemeral-lake or sabkha environments in the central Polish Rotliegend Basin (Poborsky & Wagen, 1975).
5.3 Late Permian (Zechstein) tectonic evolution

The Late Permian plate-tectonic setting of Europe (Figure 3.10) was dominated by the ongoing northwest subduction of the Paleotethys Ocean beneath the southern margin of Europe (Shanmugam & Brotz, 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, 1990a, 1990b; Trouailh et al., 2002; Gnoeddeveld et al., 2003). This transgression probably advanced the margins of the Northern and Southern Permian basins through the Faroe-Shetland Trough, the fault-controlled basins of the Irish Sea and the Slesvig-Yale of Japan 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) evaporitic banks moved west–east across the grabens (P. Ziegler, pers. comm., 2006).

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 SPM during late Rißelvik times have been recorded (Ziegler & Schmitz, 2000), 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 basin 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 (Faul, 2000).

Late Permian evolution of the SPM continued to be dominated by thermal relaxation of the lithosphere (Duduk et al., 1999; van Wees et al., 2002; Schenk et al., 2003; Schenk-Wendtner et al., 2004) 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 basin areas were starved of sediment; basin areas were filled by thick halite during sea-level lowstands. Marine flooding along west–east extension is reported from the southern margins of the SPM in the Netherlands and Germany (Ziegler, 1990a; Gelski & Böhlig, 1997; Gelski et al., 2004a). Post-depositional salt mobilization marks pre- and synsedimentary faulting in basin areas (Schenk-Wendtner et al., 2003). Fault system development of the Zechstein and its relay into the Mid North Sea and Rüngshahls-Pyoyn Men high indicates that early extensional faulting in the Northern and Southern Permian basins (Pastiak, 1997; Taylor, 1998; Claussen & Pöslener, 1999). In southern Germany and Poland, the Zechstein Sea intercontinentally inundated only the largest of the relatively slowly subsiding intramontane troughs (Ziegler, 1990a; Freudenberger, 1994; Dallad et al., 1999). Differences observed in Late Permian subsidence rates between both parts of the SPM and its flanking areas are attributed to differences in the magnitude of thermal destabilization of the lithosphere during the Permo-Carboniferous tectono-magnetic cycle (Wilson et al., 2002; Ziegler et al., 2006).

Cyclic erosion led to the deposition of the more than 1500 m-thick halite-dominated Zechstein sequence in the salt part of the SPM area in northern Germany (Ziegler, 1990a). The northern margin of the basin was rather straight along the northern slope of the Rüngshahls-Pyoyn Men high (Figure 3.11). In the south, the basin margin was differentiated into several north–north-west-trending swells and lows, such as the Emn Lohe, the Butte Trough and the Schachfeld-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 faster developments. The Z2 (Staunton) cycle deposited the thickest sequence of evaporites; later cycles overstepped farther onto the margins of the basin, as recognized in Denmark, and as many as seven cycles have been recognized in the basin axis (Ziegler, 1990a; see Chapter 6). Due to later tectonic movements, little is known about contemporaneous sinking. West–north-west-trending grabens are reported from the Z2 (Rüngshahls-Pyoyn Men high) (Schuldt, 1993) in the eastern Netherlands (Gelski, 2005) and the Lower Rhine Embayment, as well as the Z2 cycle on the Butte Swell. A later (Tubantian II) phase of faulting predates the Z4 (Aller) cycle in the Netherlands (Geluk, 2005) and Germany (F. Kordel, pers. comm., 2005). The fill of the Mid-Polish Trough includes a thick Zechstein salt layer within the north-western and central segments of the basins (Harms & Pöslener, 1997; Dallad et al., 1999). As elsewhere in the SPM, the tectonic motif was one of thermal instability and broad overstepping of the Rüngshahls-Pyoyn Men basin margins (P. Ziegler, pers. comm., 2006).

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

6.1 Early Triassic rifting and subsidence

Rifting intercalated between Greenland and Scandeswacia during the Early Triassic (Figure 3.12) and propagated into the North Sea as well as the North Atlantic domain (Ziegler, 1988, 1990b; Robert et al., 1995). The North Sea rift system, connecting the Viking and Central grabens and the Words 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.

[Diagram of Triassic tectonic evolution]
of Pennsylvania to the north and the Anglo-Brabant and Variscan margins to the south. The northward drift of the SPB to about 35°N by Early Triassic times (Figure 3.13b) was the result of Triassic–Cretaceous 60° counter-clockwise rotation of Pangaea. Continued northward subsidence of the Paleotethys Ocean basin was accompanied by the northward drift of the Cimmerian Superterrane, which had collided in the northern margins of Gondwana during late Early Permian times. Bal-bak of the Paleotethys subsidence zone resulted in back-arc rifting and the opening of the Early Triassic Eisle (Molasse) basins (Stampfli & Borel, 2002) (Figure 3.14e).

The tectonic development and thickness of the Rhaetian translatent series (see Figure 3.1) are evidence that thermal subsidence of the SPB persisted during the Early Triassic (Jager, 1990a; Geib, 2005). However, its broad source-shaped subsidence pattern was interrupted by the development of the northernly trending Central, North and Gluckstadt 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 (Steinhaff & Strauss, 1984; Kreuzer & Knepper, 2002; Mayntz et al., 2005; Schuck et al., 2005a), on which Buntsandstein and Muschelkalk reflections are apparently unaffected by neotectonic faulting (e.g. see Figure 3.14). Contemporary tectonic activity is manifested in north-west-trending structures parallel to the main basement faults.

The so-called ‘Hardhege Unconformity’, which affects the margins of the North German Basin and for which a tectonic origin is often proposed (Beulier & Schüller, 1987; Pöhling, 1991), may be the result of deformation of the crust in response to the build-up of intraplate stress (Geib, 1990b). The Londen-Brabant and Bohemian margins remained important sources of clastic detritus, whereas the Rheinland Haard gradually subsided beyond the eustatic base level, mainly in response to thermal re-equilibration of the epeiric ammonite-limestone system (Jager et al., 2004).

6.1.1 North German Basin

The Buntsandstein series is up to 500 m thick in the central North German Basin (see Figure 6.5) (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 northern margins of the SPB are reflected in the thickness variations of the Buntsandstein series. These swells and troughs were more pronounced during the Buntsandstein time, including the Netherlands Swell, the Borkum High, the Eem Lagoon, the Dutch Shelf, the Solingen and West troughs, the Schelphol–Aldeneik Swell, the Thuringian and West Brandenburg basins and the East Brandenburg High (Figure 3.13a). Pulsed uplift of the swell caused erosion on their crests, which is evident in the Buntsandstein facies trends (Eckhardt, 1986; Ziegler, 1990a). During the more significant basin-Hardhege unconformity (Wolburg, 1982; Schüller, 1988), erosion cut down to Lower Buntsandstein or even Zechstein levels (Wolburg, 1983; Baldschuhn et al., 1996; Pöhling, 1991). This unconformity is not as well expressed in the intervening troughs. However, there is no clear expression of such shoulders, suggesting that the causal mechanism is unrelated to tectonics on individual graben systems. It is more likely that the build-up of regional terminal stresses gave rise to broadscale warping of the crust and lithosphere (Geib, 1990a). In the centre of the basin, north-north-east to west-north-west-trending grabens started to subside during deposition of the Rhaetian translatent series, namely the Eem–Horn and Glückstadt grabens, the Nieder-Bremer Becken–Trough, the Krummenberg–Bremer Becken–Trough, and the Alster–Kamener–Bremer graben systems (Figure 3.14a). Some of the west-north-west-oriented basement trends also became active, such as the Alster–Helmian lineaments (Figure 3.13b). Subsequent (Late Triassic) structure was more locally intense that the Buntsandstein erosive sequence was torn apart into ramps and the Grabens became tectonically eroded in areas such as the Eem–Horn and Nieder-Bremer Becken grabens and the Alster Lineament. The incision pattern between the Solingen Formation (Hardhege Unconformity) clearly illustrates the structural differentiation of the SPB (Geib, 2001; see Chapter 9).

6.1.2 Southern North Sea Basin

During the Buntsandstein 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 (NGI-TRA, 2004). The lower-upper grabens of the Main Buntsandstein Formation overlie the fine-grained Lower Buntsandstein clastics. The predominantly sandy Solingen Formation overlies earlier Triassic strata above the Hardhege Unconformity. As no uplift of the grabens 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 Frentham 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 domed structural style in pre-Pennsian strata. The Lower Aal Beach High, Amundsen Block and Schill Grand High formed intermediate platforms (Figure 3.13a) between the grabens and persisting highs such as the London-Brabant High. The latter was bordered by rivers flowing northwards from Armorica in the Wessex, Campine and Central Netherlands basins (see Chapter 9).

6.2 Polish Basin

Rapid subsidence of the Polish Basin during Zechstein times is attributed to a distinct extensional pulse superimposed on its thermal subsidence (Dallmann et al., 1974; Stepnowski et al., 2002). Salt movements in the Mid-Polish Trough were initiated during the Early Triassic, significantly modifying local subsidence patterns (cf. Eyskens 2004a, 2009, 2006a). The Mid-Polish Trough, Norwegian-Danish Basin and the cooling Basen Group all began to subside rapidly, implying linked reactivation of the Teutoburg and Sorgenfrei segments of the Tornquist Zone (Ziegler, 1990a). Buntsandstein strata are several hundred metres thick in the Norwegian-Danish Basin (Figure 3.13c, d) (Eyskens et al., 1990; Cacas and Pedersen, 1988). The thickened Buntsandstein is observed in the northern part of the Mid-Polish Trough, where Upper Permian sediments are up to 50 m thick, enhanced during the Triassic (Pomykay et al., 1987). Synsedimentary fault movements were restricted to north-west trending structures (Eyskens, 1990; Eyskens et al., 2004a, 2006a), parallel to the main inferred basement fault systems (Eyskens et al., 2006a). Subsidence is attributed to the Triassic evaporation of the Baltic Sea (Eyskens et al., 2006a). In the Polish offshore sector, thickening of the Triassic succession towards a basement-fault zone indicates localised extension and subsidence (Figure 6.2.1) at the northern end of the Mid-Polish Trough. The Permian succession of the Alpine sector underlain by thick Zechstein salt, faults in the Triassic are decoupled from those below the thick East-north-east-trending subsidence axes in central and southern Germany, such as the Thüringian Basin, principal rift structures, the Horn and Glückstadt grabens and the Emsland Trough (Westdorf graben). These structures were evident during deposition of the Grabfeld and Weser formations (Lower and Upper Gipskeuper), when separated by a period of tectonic quiescence during the deposition of the Stuttgart Formations (Schludermanns) (see Chapter 9). Both extensional pulses affected the Horn and Glückstadt grabens and several major north-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 flexural collapse. Displacement of Permian salts into the Mesozoic overburden started during Late Triassic times; salt diapirs continuously erode basement faults and so mark the border faults of grabens. Nin-synclines, which subsided in response to salt migration into diapirs, can contain very thick Kapfer strata making it difficult to assess the magnitude of tectonic subsidence. About 70% of the known salt domes in northern Germany entered the diapirc phase during mid-Jurassic times. The first structural trends had evolved in Rotliegend and basal Jurassic reservoirs and had probably been formed from Paleozoic source rocks.

