Turonian sandstones of Saxony, south of Dresden (photo: Thomas Voigt).

Chapter 11 Cretaceous

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1 Summary and introduction

The transition from the Jurassic to the Cretaceous Period took place against a background of abating rifting and a gradual change to a period of regional tectonic subsidence controlled by lithospheric cooling following the Jurassic rifting events. The transition is defined by a main unconformity of early Berriasian age (Late Cimmerian Unconformity = Base Cretaceous Unconformity), which may truncate Jurassic or even older sediments (**Figure 11.1**). After a main phase of condensed sedimentation, emergence and erosion during the Kimmeridgian and Volgian (Tithonian), and subsequent nonmarine 'Wealden' type sedimentation during the Berriasian, the latest Berriasian to Hauterivian was an interval of continuous transgression (e.g. Rawson & Riley, 1982; Mutterlose & Bornemann, 2000).

The Boreal epicontinental basins of northern Europe expanded throughout Cretaceous times, with seaway connections to the Tethys Ocean either via the Polish Trough or via the proto-Atlantic along the present-day area north-east of Greenland and the northern North Sea. The southern part of the SPB area was occupied throughout the Cretaceous by the emergent Rheno-Bohemian Massif. The massif formed a continuous landmass that included the British Isles and the Shetland Plateau, which created a restricted seaway northwards through the Viking Graben, as sea-floor spreading was only incipient in the North Atlantic north of the Charlie Gibbs Fracture Zone. The Barremian was an overall regressive interval, whereas the mid-Cretaceous was characterised by continuous transgression (Aptian to Turonian) peaking in late Cenomanian to early Turonian times. The transgression was the result of both global sea-level rise and regional subsidence, which reduced the land area of the Rheno-Bohemian Massif and the south-eastern London-Brabant Massif as they became inundated by the sea, as did most of northern Europe (see **Figures** 3.20 & **11.8**). Sea level remained high until Campanian times and was followed during the late Campanian by a long-term regression that continued into Maastrichtian and Danian times. This general development was punctuated by a number of smaller-scale, regressive-transgressive cycles of which those in the Barremian, Aptian and Albian are of particular importance (see below). The full palaeogeographical extent of the Cretaceous sea is not certain, as the present-day distribution of Cretaceous rocks is largely controlled by syn- and post-depositional erosion. Cretaceous sediments were also removed by local erosion within the greater SPB due to inversion of former Late Jurassic to Early Cretaceous depocentres. Erosion is particularly prominent along the southern margin of the SPB, as seen in the Roer Valley Graben, Central Netherlands Basin, Lower Saxony Basin, Lausitz High and the Mid-Polish Anticlinorium, which are prominent examples of inversion tectonism (**Figures 11.2** to **11.7** and see **Figure 11.11**).

Palaeoclimatic conditions were warm and humid during earliest Cretaceous times, as indicated by coal seams and dinosaur remains. The Valanginian to Hauterivian interval was a period of cooler conditions with potential ice-cap formation during mid-Valanginian times (e.g. Kemper, 1987). The Barremian was a time of relatively warm and arid climate with cold interludes during the latest Barremian, early Aptian and early Albian. The Albian was otherwise an interval marked by the onset of greenhouse conditions, which culminated in a thermal maximum in the mid-Turonian as a consequence of a further increase in CO_2 emissions, perhaps related to high rates of oceanic-crust generation (Wilson et al., 2002). These conditions caused the climatic belts to migrate northwards, leading to reduced polar ice and a subsequent distinct rise in sea level augmented by an increase in the volume of the mid-ocean ridges. Overall, the later Cretaceous was a time of slow, long-term cooling from the late Turonian onwards (Jenkyns et al., 1994).

Long-term studies of the extensive Cretaceous outcrops in the UK have led to the development of detailed, local lithostratigraphic schemes that were not generally extended offshore (Rawson & Curry, 1978; Rawson, 2006). The first formal lithostratigraphic nomenclature for the offshore Cretaceous of the UK was proposed by Rhys (1974), who recognised two units of group status, the Cromer Knoll (Ryazanian-Albian) and Chalk (Cenomanian-Danian) groups. The Cromer Knoll Group (the Spilsby Sandstone, Valhall, Carrack and Rødby formations) is a thick clastic sequence of calcareous marine mudstones and sandstones with intermittent chalk lithologies (Lott & Knox, 1994). The Chalk Group is essentially synonymous with the pelagic, chalk-dominated sedimentation of the Late Cretaceous, which continued into the Early Paleocene (the Danian Ekofisk Formation; Lott & Knox, 1994). A recent review of Cretaceous sedimentation and stratigraphy in the UK has been published by Rawson (2006).

The formal Cretaceous stratigraphy of the Netherlands is defined in Van Adrichem Boogaert et al. (1994). A concise description of the Cretaceous geology of the Netherlands is given by Herngreen & Wong (2008). Additional stratigraphic details were published by Herngreen et al. (2000) and DeVault & Jeremiah (2002). The sequence stratigraphy of the Berriasian to lower Aptian interval has been described by Hoedemaeker & Herngreen (2003). For the Santonian to Maastrichtian and Danian intervals along the southern border of the North Sea Basin, reference is made to Vandenberghe et al. (2004). Studies of the Upper Cretaceous stratigraphy include Felder (1975), Gras & Geluk (1999) and Felder & Bosch (2000) for the onshore area, and Herngreen et al. (1996), Van der Molen (2004) and Van der Molen & Wong (2007) for the offshore sector. The formal lithostratigraphy of the Santonian to Danian interval in the Limburg Province of the south-east Netherlands has been described by Felder (1975) and Felder & Bosch (2000), who distinguished five formations, each of which is subdivided into several members (**Figure 11.1**). Additional detail regarding stratigraphic ages in this area has been added by Brinkhuis & Smit (1996), Herngreen et al (1998) and Jagt (1999).

2 Stratigraphy

Summaries of the geological development of the Cretaceous strata in each country in the SPB area are given below. A more comprehensive summary of the northern European Cretaceous development was published by Ziegler (1990a) and Voigt et al. (2008).

2.1 Berriasian

The Early Cretaceous basins of north-west Europe follow the structural patterns established during the Late Jurassic, with locally high sedimentation rates in the rapidly subsiding depocentres such as the Lower Saxony Basin (including the sub-basins in eastern Germany), the Central Netherlands Basin, the Central Graben, the Farsund Basin and the Mid-Polish Trough (Pożaryski, 1977b; Ziegler, 1990a; Kockel, 1991; Jensen & Schmidt, 1993; Baldschuhn & Kockel, 1996; Mutterlose & Bornemann, 2000; Senglaub et al., 2005). The varied Upper Jurassic sediments (carbonate-dominated in places; organic-rich shales in the North Sea) were replaced during the Berriasian by the siliciclastic sediments common to all of these basins (**Figures 11.1** & **11.8**). Lower Cretaceous successions are locally up to 2000 m thick due to differential subsidence (**Figure 11.4**). The basins were later inverted under the influence of transpressional tectonics as described below. In earliest Cretaceous times, open-marine conditions only existed in the present-day offshore areas where Late Jurassic depocentres continued to subside, but with reducing effect into the Early Cretaceous (**Figure 11.8**). There are important examples of deep-marine conditions in the central parts of the North Sea where the important source-rock facies of the organic-rich Farsund and Kimmeridge Clay formations were deposited (Schlanger & Jenkyns, 1976; Ineson et al., 2003; **Figure 11.1**). However, other depocentres such as the Lower Saxony Basin and the Mid-Polish Trough are characterised by nonmarine 'Wealden' type sediments (Mutterlose & Bornemann, 2000) represented by sandstones with dinosaur tracks and claystones containing freshwater ostracods and coals (see **Figure 11.13**).

In the Norwegian, UK, Dutch and Danish sectors of the North Sea, the base of the Lower Cretaceous is taken as the interface between the Cromer Knoll Group and the Farsund Formation (Norway and Denmark) and the equivalent Kimmeridge Clay Formation (UK), which occurs in the Ryazanian and therefore is not the actual base of the Cretaceous. This transition corresponds to a distinct change from organic-rich claystones below to organic-poor carbonaceous claystones and siltstones above, which is assumed to represent a synchronous, albeit poorly understood, basinwide change in the marine environment (Rawson & Riley, 1982). In large areas of the central North Sea, this change corresponds to a conspicuous and easily traceable reflector usually referred to as the Base Cretaceous Unconformity, although there may be Berriasian sediments below the reflector. In other areas, for instance in large parts of the Danish Basin, sediment changes across the Jurassic-Cretaceous boundary do not give rise to significant seismic reflectors and are therefore difficult to map. In this basin, the transition is found mainly between the Upper Jurassic Børglum Formation and the Lower Cretaceous Vedsted Formation, and so is also assigned a Ryazanian age (**Figure 11.1**). Both formations consist of dark grey claystones, although there are sporadic silty layers in the Vedsted Formation (Larsen, 1966; Michelsen et al., 2003). There is a marked lithological difference to the north-east where the Vedsted Formation overlies deltaic to shallow-marine siltstones and fine-grained sandstones of the Frederikshavn Formation (**Figure 11.1**).

Offshore of the Netherlands, the 'Wealden' type rocks are included in the mainly Jurassic Scruff Group. Onshore, there are continental deposits that are part of the Jurassic to Cretaceous Schieland and Niedersachsen groups. The subsequent transition to marine conditions in the whole area led to the widespread deposition of the fine-grained siliciclastics of the Rijnland Group. As sedimentation at the onset of the Cretaceous largely followed a pattern that was established during the Late Jurassic, these basal Cretaceous sediments are often assigned stratigraphically to the upper parts of mainly Upper Jurassic formations.

Alternating continental sandstones, claystones, dolomitic limestones and coals were deposited in the Vlieland, Broad Fourteens, Central and West Netherlands basins and the Roer Valley Graben, which now form basins beneath present-day Europe. Different formations have been defined in each basin, but all are part of the Delfland Subgroup (Schieland Group). Brackish-water conditions with repeated marine influences persisted into Early Cretaceous times in the Dutch part of the Lower Saxony Basin. This led to deposition of calcareous claystones with sporadic limestone beds in the youngest section of the Niedersachsen Group (**Figure 11.1**). In contrast, there was an open-marine environment in the Central Graben, Terschelling Basin and northern part of the Vlieland Basin, where mainly glauconitic sandstones and a thick succession of dark-coloured (locally bituminous) claystones were deposited to form the upper part of the Scruff Group.

2.2 Lower Cretaceous

Differential subsidence, local tectonics and sea-level changes during the latest Berriasian (Ryazanian) led to the establishment of a marine environment across most of the SPB and adjoining Norwegian-Danish basins. Claystones, marls, siltstones and glauconitic sandstones were the predominant facies from latest Berriasian to Albian times; chalk lithologies and laminated black shales are found regionally in rocks of Barremian age. These facies form the Rijnland Group deposits of the Netherlands, whereas they are mainly dark-coloured mudstones in the Cromer Knoll Group in the UK and Danish North Sea, and in northern Germany and Poland. The present-day thickness of these rocks varies widely due to locally significant differential subsidence, as well as later erosion during Late Cretaceous and Cenozoic tectonic inversion of these Jurassic-Cretaceous basins (**Figure 11.4**).

Following an end-Jurassic depositional hiatus, rising sea levels during the Early Cretaceous led to open-marine deposition in the Southern North Sea Basin, which extended northwards into the Central Graben across the then submerging Mid North Sea High, and eastwards into the present-day Netherlands. An exception is the former Sole Pit Basin, where Lower Cretaceous sediments are absent due to periodic inversion that began in latest Jurassic times, making it difficult to unravel the history of its remnant successions (Cameron et al., 1992). Much of this area underwent gradual subsidence; however, local fault reactivation, notably along the Dowsing-South Hewett Fault Zone, and intermittent halokinetic activity created both condensed sequences and local depocentres where substantially thick marine successions accumulated during the Early Cretaceous (**Figure 11.4**).

Sedimentation in this northern offshore area and contiguous Cleveland Basin was initially dominated by glauconitic, siliciclastic sandstones (Spilsby Sandstone Formation) and subsequently by dark grey, variably calcareous, fossiliferous and glauconitic, marine-mudstone deposition with sporadic volcanic tuffaceous bands (Speeton Clay and Valhall formations; **Figure 11.1**). Sedimentation culminated with the deposition of variegated silty and sandy phosphatic mudstones (uppermost Speeton Clay and Carrack formations)and the late Aptian to Albian red calcareous mudstones and chalks of the Hunstanton and Rødby formations. In contrast, this northern marine basin was separated from the more stable East Midlands Platform area to the south by the Market Weighton High, over which an attenuated Lower Cretaceous succession of interbedded shallow-marine, ferruginous sandstones, limestone and ooidal ironstones was deposited (**Figure 11.1**).

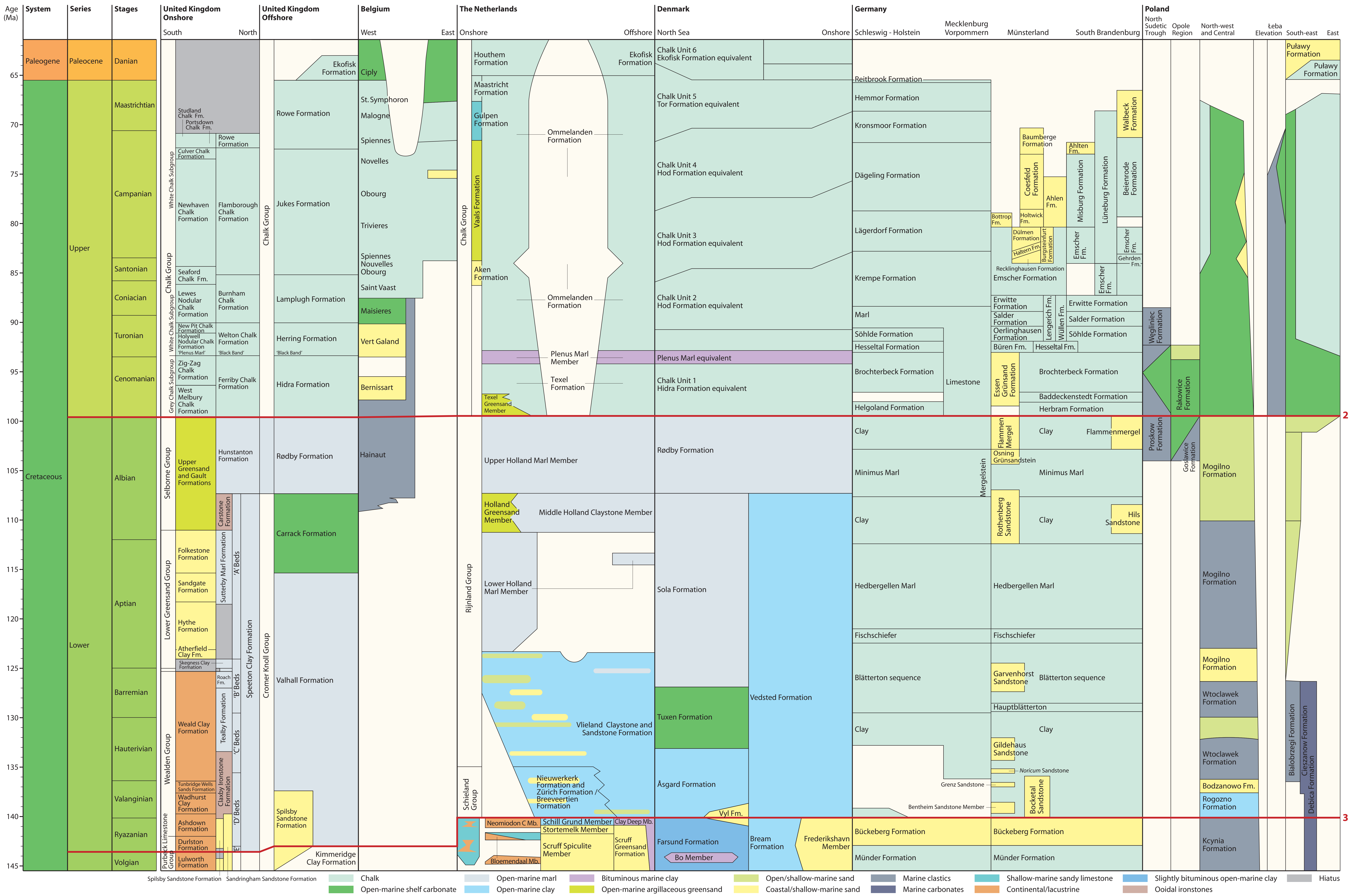


Figure 11.1 Tectonostatigraphic chart of the Cretaceous. The red lines at the base of the Cenomanian (2) and Ryazanian (3) are the lithostratigraphic horizons used to map the depths to the base of the Upper Cretaceous and near the base of the Ryazanian. See Figures 1.5, 11.2 & 11.3.

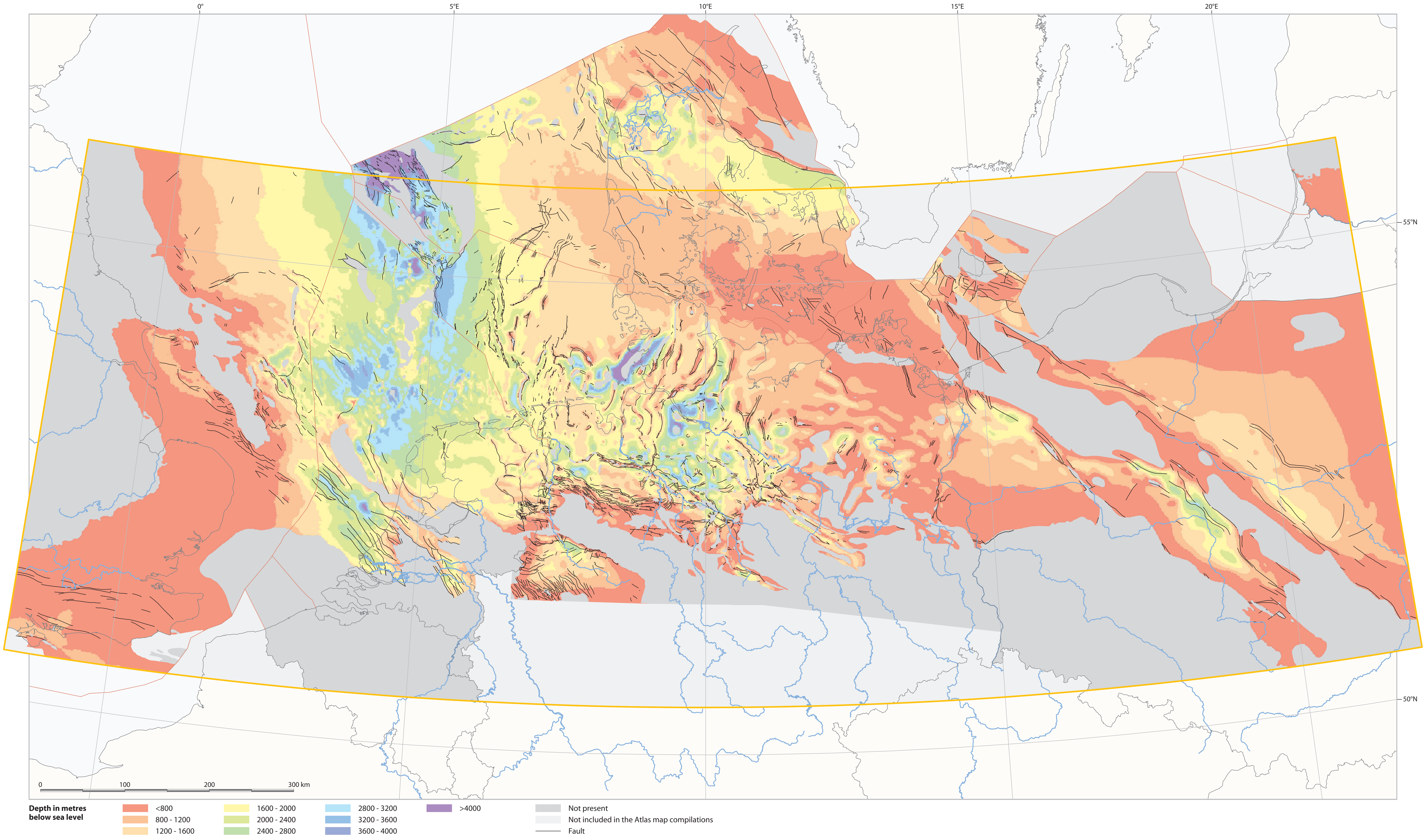


Figure 11.2 Depth to near base of the Lower Cretaceous (approximately near base Ryazanian). This lithostratigraphic horizon is shown as Horizon 3 on Figures 1.5 and 11.1.

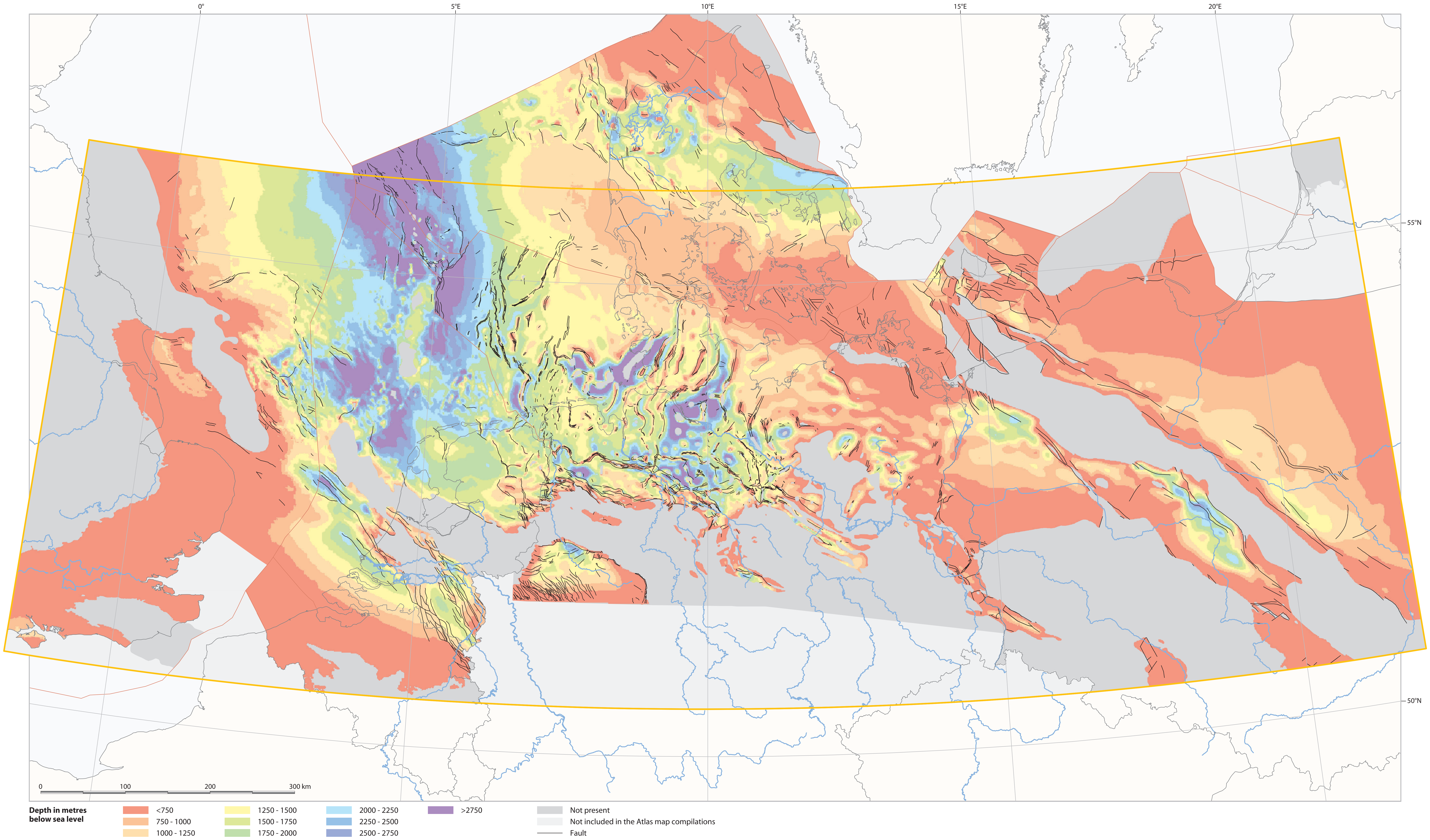


Figure 11.3 Depth to the base of the Upper Cretaceous (base of the Chalk Group; base Cenomanian). This lithostratigraphic horizon is shown as Horizon 2 on Figures 1.5 and 11.1.

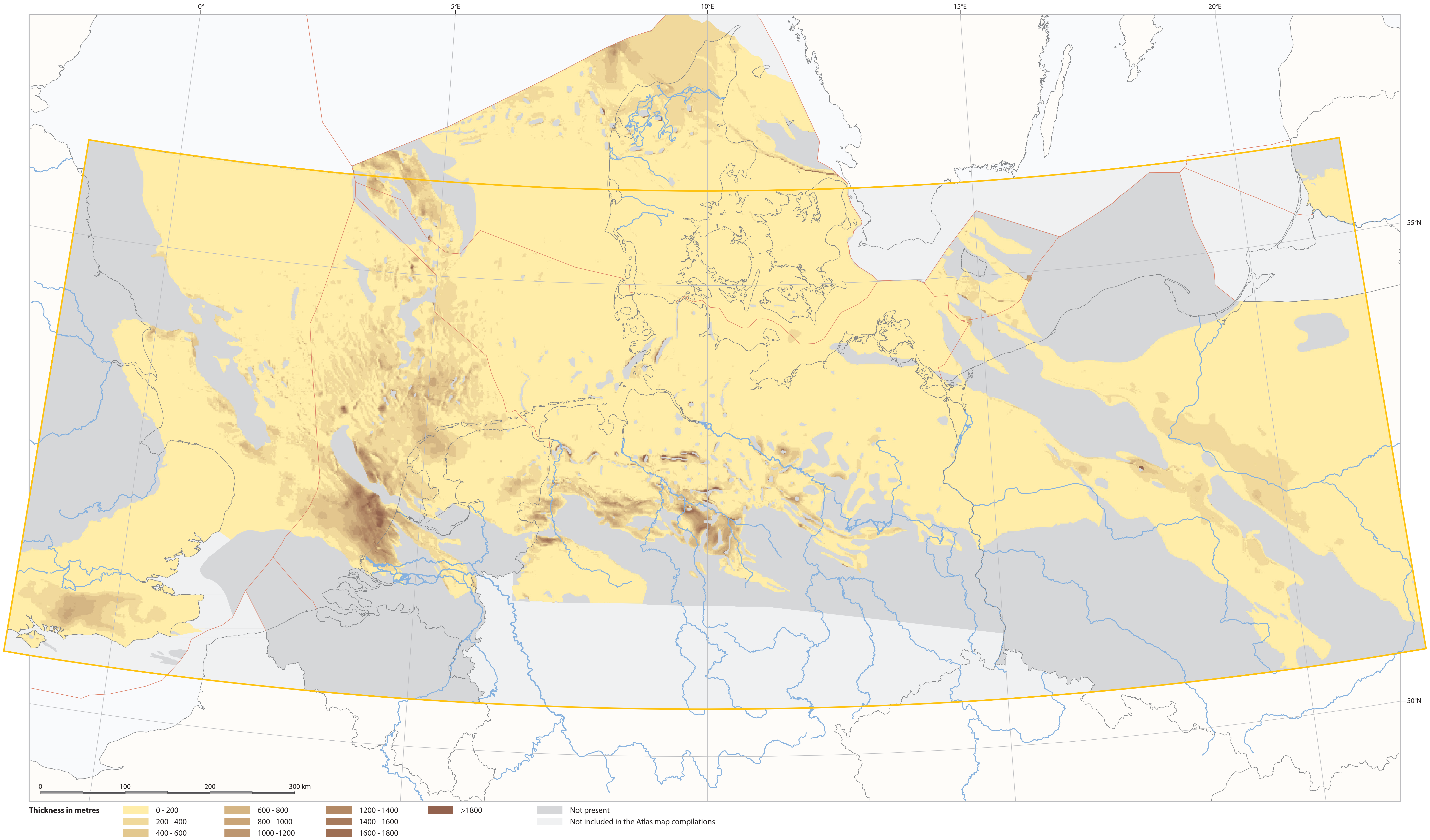


Figure 11.4 Thickness of the Lower Cretaceous.

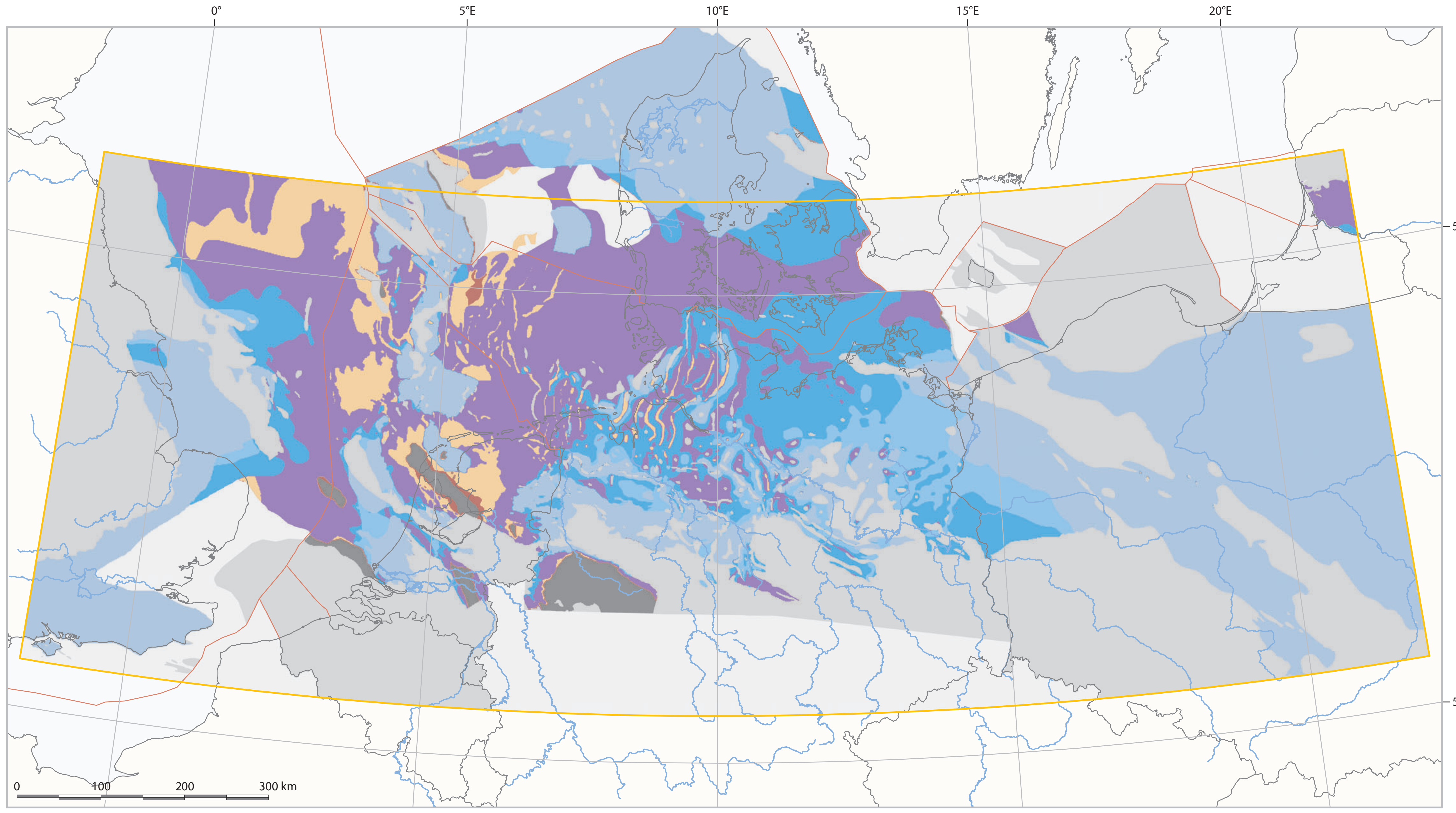


Figure 11.5 Subcrop at the base of the Lower Cretaceous.

Farther south, Early Cretaceous sedimentation until Aptian to Albian times was limited in part by the emergent Anglo-Brabant Massif. However, south of this massif in the Weald Basin, largely nonmarine sediments more than 800 m thick (limestones, mudstones, sandstones and claystones of the Purbeck Limestone and Wealden groups) were deposited during the Volgian (Tithonian) to Barremian. Eustatic sea-level rise and transgression during Aptian and Albian times gradually re-established links between the two main basin areas with the deposition in the south of glauconitic marine clays and sandstones of the Lower Greensand Group and Gault / Upper Greensand formations (Selborne Group) (**Figure 11.1**; Rawson, 2006).

In the Netherlands, the lower section of the Rijnland Group is the Vlieland Sandstone Formation, which consists of continental to shallow-marine, fine- to medium-grained, glauconitic sandstones. These sandstones form the principal reservoir of most Dutch oilfields. Depending on location, the age of the formation ranges from Ryazanian to Aptian. The overlying succession is mainly Valanginian to early Aptian grey, locally calcareous, claystones of the Vlieland Claystone Formation, which were deposited in a shallow- to fairly deep-marine (middle- to outer-neritic) setting.

The upper part of the Rijnland Group is formed by the grey and reddish brown marls and marly claystones of the Holland Formation. These sediments were deposited during Aptian to Albian times in a relatively deep-marine environment (middle to outer-neritic). Locally, bituminous sediments indicate periods of stagnant basin-floor circulation.

South of the London-Brabant Massif in western Belgium, the Wealden deposits are lacustrine claystones with lignites and fluviatile sandstones. They are preserved as paleokarst fill in Lower Carboniferous limestones, where deep vertical pipes cut through the Upper Carboniferous Coal Measures (containing the Iguanodon fauna from the Bernissart colliery) were formed by dissolution of evaporites in the underlying limestones (**Figure 11.12**). The Wealden deposits are also preserved in the rapidly subsiding northern margin of the Mons Basin (including the classical Hautrage quarry), either due to pull-apart faulting of the basin between the border fault of the Brabant Massif and the Nord-Artois Fault Zone, or in combination with deep-evaporitic dissolution (Delmer, 2004; Vandycke, 2002). These continental deposits pre-date the Albian and Cenomanian marine transgressions and are dated as Barremian to Albian in age (Yans et al., 2006).

The Lower Cretaceous deposits of the Danish North Sea are formed by the Cromer Knoll Group. Most of the succession consists of the upper Ryazanian to upper Hauterivian Åsgard Formation, which is in part equivalent to the Valhall Formation in Norway. The succession is mainly grey calcareous claystones with subordinate marlstone; the formation may be up to 600 to 900 m thick in the Central Graben area.

Although there are silty to sandy intercalations (the Leek Member) near its base, the formation has no reservoir potential (Jensen et al., 1986; **Figure 11.1**). The Vyl Formation developed locally during late Ryazanian to early Valanginian times as a submarine-fan deposit, possibly sourced from the Ringkøbing-Fyn High (Michelsen et al., 1987). The sediments are mainly siltstones with subordinate sandstones, silty claystones and marlstone layers. The Vyl Formation is thought to have some reservoir potential, but this has so far remained unproven due to the poor reservoir properties associated with the heterogeneous lithology of the unit. Similar local sandy developments sourced from adjacent highs are known from the UK and Norwegian North Sea, such as the Devil's Hole Formation (Hesjedal & Hamar, 1983) and the Ran Sandstone unit in Norway (Isaksen & Tonstad, 1989).

A layered succession of chalk and argillaceous chalk interbedded with calcareous claystones, known as the Tuxen Formation, was deposited in late Hauterivian to Barremian times. The pelagic to hemipelagic sediments of this formation are up to 75 m thick and were generally deposited under oxygenated conditions. However, a 2 m-thick black organic-rich calcareous claystone, the Munk Marl Bed, was deposited during a late early Barremian anoxic event (equivalent to the Hauptblätterton of the Lower Saxony Basin). Despite the low permeability of this formation (gas permeabilities typically about 0.4 mD), it is an increasingly important oil reservoir (e.g. the Valdemar and Adda fields) together with the overlying Sola Formation (Jakobsen et al., 2005). The base of the Tuxen Formation is locally developed as a minor hiatus, the distribution of which suggests that it represents a minor inversion tectonic pulse that was a precursor to Late Cretaceous inversion tectonics (Vejbæk & Andersen, 1987, 2002; Kühnau & Michelsen, 1994).

The Sola Formation was deposited under highstand conditions in the Central Graben during Aptian and early Albian times. It is up to 80 m thick and is a heterogeneous layered succession of chalk that may attain a reservoir quality similar to the Tuxen Formation, with interbedded argillaceous chalk and organic-rich claystone (Albian Shale). These laminated organic-rich calcareous mudstones are analogous to the Fischschiefer in the Lower Saxony Basin and are found in the upper part of the formation. The mudstones correspond to a recognised global anoxic event (Arthur et al., 1990) when oxygen levels were more variable than during deposition of the underlying Tuxen Formation.

Anaerobic conditions prevailed outside the Central Graben area in the Danish Basin, although with negligible preservation of organic material during deposition of the Børglum Formation until the late Ryazanian, when deposition changed to the more proximal muddy to silty Vedsted Formation that continued until Albian times (Sorgenfrei & Buch, 1964; Larsen, 1966; Hesjedal & Hamar, 1983). The Vedsted Formation consists of marine, dark grey, silty claystones that are locally glauconitic, and is rather similar to the underlying Børglum Formation albeit with a higher silt content. Oxygenated conditions in the water are indicated by the fossil content. The Vedsted Formation extends far beyond the underlying Upper Jurassic onto the Ringkøbing-Fyn High, where to the south it lies on progressively older rocks, indicating the Early Cretaceous transgression (**Figure 11.5**). The Vedsted Formation is up to 200 m thick, but the very gradual thickness variations indicate some tectonic activity during deposition (**Figure 11.4**). In the Sorgenfrei-Tornquist Zone, synsedimentary transtensional tectonic activity is reflected in Lower Cretaceous thicknesses exceeding 700 m, most of which are Vedsted Formation strata (Larsen, 1966; Rasmussen, 1978; Mogensen & Korstgård, 2003; **Figure 11.4**). Farther east on Bornholm, the marginal facies are found in a discontinuous series of formations. Fluvial sands and gravels of the lower Berriasian Rabekke Formation crop out on the island and are overlain by clays and coals deposited in a nearshore-swamp environment (Gravesen et al., 1982). Deposition continued in the late Berriasian with the fine- to coarse-grained coastal quartz sands of the Robbedale Formation. The upper Berriasian to Valanginian Jydegård Formation is also exposed on the island and consists of lagoonal-lake clays laterally passing into back-barrier and channel sands (Noe-Nygaard & Surlyk, 1988). These varied formations show clear evidence of syndepositional tectonic activity. A significant hiatus during Albian to mid-Cenomanian times is represented by a conglomerate that forms the base of the 85 m-thick Cenomanian Arnager Greensand Formation and marks the end of Early Cretaceous deposition in the area.

The Rødby Formation forms the topmost unit of the Lower Cretaceous succession in and around the Central Graben. This lower to upper Albian formation, which extends into the Cenomanian, is a transgressive sequence of pelagic marlstones, limestones and claystones that may reach thicknesses up to 50 m. Oxygenated depositional conditions prevailed, as indicated by their reddish colours and low organic content. The sediments become increasingly calcareous, which with their uniform thickness reflects the transition to a more regional subsidence pattern and onset of Late Cretaceous chalk deposition (Sorgenfrei & Buch, 1964; Larsen, 1966).