Seismic-reflection data show that widespread synsedimentary normal faulting and salt mobilisation took place in the Central Graben (Michaelsen et al., 1985, 1993; Sembler and Meynen, 1993) and Horn Graben (Best et al., 1981; Sembler and Meynen, 1993) during the Late Triassic, although this is poorly constrained. Differential subsidence continued in the principal rift structures, the Horn and Glückstadt grabens and the Emsland Trough (Weser 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 eastern Dutch Central Graben and Broad Fourteenth Basin (see Figure 6.5). Development of salt walls at the bounding faults of the Central Graben suggests that downfaulting was well underway in the Triassic times (Remmelts, 1985, 1984). 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 wells drilled at this time (see Figure 9.5) piercing salt domes and kin-synclines developed later (see e.g. Jepsen, 2007). Only minor thickness variations are visible across faults in the West Netherlands Basin to the south of the Zechstein Basin.

Connections with the Tethyan domain was established via the East Carpathian, Silurian–Mississippian area and the Saxon–Moravian synclinorium (formally referred to as the Western Ganges by Sbar et al. (2008)) (see Chapter 9). Progressive basin faulting in the central (Kazanian) part of the Mid-Polish Trough resulted in further salt movements envisaged by local thickness variations of the Mid-Triassic succession (Eyskens, 2004a, 2006a). Within the southern (Ostsee Flach Mass) segment of the Mid-Polish Trough, basin faulting along the Nowy-Most-Hol Fault Zone resulted in localised Triassic sedimentation within the salt depositional regime (Hakenkamp et al., 1987; Eyskens, 2009a); (see Figures 3.22.1 and 3.42).

6.3 Late Triassic subsidence and rifting

The Late Triassic megashear setting of central and western Europe (Figure 3.15a, b) is dominated by an arcing Atlantic–North Atlantic seismotectonic system that propagated southwards into the Central Atlantic domain during the Late Triassic (cf. see Figures 3.4, 3.4a), and was followed by contemporaneous uplift of the flanks of the Norwegian–Greenland rift as shown by the increase in crustal inflows from the SPB from northern sources (Ziegler, 1988, 1989a).

b. continued northward subsidence of the Palaeokratos Ocean beneath European, contemporaneous opening of the North Sea and collision of the Boreal Tethyan and Cimmerian sequences with the Palaeokratos arc-trench system caused by closure of the Koe Basin during the early Cimmerian Orogeny (Tangberg & Brosi, 2002; Mikulic et al., 2001). The result was building up of intracontinental compressive stresses in the European Platform controlled its uplift and the westwards deformed swelling of the basin of the Early Triassic (Ziegler, 1990a). The basin of the Early Triassic was produced by dextral transtension and are anastomosing in plan with a negative flower structure in profile (Van Hoorn, 1987).

The North Sea, Horn and Glückstadt grabens remained active during the Late Triassic. Similarly, crustal extension persisted along the Sorgenfrei-Tornquist Zone. Basen and Glückstadt graben and the Mid-Polish Trough (Ziegler, 1990a; Scholl and Wandel-Weber et al., 2008). The absence of significant Triassic salt walling in the Triassic, however, suggests that its development was not associated with thermal destabilisation of the lithosphere (Ziegler, 1990a), or with major volcanic activity and the absence of regional thermal downwelling of the North Sea rift suggests that its development was not associated with thermal destabilisation of the lithosphere (Ziegler, 1990a). Important basin–fracture and base–basement contacts are mainly along the SPB margins and are attributed to intracontinental stresses operating on a lithospheric scale (Cristol, 1988; Ziegler, 1990a).

6.3.1 Northern German Basin

Late–Early–Middle Jurassic extension is reflected in the continued subsidence of the northwards-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 Formations (Schludermanns) (see Chapter 9). Both extensional pulses affected the Horn and Glückstadt grabens and several major north-west–north-east-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 flexural collapse. Displacement of Permian salts into the Mesozoic overburden started during Late Triassic times; salt diapirs continuously erode basement faults and so mark the border faults of grabens. Nin-synclines, which subsided in response to salt migration into diapirs, can contain very thick Kapfer strata making it difficult to assess the magnitude of tectonic subsidence. About 70% of the known salt domes in northern Germany entered the diapirc phase during mid-Jurassic times. The first structural trends had evolved in Rotliegend and basal Jurassic reservoirs and had probably been formed from Paleozoic source rocks.

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6.3 Late Triassic subsidence and rifting

The Late Triassic megashear setting of central and western Europe (Figure 3.15a, b) is dominated by a). a pronounced activity along the Arctic–North Atlantic seismotectonic system that propagated southwards into the Central Atlantic domain during the Late Triassic and was followed by contemporaneous uplift of the flanks of the Norwegian–Greenland rift as shown by the increase in crustal inflows from the SPB from northern sources (Ziegler, 1988, 1989a).
6.3.3 Polish Basin

Keuper sediments up to 2000 m thick were deposited in the north-west-trending Mid-Polish Trough (Czerwinski & Pajchtera, 1971), where a strong structural movement of Zechstein salt. This is documented in the Permian sequence by localized syndepositional collapse developed within the Upper Triassic succession (see Figure 3.30). The Early Cretaceous (130–90 Ma) transgression translated the Keuper sequence down to the Muschelkalk level and was locally accompanied by normal faulting (Ziegler, 1990a). Basement faulting was most intense in the central (Kujawy) of the basin and eventually led to development of the spectacular Kłodawa salt diapir. This salt structure extruded onto the basin floor (see Figure 3.30). The Kłodawa salt diapir is one of the largest salt structures in the world, and its extrusion onto the basin floor has been studied extensively (Michalak, 1998).

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

7.1 Jurassic to Early Cretaceous transgression setting of the Southern Permian Basin

Rifting in the Central Atlantic culminated in crustal separation and the onset of sea-floor spreading during Jurassic times. This was followed by northward propagation of rifting into the North Atlantic, Norwegian-Greenland Sea and Labrador-Boffin Bay domains (Figure 3.30). Moreover, rifting accelerated in the Alpine Tethyan domain in which sea-floor spreading started during Mid-Jurassic times (Ziegler, 1988; Stampfli & Beres, 2002, 2004). Extensive, the Alpine Tethyan spreading system linked up with that of the Betic-Rif basin, isolating the Carpathian and Torn Kilen and partly the Dacia Block (Schenal et al., 2004). Following collisions of the Gondwana-derived Carpathian Tethys with the southern margin of eastern Europe, which gave rise to the early Cimmerian orogeny, northward subduction of Betic-Rif was accompanied by intermittent back-arc extension and compression in the Black Sea domain. This gave rise to the Mid-Jurassic, mid-Cenomanian orogeny, and at the Cimmerian–Cretaceous transition, to the late Cenomanian orogeny (Miklusko et al., 2001; Stampfli & Beres, 2002, 2004).

The Mid-Jurassic onset of sea-floor spreading in the Alpine Tethyan Ocean underlies a fundamental reorganization of the stress field that had dominated the evolution of western and central Europe during Triassic to Mid-Jurassic times. While the relaxation of tectonic stresses in the Alpine Tethyan domain, the stress field of western and central Europe became dominated by stresses controlling the evolution of the Aztec-North Atlantic rift system and the related backarc, seaward extension of the Lower-Saxonian Armor to relative Motion (Ziegler, 1988, 1990a; Torsvik et al. 2002).

During the Early Cretaceous, the Central Atlantic Ocean sea-floor spreading axis gradually propagated northwards into the North Atlantic domain (Figure 3.30). By mid-Aptian times, crustal separation became effective between Ibiza and the Grand Banks, and in the Bay of Biscay. Ibiza, including the Aztec Biscayano Tectonic, became isolated from Europe, causing the Yacimiento to open. At the same time, activity decreased in the North Sea rift system and concentrated on the North Sea-Rhenish rift system. In the Middle Jurassic, the northern Labrador Sea-Baffin Bay rift (Ziegler, 1988; Stampfli & Beres, 2002, 2004; Schmidt et al. 2008).

7.2 Early Jurassic rifting

The North Sea rift system remained active during the Early Jurassic, as evidenced in the Viking and Dutch Central grabens. At the southern end of the North Sea rift system, crustal extension was compensated by activations of a system of west-north-west trending transtensional basins along the southern margin of the Iberian Peninsula (Figure 3.17). Similarly, tectonic activity along the Moropus–Triauton and Tramoni–Triauton transtensional basins persisted during the Early Jurassic. Continental regional thermal subsidence of the Northern and Southern Permian basins during the Rhaetian and Bettanini, combined with a latitudinal sea-level rise, controlled the development of a wide, shallow-marine basin. The basin was open to the Atlantic via the English Midlands, the continental basins of the Irish Sea and Atlantic seaboard, and to the Tethyan Sea via the Thuringian and Euxinian depressions and the Helvetic Shelf. The Viking Graben was also transposed during the Cenomanian. Glaicenses were shed into this basin, regionally widening basins from the Pontosurian Shelf, East European Platform and Bohemian Massif (Figure 3.17). Stagnant-water stratification led to the deposition of the Frasnian Stage Formation during the Tournaisian, the principal source rock for the oil provinces of the southern North Sea and northern Germany (Ziegler, 1988a).

7.2.1 North German Basin

The German part of the STB area underwent major changes during Early Jurassic times. However, these changes are difficult to assess due to Mid-Jurassic to Early Cretaceous deep-sea stratigraphic hiatus of Lower Jurassic strata, particularly in northern Germany. The Norn and Glückstadt grabens gradually became...
intranoe during the Early Jurassic (Figure 3.17), although they are still expressed on thick ramps due to continuing salt withdrawal from syn-rift episodes into the diapiric straddling the grabens margins and/or compaction-driven subsidence (see Figure 10.3). The Den Low and Brunsbüttel–Gifhorn Trough also became inactive at this time (Betz et al., 1987).