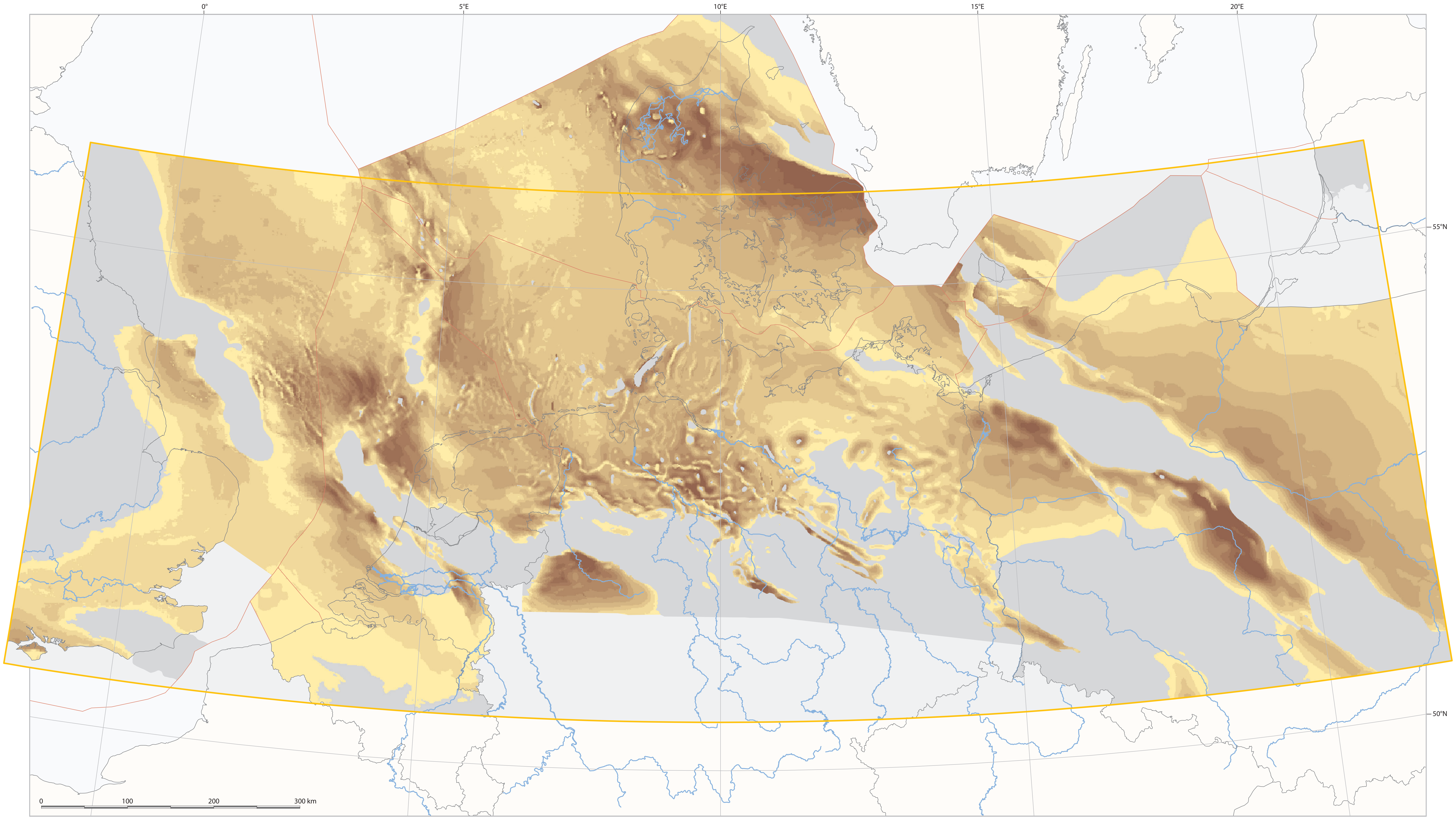
The Lower Cretaceous succession in the Mid-Polish Trough is dominated by siliciclastic shelf deposition with subordinate deltaic, fluvial, lacustrine, lagoonal and marine calcareous deposits; the latter are found mainly in the south-west of the trough (Marek, 1989; Marek & Pajchlowa, 1997; Leszczyński, 1997a). Deposition was controlled by both eustatic events and regional extensional tectonics; the latter centred in the Mid-Polish Trough (Pożaryski, 1977b; **Figures 11.8** & **11.9**).

Early Berriasian times were dominated by 'Wealden' type sediments consisting of the brackish, hypersaline to freshwater deposits of the Kcynia Formation. Sediments deposited during this interval are well developed in the Kuiavian Sub-basin of the Mid-Polish Trough, but thin towards its margins (Marek, 1989; Marek & Pajchlowa, 1997; Leszczyński, 1997a).

Late Berriasian to Barremian strata are represented by marine sandstones and mudstones with carbonates in the south-east. Larger hiatuses are due to erosion over local structural highs, related mainly to rising salt structures, and marginal areas. The interval was initiated by a late Berriasian to early Valanginian transgression that led to deposition of calcareous-facies sediments followed by up to 160 m-thick dark-coloured siliciclastic deposits restricted to the Mid-Polish Trough. Further transgression during the late Valanginian caused basin expansion to the south-west and north-east from the Mid-Polish Trough, characterised by the deposition of 50 to 70 m-thick siliciclastic sediments and more calcareous sedimentation to the south-east. An initial regression during Hauterivian to ?early Barremian times was followed by a transgression that resulted (in the Hauterivian) in increased marine influence and renewed deposition of dark-coloured siliciclastic sediments grading south-eastwards into carbonates with an average thickness of 120 m, but reaching more than 200 m in local tectonic depocentres. Deposition during Barremian, Aptian and Albian times was dominated by sandstones, such as those forming the Mogilno Formation, and fissile mudstones that were partly nonmarine (Raczyńska, 1979); maximum thicknesses are about 200 m.

There was a major transgression in mid-Albian times, which started with siliciclastic sand deposition followed by mainly Upper Cretaceous carbonate sediments up to 2500 m thick. Local sandstones and fissile mudstones were derived from erosion along inversion structures in the Mid-Polish Trough (Leszczyński, 1997b). A formal lithostratigraphic subdivision of the Upper Cretaceous has not yet been established in the Mid-Polish Trough.

The Lower Saxony Basin, which here includes the neighbouring Prignitz and Altmark-Fläming sub-basins, is bounded by the Pompeckj Swell to the north, Rhenish Massif to the south, East Netherlands High to the west and the East Brandenburg High to the east (**Figure 11.8**). Sedimentation was influenced by salt movement along east–west-striking anticlines, by tectonic movement along north-west–south-east-striking faults and by sea-level changes. Berriasian to Albian shallow marginal-marine siliciclastics are up to several hundred metres thick in the western, southern and easternmost part of the Lower Saxony Basin. In particular, marine sandstones and siltstones (Osning Sandstone) are up to 1000 m thick and were deposited along the southern basin margin throughout Valanginian to Aptian times (**Figure 11.14**). Within the basin, these facies interfinger with up to 2000 m-thick dark-coloured mudstones. The



Thickness in metres					
0 - 200	600 - 800	1200 - 1400	>1800	Not present	
200 - 400	800 - 1000	1400 - 1600		Not included in the Atlas map compilations	
400 - 600	1000 -1200	1600 - 1800			

Figure 11.6 Thickness of the Upper Cretaceous.

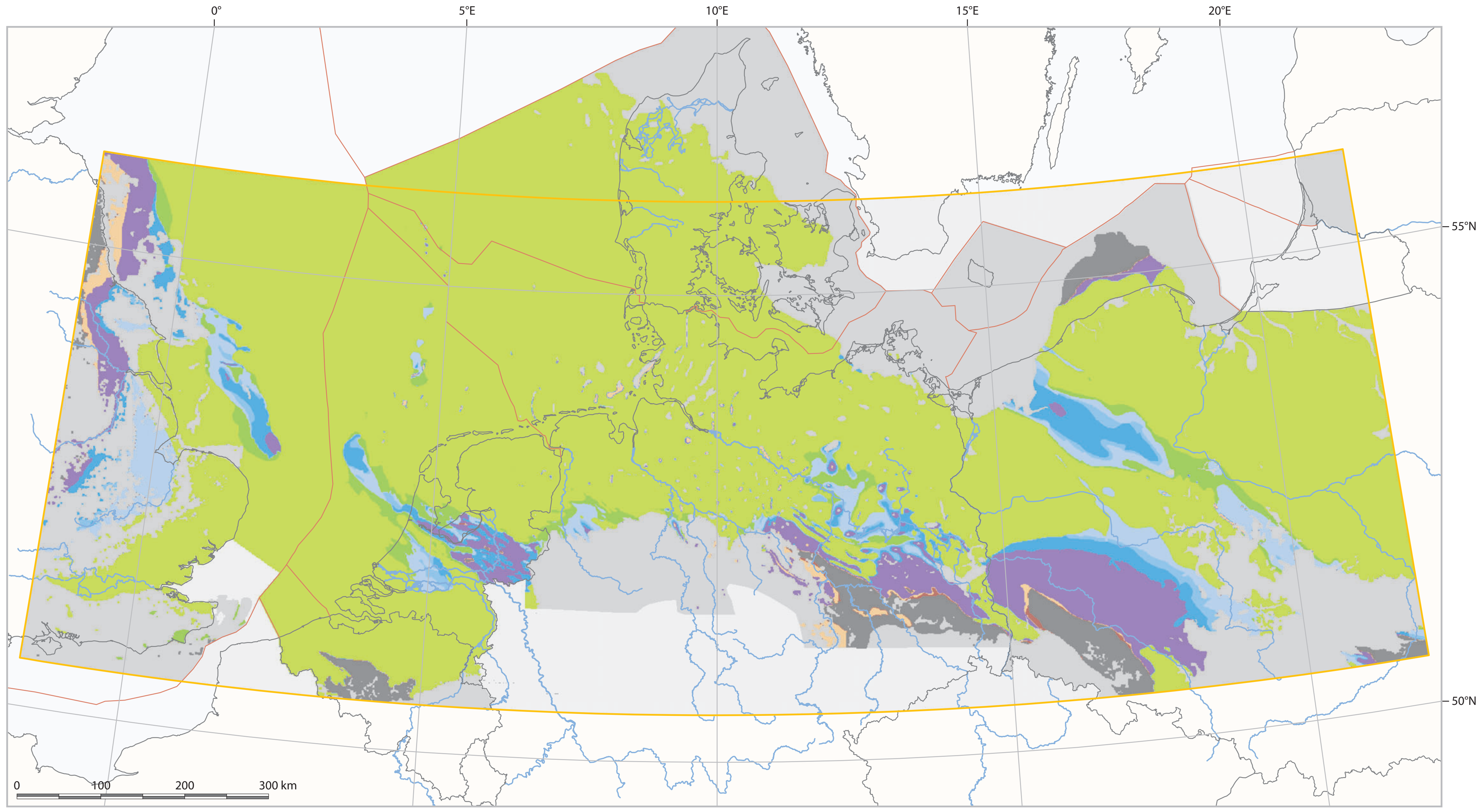


Figure 11.7 Subcrop at the base of the post-Chalk Group (near the base of the Tertiary).

following data have been compiled from Schott et al. (1967, 1969), Michael (1974), Kemper (1979), Betz et al. (1987), Mutterlose & Bornemann (2000) and Mutterlose et al. (2008).

The widespread regression during Berriasian times resulted in nonmarine conditions in the Lower Saxony Basin when 'Wealden' type sediments were deposited in a brackish-lacustrine environment, although several short-lived marine incursions have been described for the Bückeberg Formation. Mudstones up to 400 m thick accumulated along the western basin margin, with claystones and mudstones up to 300 to 700 m thick in the basin centre farther north and north-east, and finally fluviodeltaic sandstones (Obernkirchen Sandstone) in the south-east (**Figure 11.13**).

The opening of the Carpathian Seaway via the Polish Trough and the following major transgression, brought a return to marine conditions in the earliest Valanginian, starting with the fully marine *Platylenticeras* Beds. The Ems, Hoya and Gifhorn channels linked the Lower Saxony Basin to the southern North Sea via the Pompeckj Block (**Figure 11.8**). Dark-coloured claystones up to 350 m thick, including several intercalated sandstone horizons (Bentheim Sandstone, *Dichotomites* Sandstone, Grenz Sandstone), were deposited at the western margin of the Lower Saxony Basin. The Bentheim Sandstone is an important oil reservoir that still supplies crude oil. Shallow-marine sandstones were deposited along the southern basin margin, whereas about 300 m of marine clays accumulated in the basin centre.

Expansion of the area of the Lower Saxony Basin took place during the early Hauterivian transgression, peaking in the *Endemoceras amblygonium* ammonite zone. The western basin margin is dominated by up to 500 m-thick medium-grey claystones with two intercalated sandstone horizons, the *Noricum* Sandstone (middle-lower Hauterivian) and the Gildehaus Sandstone (upper-lower Hauterivian).

An early Barremian regression again led to 'Wealden' type brackish-lacustrine conditions in central and southern Poland. Finely laminated organic-carbon rich sediments, known as the Hauptblätterton and Blätterton, are typical of the Barremian succession of the Lower Saxony Basin. These organic-rich sediments (6-8% total organic carbon (TOC)) were deposited under anoxic conditions reflecting stable stratification of the water column under very warm temperatures (Mutterlose et al., 2008). Barremian sediments are found in the Brechte Syncline along the western basin margin where they are up to 400 m thick.

Significant palaeogeographical changes are associated with the early Aptian transgression. A new seaway between the Tethys and Boreal oceans opened via the Hampshire and Paris basins (**Figure 11.8**). Aptian strata are at least 200 m thick and are dominated by claystones and marls with dark-coloured claystones (*bodei* Clay) overlain by the Fischschiefer (equivalent to the Ocean Anoxic Event 1a; Mutterlose & Böckel, 1998). The latter, a finely laminated organic-carbon rich sediment, is followed by a succession of hemipelagic marls, the *ewaldi* Marl, the *clava* Marl and the *inflexus* Marl. Dark-coloured clays (*jacobi-nolani* Clay) are also typical of the uppermost Aptian succession. Up to 220 m of sediment (Rothenberg Sandstone) was deposited adjacent to the Rhenish Massif in the south-west of the basin (**Figure 11.1**).

Another major expansion of the Lower Saxony Basin took place during Albian times. Earliest Aptian black clays (*schrammeni* Clays) are widespread, whereas pale-coloured marls were deposited during mid- to late Albian times. Rising sea level caused the coastline to shift south and south-eastwards. The coastline was displaced even farther to the south during the late Albian, by some distance onto the Rheno-Bohemian Massif. Albian clays and marls are widespread in the basin centre and are up to 250 m thick. Farther east, the lower Albian sediments are a 40 m-thick sequence of glauconitic, sandy clays and sandstones (Hils Sandstone). The middle Albian succession is a 10 to15 m-thick clay sequence (*minimus* Clay) and the upper Albian an 80 m-thick sequence of biosiliceous marls (Flammenmergel).

2.3 Upper Cretaceous – the Chalk Group

Late Cretaceous tectonics were characterised by a regional change in the stress-field caused by onset of subduction along the northern Tethys Ocean in southern Europe, concurrent with the opening of the Bay of Biscay by sea-floor spreading (**Figure 11.8**). This resulted in a compressional phase in the SPB area associated with north-west–south-east-oriented strike-slip faulting. The compression also caused inversion of former Early Cretaceous depocentres, and along major zones of weakness such as the Sorgenfrei-Tornquist Zone and its continuation in the Polish Trough, the Roer Valley Graben and the Elbe Fault system (Pożaryski, 1977; Ziegler, 1990a; Krzywiec, 2006b). Early inversion was associated with frequent reverse faulting and associated shortening and crustal thickening (Vejbæk & Andersen, 2002). Later phases of the inversion were characterised more by stress relaxation, such that grabens where there had been lithospheric thickening during compression were subjected to upwarping and doming with little or no associated reverse faulting (Nielsen et al., 2005). Another important feature of these inverted grabens is their flanking depocentres. Prominent examples are the more-or-less continuous depocentre straddling south-western Scania, east Sjælland and northern Jylland in Denmark. The Pomeranian-Warsaw-Lublin-Lviv depressions form similar marginal depocentres to the north-east and likewise the Szczeci-Łódź-Miechów depressions to the south-west of the Mid-Polish Anticlinorium. The phases of inversion are often better resolved in the sedimentary record of these flanking depocentres as Upper Cretaceous deposits may be entirely removed from the inversion centres.

The Upper Cretaceous succession is dominated by pelagic deposits, mainly in the form of chalk. Thickness changes are a consequence of both tectonic activity that caused erosion and mass flows due to slope instability, and long-duration deep-sea currents that led to significant erosion and channel formation in contourite systems (Surlyk et al., 2003; Esmerode et al., 2007; Surlyk & Lykke-Andersen, 2007). Chalk deposition continued into Danian times in most of the North Sea area and is included in the Chalk Group. For this reason, Danian strata form the top of the mapped sequences, and therefore their description is included in this chapter.

The Upper Cretaceous Chalk Group succession of the UK is characterised mainly by chalk-dominated lithologies, although outcrops show significant lithological variability from north to south. North of the Anglo-Brabant Massif onshore (the 'Northern Province'), the chalk succession is generally dominated by hard chalk lithologies within which there is evidence of several subtle lithological changes. These changes have been used to subdivide the group into five formations (**Figure 11.1**). The lowermost grey chalk beds are distinctly more argillaceous and are flint-free, forming the lower to middle Cenomanian Ferriby Chalk Formation (21-33 m thick). This formation is terminated by a thin (<0.5 m thick), but very distinctive variegated unit of limestones, siltstones and argillaceous limestones known as the Black Band Member. This unit forms the basal interval of the overlying upper Cenomanian to Turonian Welton Chalk Formation (53 m thick), which is characterised by massive-bedded, white chalk with interbedded terrigenous, argillaceous seams. Nodular flints first appear in sediments a few metres above the base of this sequence. The overlying Turonian to Coniacian Burnham Chalk Formation (130-150 m thick) is characteristically thinly bedded with laminated beds, argillaceous seams, and an abundance of tabular-flint developments. The rest of the group comprises two formations, the Flamborough Chalk and concealed Rowe formations. The Santonian to Campanian Flamborough Chalk Formation (280 m thick) is the thickest unit onshore and is characterised by an absence of flint and increasing abundance of argillaceous seams. The topmost unit of the Chalk Group onshore in this 'Northern Province' is the upper Campanian Rowe Formation (70 m thick), which is buried beneath a thick Quaternary succession. The formation marks a return to deposition of white chalk lithologies with flint bands. There are no marginal facies developments known within this Northern Province of the UK Chalk Group.

The Chalk Group is widespread in the southern North Sea, where it subcrops thin Pleistocene sediments in the west and is gradually buried eastwards beneath a thickening Tertiary and Pleistocene cover in the Cleaver Bank Basin and beyond. The group varies greatly in thickness, locally reaching more than 1200 m in rim-synclines adjacent to major halokinetic developments east of the Sole Pit inversion. It thins northwards over the Mid North Sea High, thickens in the Central Graben, and then thins again southwards across the Anglo-Brabant High into the eastern English Channel (Cameron et al., 1992). Thinning, condensed and onlap relationships over some diapiric structures and tectonic highs are apparent from many wells in the area. The group is absent over much of the Sole Pit area due to Tertiary inversion and erosion, but thickens eastwards into the Cleaver Bank Basin where there are widespread Maastrichtian chalks and sporadic remnants of Danian chalky limestone units. The Chalk Group continues to thicken eastwards into the Dutch sector (**Figure 11.6**).

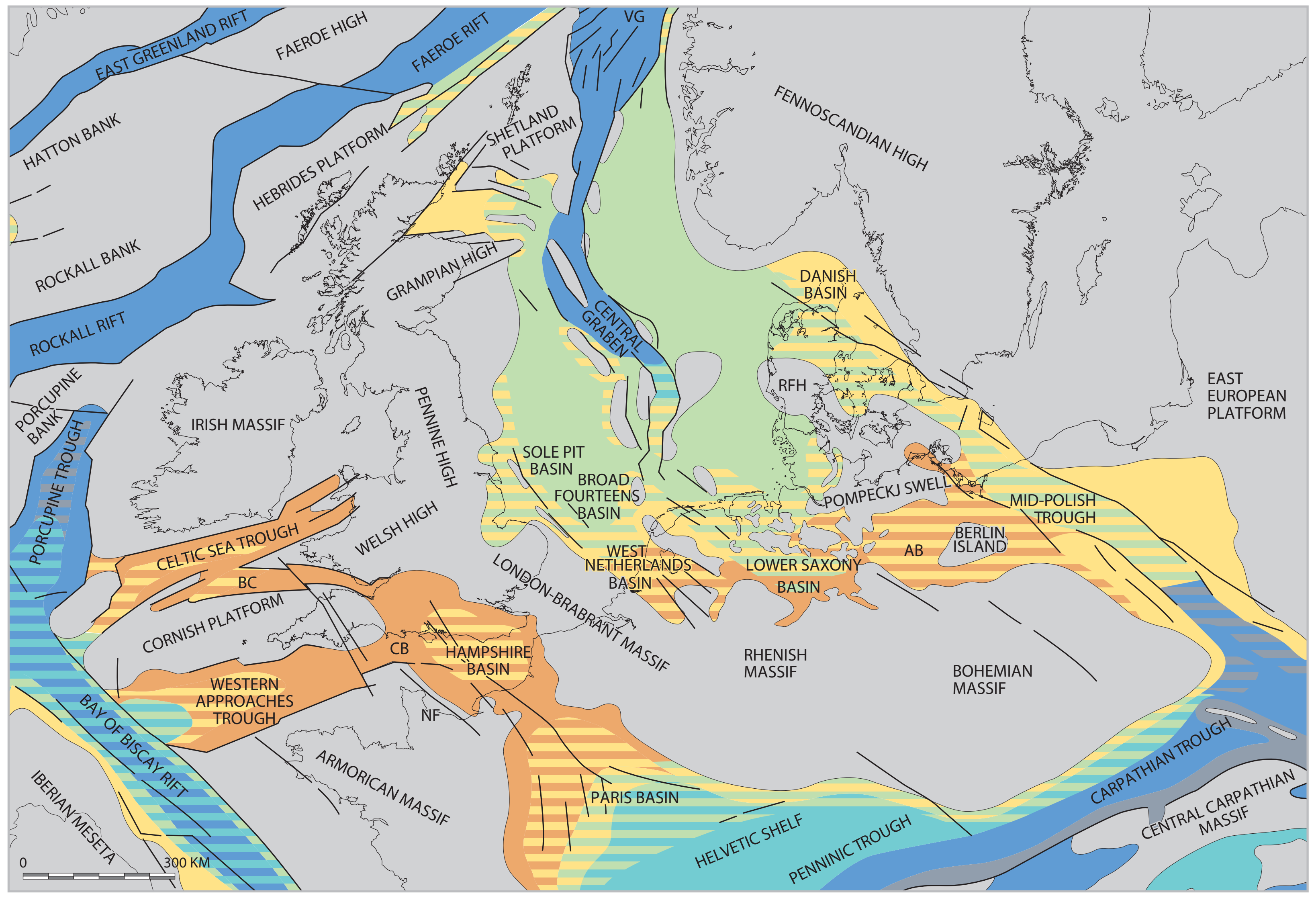
The Chalk Group offshore is also subdivided into five formations on the basis of subtle lithological changes, mainly evident from geophysical-log responses. These units can be broadly correlated with the onshore succession, but this can be problematic due to significant thickening offshore (**Figure 11.1**). The basal Cenomanian Hidra Formation (40 m thick) is a sequence of flint-free, micritic, variegated chalky limestones with more common interbedded argillaceous chalks and mudstones in the lower part of the unit. The overlying upper Cenomanian to Turonian Herring Formation (60 m thick) consists of hard, dense, chalky limestones with interbedded argillaceous chalks and mudstones and abundant flints. The limestones may be glauconitic and sporadically pyritic. The interbedded mudstones, which are best developed in the basal Black Band unit, are soft to hard, dark grey to black or variegated, carbonaceous, noncalcareous and pyritic. The Turonian to Coniacian Lamplugh Formation (200 m thick) deposits are chalky limestones, typically white to grey, soft to moderately hard, commonly argillaceous and characteristically flint-bearing. The very thick Santonian to Campanian Jukes Formation (500 m thick) typically consists of moderately hard, white to greyish white, variably argillaceous chalky limestones and is generally flint-free, although harder limestone bands have some nodular-flint developments. The Rowe Formation deposits are chalky limestones, typically white to greyish white, friable to moderately hard, commonly argillaceous and flint-bearing. The abundant flint bands throughout this unit are a characteristic feature and there are terrigenous clays in thin dark-coloured beds and seams. Chalk sedimentation continued into Danian times in parts of the western North Sea and these flint-free chalks are preserved as erosional remnants that form part of the Ekofisk Formation.

In the southern onshore area or 'Southern Province' of the UK, the Chalk Group is a succession of softer chalk beds with more persistent hardground developments and argillaceous seams, which have been used to subdivide the group. Proximal-facies developments are found around the Anglo-Brabant Massif and Cornubian Massif to the east of the SPBA area in the Wessex Basin. These massifs were periodically emergent during the Late Cretaceous and affect both Chalk Group thicknesses and facies developments in their vicinity. For example, flint developments increase in the shallower-water successions marginal to the massifs. South of the Anglo-Brabant Massif, the group thickens to more than 400 m in the Weald Basin and comprises ten different formations (**Figures 11.1** & **11.6**). A detailed description of these individual formations is beyond the scope of this Atlas, but has been summarised by Mortimore et al. (2001), Rawson (2006) and references therein. In general, there are two major subdivisions within the group, the Grey and White Chalk subgroups, which broadly correlate with the various UK chalk provinces (**Figure 11.1**). The Cenomanian Grey Chalk Subgroup (90 m thick) is a sequence of flint-free argillaceous chalks and paler bioturbated white chalks, which is equivalent to the Ferriby Chalk and Hidra formations of the Northern Province and contiguous Southern North Sea Basin. The remaining nine formations comprise the overlying upper Cenomanian to Campanian White Chalk Subgroup (>500 m thick), which commences with the argillaceous, thinly bedded Plenus Marl Member (**Figure 11.1**). The chalk of this subgroup is flint-free in its lower part, but variably flinty in its upper beds and has a wide range of chalk lithologies that typically include rhythmically bedded, shell-detrital chalk and nodular chalks in the lower part. These beds are commonly separated by omission surfaces and hardground developments that can be correlated across much of the southern area and, in some places, into the northern basins beyond. In contrast, the upper part of the subgroup is characterised by couplets of clay-free chalks (with flint developments) and marly (clay-rich) chalks; some of these marl beds can also be correlated throughout the basin.

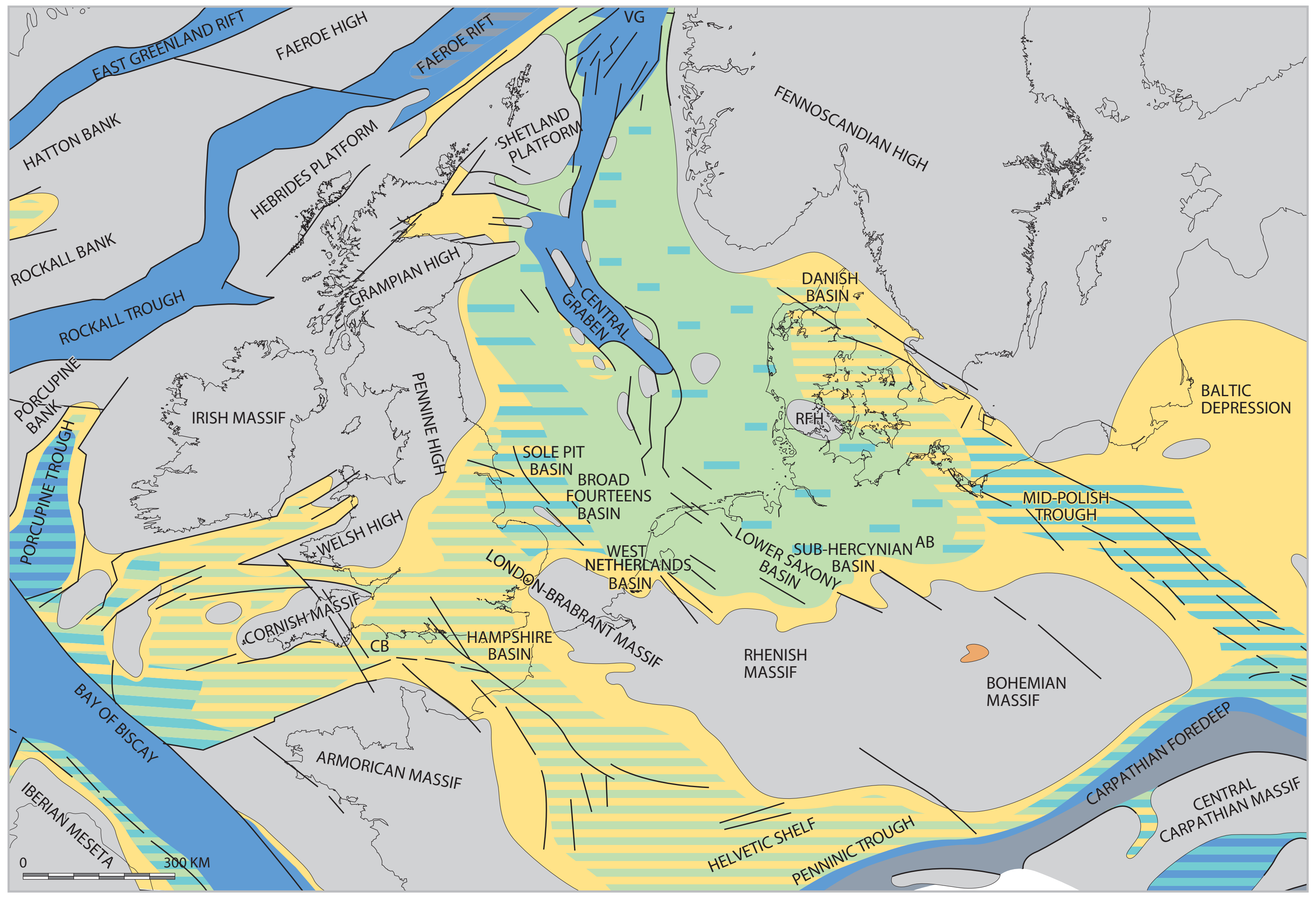
In the south-east Netherlands and Belgian Campine Basin flanking the Roer Valley Graben, the Cenomanian to Danian Chalk Group is developed as a nearshore facies consisting of chalk, marls and recurring greensand intercalations (**Figure 11.12**). These sediments grade to clean chalk towards the centre of the basin. The Chalk Group is up to 1800 m thick, but is thin or absent over basement structures that were inverted during the Late Cretaceous and Early Cenozoic.

The Upper Cretaceous outcrops in the Limburg Province of the south-east Netherlands and neighbouring parts of Germany and Belgium, have long been the subject of stratigraphic and paleontological investigations, notably resulting in the definition of the Maastrichtian Stage by the Belgian geologist Dumont in 1849 (Felder, 1975; Felder & Bosch, 2000). However, the oldest Upper Cretaceous sediments are nearshore to estuarine siliciclastics. These sediments are mainly fine-grained, partly glauconitic sands and partly calcareous clays of Santonian to early late Campanian age. The upper Campanian succession marks the change from a nearshore to fully marine environment and the deposition of chalk with varying amounts of glauconite and flint nodules. The chalks graded into calcarenites as the sedimentary basin shallowed, testifying to the rich benthic life during the early late Maastrichtian. The calcarenites became coarse-grained with intercalations of algal boundstones in the Danian succession (Bless et al., 1993). The London-Brabant Massif formed a barrier to stepwise extension of the Cretaceous deposits from the south-west. Sedimentation was still restricted to the more rapidly subsiding Mons Basin during the Albian. Several transgressive pulses encroached onto the London-Brabant Massif from Cenomanian times onwards, which became almost entirely flooded and covered by chalk during the Santonian to Campanian sea-level highstand. The London-Brabant Massif became emergent again during a relative sea-level fall in the Maastrichtian and then developed as the southern limit of the Cenozoic North Sea Basin (Felder et al., 1985; Vandenberghe et al., 2004; Dusar & Lagrou, 2007).

The base of the 'clean' chalks is formed by the Cenomanian Texel Formation, which is the stratigraphic equivalent of the Hidra Formation in the UK and Norwegian North Sea sectors (**Figure 11.1**). The formation is a sequence of pale-grey chalks and marly chalks with calcareous clay intercalations. Greenish glauconitic sands and marly chalks are locally interbedded at its southern limit. The formation is consistently overlain by the Plenus Marl Member, a black mudstone layer that can be traced throughout north-west Europe. This marker horizon is characterised in well logs as a conspicuous spike of high gamma-ray and density responses and low sonic values. The overlying Ommelanden Formation spans the entire Turonian to



a.



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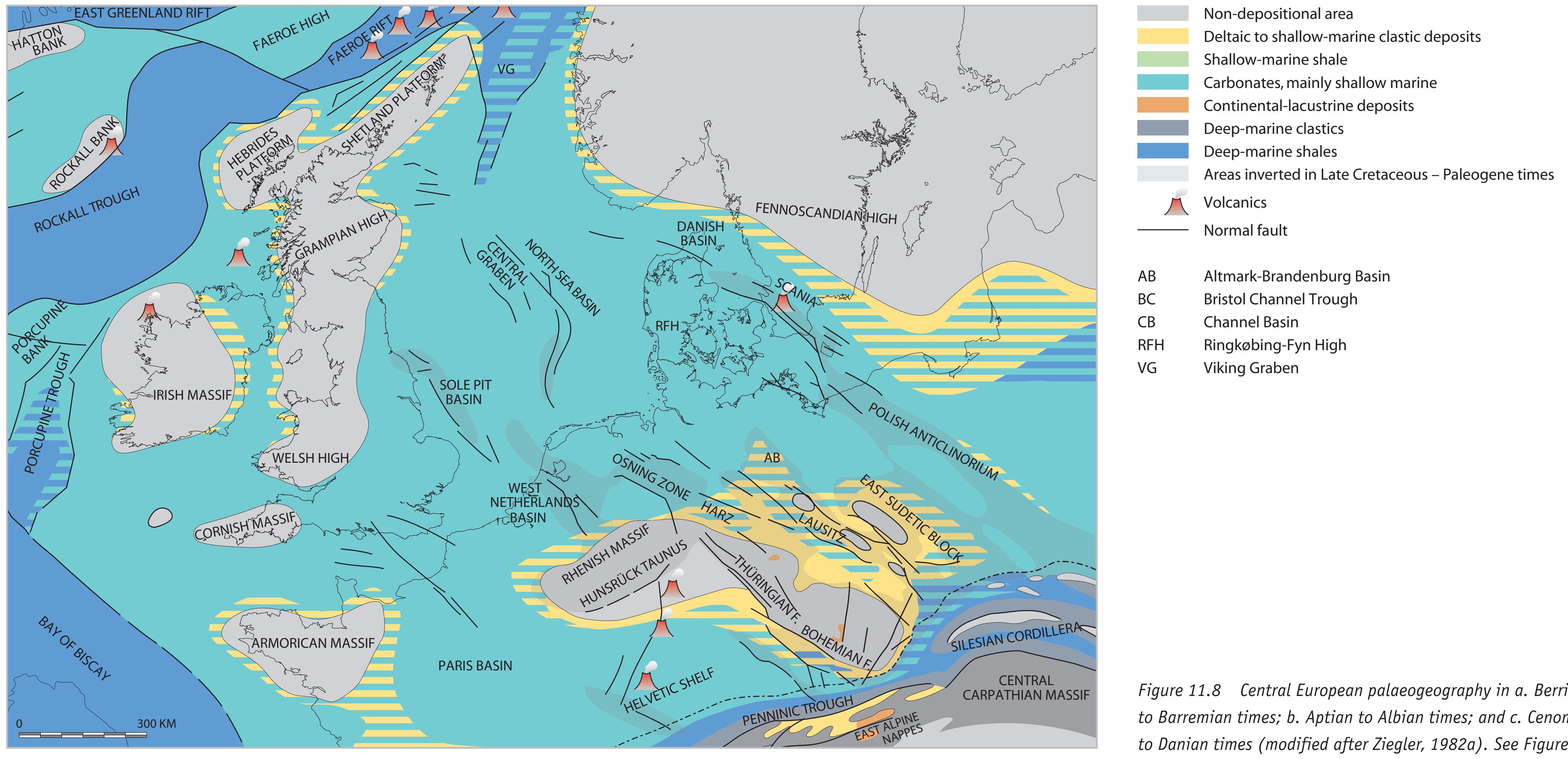
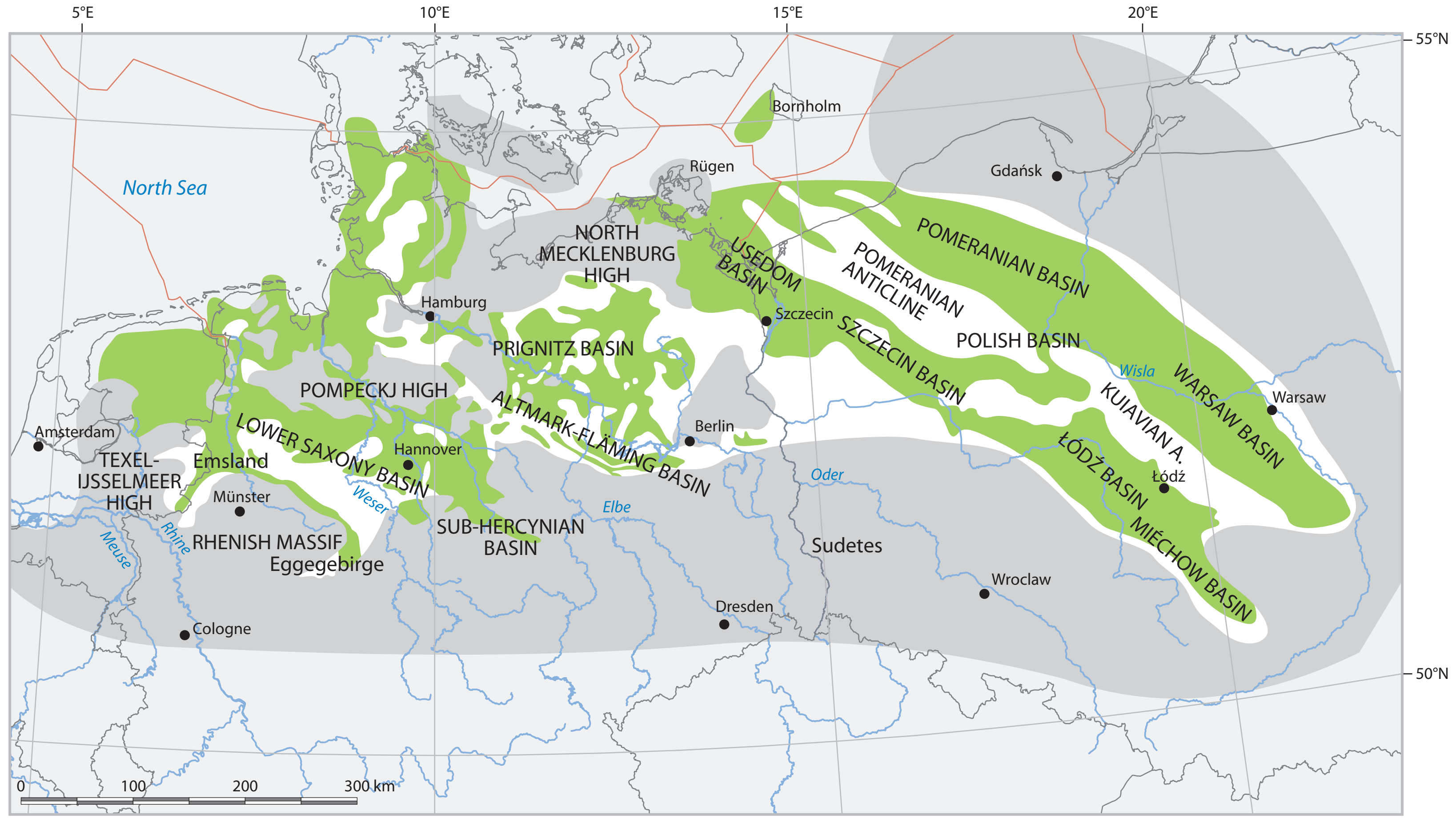
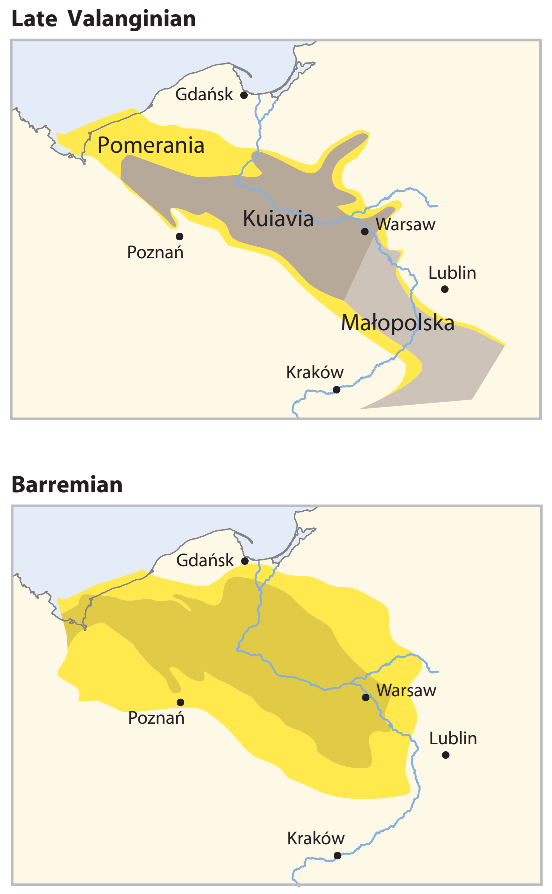


c.