Conversely, new north-west–north-trending subsidence centres or sub-basins developed mainly along the western margin of the SPB (Figure 3.17). These include the Lower Saxony Basin (Maurer & Schröder-Winter, 2001), the Sub-Magnesian Basin (including parts of the Lower Mountain), the Prerow Basin (including the late Flöckinger High (Kunzepck et al., 1995)), the Grimmense Basin (Kunzepck & Kunzepck, 2005) and the South–west Brunsbüttel–Lübeck Basin (including the last Kunzepck Block (Figure 3.17)). Local elongated and narrow subsidence centres also developed as grabens or half-grabens above major basement–fault zones. Until Triassic times, these sub-basins did not represent individual and isolated palaeoenvironmental units with local facies development, but were covered by the Tethyan sea, as were the platform areas between. Sediment influx was strongly influenced by subsidence of these sub-basins. Some, for example the Lower Saxony and South-west Brunsbüttel–Lübeck Basins, are transpressional in origin; others, such as the Prerow or Grimmen Basins, are shelf boundary depressions bordered by narrow escarpments (Correa & Kunzepck, 2002). Transpressional 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.

7.3.2 Southern North Sea Basin

The Lower Jurassic series was deeply transected in the central North Sea during Mid- to Late Jurassic times, consequently limiting 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 Platforms Basin. Differential fault-controlled subsidence continued in the Dutch Central Graben. The Greaser Bank High, Ameland Bank and Skirll Grand High remained platforms during much of the Early Jurassic and probably accumulated sediments hundreds of metres thick. Differential subsidence of the north-west–transpressional Thule Trough, Broad Platforms and West Netherlands basins is indicated by the thickness of the Lower Jurassic series, which is about 1000 m, 500 m and 750 m thick respectively. Subsidence of the Slepe High Basin was controlled by movements along the Draugen–North Hewett and Flamborough Head fault zones (Holliday, 1985), which also caused uplift of the Humber Weights High. Subsidence of the Weald Basin was controlled by westward moving faults that represent extensional reactivation of Variscan compressional structures (Whittaker, 1985; Lake & Eames, 1987).

7.3.3 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 extension by Tournaisian times (Ziegler, 1984a). Lower Jurassic clastic sediments up to 1100 m thick were deposited without significant interruptions from the British Isles to the Bohemian Massif and the East European Platforms. Uplift of the latter (and invasion of the Denesan Basins) was paralleled by the early Cimmerian Orogeny in the Black Sea Domain (Glennie, 1986). A thickness of up to 750 m of Lower Jurassic channel and delta facies is preserved in the Norwegian-Danish Basin to the west of the SPB area. Tectonic activity along the Jylland-Penrose Zone increased during the Middle Jurassic (Berger & Bergström, 1987).

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

Uplift of the central North Sea started towards the end of the Aalenian, presumably in response to the impingement of a transtensional plate on the lithosphere, which continued during the Bajocian and Bathonian (Ziegler, 1986a; Underhill & Partington, 1993; Surlyk & Sorensen, 2001). Development of this large thermal dome (700–1000 km in radius) was transected by the Central Graben, caused broad transpression of Lower Jurassic and Triassic sediments and the development of the regional mid-Cimmerian crenulation 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 North Fjord grabens, causing delitral components to propagate into the Viking Graben, the grabens of the Atlantic miogeosyncline, and the SPB area. This resulted in blocking of the Arctic seas and the fully connected Tethys and Central Atlantic oceans (Ziegler, 1986a, 1990a). However, crustal extension across the North Sea rift system persisted during this upthrust due to strain locally to the Carboniferous level (Surlyk & Sorensen, 2001) such that there is a strong contrast between the thickness of the Bajocian to Bathonian strata in the axial part of the basin (500 m) and its flanks (1000–1500 m). Strain localization on the volcanic dome in the Central Graben, the Mid North Sea High, and local areas of uplift such as the Cleaver Bank, Broad Platforms, Wintershall and Finnebogas fields. Sediments of dolomite–carbonate facies were deposited on the East Midlands Shelf.

7.3.5 Polish and Norwegian–Danish basins

Sedimentation became restricted to the differentially subsiding axial trough (Eustatic Depression) during Aalenian to Bajocian times and the Mid-Polish Trough became isolated from other basins within the SFB. A connection to the Tethyan Ocean opened via the graben platform. A eustatic sea-level rise during the Late Bajocian resulted in widespread erosion across the basin margins (Paetkau & Schrader, 2002) such that there is a strong contrast between the thicknesses of the Bajocian to Bathonian strata in the axial part of the basin (500 m) and its flanks (1000–1500 m). Strain localization on the volcanic dome in the Central Graben, the Mid North Sea High, and local areas of uplift such as the Cleaver Bank, Broad Platforms, Wintershall and Finnebogas fields. Sediments of dolomite–carbonate facies were deposited on the East Midlands Shelf.

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 transpressional subsidence of north–west–oriented basins and transpressional uplift of normal highs along the southern SPB margins (Figure 3.18a). This resulted in the development of east–west–oriented fault systems with the development of the North Sea rift system (Ziegler, 1986a; Schrader et al., 2001).

A eustatic sea-level fall initiated at the Jurassic–Cretaceous transition, followed by a series of selected deflections of the lithosphere, led to Jurassic transtensional extension 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 or Albanian (Ziegler, 1986a; Torrens et al., 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.

7.4.2 North German Basin

North-west-oriented sub-basins along the southern SPM margin have an exo-terrestrial relationship (Figure 3.18b). The Friesland Platform of the Lower Saxony Sub-basin and Altmühlen-Brandenburg sub-basin is linked to that of the sub-basins further west in the Dutch sector.

Previously exposed areas of the SPM were gradually transgressed during the Late Jurassic, whereas underlying the late Hauterivian unconformity in the Danish Central Graben (Vejbæk & Andersen, 1987, 38 Petroleum Geological Atlas of the Southern Permian Basin Area)

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-Harzian and Altmühlen-Brandenburg sub-basins, which became deeply eroded into Lower Triassic and, locally, to Zechstein levels (Ziegler, 1990a). Metallic hydrothermal vein deposits formed in areas such as the Harz Mountains and southern Lower Saxony Basin, contemporaneous with crustal diapirs and magmatism (Boit et al., 1989). A new pulse of sub-basinal extension started in late Oxfordian times accompanied by rapid subsidence of the Lower Saxony and Sub-Harzian sub-basins (Jüttner, 1980, 1987; Bets et al., 1995; Nalpas, 1995, 1999; Scheib & Schenk, 2002). The intervening troughs and flanking highs were uplifted during transpressional deformation, whereas deep crustal fractures triggered increased open-space activity (Dick, 1998c). Metallic hydrothermal vein deposits formed in areas such as the Harz Mountains and southern Lower Saxony Basin, contemporaneous with crustal diapirs and magmatism (Boit et al., 1989). A new pulse of sub-basinal extension started in late Oxfordian times accompanied by rapid subsidence of the Lower Saxony and Sub-Harzian sub-basins (Jüttner, 1980, 1987; Bets et al., 1995; Nalpas, 1995, 1999; Scheib & Schenk, 2002).

Recent events (Ziegler, 1990a) that triggered the Main Central Trough and the Zechstein depression were the margin of the Friesland Platform, which had been little affected by mid-Cimmerian uplift. Late Jurassic to Early Cretaceous dextral transtension was the controlling mechanism for their formation (Jüttner, 1980, 1987; Bets et al., 1995; Ziegler, 1990a; Nalpas, 1995; Scheib & Schenk, 2002). The intervening troughs and flanking highs were uplifted during transpressional deformation, whereas deep crustal fractures triggered increased open-space activity (Dick, 1998c). Metallic hydrothermal vein deposits formed in areas such as the Harz Mountains and southern Lower Saxony Basin, contemporaneous with crustal diapirs and magmatism (Boit et al., 1989). A new pulse of sub-basinal extension started in late Oxfordian times accompanied by rapid subsidence of the Lower Saxony and Sub-Harzian sub-basins (Jüttner, 1980, 1987; Bets et al., 1995; Nalpas, 1995, 1999; Scheib & Schenk, 2002).

The Pompeckj Swell to the north was uplifted during the Kimmeridgian, subjected to erosion, and gradually transgressed again during the late Early Cretaceous. Correspondingly, the Late Cretaceous unconformity is evident in large areas of northern Germany (Böhles et al., 1991). 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 Krummenieder and Thüringer basins are up to 700 m thick in the Lower Saxony Basin and developed in carbonate and evaporite facies. These are overlain by Bunterian limestones and Aalenian shales and sands that are up to 500 m thick. Valanginian to Albian marine shales and deltaic sands are up to 1000 m thick; the sands are important hydrocarbon reservoirs (Ziegler, 1990a; Scheib & Schenk, 2002).

Tertiary sediments are up to 4000 m thick, and a thin, younger, Upper Jurassic sequence rests on the Triassic, whereas the Lower Cretaceous sequence in the northern Central Graben (Herngreen & Wong, 1989). In the southern graben, the provenance of sedimentation was dominated by the Central Trough. Deltaic sands and clays of the Terschelling Basin and the Schill Grund were deposited in the Lower Cretaceous and Early Cretaceous times. Volgian to Ryazanian shales are kerogenous in the northern Central Graben (Herngreen & Wong, 1989). This in turn caused the provenance of clastic sediments was the Central Trough-Broad Fracture Zone, which was uplifted during Cenomanian times. Adjacent highs such as the Fracture Zone were uplifted and eroded at the same time. The Sylt Island High formed a stable platform area in the western flank of the Central Graben. The Sylt Graben and the Terschelling Basin subsided more slowly than the Central Graben during the Late Jurassic and accumulated thicknesses. Salt walls developed along the main bounding faults of the Central Graben and in the Outer Basin Fracture (Figure 3.19a), which had been little affected by mid-Cretaceous uplift. Late Jurassic uplift of the Fracture Zone resulted in erosion down to Lower Triassic and, locally, to Zechstein levels (NITG-TNO, 2004).

Further south, the Sole Pit, West Netherlands, Central Netherlands, Rose Valley and Weldand basins (like the Lower Saxony Basin) are a trend north-west–south-east and probably developed by transtensional reactivation of pre-existing basin axes (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 Bredius Fault-Pad Between zone between the Terschelling Basin and the Sylt Island High (De Jager, 2005). During Cenomanian-Callovian, the uplifting of intervening highs such as the Broad Fracture Zone, Winterswijk and Fracture Zone highs (Figure 3.19a) shed starts into the adjacent rapidly subsiding basins. A similar tectonic regime has been suggested for the Vielwand Basin (Heremans et al., 1991), which linked the Central Graben and Lower Saxony Basin during the Krummenieder transgression. The Zechstein alkaline eruption complex (Krummenieder) developed during the late Cenomanian rifting phase; the associated thermal anomaly allowed the Late Jurassic subsidence of the Weldand Basin. In the Terschelling Basin, tectonic events were slightly delayed relative to the Dutch Central Graben; uplift occurred before the end of the Miocene 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 Fracture Zone and the Central Netherlands Basin (Figures 3.19a & b) have the same structure, started to subside rapidly during the Krummenieder and developed into progradational fault-bounded basins during the Late Jurassic and Early Cretaceous. The margins of these basins were overprinted by post-rift deposits during the Valanginian-Barremian transgression. Margin clastic wedges were deposited in the Central Netherlands Basin, while at the same time continental sedimentation took place in the Rose Valley Basins. The Zechstein Shelf is 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
strongly eroded (Late Cretaceous 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 halite. The Clearance Bank High, the Ameland Bank and Schill 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 Cretaceous rifting phases. Upper Jurassic and Lower Cretaceous eye-grabens are consequently missing from these highs. Lower Cretaceous uplifts of the Haarlemmer High led to erosion of most of its Middle Jurassic to Permian cover. The Twente-Eemnes and Winterbergen highs (Figure 3.19b) were eroded even more severely, down to the Westphalian. The thick Cenozoic group (latest Rassen to Alban) increase, 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 Clearance Bank and Winterbergen highs (Figure 3.19c). A minor basal Emsian unconformity and a stronger Late Cretaceous unconformity are evident in the Clearance Bank High (Zangerl, 1990a). The hanging wall of the Deventer-South Rewaert Fault Zone was a lens for deposition of relatively thick Lower Cretaceous strata, locally up to 3000 m thick (Cameron et al., 1982). The basal Cretaceous Spilsby Shale Formation is apparently continuous with the underlying Upper Jurassic clays (Glennie & Boegner, 1981). The Lower Cretaceous is not preserved in the axial zone of the Sole Pit Basin, but there is evidence of substantial amounts of overthrusting of these rocks in the Middle Jurassic, and partially even the entire basin (Zangerl et al., 1982). The basal Cretaceous Spilsby Shale Formation is apparently continuous with the underlying Upper Jurassic clays (Glennie & Boegner, 1981). The Lower Cretaceous is not preserved in the axial zone of the Sole Pit Basin, but there is evidence of substantial amounts of overthrusting of these rocks in the Middle Jurassic, and partially even the entire basin (Zangerl et al., 1982).

Figure 3.19b

Sedimentation was restricted to axial parts of the Mid-Polish trough during the early Cenozoic sea-level lowstand. This area of the Telychian Basin was separated from the North German Basin by a land barrier joining the Lusatia Block to the Pommerland Trough and the eastern Rhenish-Penn Highland. The basin was overdepressed by the late Cenozoic transpression, linking the Arctic and Tethyan faunal provinces until latest Tortonian times (Zangerl, 1990a). A pulse of extension-related accelerated tectonic subsidence affected the Polish Basin in Funtensee to a minor extent (Heublein et al., 1990; Steppen et al., 2003). Sedimentation was restricted to axial parts of the Mid-Polish trough during the early Cenozoic sea-level lowstand. This area of the Telychian Basin was separated from the North German Basin by a land barrier joining the Lusatia Block to the Pommerland Trough and the eastern Rhenish-Penn Highland. The basin was overdepressed by the late Cenozoic transpression, linking the Arctic and Tethyan faunal provinces until latest Tortonian times (Zangerl, 1990a). A pulse of extension-related accelerated tectonic subsidence affected the Polish Basin in Funtensee to a minor extent (Heublein et al., 1990; Steppen et al., 2003).
Chapter 3 — Tectonic evolution

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 Cenomanian when the Pyrenean subduction system was active. Crustal subduction was achieved at the same time in the Labrador Sea (Figure 3.20c). During the Late Cretaceous, crustal separation between Greenland and Europe focused on the Baffin-Faroe Trench, the Nares-Greenland rift, and on the Norwegian-Greenland Sea rift. Crustal extrusion along 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, 1990b; Teixell et al., 2002; Stampfli & Borel, 2004).

Africa–Asia converged with Eurasia in a counter-clockwise rotational motion related to Cenomanian–Turonian crustal separation between Africa and South America in the Equatorial South Atlantic (Dewey et al., 1989; Ziegler & Stampfli, 2001; Rosehauser et al., 2002). The ensuing Late Cretaceous plate interaction in the Alpine–Mediterranean domain reflects the build-up of regional compressional stresses and the deformation of the weaker elements of its continental and oceanic lithosphere. In the Alpine Foothills domain, this involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain, this involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain, which involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain, which involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain, which involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain, which involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain, which involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain, which involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain, which involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain, which involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain, which involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain, which involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain, which involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain, which involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain, which involved Cenomanian–Turonian activation of the Piemont–Penninic–Vahic subduction system along the northern margin of the Austro-Alpine–Alpine-Mediterranean domain.

The Pyrenean and Alboran subduction systems in the Iberian domain largely experience pulsating phases of inversion that continued well into the Cenozoic. The first of these pulses, the Sub-Hebridean Phase, peaked during Cretaceous times. Deformation was mainly localized on north-west–north-east striking faults of continental crustal weakness (Figure 3.21a, 3.21c, 3.21e, 3.21f, 3.21g, 3.21i, 3.21k, 3.21n) and to a lesser extent on north-west–north-east trending grabens such as the Dutch Central Graben (Figure 3.21b). Extension of the North Sea rift system continued during the Cenomanian, Turonian and early Cenozoic times. Compressional stresses related to convergence of Africa–Asia with Europe (Figure 3.20c) started to build-up during the Late Turonian, leading to reactivation of pre-existing crustal discontinuities in the SPB area.

The Sub-Hebridean inversion phase (Campanian to Maastrichtian)

The lithosphere of the SPB area was subjected to compressional intraplate stresses starting in the late Turonian. Tectonic stresses related to convergence of Africa–Asia with Europe (Figure 3.20c) started to build-up during the late Turonian, leading to reactivation of pre-existing crustal discontinuities in the SPB area.

1. The Sub-Hebridean Phase: peaked during Cretaceous times. Deformation was mainly localized on north-west–north-east striking faults of continental crustal weakness (Figures 3.21a, 3.21c, 3.21e, 3.21f, 3.21g, 3.21i, 3.21k, 3.21n) and to a lesser extent on north-west–north-east trending grabens such as the Dutch Central Graben (Figure 3.21b). Extension of the North Sea rift system continued during the Cenomanian, Turonian and early Cenozoic times. Compressional stresses related to convergence of Africa–Asia with Europe (Figure 3.20c) started to build-up during the late Turonian, leading to reactivation of pre-existing crustal discontinuities in the SPB area.

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The lithosphere of the SPB area was subjected to compressional intraplate stresses starting in the late Turonian. Tectonic stresses related to convergence of Africa–Asia with Europe (Figure 3.20c) started to build-up during the late Turonian, leading to reactivation of pre-existing crustal discontinuities in the SPB area.
example in the Holy Cross and Sudetic mountains (Lamarche et al., 1999, 2002), Ziegler (1982a) has noted a major post-palaeozoic compressive component during inversion of the central European foreland. Development of major anticline, such as the Goringen Zone of the Lower Saxony Basin and the Polish Anticline (Figures 3.21a & b), involved shortening of the thick sedimentary fill. Zechstein salts have a distinct decoupling effect as indicated by upwarping of the Mesozoic fill of the Central North Sea Graben into a broad anticline (Figures 3.36e & 3.42f), separating the West and Central Netherlands basins. The inversion axes were overstepped by the Maastrichtian to Danian Chalk Group, reflecting 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. 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 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.

### North German Basin

The North German Basin was strongly affected by interplate compressional stresses from Cretaceous to end-Campanian times (Brown, 1981; Bokhorst et al., 1985, 1991; Böse et al., 1991; Eickel, 1991; Kingsley & Vos, 2000). Seismic sections across the Lower Saxony and Altmärk-Brandenburg basins (Figures 3.3a & 3.4b) demonstrate that their inversion axes coincide with their north-west trending Late Jurassic to Early Cretaceous depocenters (Lühe et al., 2007; Schenk-Wendeler et al., 2004). The section cutting the Lower Saxony Basin clearly shows that the area throned little Zechstein salt 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 (Böse et al., 1987; Muser & Schenk-Wendeler, 2005; Lutze et al., 2007). Similar phenomena are seen in the Sub-Baltic Basin (Kloos et al., 2005; Geijer, 2004; Vos 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 show that this diapiric effect contributes to the uplift by: (1) diachronous deformation (folding and thrusting?) during the Cretaceous to Early Cenozoic and (2) uplift of the inversion structures in Late Cretaceous to Palaeocene time. Phase 1 was observed during the Turonian to Coniacian by downdropped clinals and disillusioned clinals on the basin scale (Thiele, 1999). During the Late Cretaceous, the bounding normal faults of the former basins reactivated as reverse faults during inversion (Figures 3.3a & 3.4b). The sedimentary fill within the basins and grabens was thrust over the former graben shoulders (Figures 3.22e & 22f) onto the adjacent stable uplifted Pacific Ocean to the north and North Atlantic Ocean to the north-west, forming the foreland basin (Figures 3.21 & 22). Inversion is only expressible during the latest Cretaceous and in major fault zones transecting these blocks. Sub-horizontal block thrusts extending over 4000 m occur within the Lower Saxony Basin (Figures 3.22 & 3.4b). stretching over 4000 m occur within the Lower Saxony Basin (Figures 3.22 & 3.4b).

The central Lower Saxony Basin was deeply eroded prior to the late Cretaceous transtensional movements occurred in the north-eastern part of the basin towards the Ems-Dollard Fault (Figures 3.9 & 3.22b). These transtensional movements created a broad north-west-trending depocentre (Figures 3.22c & 3.35). The area south-west of the Ems-Dollard Fault was affected by a post-Late Cretaceous to Early Palaeocene extensional movement (Figures 3.9 & 3.35). This post-Late Cretaceous to Early Palaeocene extensional movement created a broad north-west-trending depocentre (Figures 3.22c & 3.35). The area south-west of the Ems-Dollard Fault was affected by a post-Late Cretaceous to Early Palaeocene extensional movement (Figures 3.9 & 3.35). This post-Late Cretaceous to Early Palaeocene extensional movement created a broad north-west-trending depocentre (Figures 3.22c & 3.35).