Figure 11.8 Central European palaeogeography in a. Berriasian to Barremian times; b. Aptian to Albian times; and c. Cenomanian to Danian times (modified after Ziegler, 1982a). See Figure 3.21 for Late Cretaceous tectonic evolution.





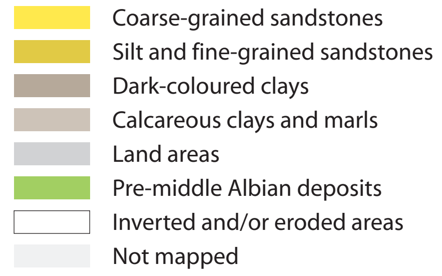


Figure 11.9 Aptian-Albian palaeogeography of northern Central Europe. The inset maps show Valanginian transgression and Barremian regression in the Mid-Polish Trough (modified after Voigt et al., 2008).

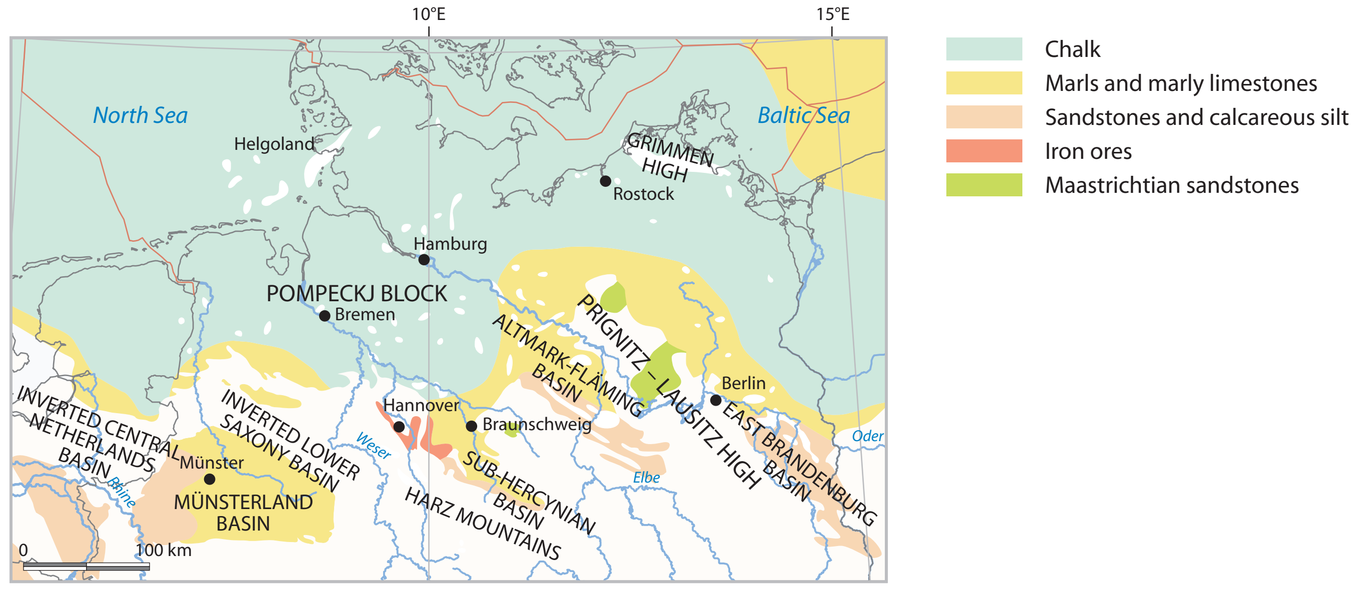


Figure 11.11 Early Campanian palaeogeography of northern Central Europe (modified after Voigt et al., 2008).

Maastrichtian interval (**Figure 11.1**). The formation is an apparently monotonous succession of chalks, chalky limestones and argillaceous chalks. An informal three-way subdivision is often used, comprising the Turonian 'lower Ommelanden' interval, which consists of hard chalk (and so has a high sonic response), a more marly Coniacian to Santonian 'middle Ommelanden' interval and a Campanian to Maastrichtian 'upper Ommelanden' interval of clean chalks with abundant flint layers. Herngreen et al. (1996) recognised several unconformities in this interval from biostratigraphic and well-log correlation studies on wells from the northern Netherlands. Additionally, nine seismic sequences were recognised in the Ommelanden Formation on regional seismic data from the Dutch offshore sector, supporting a further formal subdivision of this interval (Van der Molen & Wong, 2007).

The Danian Ekofisk Formation is a succession of white, chalky limestones with rare nodular and bedded flint layers. The Cretaceous-Cenozoic boundary is visible in places as a gamma-ray peak on well logs, thought to be the equivalent of the Danish 'Fish Clay' unit.

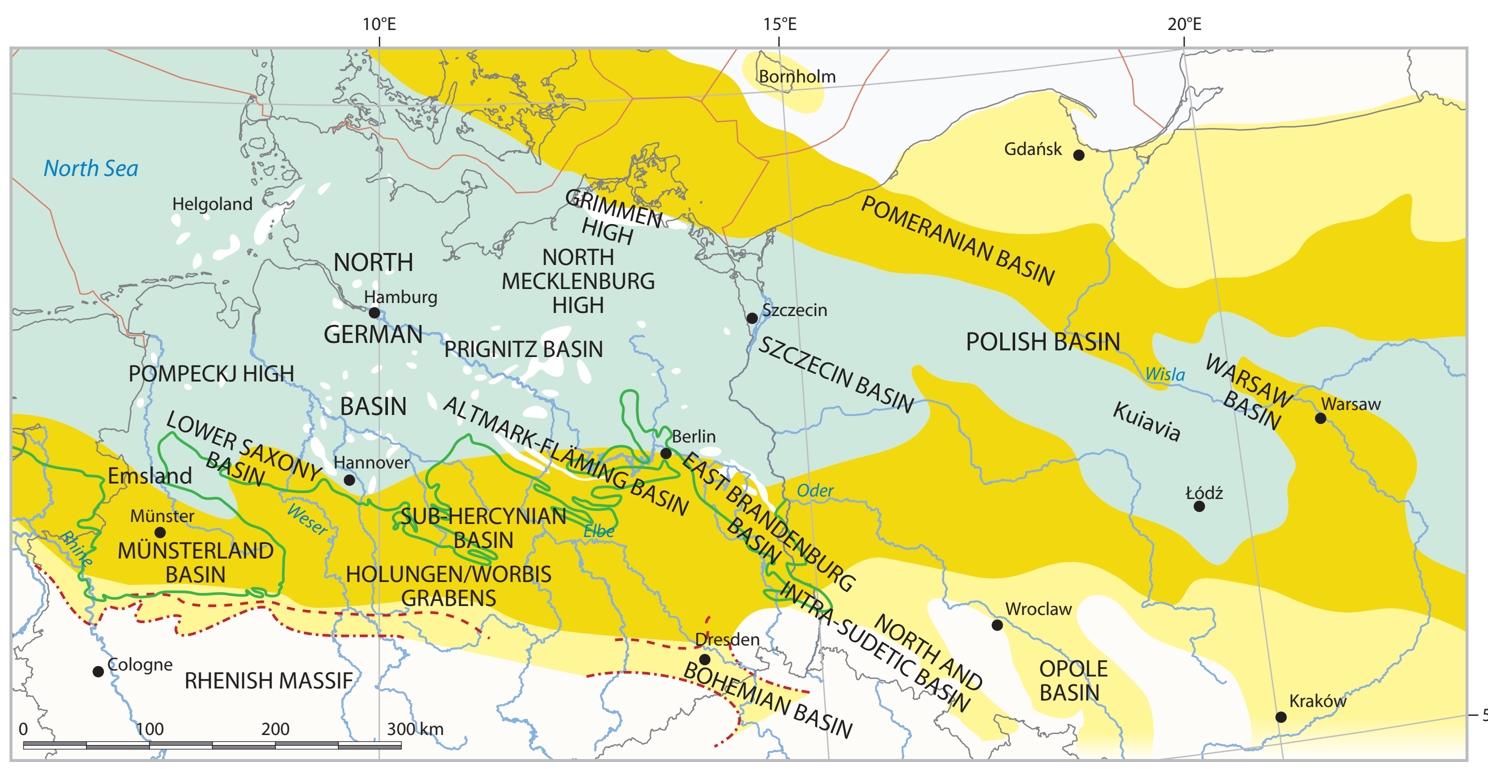
The Upper Cretaceous lithostratigraphy of the Danish North Sea follows the scheme established by Deegan & Scull (1977), which was also applied to the Norwegian North Sea. However in the Danish Central Graben, an informal subdivision into Chalk Units 1 to 6, as portrayed in the lithostratigraphic scheme in this Atlas, has proved convenient for dividing the succession according to its reservoir properties (Lieberkind et al., 1982; **Figure 11.1**). Deposition commenced with the Cenomanian Hidra Formation consisting of fine-grained, chalky, bioturbated limestone interbedded with dark grey mudstones. The clay content of this formation can be significant, although it is cleaner in the basin centres. The formation is equivalent to Chalk Unit 1 and may be up to 170 m thick, although it is generally less than 70 m. Water depths during deposition are estimated to have been several hundred metres in the Danish Central Graben and the central Danish Basin, well below the photic zone, as they were during the remainder of Late Cretaceous time. In contrast, water depths were shallower in the eastern part of the Ringkøbing-Fyn High for the duration of the Late Cretaceous, as indicated by more common hardground developments, flint bands (suggesting a greater abundance of siliceous sponges) and the particularly high proportion of benthic fauna.

The Hidra Formation is overlain by the uppermost Cenomanian to lower Turonian black mudstones of the Blodøks Formation, which grade upwards into grey mudstones and clayey limestones with a combined thickness of up to about 30 m. The formation correlates with the Herring Formation and the lower mudstone-rich sediments correlate with the lower part termed the Black Band Bed in the UK North Sea (Johnson & Lott, 1993), which represents a period of low oxygen levels at sea bed. The interval is also referred to as the Plenus Marl Formation, or as the Turonian Shale in the informal Danish unit subdivision. In eastern Denmark, the middle Cenomanian Arnager Greensand Formation is exposed on Bornholm and is up 85 m thick (Ravn, 1925). This formation has a basal phosphatic conglomerate containing reworked fossils from two Albian and one lower Cenomanian zone, which indicates several phases of erosion and deposition.

The Blodøks Formation is overlain by the upper Turonian to upper Campanian Hod Formation, which includes the upper part of the above-mentioned Turonian Shale and Chalk Units 2, 3, and 4 of the Danish scheme, which may be up to 700 m thick. The succession is chalk with a variable, though relatively high, clay content (still mainly classifying as chalk). The Hod Formation is divided into three units separated by major unconformities comprising respectively the Turonian, Turonian-Santonian, and Santonian to early Campanian intervals. There are fairly clean chalks in the lower and upper units, which may be important reservoirs (e.g. in the Valhall and Hod fields in the Norwegian North Sea; Farmer & Barkved, 1999), whereas the middle unit is more clay-rich. However, clay content also varies laterally and, together with the internal unconformities, complicates seismic interpretation such that reservoir quality is difficult to predict. Equivalents of the lower part of the Hod Formation are exposed on Bornholm with the Coniacian Arnager Limestone Formation unconformably overlying the Arnager Greensand Formation mentioned above (Noe-Nygaard & Surlyk, 1985). This formation also has a basal conglomerate with reworked fossils, indicating several phases of deposition and erosion, and is up to 20 m thick. This is followed by the Santonian Bavnodde Greensand with up to 70 m-thick argillaceous sandstones containing glauconite. Campanian proximal deposits are exposed in the calcarenites found in several localities in Skåne (Erlström & Gabrielson, 1992; Erlström, 1994; Erlström & Guy-Ohlson, 1994). Although it is not exposed, the Lund Sandstone represents a more spectacular Campanian proximal facies and comprises up to 700 m-thick clean quartz sandstones. The sandstone appears to be directly sourced from the Sorgenfrei-Tornquist Zone against which it is juxtaposed. However, the advanced chemical and mechanical maturity of the deposits suggests a source area some distance farther to the east, which has now been removed by erosion (Erlström, 1990).

The Tor Formation (overlying the Hod Formation) is mainly of Maastrichtian age, but may also include upper Campanian deposits. This formation has the cleanest chalk, with typical clay contents of only 2 to 3%. Furthermore, the chalks are formed mainly by coccoliths, a species that provides a large grain-size framework (albeit still microscopic), which gives the best reservoir properties for a given porosity when compared to other chalk formations (e.g. Brasher & Vagle, 1996). The Tor Formation is the main reservoir in many hydrocarbon fields in the Chalk Group, as for example the Valhall (Farmer & Barkved, 1999), South Arne (Mackertich & Goulding, 1999) and Dan fields (Jørgensen, 1992) to name only a few. The top of the formation is typically represented by a hardground that also marks the Cretaceous-Cenozoic boundary. Maastrichtian pelagic chalk is exposed in eastern Denmark on the island of Møn (Møns Klint; **Figure 11.16**) and in the lower part of the Stevns Klint on the east coast of Sjælland (**Figure 11.18**). These localities represent somewhat shallower water deposition compared to the central North Sea, as indicated by, for example, their higher content of benthic fauna (see above) and also possibly by their increased flint content, often arranged in bands about 0.5 m below omission surfaces. Basinal facies similar to those in the North Sea are exposed in northern Jylland near Ålborg.

Danian sediments are represented by the Ekofisk Formation, equivalent to Chalk Unit 6 in basinal settings; their maximum thickness is 350 m (Stenestad, 1972). The Danian interval was characterised by a prominent regression culminating in the early Danian, followed by a gradual overall transgression. Deposition started in large parts of the Danish Basin with a 2 to 5 cm-thick clay unit called the Fish Clay, which characterises the ecological crisis that marked the Cretaceous-Cenozoic boundary. While chalk deposition continued in the central North Sea Basin and the axial parts of the Danish Basin, large parts of the more-marginal areas of the basins, including the Ringkøbing-Fyn High margins, became dominated by bryozoans; rich chalk sedimentation developed as bryozoan limestone in places. There are local coralliferous limestones on eastern Sjælland that were presumably deposited at water depths deeper than the photic zone (Faxe limestone quarry; Willumsen, 1993). Calcarenites are also common in proximal locations along the Sorgenfrei-Tornquist Zone and in the upper part of the Danian succession, possibly reflecting contemporary inversion movements. The Ekofisk Formation chalks form important reservoirs in the North Sea, especially in the upper part. The Ekofisk reservoir is somewhat inferior in quality to the Tor Formation. This is due to the smaller grain sizes and markedly higher clay contents, especially in the lower part of the succession, which reflects the culmination of the regression. On some parts of the central Ringkøbing-Fyn High, the



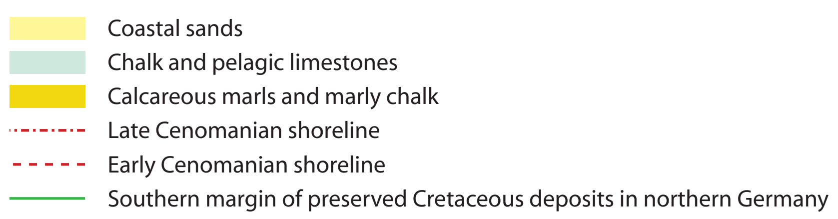


Figure 11.10 Mid-Cenomanian palaeogeography of northern Central Europe (modified after Voigt et al., 2008).

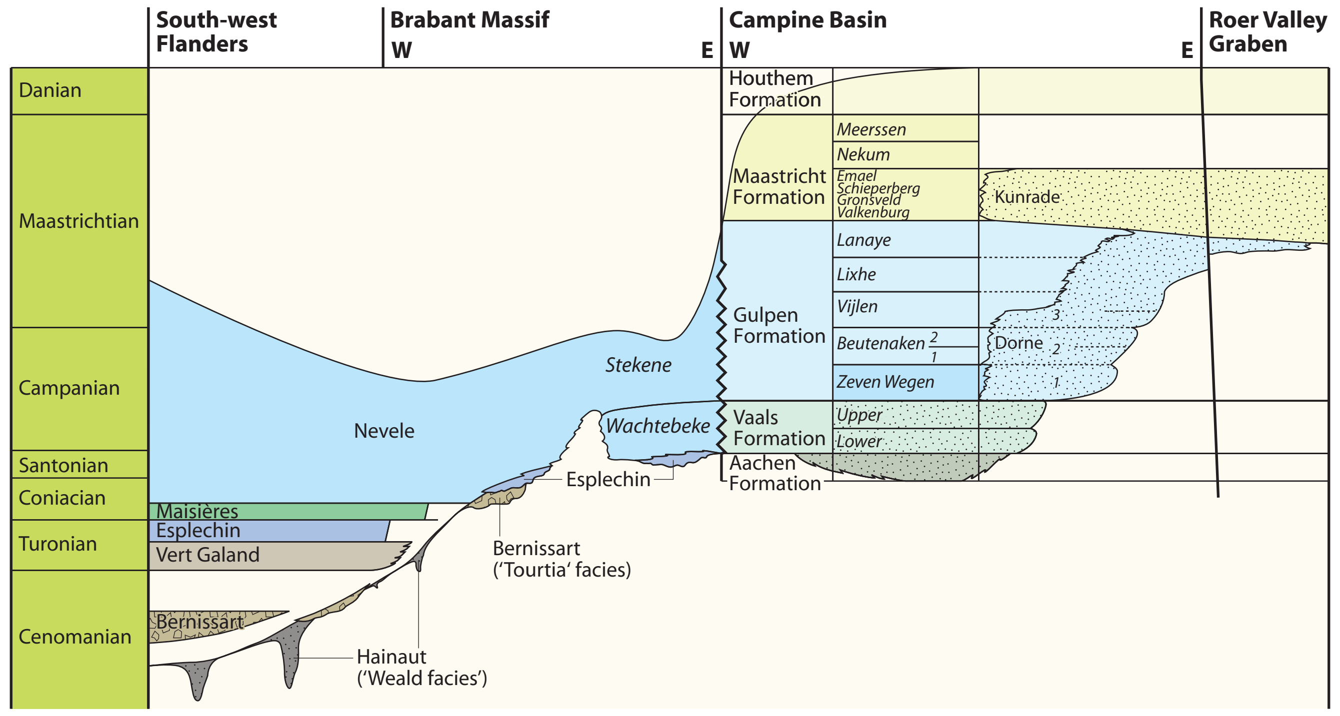


Figure 11.12 Stratigraphic scheme from the London-Brabant Massif to the Roer Valley Graben in northern Belgium.