The inversion in the Lower Saxony Basin is clearly demonstrated by the Late Cretaceous to Early Palaeocene extensional movements along the Ems-Dollard Fault, which affect the area south-west of the Ems-Dollard Fault (Figures 3.9 & 3.35). The area south-west of the Ems-Dollard Fault was affected by a post-Late Cretaceous to Early Palaeocene extensional movement (Figures 3.9 & 3.35). This post-Late Cretaceous to Early Palaeocene extensional movement created a broad north-west-trending depocentre (Figures 3.22c & 3.35). The area south-west of the Ems-Dollard Fault was affected by a post-Late Cretaceous to Early Palaeocene extensional movement (Figures 3.9 & 3.35).

### Southern North Sea Basin

In the southern North Sea area, early Late Cretaceous thermal subsidence was followed during the Late Cretaceous to Eocene by the development of major regional undulations in the sedimentary fill, which were upwarped during the Early Eocene (Lamarche et al., 1999; Kossow & Krawczyk, 1999; Kossow et al., 2002). In contrast, compressional deformation in the central North Sea and the western part of the British Isles was less pronounced during the Cretaceous and Early Palaeocene and was accompanied by a new phase of salt movement during which north-west-trending salt diapirs formed parallel to the uplifted blocks. Such phenomena are seen in the Sub-Baltic Basin (Kloos et al., 2005; Geijer, 2004; Vos 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. Such phenomena are seen in the Sub-Baltic Basin (Kloos et al., 2005; Geijer, 2004; Vos et al., 2004).

Two phases of inversion can be observed: (1) diachronous deformation (folding and thrusting?) during the Cretaceous to Early Cenozoic and (2) uplift of the inversion structures in Late Cretaceous to Palaeocene time. Phase 1 was observed during the Turonian to Coniacian by downdropped clinals and disillusioned clinals on the basin scale (Thiele, 1999). During the Late Cretaceous, the bounding normal faults of the former basins reactivated as reverse faults during inversion (Figures 3.3a & 3.4b). The sedimentary fill within the basins and grabens was thrust over the former graben shoulders (Figures 3.22e & 22f) onto the adjacent stable uplifted Pacific Ocean to the north and North Atlantic Ocean to the north-west, forming the foreland basin (Figures 3.21 & 22). Inversion is only expressible during the latest Cretaceous and in major fault zones transecting these blocks. Sub-horizontal block thrusts extending over 4000 m occur within the Lower Saxony Basin (Figures 3.22 & 3.4b) and the southern end of the Glückstadt Graben (Bremen Graben).
Chapter 3 — Tectonic evolution

[Various diagrams and images related to tectonic evolution, including geological cross-sections and stratigraphic columns.]

f. Cross-section across the Osning Lineament-Nordwestfalen-Lippe Swell at the southern margin of the Lower Saxony Basin. Gently dipping thrusts are detached in both Permian and Triassic salt; Basin (Kockel, 2003). The Zechstein salt is thick and there is a decollement at this level and in the Upper Buntsandstein.

e. Cross-section across the Barenburg, Sulingen, Scholen inversion structures at the northern margin of the Lower Saxony the northern step-faulted margin of the central Lower Saxony Basin during the late Early Cretaceous; Coniacian-Santonian marginal trough;

k. l. Seismic section showing the detailed structure in the vicinity of the Wassenaar field. See Figure 3.21 for locations. the inversion-related wrenching was active until mid-Paleocene times (Laramide Phase); variations and local unconformities document several stages of inversion tectonics; the first was late Turonian in age. The Triassic-Jurassic cover is characterised by localised thickness increase across this structure, suggesting that it was deposits forming the basal part of the post-tectonic succession; of structures resulting from inversion of the Mid-Polish Trough. Basement-controlled, extensional-fault 2004) suggest that there are Eocene sediments in the entire Pomeranian segment of the Mid-Polish Trough, apart from and adjacent to the diapir during the Sub-Hercynian Phase. The entire Upper Cretaceous succession is folded and truncated structure and is characterised by intraformational local unconformities, which reflects synkinematic sedimentation above and adjacent to the step during the Sub-Hercynian Phase. The entire Upper Cretaceous succession is folded and truncated across the Bowmen salt structure as well as across the entire invertible Wulff-Falrich salt, and is unconformably overlain by Jurassic deposits, suggesting a post-Maastrichtian final inversion pulse. Paleogene palynofacies maps (Prousch, 2006) suggest that there are Eocene sediments in the entire Permian segment of the Wulff-Falrich salt, pertinent to its eastward parts. This suggests that the final inversion pulse took place during the Palaeocene (Lower Eocene), with Eocene deposits forming the post-autochthonous part of the post-orogenic foredeep.

l. Seismic section across the inverted Mid-Polish Trough and the Dower salt structure. Example from the SW flank of the Permian segment of the inverted Mid-Polish Trough showing the compressionaly reactivated Dower salt structure (based on Krzywiec, 2005). The Triassic-Cretaceous section above this salt structure thins towards the Bowmen salt structure and is characterized by intraplate local unconformities, which reflects synkinematic sedimentation above and adjacent to the step during the Sub-Hercynian Phase. The entire Upper Cretaceous succession is folded and truncated across the Dower salt structure as well as across the entire invertible Wulff-Falrich salt, and is unconformably overlain by Jurassic deposits, suggesting a post-Maastrichtian final inversion pulse. Paleogene palynofacies maps (Prousch, 2006) suggest that there are Eocene sediments in the entire Permian segment of the Wulff-Falrich salt, pertinent to its eastward parts. This suggests that the final inversion pulse took place during the Palaeocene (Lower Eocene), with Eocene deposits forming the post-autochthonous part of the post-orogenic foredeep.

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 Eocene, particularly along the northern SPF margin during the Laramide pulse of interplateal compressions (Ziegler, 1990a). Significantly, this change in depositional regime did not occur during the Sub-Hercynian Phase of interplateal compressions and basin inversion, which apparently was not as intense as the Danian Phase. Related basinwide reorganisation led to the erosion of the pre-Sub-Hercynian and Laramide inversion structures and the development of the Late Palaeocene (Landen, Tornquist) and Early Eocene (Lundeström) a unconformities.

During the Cenozoic, the western SPF area was incorporated into the intramountainally subsiding North Sea thermal sag basin, as indicated by the thickness of the North Sea Group deposits (Landen, 2006a; Ziegler & Dèzes, 2007). The Cenozoic depositional history of the North Sea Basin has allowed the establishment of a detailed sequence stratigraphic framework (Lundeström, 1993, 2004). Tectonic evolution of the SPF area reflects repeated fluctuations in the magnitude and orientation of the interplateal stress field and related sub-salt compressions of pre-Cenozoic structural elements (Reichert, 2006). These changes in stress field were controlled by the interaction of the Alpine-Carpathian and Pyrenean orogenies with their northern foreland and, following earliest Eocene crustal separations between Greenland and Europe, by increasing ridge-push forces exerted by the Arctic-North Atlantic sea-floor spreading axis (Ziegler, 1998; Gillon & Gilliotte, 1999). Trenkler et al., 2002; Díez et al., 2004; Jørgensen et al., 2005). Continental palaeomagnetism indicates that the SPF area was part of the Eurasian plate during the Palaeocene, with its eastern margin along its southern flank and by the Late Miocene and Pliocene ongrowing of the Fennoscandian Shield and the Pli-Quaternary subsidence acceleration of the North Sea and North German basins (Scheck-Wenderoth & Lamarche, 2005; Steinskog et al., 2006; Ziegler & Díez, 2007). Development of the European Cenozoic rift system had only an indirect and marginal affect on the SPF area (Ziegler, 1999b). Finally, major Maestrichtian and Danian unconformities are preserved in the Lower Saxony Basin, which led Baldschun et al. (1995) and F. Ricken (pers. comm., 2007) to infer that inversion movement had ceased during
the Campanian, the youngest preserved Upper Cretaceous series. In contrast, Bates et al. (1987) made a case for strong Laramide deformation. The Pyrenean (end-Eocene) pulse caused broad uplift of the Pyrenean (end-Eocene) pulse led to further significant uplift in the West Netherlands Basin (De Jager, 1995), in the Ringkøbing-Fyn High and Central Graben in the Danish sector (Rasmussen, 2009), and in the Moerdijk-Zeelandic basins (Wijnands, 1983) of the UK sector, but apparently little in the Netherlands or Germany. Simpson et al. (1989) estimated that the North Sea 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 Latro Eocene change in convergence direction of the Adriatic indenter with the European foreland, from a northeasterly direction during the Paleocene and Eocene, to a north-westerly direction during the Eocene and Miocene (Ziegler, 1987, 1990a; Ziegler et al., 1995; Schmid & Basu, 2003; Dierd et al., 2004; Ziegler & Dierd, 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 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 barely affected the Broad Fourteens Basin (De Jager, 2007), hardly affected the Broad Fourteens Basin (Figures 3.35b). Uplift attributable to the Pyrenean and Saxon 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 graven shoulder of the Campine Basin where Zechstein salt is almost absent (Figure 3.36d e). Sub-hemispheric 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 anamorphosis, and generally appears to support the regional model of fault displacement compatible with east-west late Cenozoic.

**Aquitanian (23 Ma)**

![Aquitanian structure](image)

**Selonian (59 Ma)**

![Selonian structure](image)

**Figure 3.23 a. Palaeogeographic map for the Paleocene (Aquitanian, 23 Ma). b. Palaeogeographic reconstructions after C. Science, 1993, edited by Shell.**
The latter is a deeply eroded and broken-up basin (Figure 3.15c) in which Zechstein deposits rest on Permian to Triassic sediments. Carboniferous 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.

As a consequence of the strong inversion of the Mesozoic West and Central Netherlands basins, their Zechstein zones are at most only 500 m thick compared to 2000 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 split of the West Netherlands Basin during the Pyrenean phase formed on the Mid-Netherlands Fault Zone (Figure 3.3b), forming the Eijgendijke High (Figure 3.23) in Late Eocene times. The Vormer Trough is an asymmetric basin to the south-west of this high, which was filled with Paleocene and Eocene deposits (Nets-TNO, 2004). In the Lawkermee Trough, fault maps at the Cretaceous and Cenozoic levels indicate that both north-north-west and south-north-east trending faults were reactivated during the Cenozoic (De Jager, 2007). Paleocene strata in this basin are up to 1000 m thick; the whole Cenozoic sequence is more than 1750 m (Nets-TNO, 2004). Subsidence along the eastern and western (Hantum Fault Zone) boundary faults of the trough was accompanied by Zechstein salt flowing towards the basin margin (Nets-TNO, 2004). Reverse faults and low-angle thrusts have resulted in much more shortening than in the Boes Valley Graben, where Zechstein deposits are up to 2000 m thick (Hulsker et al., 1994). Van den Berg et al. (1996). Significant post-Zechstein subsidence of this graben was associated with propagation of the Lower Rhine rift system into the Netherlands.

The Sole Pit Basin, which was heavily affected by the Laramide inversion phase, was strongly inverted during the Tertiary Phase, involving frontal strike-slip on the Drummond-South Bennett Fault Zone (Figures 3.3a to c) (Kieneke & Boegner, 1981; Van Horne, 1987). This was followed by erosion of 200 to 500 m of Paleogene sediments.

**Data depth range criterion:**

- **A**: L, depth ≥ 5 km
- **B**: L, depth ≤ 5 km and ≥ 3 km
- **C**: L, depth ≤ 3 km

**Stress regime:**
- Normal faulting
- Strike-slip faulting
- Thrust faulting
- Unknown regime

Figure 3.17: Present-day stress field for the SPE area, displaying the present-day orientation of maximum horizontal stress (C). Symbols stand for the stress indicator, and the length of the cross represents the data quality with A being the highest quality. Extracted from the World Stress Map Database (Kienker et al., 2005). Available online at www.worldstressmap.org.
stress (Cameron et al., 1982). The same phase affected the complex Cleveland High, where according to Ziegler (1988a) this phase had a far greater effect than the preceding Laramide phase. Estimates for the total uplift of the basin Permian strata achieved by the two phases of inversion range from 1200 and 1300 m (Ras, 1973) to as much as 2500 m (Bolek & Bödker, 1987). The structure of the Silver Pit Basin is dominated by numerous north-west-trending salt pillows and walls. Widespread episodes of Zechstein salts was triggered by the Pyrenean reactivation of basement faults under a dextral transpressional regime. North-west to north-east north-west directed compression during Laramide and/or Pyrenean inversion has been deduced from seismic data in the northern North Sea (Boyle et al., 1991), the Weald Basin (Whittaker, 1985; Hansen et al., 2002), and also from palaeo-strain studies in the Chalk Group of south-east England and north-east France (Vandenberg & Bergeert, 2001). The structural effects of the Zechstein Phase were significant in the Weald Basin, for example, in the Isle of Wight (Figure 3.24) and Boy's Bank monofaults.

The different inversion events probably caused significant remigration of gas in the Rotliegend reservoirs (Figure 3.19). The different inversion events probably caused significant remigration of gas in the Rotliegend reservoirs (Figure 3.19). The different inversion events probably caused significant remigration of gas in the Rotliegend reservoirs (Figure 3.19). The different inversion events probably caused significant remigration of gas in the Rotliegend reservoirs (Figure 3.19). The different inversion events probably caused significant remigration of gas in the Rotliegend reservoirs (Figure 3.19).

Section 3.2.2 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 Jura to Early Cretaceous transpressional 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). Displacement is geographically limited, for example some salt domes to the west and east of the Blacklawian Uplift occur during 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 downwarp development along the flanks of actively growing major salt walls and salt diapirs, subsidence of the North German Basin and the North Sea Basin accelerated during Plio-Quaternary times (Schock-Weserlack & Lanzkins, 2001; Ziegler & Déeses, 2007).

Section 3.2.3 Polish Basin and Forecarpathian Border Zone

During the Cenozoic, the south-western part of the inverted Mid-Polish Basin was overridden by the Outer Carpathian Foredeep Basin in response to thrust- and slab-loaded deflection of the foreland lithosphere (Krupiewicz, 2001; Osyczynke, 2006). The Mid-Polish basin inversion anticline (Figures 3.42a & b) extends northwards to the Bohemian (Prague) and Bohemian-Lower Lusatian (1979; Bergström et al., 1990) where it links up via the inverted Rønne Graben with the Laramide structures of Scania and the Moravian-Danish Basin (Liljeström et al., 1987; Bergström & Bergström, 1987; Tornqvist et al., 1997; Petersen et al., 2003a). The Bohemian Block and Ossolineum Basin (see Figure 3.42c) were also uplifted at this time (Krupiewicz, 1979; BABE Working Group, 1993). The Zemekha Ridge was uplifted by at least 2500 m on a steep reverse fault (Bergström & 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 presently topographic relief of the Mid-Polish Anticline was reduced by erosion during the Eocene and was completely overstepped by mid-Oligocene times (Ziegler, 1990a), with no evidence for Pyrenean or Alpine inversion (Janssens et al., 2008).

Figure 3.19. Mesozoic-Cenozoic subsidence and inversion: a. Secant-strain superimposed on map showing the depth to the Hekla in north-east Europe (Cootmans, 1966). Basins uplift and subsidence sequences are indicated by circled plus and minus symbols respectively (cf. Jopson & Dalziel, 2005). Location of inverted basin structures are also shown (modified after Bolek & Bödker, 1984). Note the concentration of contractile strain in areas of crustal thickening, the contrast between the Baltic and Fennoscandian massifs, as well as in areas of crustal thinning along the NE Atlantic rifts margins. A notable exception is the area of the southeastern active foreland in Poland. The west Netherlands Basin is modified from Overeem & Van den Haak (2005). b. A. Mid-Netherlands Basin inversion, left: extensional stage for the basin inversion (transpressional and structural analysis), right: reconstructed geotectonic cross-section in Late Cretaceous times, ended down to the Lower Cretaceous stratigraphic boundary. The colour-coded surface is the base of the Jurassic derived from 3D seismic interpretation in agreement with the reconstructed geotectonic cross-section (modified after Bödker, 2004). c. Sub-surface curves and modelled solutions for the Sole Pit Basin (modified after Bödker & Bagge, 2005) and the West Netherlands Basin (modified after Bödker et al., 2000). See Figure 3.26b for locations. The West Netherlands Basin is mainly marked by Late Cretaceous to Early Cenozoic inversion (Sub-Hercynian to Laramide phases), constrained by the erosional activity in Figure 3.24d. d. 1984 inversion Array. Modified after de Jonge (2007).
The EECU was activated during the Mid- to Late Eocene and propagated southwards during the Oligocene from the Upper Rhine Graben into the Lower Rhine-Ruhr Valley (Figure 3.36a) and Rheno-Hessoan grabens (Ziegler, 1990a, 1994; Geluk et al., 1994; Díezes et al., 2004). Evolution of the EECU was marked 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 Adriatic-Hellenic Massif and partly the Bohemian Massif (Ziegler & Díezes, 2006, 2007). The Podravina Basin Fault (Figure 3.36b) has a post-salt throw of up to 1300 m (De Jager, 2003) and separates the River Valley Graben from the Peel Block to the north-east. On the south-western graben margins, bounding faults step up more gradually to the London-Brabant Massif (Geluk et al., 1994; 2002-2004). Recent earthquakes along the bounding faults of this system (Roozenbeek event) indicate that tectonic activity still occurs (Gunawarman et al., 2001; Van Balen & Berghout, 2007). Evidence for recent faulting is very rare elsewhere, except on the coastal areas of the Netherlands where faults dissect early Pleistocene deposits. Minor earthquakes have been recorded in northern Germany with hypocentres at basement faults at depths of about 8 to 17 km (Schröcker, 1986; Kaiser et al., 2005). A system of narrow and relatively shallow grabens developed on the Fehm-Rügen Monocline along the south-eastern SBB margins during the Late Eocene and Oligocene. This graben system propagated southwards into the Polish Lowlands at the end of the Oligocene and remained active until Mid-Miocene (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 3.32). Widespread Oligocene exhumation activity preceded the subsidence of the shallow Eger Graben, which ended with the mid-Mediterranean upwelling of the Bohemian Massif. The fault systems of the Bohemian Massif started to reactivate during the Late Miocene (Figure 3.28) in response to the build-up of the present-day compressional stress field. This resulted in accretion of its marginal high, such as the Thuringian-Böhmen and Branden Forest and the Spessartge, Lužicko, Sudetic and Morave-Silesian blocks, and a reorganisation of its drainage systems (Mallerob, 1986, 1987; Stad et al., 1984; Ziegler & Díezes, 2006). Recessions of Lower Oligocene marine sediments on the central Ries Mountains (König & Blumenstengel, 2005, quoted by Reicherter et al., 2006) testify to their uplift by some 100 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 times, and the Mid North Sea and Kingston Deep high were reactivated (Ziegler, 1990a; Díezes et al., 2007). Regional post-salt thermal subsidence of the North Sea Basin continued during the Paleogene and Neogene (Krohn et al., 1989; Ziegler, 1994). Progressive uplift of the Archean and the Rheno-Hessoan basins resulted in increased clastic influx into the North Sea Basin during the Miocene. Major delta systems have propagated westwards into the deeper-water North Sea Basin since Miocene times (Cameron et al., 1993). The most important delta, the Eidersand Delta (Trouw et al., 2002), developed due to gradual uplift of the Fenscosanield Shelf. Shallow-water conditions were established throughout the southern North Sea area by mid-Pliocene times (~1.7 Ma) (Gelder & Lewis, 2003). Quaternary deposits are up to 1600 m thick in the northern part of the Dutch offshore sector. Basinal curvatures show a sharp increase in the rate of subsidence during the last few million years (De Jager, 2003), considered to be the fossil response of the lithosphere to the build-up of the present-day north-north-west directed compressional stress field (Krohn et al., 1989; Van Meer & Grootes, 1995). Subsidence in the south-eastern North Sea was rather rapid, especially during the early Pleistocene, contemporaneous with rapid uplift of the Fennoscandian Shelf in the eastern flank of the North German Basin, such as the Rues, Rhein-Maas 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 Rheno-Maas High during the Cenozoic (Van Ylen-Laurec et al., 1988; Gelders & Lewis, 2003; Ziegler & Díezes, 2007).