Danian sequence represents a major hiatus with only latest Danian deposits in the form of bryozoan-limestone developments, indicating continued transgression during the Danian interval (Thomsen, 1995). The top of the Danian sequence is erosive in large areas of the Danish Basin due to a sea-level fall that was concurrent with the end of chalk-dominated deposition in north-west Europe. There is no hiatus in the central North Sea; however, carbonate deposition still came to an end over a relatively short period.

The Upper Cretaceous of the Polish Lowlands shows similar responses to inversion tectonism and eustacy seen in other parts of the SPB area, although a formal lithostratigraphic subdivision has not been presented so far. Inversion commenced in late Turonian to Coniacian times (Dadlez et al., 1995; Leszczyński, 2002b; Gutowski et al., 2003; Krzywiec, 2006b). The transgressive late Albian to Cenomanian depositional phase therefore had a depocentre with up to 150 m of sediment in the axial parts of the Mid-Polish Trough (and Miechów Depression) where pelagic limestone (chalk) deposition indicates their remoteness from coastal areas. However, there was siliciclastic input to the north-east and south-west of the trough in the form of calcareous marls, which became thinner with distance from the trough. The sea also extended into the Sudetic and Opole basins where there are proximal deposits of glauconitic to marly sands with phosphatic nodules. Thinning sediments with associated condensed sequences, hardgrounds and erosion are seen south-eastwards towards the present-day Carpathians.

A large areal expansion of pelagic limestone and chalk sedimentation took place from the late Turonian until Maastrichtian times, eventually covering almost the entire Polish Lowlands (e.g. Leszczyński, 1997b). Siliciclastic deposition in the form of siltstones and sandstones continued until mid-Santonian times in the Łeba High and as argillaceous limestones and marls in north-easternmost and south-westernmost parts of present-day Poland respectively. The focus of pelagic limestone and chalk sedimentation also extended laterally southwards with increasing thicknesses and a consequent shift of the area of condensed sedimentation closer to the Carpathians. Maximum subsidence is inferred (due to later removal by erosion) to have been centred along the Mid-Polish Trough until Santonian times, whereas depocentres shifted to the flanking synclines along the margins of the Mid-Polish Trough during Santonian to Maastrichtian times as a consequence of the main phases of inversion (Krzywiec, 2006b). Uppermost Maastrichtian and Lower Paleocene (Danian) sediments are found only to the north-east of the Mid-Polish Trough because the Szczecin-Łódź-Miechów syncline was also uplifted during the later phases of inversion

Figure 11.13 The Obernkirchen Sandstone ('Wealden' facies of the Berriasian) showing ripple marks and a dinosaur footprint (Münchehagen, N Germany; photo: J. Mutterlose). See Figure 11.24 for location.

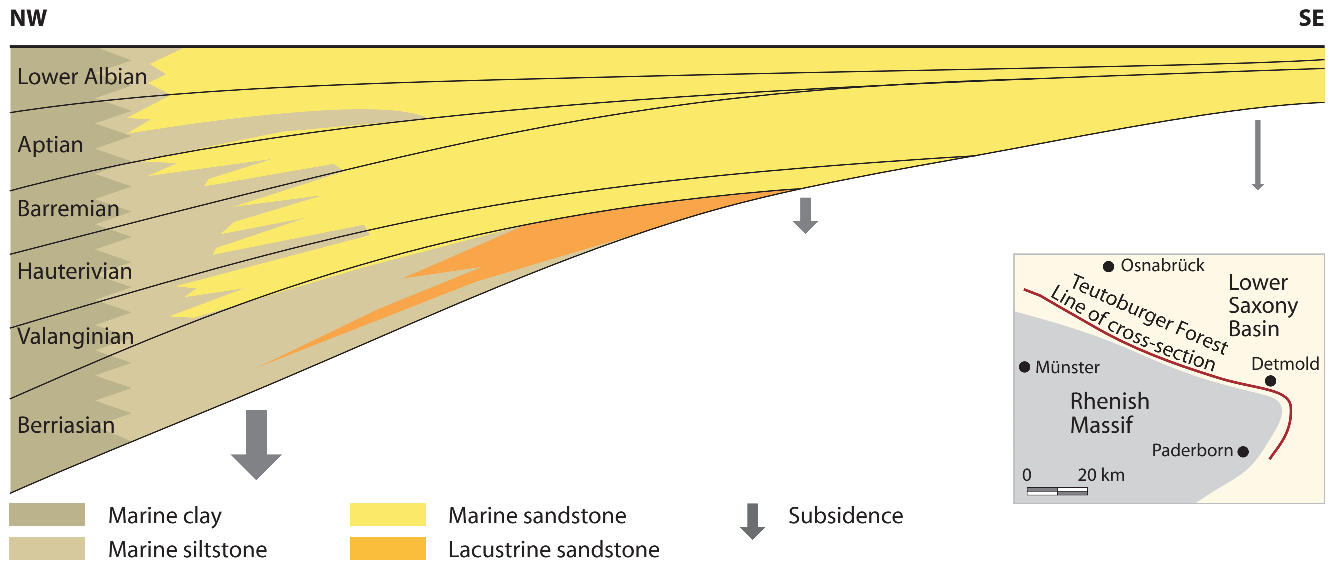


Figure 11.14 Schematic cross-section of the Osning Sandstone in the Bocketal area of the Teutoburger Forest (from Mutterlose & Bornemann, 2000).

(Jaskowiak-Schoeneichowa & Krassowska, 1988; Marek & Pajchlowa, 1997). These deposits are pelagic-limestone facies proximal to the Mid-Polish Trough grading into siltstones and sandstones to the north (Gdańsk area) and the east (east of Warsaw to the Lublin area).

In the area of present-day northern Germany, the various sub-basins coalesced into one broad shelf flanked by the Rheno-Bohemian Massif to the south, as a consequence of mid-Albian transgression. Middle to upper Albian marls overlie Jurassic rocks at the southern margin of the Pompeckj Block; towards the Schleswig-Holstein area they overlie Triassic strata, whereas in the Mecklenburg area, upper Albian glauconitic sandy phosphatic deposits followed by marls overlie a variety of pre-Cretaceous to Lower Cretaceous sediments (**Figure 11.7**). Farther south, along the northern margin of the Rhenish Massif, for example, at the southern margin of the Münsterland Basin, conglomerates and greensand deposits in nearshore facies of mid- to late Albian age rest on Paleozoic rocks and pass laterally into marls (Flammenmergel) and marly limestones. These marginal complexes of glauconitic sandstones, marls and marly limestones extended diachronously farther south during the Cenomanian transgression.

Until early Coniacian times, deposition was governed by thermal subsidence and eustacy. A simple zonation scheme can be distinguished during this time, with a distal zone to the north contiguous with the Danish Basin, a medial zone, and a proximal zone close to the Rheno-Bohemian Massif to the south. There was strong diversification of this pattern linked to the main phases of the inversion during Coniacian to Maastrichtian times.

The distal zone of the early Late Cretaceous includes, as before, the flooded Pompeckj, North Mecklenburg and East Brandenburg highs and is continuous with the Danish Basin. In the Helgoland area, Cenomanian to Turonian deposits are about 30 m thick and consist of marly limestone, bioclastic calcarenites and chalk; the sequence becomes more calcareous upwards reflecting its transgressive development. In the upper Cenomanian to lower Turonian succession, the black shales are the result of a global anoxic event, whereas during mid- to late Turonian times there was a return to deposition of chalk with flint. A hiatus is seen in the lower Coniacian succession of the Helgoland area (Wood & Schmid, 1991). In the eastern part of the Pompeckj High, Cenomanian sediments are thicker (reaching 70 m) and more argillaceous and are likewise followed by deposition of limestones and thin black shales. The upper Turonian is a succession of limestones with marly intercalations containing flint. Farther east (Mecklenburg, Prignitz-Lausitz High area), the Cenomanian succession is up to 30-40 m thick and overlain by 150-170 m of Turonian sediments

Figure 11.15 Cenomanian-Turonian succession of the Halle section (Westphalia, N Germany) showing the black shales (Oceanic Anoxic Event 2; photo: J. Mutterlose). See Figure 11.24 for location.

Figure 11.16 Exposure at Møns Klint, Denmark showing Lower Maastrichtian chalk tectonically stacked by thrusting due to glacial compression. Photo: Peter Moors (GEUS). See Figure 11.24 for location.

that may be up to 400 m thick in salt rim-synclines. The late Albian to Cenomanian transgression also encompassed the western Pomeranian area where upper Albian glauconitic pebbly sandstones are up to 10 m thick and form the basal sequence of the Upper Cretaceous followed by up to 30 m of greenish and reddish marls. A complete Upper Cretaceous succession more than 600 m thick is found in the Rügen area between the Grimmen High and Pomeranian Kuiavian Anticline. The succession consists of rather monotonous chalks similar to the Danish sequence, but without hardgrounds, suggesting a somewhat deeper-water position in the basin; the chalks are exposed in glacial thrust sheets on the northern coast of Rügen (**Figure 11.17**). The Grimmen High is assumed to have been uplifted during Maastrichtian times.

The medial zone includes the Lower Saxony and Münsterland basins, which can be considered to be erosional remnants following inversion uplift. There was a marked increase in carbonate content in the basin sediments following the mid-Albian transgression, and the middle to upper Albian marls and clayey marls (the Flammenmergel) may reach up to 300 m in thickness. The Albian-Cenomanian transition is a minor hiatus and the Cenomanian greenish silty marls have a thin glauconitic unit at the base. The overlying succession consists of marl-limestone interbeds, with marls becoming less prominent upwards. The middle to upper Cenomanian strata are almost pure limestones (clay content ~5%) reflecting the impact of the transgression. Thicknesses reach 120 to 150 m, thinning towards the basin margins, but with the upper pelagic units having a more constant thickness. Late Cenomanian to early Turonian regression led to further depositional facies variation, with black mudstones alternating with grey limestones deposited in the Lower Saxony and Münsterland basins (**Figure 11.15**). Transgressive encroachment in the Münsterland area is marked at the base of the succession by glauconitic sands (Bochum Greensand). This is followed by middle Turonian grey to white limestones (Weisspläner) in a deeper-water setting and condensed, red, nodular-limestone developments on swells; the combined lower to middle Turonian succession is up to 50 m thick. The upper Turonian white limestone deposits are between 4 and 120 m thick. Sediments up to 400 m thick are associated with marly intercalations and are seen locally in the Münsterland area. Early Coniacian inversion activity resulted in the deposition of marly-silty fossiliferous limestones grading into calcareous marls known as the Emsher Marls, which lie unconformably on upper Turonian strata in the Sachsen-Anhalt area.

Figure 11.17 Lower Maastrichtian chalk tectonically stacked by thrusting due to glacial compression on the northern coast of Rügen Island, NE Germany. Photo: Dieter Stellmacher. See Figure 11.24 for location.

The proximal zone comprises marginal facies found at the southern edge of, for example, the Münsterland, East Brandenburg and Sub-Hercynian basins, the Holunger Graben in Thüringia and other minor accumulations at the northern margin of the Rhenish Massif. The extent of the initial transgression is shown by middle to upper Albian nearshore, conglomeratic, sandy strata and greensands, which overlie Paleozoic basement in the southern Münsterland Basin (Wünnenberg and Rüthen beds) (**Figure 11.1**). Sandstones, glauconitic sandstones, glauconitic marls and marly limestones encroached farther south during the Cenomanian transgression. Turonian glauconitic marls and limestones are found as far south as the Dortmund area; the uppermost Turonian strata have no preserved proximal facies.

Coniacian to Maastrichtian times were dominated by inversion tectonics in the area between the Rhenish Massif and the inundated Ringkøbing-Fyn High. Inversion activity started during the late Turonian and continued into Santonian times. Complete inversion, during which the Upper Cretaceous deposits were fully removed, is seen in the Lower Saxony Basin, Prignitz-Lausitz High, the Grimmen High near Rügen, and perhaps most prominently with the uplift of the Harz Mountains. Here, with an estimated 7000 m of uplift relative to the adjacent Sub-Hercynian syncline, the inversion removed the Mesozoic succession completely, exposing and eroding Devonian and Carboniferous rocks (Chapters 5 & 6). These inversion structures are flanked by generically related, rapidly subsiding marginal troughs as seen in the Münsterland Basin and the Sub-Hercynian, Altmark-Fläming and East Brandenburg basins. Tectonic activity abated during the Campanian to Maastrichtian. In the distal zone mentioned above, pelagic chalk deposition prevailed from Turonian to Maastrichtian times, with maximum water depths estimated at 150 to 200 m, whereas areas close to inversion structures show evidence of a change from pelagic deposition to siliciclastic domination expressed by the dark-coloured Emscher Marl from Coniacian through to the end of Cretaceous times. Maximum chalk thicknesses of about 1000 m are seen in rim-synclines in the Lower Saxony Basin, Schleswig-Holstein, south-west Mecklenburg (Baldschuhn et al., 1977) and offshore of Rügen (Petzka & Reich, 2000a).

In the Münsterland Basin, clastic deposition gradually began to dominate over the carbonates during Cenomanian to Coniacian times. In the west, the inverted Central Netherlands Basin, with erosion of Carboniferous to Triassic rocks, contributed to the Coniacian Emscher Greensand and the Santonian marine sandstones of the Recklinghaus and Haltern beds that grade laterally into the Emscher Marls. Upper Santonian to lower Campanian calcarenites were also deposited, followed by lower to middle Campanian sandy marlstones of the Holtwick and Coesfeld beds. Glauconitic calcarenites of the Baumberge Beds were deposited as a shallow-marine facies during the late Campanian.

The Sub-Hercynian and Altmark-Fläming Cretaceous basins had very similar developments characterised by uplift of the adjacent inversion structures. Deposition commenced during the Turonian with hemipelagic limestones and continued until Coniacian times. Conglomerates, sandstones and calcareous siltstones were deposited during the Coniacian to early Campanian interval. Hemipelagic limestones up to about 300 m thick were deposited in both basins during the Cenomanian to Turonian. Sedimentation during the Coniacian to Santonian changed to silty marls in a nearshore environment, with a thickness of about 600 m in the Altmark-Fläming Basin. However, in the Sub-Hercynian Basin, the sediments are conglomerates, sandstones and calcareous siltstones deposited during the Coniacian to early Campanian interval, including the middle Coniacian Halberstadt (up to 150 m thick), middle to upper Santonian Sudmerberg (some 180 m thick), Salzberg (250 m thick) and upper Santonian Heidelberg formations (450 m thick); all were deposited in nearshore to coastal-plain environments. The clastic sediments are the result of contemporaneous uplift and erosion of the Harz Mountains. The uppermost deposits of the latest Santonian to early Campanian Heimburg and Blankenberg formations form narrow fringes to the Harz area and contain clasts of Devonian slates and Carboniferous greywacke sandstones (**Figure 11.11**). Laterally, these deposits pass into marl-dominated units such as the Emscher Marls. Abating inversion tectonism was coeval with deposition of the upper Campanian to Maastrichtian, marine to deltaic, glauconite sands of the Oebisfelde and Steinförde formations in the Altmark-Fläming Basin, which straddle the former boundaries of the basin and are up to 600 m thick.

Figure 11.18 Exposure of the Cretaceous-Paleogene boundary at Stevns on the east coast of Sjælland, Denmark. The Danian bryozoan limestone is the light grey rock overhanging the white Maastrichtian chalk (photo: O.V. Vejbæk). See Figure 11.24 for location.

Figure 11.19 Maastrichtian chalk exposed in the Kroonsmoor limestone quarry, northern Germany (photo: A.S. van der Molen). See Figure 11.24 for location.

3 Chalk reservoirs

The North Sea chalk reservoirs are characterised mainly by their high-porosity/low-permeability relationship. Superior reservoir quality in terms of favourable porosity-permeability properties is typically found in the upper Maastrichtian Tor Formation, whereas the Danian Ekofisk Formation is of poorer quality and the Barremian Tuxen Formation is the least favourable reservoir (**Figure 11.20**). The Campanian Hod Formation is of variable quality, broadly comparable to the Tor Formation, but generally closer to the Ekofisk Formation. Reservoir quality in the Tor Formation is determined by its effective stress and so also by overpressure, compaction and retardation due to early hydrocarbon invasion and to a lesser degree by subtle variations in depositional environments (Anderson, 1999; Surlyk et al., 2003). In the Ekofisk, Hod and Tuxen formations, reservoir quality is also controlled by their clay content. Clay can be significant in these formations because it increases the residual water saturation, lowers the permeability, and causes an increase in capillary entry pressures (**Figure 11.23**). Large distances between oil-water contacts due to capillary forces are a special trait of chalks. Compaction is dominated by pressure-dissolution from stylolites and reprecipitation at depths of about 900 m and deeper (Scholle et al., 1998; Fabricius, 2003; Fabricius & Borre, 2007; **Figure 11.21**). *In situ* hydrocarbons may therefore retard compaction inflicted by further effective stress increases by blocking the dissolution-reprecipitation process. A porosity of 45% is seen at depths of around 3000 m, with an overpressure of about 150 bars preserved by hydrocarbons where water-bearing chalks have porosities less than 25% (**Figure 11.22**). Porosity decay requires increases in effective stress, so if effective stress increases are halted by overpressuring before hydrocarbon charging, then porosity-preserving effects from hydrocarbons will be negligible.

Traps within the Chalk Group range from inversion-generated anticlines (such as the Valhall, Roar, Tyra and South Arne fields), traps over salt domes with some degree of inversion overprint (e.g. the Dan, Ekofisk and Svend fields) to salt diapirs (e.g. the Skjold and Harald fields). Stratigraphic traps may also play a major role (e.g. Halfdan and Adda fields). Trapping in chalk is often associated with dynamic-fluid phases because the fluids require geological timescales to reach equilibrium due to the low permeability of the chalk (Megson & Hardman, 2001; Dennis et al., 2005; Vejbæk et al., 2005).

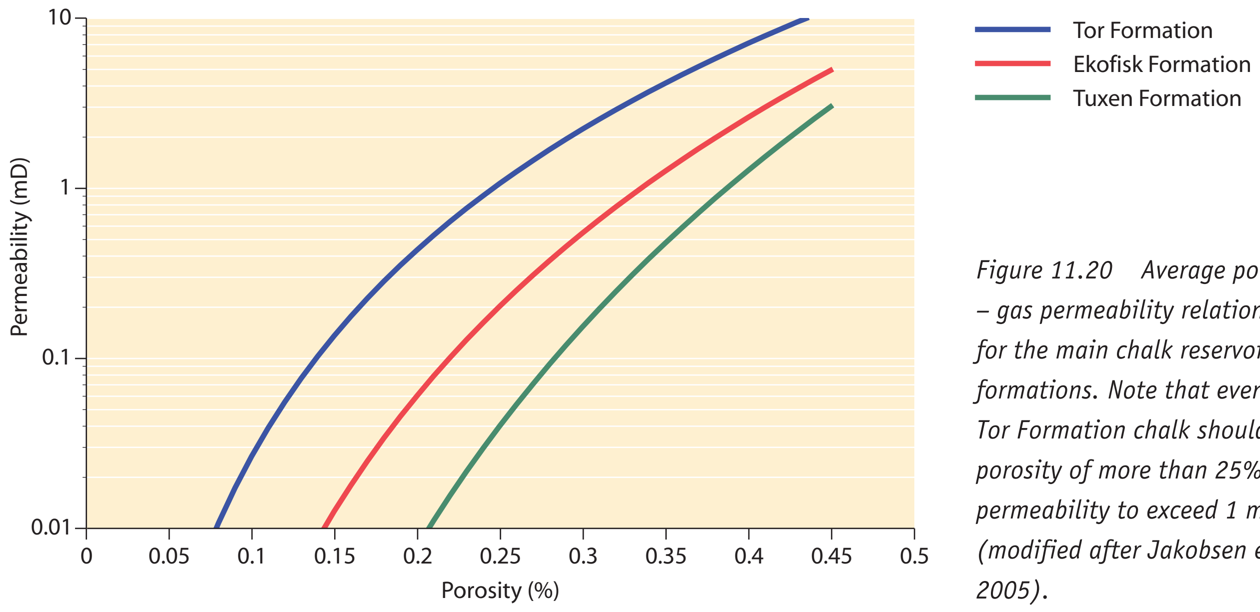


Figure 11.20 Average porosity – gas permeability relationships for the main chalk reservoir formations. Note that even the Tor Formation chalk should have porosity of more than 25% for permeability to exceed 1 mD (modified after Jakobsen et al., 2005).

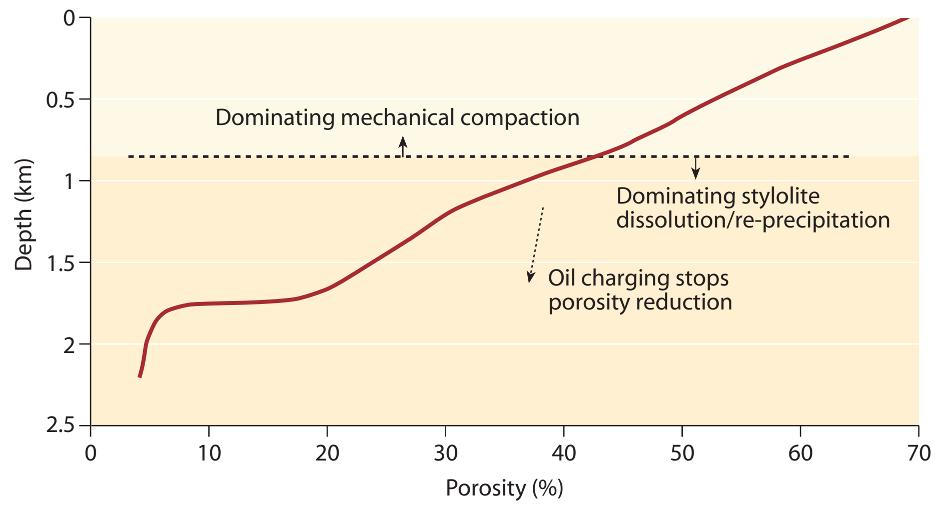


Figure 11.21 Idealised porosity-depth trend for chalk in the absence of overpressuring. The depth here is a proxy for effective stress and therefore changes if overpressured due to rapid burial. Further deviation will be caused by early oil invasion as it retards chemical compaction (see Fabricius, 2003; Japsen et al., 2006; Fabricius & Borre, 2007).

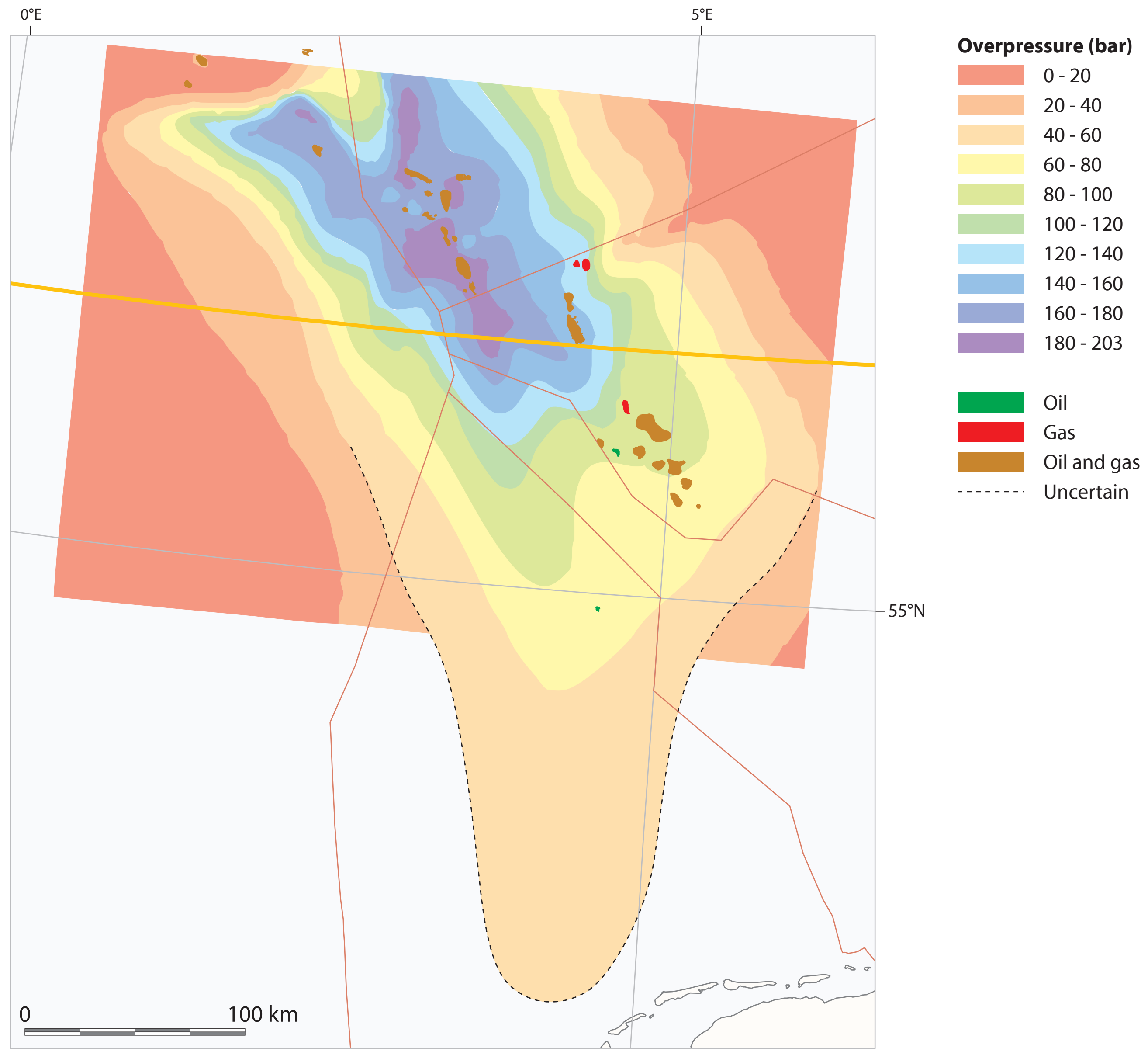


Figure 11.22 Fluid overpressure at top Chalk Group level in the central North Sea (modified after Dennis et al., 2005). The overpressure is primarily generated by rapid Neogene deposition (Osborne & Swarbrick, 1997; Japsen, 1998; Vejbæk, 2008). The overpressure therefore mimics the Neogene thickness except where continuous Paleogene sand layers locally drain the overpressure.

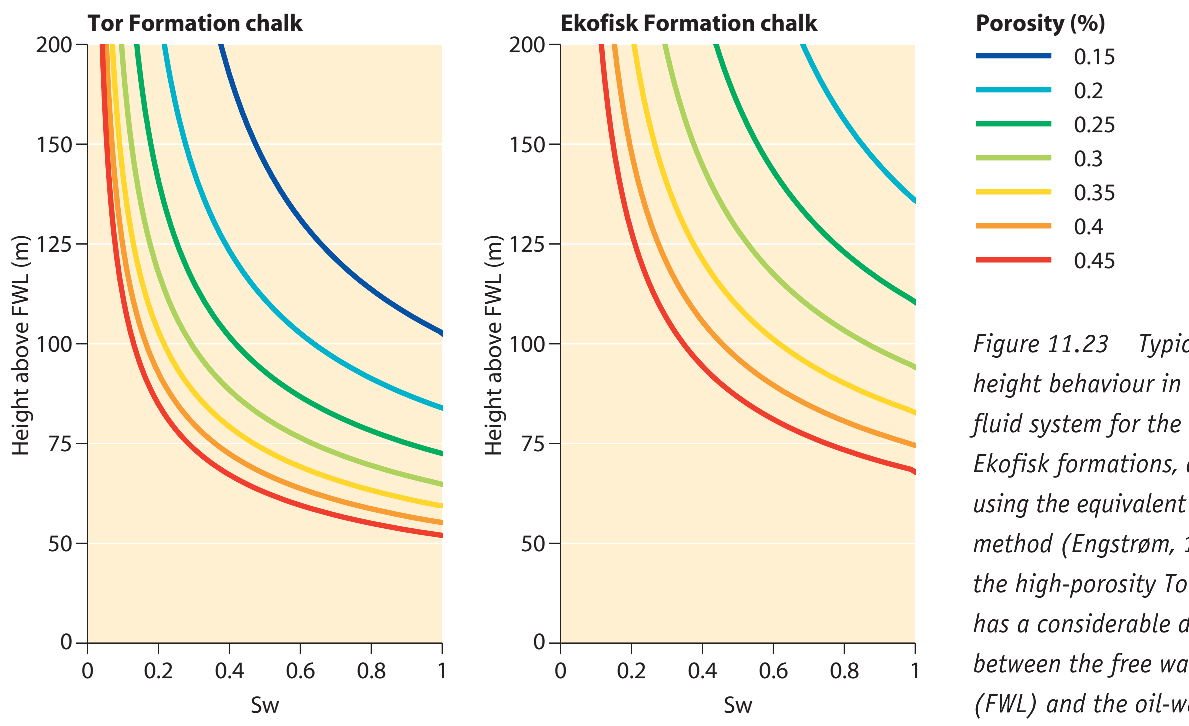


Figure 11.23 Typical saturation height behaviour in an oil-water fluid system for the Tor and Ekofisk formations, as predicted using the equivalent radius (EQR) method (Engstrøm, 1995). Even the high-porosity Tor Formation has a considerable distance between the free water level (FWL) and the oil-water contact.

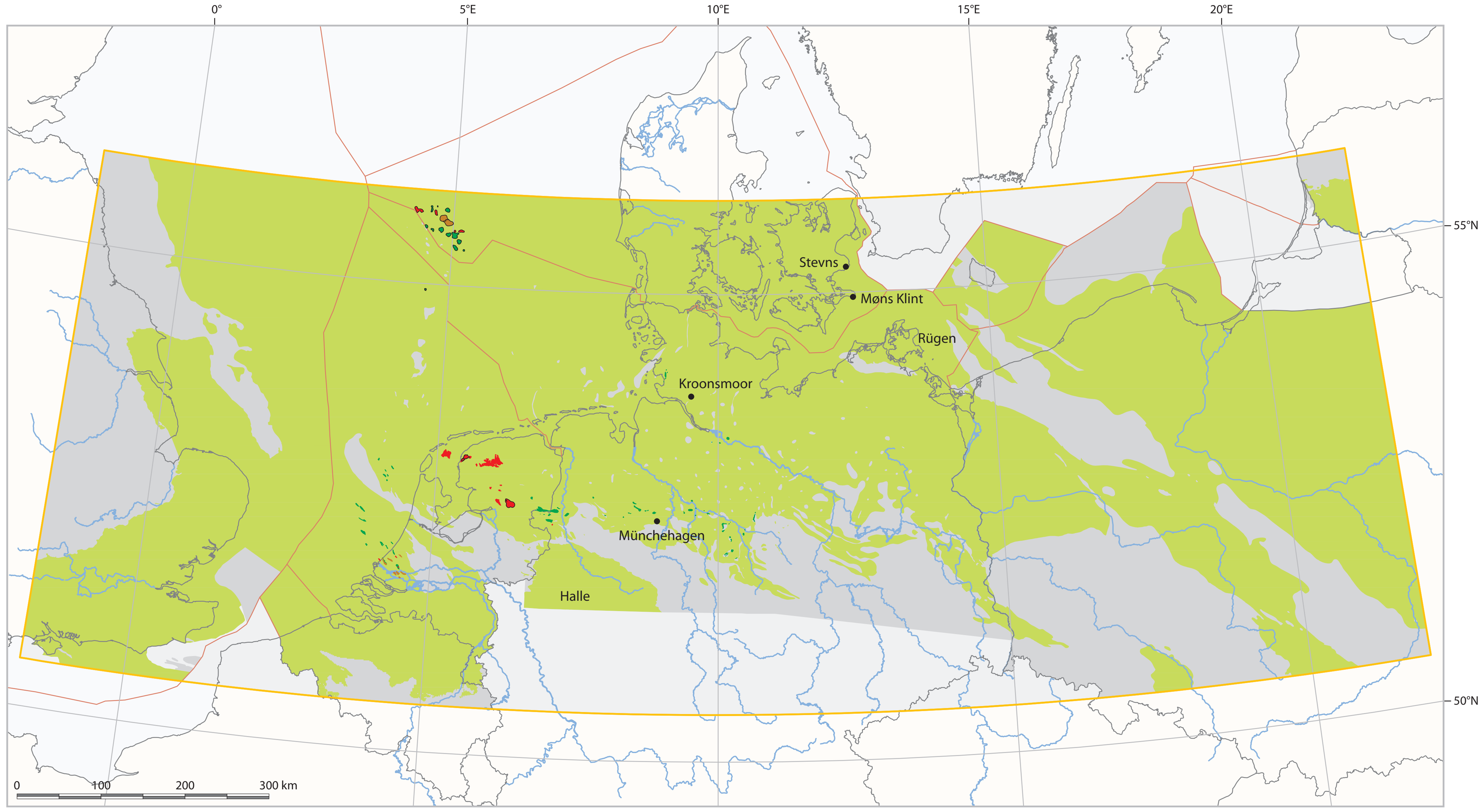


Figure 11.24 Distribution of Cretaceous hydrocarbon reservoirs. Fields with Cretaceous reservoir.

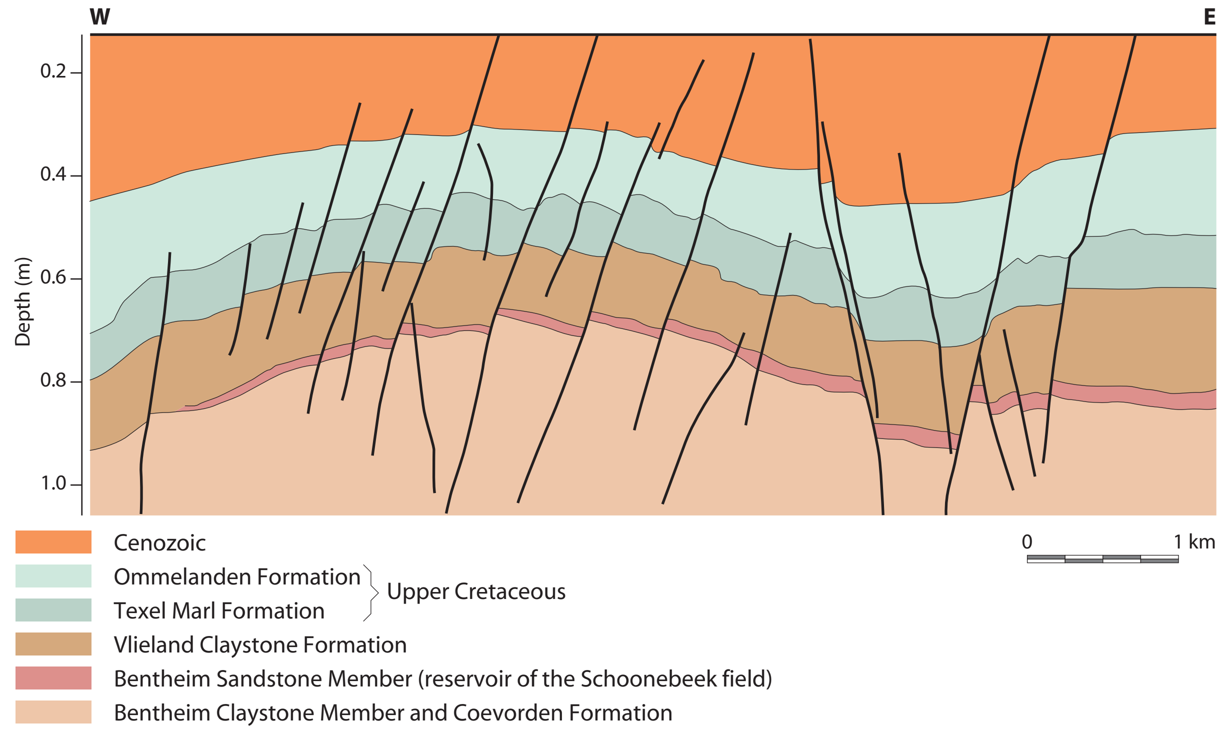


Figure 11.27 Geological cross-section through the Schoonebeek field. See Figure 11.26 for location.

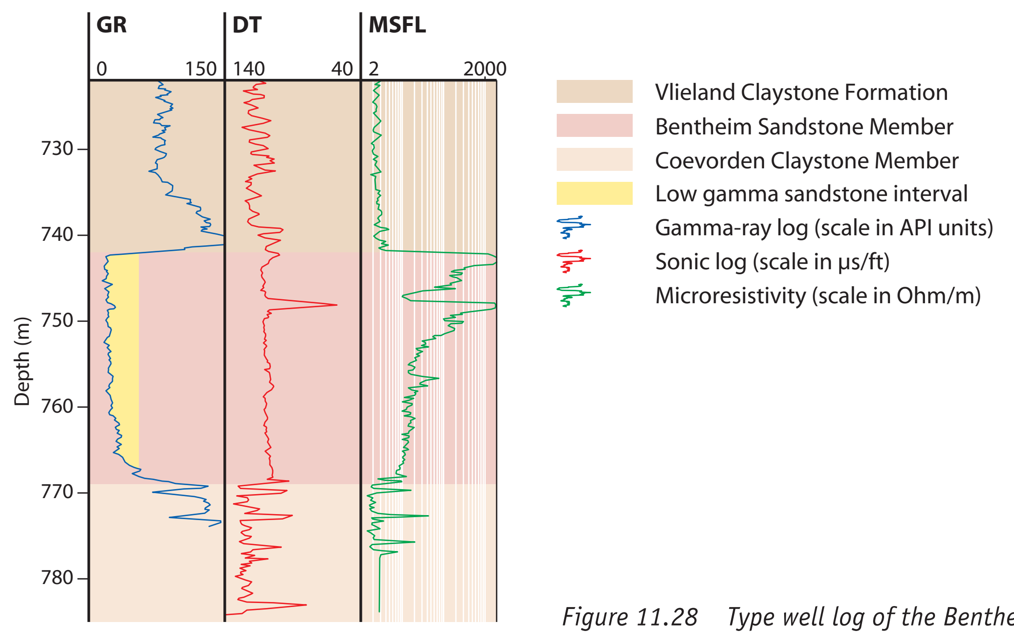


Figure 11.28 Type well log of the Bentheim Sandstone Member.

4 Hydrocarbon field examples

4.1 *Schoonebeek oilfield, onshore Netherlands*

The Schoonebeek oilfield is the largest onshore oilfield in western Europe. It is located mostly in the north-eastern part of the Netherlands, but straddles the border with Germany (**Figure 11.25**). The field was discovered in 1943. Production started in 1944 and lasted for more than half a century during which time 599 wells were drilled. The field was abandoned in 1996; however, a decision was recently taken to re-develop part of the Schoonebeek oilfield and resume production. The field contains 164 mln m^3 (1 bln bbl) STOIIP of medium to heavy, highly viscous oil (160 cp at 40°). A quarter of this (40 mln m^3 of oil) has so far been produced by a combination of primary recovery, cold- and hot-water flooding and steam-flooding techniques.

The Schoonebeek oilfield is located in the western part of the Late Jurassic to Early Cretaceous Lower Saxony Basin, which hosts more than 95 oilfields. There are several source and reservoir rock intervals in the basin (Kockel et al., 1994). The field is 16 km long and between 1 and 4 km wide, and lies at a depth of about 650 to 900 m (**Figure 11.26**). The field occupies a heavily faulted east–west-trending anticline that formed during Late Cretaceous basin inversion. The structure forms a dip closure north, east and south of the field. A depositional pinch-out forms the western boundary of the field (**Figures 11.25** & **11.27**). The conformably overlying Lower Cretaceous (Valanginian) Holland Claystone Formation forms the top seal (**Figure 11.28**). The oil in the Lower Saxony Basin is sourced from the Lower Cretaceous (Berriasian) Wealden 'paper shales' (Kockel et al., 1994) (**Table 11.1**).