0.4 Present-day stress field

The present-day maximum horizontal stress patterns is based largely on studies of earthquake focal mechanisms, analyses of borehole break-outs and hydraulic fracturing experiments at levels below the Zeelandia 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 with the Alpine-Carpathian Orogen (Göcke & Geidt, 1996; Geluk et al., 1994). At about the longitude of Nieuw-Amsterdam, the trajectories of the 300 kPa stress axes are generally oriented north-south, whereas a general north-eastly trend prevails east of the River Elbe. In the North Sea area, stress trajectories are predominantly to the north-west (Maretta et al., 2002; Beinzier et al., 2005; Jarosiński, 2006). No correlation between the present-day stress pattern and the direction of the major basement faults has been found (Krohn et al., 2005). Major deviations of the stress patterns occur along strong contrasts in the lithospheric structure and are minor the inherited attributes from deforming terranes and systems (Ziegler et al., 2004b). At sub-salt levels, the 300 kPa stress trajectories strongly deviate from those at sub-salt levels, indicating strong decoupling by the Zeelandia salt (Lamp & Lerbrink, 2004) and both regional and local influences (Eley et al., 2008).
Quantitative basin analysis shows shifting patterns of basin inversion through time. East of the Netherlands, fission-track analysis (e.g. Brun & Nalpas, 1996), and geological interpretation (e.g. Worum & Michon, 2005; locally in deep erosion of their sedimentary fill. Quantitative dating, structural style and magnitude of lithosphere between the East European Platform and Africa. Also suggest that the Late Cretaceous event was caused by the pinching of west central Europe's thin propone that the Sub-Hercynian event reflects the onset of Africa-Iberia-Europe convergence; Alpine collision model. Kley & Voigt (2008) present an alternative hypothesis, arguing that this interpretation (Krzywiec, 2005) have interpreted the Sub-Hercynian event as a consequence of convergence or early Worum & Michon, 2005). Until recently, most authors (e.g. Ziegler et al., 1995; Marotta et al., 2001; Dèzes et al., 2004; Ziegler & Dèzes, 2007). This is compatible with the accelerated Pliocene subsidence of the collisional system towards the west, i.e. between the Pyrenean Orogen and central Europe. The present-day stress field evolved from the late-Marine stress field, but intensified during the Pliocene (Dines et al., 2004; Ziegler & Dines, 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). 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 (Saxonian), Laramide (Palaeocene), Pyrenean (Late Eocene to Early Oligocene) and Tethyan (Late Oligocene to Early Miocene) pulses (Ziegler, 1994a; Wonen & NIchon, 2005). Until recently, most authors (e.g. Ziegler et al., 1995; Nucciat, et al., 2001; Erspamer, 2005) have interpreted the Sub-Hercynian event as a consequence of convergence or early collision of the Alpine-Carpathian Orogen with Europe’s northern margin in a classic orogenic foreland collision model. Kley & Vrugt (2005) present an alternative hypothesis, arguing that this interpretation is unlikely because the Late Cretaceous orogenic history of the two regions is incompatible, whereas the location of the developing Alpine chain on the Asian Plate in recent plate reconstructions (e.g. Stampfli & Borel, 1994a) means that it may be too far to the south of Europe across an ocean basin. They therefore propose that the Sub-Hercynian event reflects the onset of Africa-Iberia-Europe convergence, Alpine collision with northern Europe did not commence until Palaeocene or bronze times. Kley & Vrugt (2005) 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 been piecemeal or transpressitionally reconstructed Mesozoic transpressional basins, resulting locally in deep erosion of their sedimentary fill. Quantitative dating, structural style and magnitude of inversion have been established through re-evaluation of sections by zero-compaction and maturation modelling (e.g. Hillis, 1996; Petersen et al., 2003a), geochronological techniques including exhumation track analysis (e.g. Brun & Nalpas, 1996), and geophysical interpretations (e.g. Wonen & 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 Heem, 1987; Ziegler, 1994a, 1998; De Lui et al., 2001; De Jager, 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 Mesozoic lithosphere folding The compressional intraplate stress field controlling basin inversions 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 Palaeocene 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, 2001; Dèzes et al., 2004; Kley & Vrugt, 2005). Instant pulses of inversion are well-correlated with Alpine compression phases (Ziegler, 1994a, Ziegler et al., 1998) clearly demonstrating the key role of far-field compressional intraplate stress. Most of the Mesozoic tectonic basins and crustal-scale faults in the SPB area were inverted during the Late Cretaceous and Palaeocene (Scheck-Wenderoth & Lamarche, 2001; Figure 3.28a). The evolution of these basins has been characterised by repeated reactivation of their fault systems. Thermo-mechanical modelling of the lithospheric strength of the inverted basins in the SPB shows a thermally-strong lithosphere for a long period after riftling (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 weaknesses, such as faults, cause a permanent strong reduction of integrated strength values (Ziegler et al., 1995; Van Wees & Beekman, 2000), and may 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 Muir 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 few deformable areas in the northern part of the basin model correspond to an area of low observed seismicity and fast-like XHmax stress pattern. Muir 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.
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Figure 3.12: Comparative regional depth cross-sections through the SPB

- United Kingdom to Denmark
- The Netherlands
- Denmark

Key to the sections:
- Cretaceous
- Upper Jurassic
- Lower Jurassic
- Middle Jurassic
- Lower Triassic
- Upper Triassic
- Lower Cretaceous
- Salt-Zechstein
- Fault

Legend:
- North Sea Platform
- Central Graben
- South Graben
- North Sea Boundary Fault
- North European Platform

The Pomeranian segment of the Mid-Polish Basin, Germany: BASIN 9601 (Krawczyk et al., 2010).

The British segment of the Mid-Polish Basin, UK-NL: MOBIL7-8; UK-NL-DE-DK: SNST83-07.


The Danish segment of the Mid-Polish Basin, Denmark: MINDA 14 A & B.

The Polish segment of the Mid-Polish Basin, Poland: KRAKOW, BOSNIKOWSKA, et al. (2000).

The Polish segment of the Mid-Polish Basin, Poland: regional depth profile from the Pomeranian segment of the Mid-Polish Trough.

The Polish segment of the Mid-Polish Basin, Poland: regional depth profile from the Baltic segment of the Mid-Polish Trough. See Figure 3.11 for locations.
before dipping to 1.6 s TWT just NE of the Dowsing Fault Zone. A thick sequence of Triassic to Upper Jurassic strata was the middle of the Sole Pit inversion axis. Van Hoorn (1987) published an interpretation of part of the line to demonstrate profile crosses the East Midlands Shelf, the Mesozoic strata of which are little affected by faulting and thin towards The Midlands (Amethyst) Shelf and Sole Pit Inversion, onto the Indefatigable Shelf and Cleaver Bank High. The west end of the 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 Figure 3.34 Regional seismic profiles in time: UK-NL-DE-DK; SNST83-07, MOBIL 7-8 plus detail panels (e.g. after Badley Figure 3.33 Regional depth cross-sections: UK-NL-DE-DK; SNST83-07, MOBIL7-8.

b. 1987). The final inversion of the basin took place during the Late Triassic, when the area was affected by the Alpine orogenic movements. Van Hoorn (1987) estimates that up to 1000 m of uplift was achieved by these two phases of inversion combined. Pre-Frasian erosion resulted in the truncation of the warped Lower Jurassic strata along the eastern flank of the Sole Pit High to the vicinity of the Somerville 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 reflector is best developed 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 Bank Hinge Zone. The western margin of the Cleaver Bank High is complex, has preserved Jurassic strata, and exhibits evidence that up to 1500 m of uplift was achieved by these two phases of inversion combined. Pre-Pliocene erosion resulted in the basin margin. The extension of the profile (MOBIL 8) extends from the shoulder of the Central Graben towards the NE.
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 intrasimal seismicity, which particularly affected the non-oil-bearing areas (Jagor et al., 1994; Goesling et al., 1994, 2008). This change characterized fault-bounded depressions that developed in response to broad-scale tectonics (see Chapters 3 and 4). It is evident from the present-day field map that the North Sea Basin has a generally north–south trend, with the direction of tectonic activity changing from a broad-scale compressional regime to a subextensional regime, as portrayed by the NW-SE-trending fault system that developed during the Miocene. The North Sea Basin is affected by a broad-scale tectonic regime that is well documented by seismic surveys, and gave way to broad-scale lithospheric folding and intracontinental seismicity, which particularly affected the BFB (De Jager, 2007). From a regional perspective this basin represents the most complex inverted basin in the Netherlands, especially the NE margin. Details on the inversion process are given by Van Rijssel (1989) and Nalpas et al. (1995). Deposition in the north–south trending alluvial plain of the Rhine and Meuse rivers during the Early Eocene is well documented by seismic surveys, and gave way to broad-scale lithospheric folding, which is most pronounced in the onshore Variscan massif areas that were strongly weakened by thermal perturbations (Cloetingh & van Wees, 2005).

10.3 Salt movement

Salt diapirism in the northern North Sea and North German Basin accompanied the Tertiary development of large intracontinental basins, such as the Central, Rhine, and Schleswig-Holstein grabens and Rhine-Meuse Trough (Brakenhoff & Pletschen, 1988). Halokinesis structures are progressively younger towards the margins of these grabens and distinct palaeo-seals of haloclastites are seen in the Jurassic in both the Netherlands and Belgium (Jellinek, 1967, Breynecke et al., 2005). This may relate to footwall collapse of the grabens with time. In the north-western oriented basins at the southern end of the SPF margin (Oost Pits, Broad Freestones, Winiwarter, West and Central Netherlands, Lower Saxony, Altmark, Sub-Bremenian and Bremenian basins), phases of salt movement along north–west-trending faults accompanied their rapid Late Jurassic to Early Cretaceous subsidence and their Late Cretaceous and Palaeogene inversion. The axes of major salt structures in the Dutch Basin generally trend north–west–south–east. Salt movement started during the Triassic (Smolders & Blank, 1976, Krijpke, 2000a, 2004a), particularly the Late Triassic when intense basement faulting and lateral salt withdrawal of the Delfland Subgroup. Younger Cretaceous post-rift deposits are thin and cover the margins of these rift basins. Pre-existing faults were reactivated intrasubduction during the Late Eocene and Early Palaeogene inversion, with transpressional movements forming a series of prominent NNE–SSE trending flower structures. A minor sense of movement can still be detected clearly along many intracontinental faults: normal offsets occur at gentle structural levels and reverse offsets at shallower levels. Thrusting of the Upper Cretaceous chalk from the SPF towards the inverted basin, with several lateral accretionary wedges visible on seismic data, clearly shows that the flower structures were formed during the Late Cretaceous Sub-Antillean inversion. The West and Central Netherlands basins were both further uplifted and deeply eroded during the Palaeocene–Eocene inversion, leading to the present-day pattern of the Hauptgraben, best known for its Mesozoic history, is an oil structure that already existed during Miocene times (De Jager, 2007), see Chapter 6. Abnormal activity of the green zone is shown from subPermian and Permian times (see Chapters 7 and 8). The Hauptgraben 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 Sub-Antillean inversion of the Mid-Polish Trough (Krzywiec, 2002a, 2002b, 2004a, 2004b) led to extrusion of salt on the floor of the central part of the basin (Krzywiec, 2004a, 2004b). Salt movement continued throughout the Jurassic (Buckley, 1986) with further reactivation during the Sub-Bremanian and Late Cretaceous phases of the Mid-Polish Trough (Krijpke, 2000a, 2004a, 2004b, 2005a). Salt movement further continued onto the Baltic, reflected by the base-Tertiary depth map (see Chapter 12, Figure 12.1). Moreover, the Pleistocene build-up of the present-day field was associated with a reactivation of diapirs in the North Sea and particularly in other parts of the SPF, suggesting that inversion processes played an important role in the history of the basin (Boersma, 2000, 2005; Krijpke et al., 2008).

11 Regional cross-sections

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

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Figures 3.32 to 3.42
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situated below the Chalk Group and Cenozoic on the Zeeland Platform. Thick Carboniferous deposits, often in excess of Devonian and Carboniferous deposits marking the Variscan Foreland Basin infill north of the Variscan Mountains, are illustrated the relatively shallow position of Carboniferous and pre-Carboniferous rocks in the southern Netherlands and West Netherlands and Broad Fourteens basins and the bordering Zandvoort High and Noord-Holland Platform. The section 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 Kiekert & Verhoeff, 1980). The West and Central Netherlands basins are two inverted late Cretaceous rift basins. Although the onset of differential subsidence of these basins was as early as during the Permian, the wave refraction test place during the Late Jurassic. Differential movements caused in Early Cretaceous times, and from the Wadden Sea onwards the width of the basins were gradually overstepped. This process continued during most of the Cretaceous as can be seen by the breakup of Lower and Upper Cretaceous sediments onto the Zeeland Platform and the Brabant Massif. Extension of the rift basins took place during Cenomanian-Santonian and Paleocene times (see below). This inversion resulted in uplift of the basins, removing most of the Brabant Group and locally older deposits, and a contemporaneous rapid subsidence of the adjacent highs (third Chalk Group on the North-Holland Platform, Zechstein High and Vormsi Trough). The entire area was situated in the southern part of the North Sea Basin during Cretaceous times, as can be seen by the northwest dipping of deposits of the Lower, Middle and Upper Cretaceous Groups. Note that the pre-Genocarcinoid structures were mostly no longer active.

Figure 3.30: a. MONA LINE 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 LINE 2. Depth-migrated section across the Triassic-filled Horn Graben and Ringkøbing-Fyn High with a thin cover of Zechstein and Triassic strata; and c. MONA LINE 3. Depth-migrated section across the Triassic-filled Horn Graben and Ringkøbing-Fyn High with a thin cover of Zechstein and Triassic strata.
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Figure 3.36 Regional seismic profiles in time from Denmark:

a. Morgan LISA 1. Seismic section across the Trondelag-Fuldal-Nore Graben and Ringkøbing-Fyn High with a thin cover of Zechstein and Triassic strata;

b. Morgan LISA 2. Seismic section across the Ringkøbing-Fyn High to the southern limit of the Horn Graben;

c. Morgan LISA 3. Seismic section across the Mid-North Sea High (Olten Spill), Sleipner Rough Basin, Troll End Graben and the Holmsland and Grindsted blocks at the southern edge of the Norwegian-Danish Basin.

d. Geosections around Bornholm (Vejbæk et al., 1994).

e. Seismic sections across the Sognefjorden-Tasjord Zone close to the northern edge of the SNB (Jelzubari, 1997);

f. Geosections around Bornholm (Vejbæk et al., 1994).

See Figure 3.31 for locations.

Figure 3.36a Regional cross-section across the Lower Saxony Basin-Pomerania-Swabian-Glückstadt Graben. Montepel from profiles 40, 46, 56 and 58 in Bidzinski et al. (2016). The section starts in the eastern extremity of the Central Netherlands Basin, part of the Eem Low during Triassic times. The section crosses the Grono-Middelhagen Fault Zone and the NW boundary of the Lower Saxony Basin. Most of the Cretaceous cover was removed during strong Sub-Hercynian inversion. The saline NW boundary of the Lower Saxony Basin (Rheder Moor Lineament with significant salt movement) is shown. The Jurassic sequence is much thinner on the Pomerania-Böck (Oldenburg Swell) than in the Lower Saxony Basin, and the smaller amount of inversion results in preserved Upper Cretaceous strata. The section crosses the Bremer Line (Tertiary) marked by increasing thickness of Triassic strata and continues into the Glückstadt Graben. Note the evidence for early (pre-Jurassic) salt movement (Neanderhütte), with further movement during Alpine (Late Cretaceous-Palaeogene) inversion. See Figure 3.31 for location.


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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); See Figure 3.31 for locations.

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

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 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 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 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

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.

b. Regional seismic profile from the Elbe Fault System (modified after Sverjensky et al., 1995).

Figure 3.42 — Regional seismic profiles in time, from Germany:

a. Migrated seismic section across the central Glückstadt Graben (after Sverjensky et al., 1995) 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 Bunter shales and the location of the base Triassic. Major salt movements took place at the beginning of the Upper Triassic when the Glückstadt Graben was affected by extension. The interval seismic patterns of the Bunter. Stratigraphic 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 Pomeranian Basin (Bücker, 2002) may have also affected the Glückstadt Graben. Permian salt deposits are only observed around salt structures and thin with increasing distance from salt walls or salt blocks. Parts of the Jurassic were formed in Late Jurassic-Early Cretaceous times. The Upper Cretaceous shows 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 event of salt extrusion is shown to be synchronous with the development of the NW-SE striking Rheinsberg Trough in the Late Triassic. This is inferred from stratigraphic thinning in the reflections interpreted as Late Triassic (Bunter). Lower Triassic layered the trough centre and showing indications for supersynchronous salt movements in the reflection pattern. The salt is almost completely removed below the trough. While normal faults are present in the Mesozoic unit, 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 axes and to the NW-SE striking intrusive structures from south of Kopenhagen to the northern basin margin parallel to the ANG OIBE 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 self-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 the above (Figure 3.42c).

e. Interpreted seismic section across the central North German Basin to the northern margin (across the Lower Saxony Basin) (after Scheck-Wenderoth et al. 2008). The section illustrates continuous subsidence during Early to Mid-Triassic times (Buntermandate and Bunterlade) represented by parallel, continuous reflections. The underlying reflections are interpreted as Upper Triassic-Lower Cretaceous and show local stratigraphic thinning indicating accelerated subsidence in the Lower Saxony Basin at the southern margin of the JFP compared to the Pomerania Basin further north. The strike of the basin is determined by the salt extrusion along the large part of the basin is obvious, as well as localized, basement-cored uplift in the Lower Saxony Basin where salt is thin or absent. The salt cover is affected by folding and faulting in the basin part, whereas the salt basement shows almost no faulting (modified after Rau & Schel-Scheck-Wenderoth, 2005).

f. Detail of migrated and coherency filtered seismic line from Figure 3.42c. See Figure 3.31 for locations.
Figure 3.41 Regional depth cross-section in Poland: a. Geological cross-section across the Pomeranian segment of the Mid-Polish Trough. Thickness distribution of the Mesozoic sedimentary cover is much more symmetrical in comparison to the Jurassic example, suggesting more uniform tectonic subsidence along both flanks of the Mid-Polish Trough.

3. Geological cross-section across the Jurassic segment of the Mid-Polish Trough based on regional seismic profiles (Cleamen, 2000a; Krzywiec et al., 2000). Some hard-linked basement faulting is noted beneath the Kłodawa salt diapir focused both extension and normal faulting during basin subsidence, as well as progressive thickness increase of the Jurassic succession. Scheck-Wenderoth et al. (2008) propose that salt doming and folding related to the formation of the Holocene foreland basin is mostly due to the Late Pleistocene-H Holocene extensional tectonics and the ongoing rifting of the Baltic Sea.

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 Kromień-Fornalik-Alder, Zabrze and Kazimierz-Ostrowki fault zones (based on Krzywiec, 2004a; Krzywiec et al., 2000). Some hard-linked basement faulting is inferred fault zones responsible for subsidence and subsequent inversion of the Mid-Polish Trough. Some significant thickness variations of the Triassic and Jurassic successions, suggesting localised normal faulting beneath the Kłodawa salt structure. See Figure 3.31 for locations.

b. Profile from the Nida Trough (based on Scheck-Wenderoth et al., 2008). On the NE flank of the Holy Cross Mountains, a system of cover deformations is observed detached above the Zechstein main gas reservoirs within the Upper Turonian(?)-Maastrichtian deposits, which suggest they are related to inversion and indicate uplift of the axial part of the Nida Trough. 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 place later in Paleogene times. The Miocene sediments that partly cover the Mesozoic succession are part of the infill of the Carpathian foredeep basin.

c. Seismic profile from the SW (Pomeranian) segment of the Mid-Polish Trough, showing the structure of the inverted Polish Basin (based on Krzywiec, 2004a; Krzywiec et al., 2000). The Mid-Polish Trough underwent stronger 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 (Sudetian salt pillow and the Pomeranian Demandezault salt flank) 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 Jurassic gas reservoirs. Their formation may have been partly triggered by sub-Zechstein strike-slip movements (Kędzierski, 2000).

d. Seismic profile from the central (Kuyavian) segment of the Mid-Polish Trough, showing the structure of the inverted Polish Basin (based on Krzywiec, 2004a; Krzywiec et al., 2000). Some hard-linked basement faulting is 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 progressive thickness increase during basin inversion. Note that the NNW-SSE trend of the basin can be seen in the central part of the profile, which propagated into the Mesozoic cover.
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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 3755/1-7 (Figure 3.33a, 3.33b & 3.34a); b. Mesh 7-1 (see Figures 3.32b, 3.33b & 3.34b); c. Depth PLDA 3 (Figure 3.37b & 3.38b); d. WAMA LDA 1 (see Figures 3.30b & 3.31b); e. Profile from the central (Silesian) segment of the Mid-Polish Trough (see Figure 3.30a, 3.40a & 3.41a); f. Regional line crossing the entire SPBA area from the UK to Poland; g. BASIN 9601 (see Figures 3.32g & 3.39b); h. Profiles from the central (Kuiavian) segment of the Mid-Polish Trough (see Figures 3.32i, 3.41b & 3.42d); i. Regional line crossing the entire SPBA area from the UK to Poland; j. Mid-Polish Trough (see Figures 3.32i, 3.41b & 3.42d).