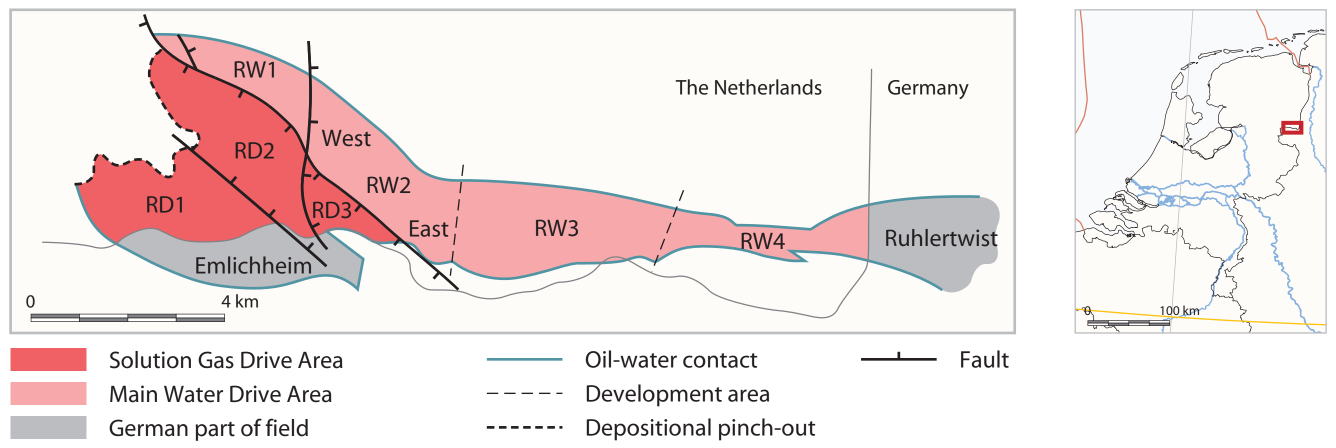


Figure 11.25 Location of the Schoonebeek oilfield.

The reservoir is formed by the unconsolidated, clean and well-sorted regressive shoreface sandstones of the Lower Cretaceous (Valanginian) Bentheim Sandstone Member, which are up to 30 m thick. A large, north-west– south-east-trending, sealing normal fault complex divides the field into two areas with separate natural drive mechanisms (**Figure 11.25**). The eastern part, known as the Main Water Drive Area or MWDA, holds two-thirds of the STOIIP (110 mln m^3). This part of the field is connected to a large aquifer, which results in a relatively high oil recovery (30% of STOIIP) but with high water cuts. The western part of the field is known as the Solution Gas Drive Area or SGDA (**Figure 11.25**) and holds the remaining one-third of the STOIIP (54 mln m^3). This part lacks aquifer support and consequently has a lower recovery (~15% of STOIIP). The high oil viscosity and resulting low primary production has led to the application of a variety of Enhanced Oil Recovery (EOR) methods over the years (Troost, 1981; Combe et al., 1990). Water broke through in the MWDA only a few years into production, which was soon reinjected to aid production. This was followed by a hot-water injection a few years later, steam injection in the 1970s, and finally high-pressure steam injection in 1981 (Holtam et al., 1990). In the SGDA, steam injection was undertaken in 1960 and 1975, followed by cold-water injection.

The recent decision to redevelop the SGDA section of the field is based on the use of a low-pressure, gravity-assisted, steam-flooding process. Sixty-nine horizontal wells (25 injector and 44 producer wells) will be drilled in the lower part of the reservoir. The injected steam will flow along the top of the reservoir to form a downward, gravity-assisted water drive. A high-resolution 3D-seismic survey was acquired in 2005 to better image the many faults in the field. This in turn helped the placement of the horizontal wells in the lower part of the 12 to 30 m-thick reservoir.

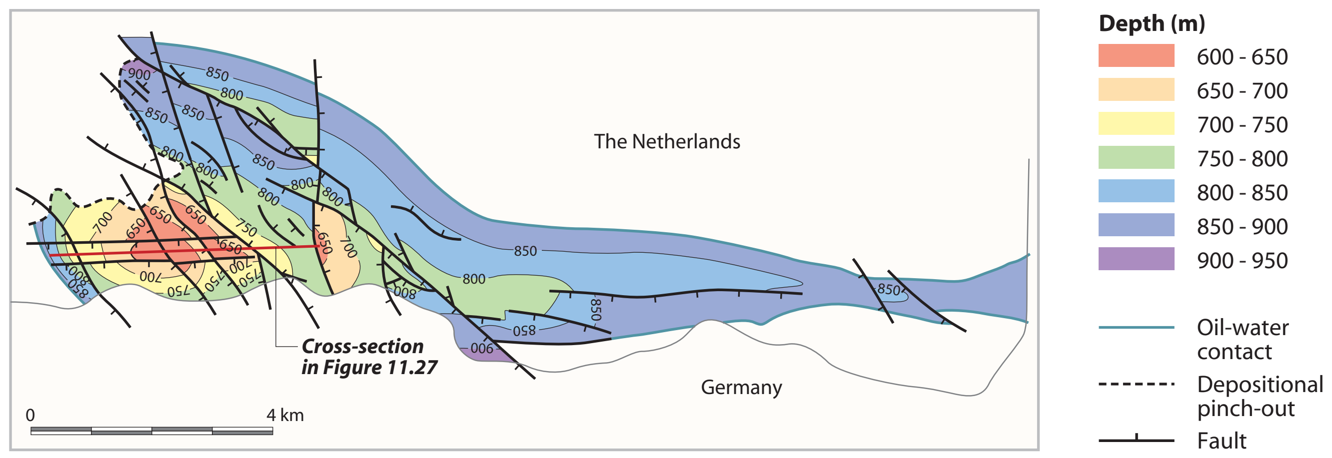


Figure 11.26 Depth-structure map of the top of the Schoonebeek field reservoir.

Table 11.1 Properties of the Schoonebeek oilfield.

Reservoir	Bentheim Sand Member (Valanginian)
Lithology	Sandstone
Depth to top (m)	630-900
GWC/GOC.OWC (m)	900 (varies for different fault blocks)
Maximum column height (m)	300
Net reservoir thickness (m)	11 293
Net to gross ratio	0.98
Porosity (%)	30
Permeability (mD)	500-4000
Fluid type	Oil
Oil gravity	25
Initial pressure (bar)	85
Temperature (°C)	40
Source rock	Wealden paper shales
Seal	Holland Claystone Formation

4.2 *Halfdan oilfield, offshore Denmark*

The Halfdan field is situated offshore of Denmark, approximately 220 km west of Esbjerg. The Halfdan was discovered in 1999 with a 9.1 km-long horizontal well (Jacobsen et al, 1999) drilled from the Dan field and the vertical Nana-1XP well completed in 1999. The horizontal well was drilled to test the down-flank potential in the direction of an acoustic impedance anomaly to the north-west. The well discovered a continuous oil column and a strongly tilted Free Water Level (FWL) from Dan to Halfdan in the Tor Formation below the structural spill-point of Dan. The height difference in FWL is ~215 m between the two. STOIIP is currently estimated to be in excess of 200 mln m^3. A small gas cap exists in the developed area.

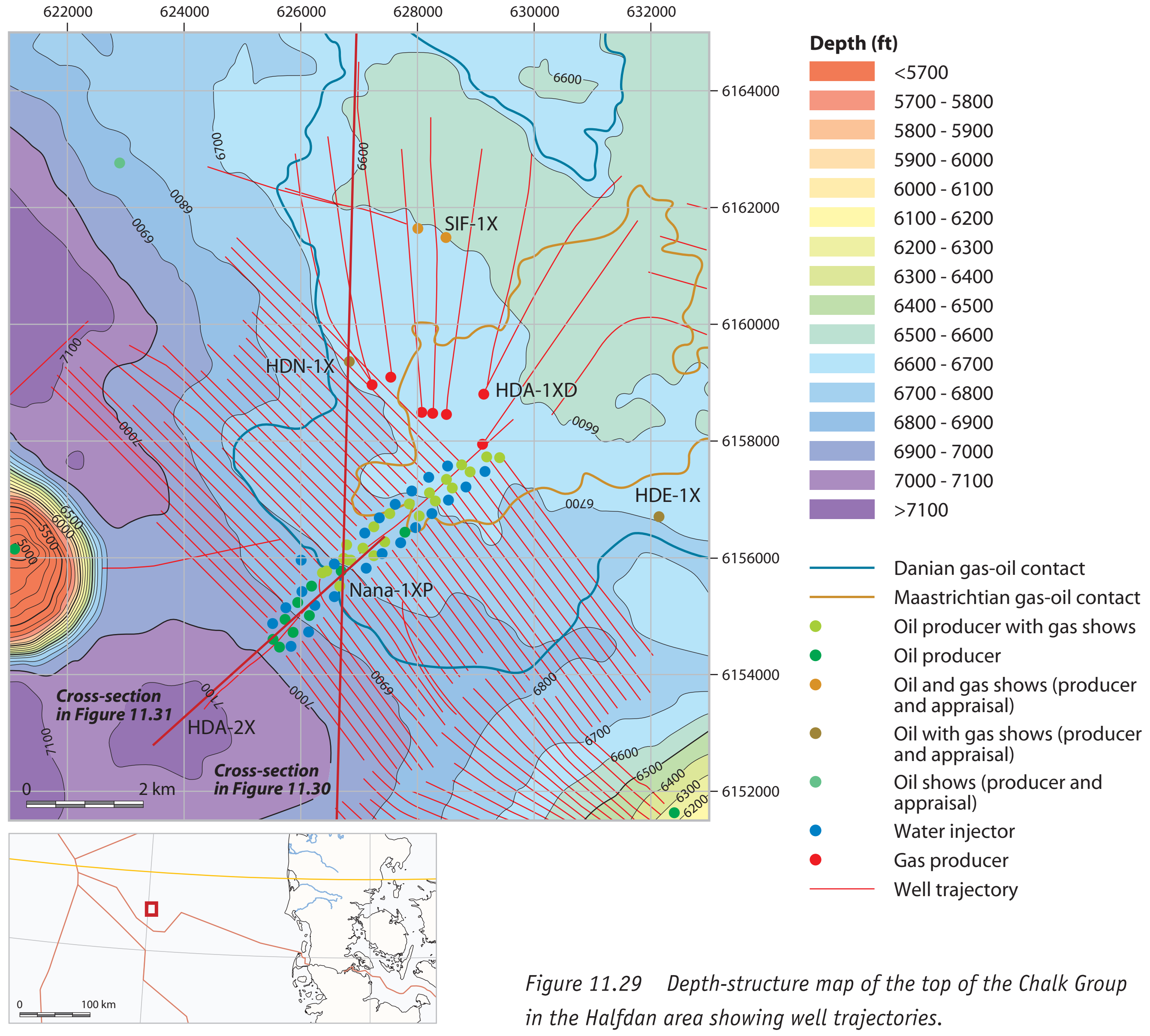


Figure 11.29 Depth-structure map of the top of the Chalk Group in the Halfdan area showing well trajectories.

The Halfdan field is a non-structurally trapped oil accumulation in Maastrichtian Tor Formation chalk located in the Danish Central Graben between the Dan and Skjold chalk fields (**Figure 11.29**). The Halfdan field is located on the slope of a north-west–south-east-oriented Upper Maastrichtian basin that thins out north-eastwards to the so-called Tyra-Igor inversion ridge where the interval is very condensed (**Figure 11.30**). Condensation is generally associated with deterioration of the reservoir properties as illustrated on the acoustic impedance section (**Figure 11.31**). There is good evidence that the oil that is currently non-structurally trapped in Halfdan is not in equilibrium, i.e. it is in motion (Albrechtsen et al, 2001). Gas-oil contacts are tilted, and pressure data from wells in the oil zone show that lateral changes within a continuous oil leg imply moving oil. The complex fluid distribution is controlled by factors such as primary depositional characteristics, charge history, regional tectonic evolution, capillary pressure effects and regional fluid dynamics. The Halfdan field is thought to have constituted an earlier closure that disappeared some 4 Ma ago, caused by late tilting of the basin. Up-dip migration of oil in the direction of the inversion structure to the north-east is arrested by the condensed high-porosity interval. Instead, oil is migrating south-eastwards to the present-day Dan field. The low permeability of the chalk and the high capillary entry pressures result in very low migration velocities, also when measured in geological time. Megson & Tygesen (2005) have used the term 'constriction trap' for this type of non-structural trap where the hydrocarbon accumulation has not reached a state of equilibrium.

The Halfdan reservoir comprises unfractured, low-permeability chalk (0.5-4 mD) of mainly Maastrichtian age (Albrechtsen et al, 2001) (**Table 11.2**). The porosity and saturation profile for the Danian and uppermost Maastrichtian chalk intervals in the vertical Nana-1XP well is shown in **Figure 11.32**. The

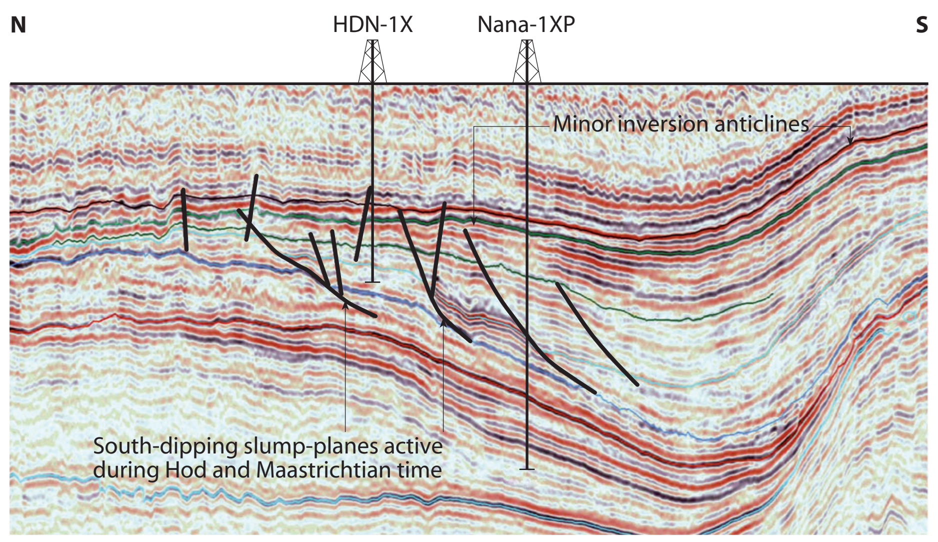


Figure 11.30 Seismic section across vertical wells Nana-1XP and HDN-1, showing northward thinning of the upper Maastrichtian section and intra-chalk slumping. See Figure 11.29 for location.

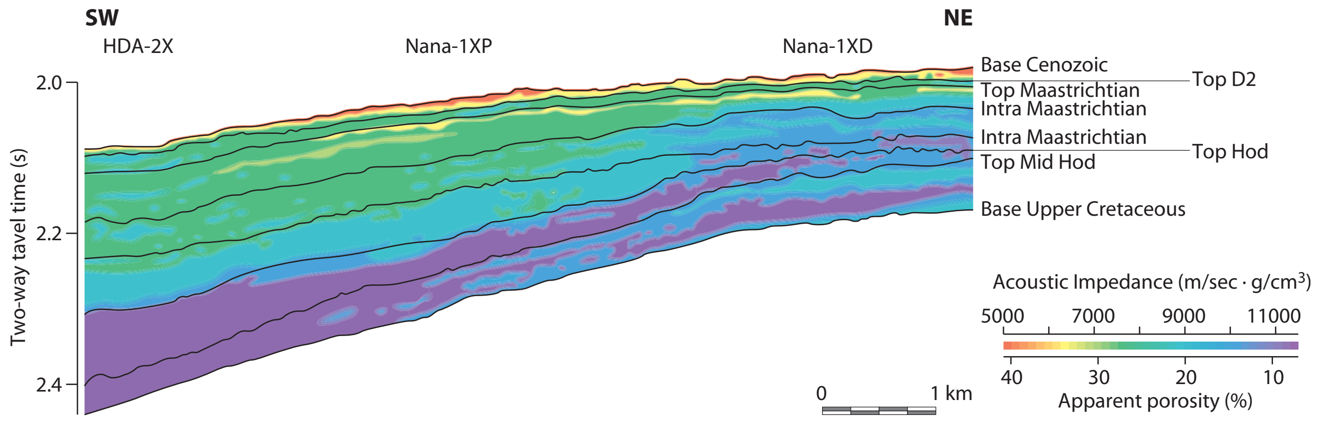


Figure 11.31 Acoustic impedance cross-section across the Halfdan area. See Figure 11.29 for location.

Maastrichtian chalk of pelagic origin exhibits a regionally correlatable characteristic cyclic pattern in the 1 to 3 m scale of high- and low-porosity intervals. In the high-porosity cycle intervals, the porosity may be up to 35 to 37%, with hydrocarbon saturations more than 90%. At a reservoir scale, the average porosities are 25 to 30%. Danian chalk forms a secondary reservoir divided into an upper unit characterised by relatively clean chalk and a lower unit characterised by impure argillaceous and siliceous chalk. The latter is interpreted to form a baffle between the Maastrichtian oil-development target and the upper Danian gas.

The so-called FAST-technology (Fracture Aligned Sweep Technology) is currently used for production in the oil-bearing reservoir. Wells with long horizontal sections are aligned in the direction of the present-day *in-situ* horizontal stress, with producer-water injector well spacing of about 200 m (**Figure 11.29**). The effect of water injection gives rise to a response on 3D-seismic data acquired 5 years after the start of production. There is a very good agreement between the location of water injectors and maximum peak amplitudes (**Figure 11.33**).

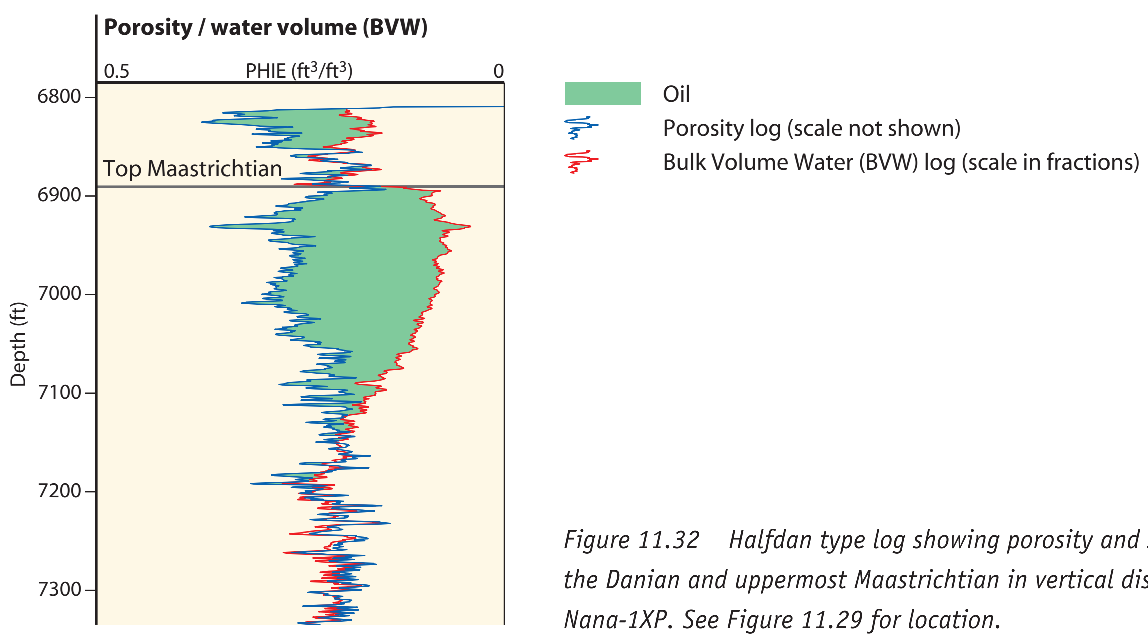


Figure 11.32 Halfdan type log showing porosity and saturation in the Danian and uppermost Maastrichtian in vertical discovery well Nana-1XP. See Figure 11.29 for location.

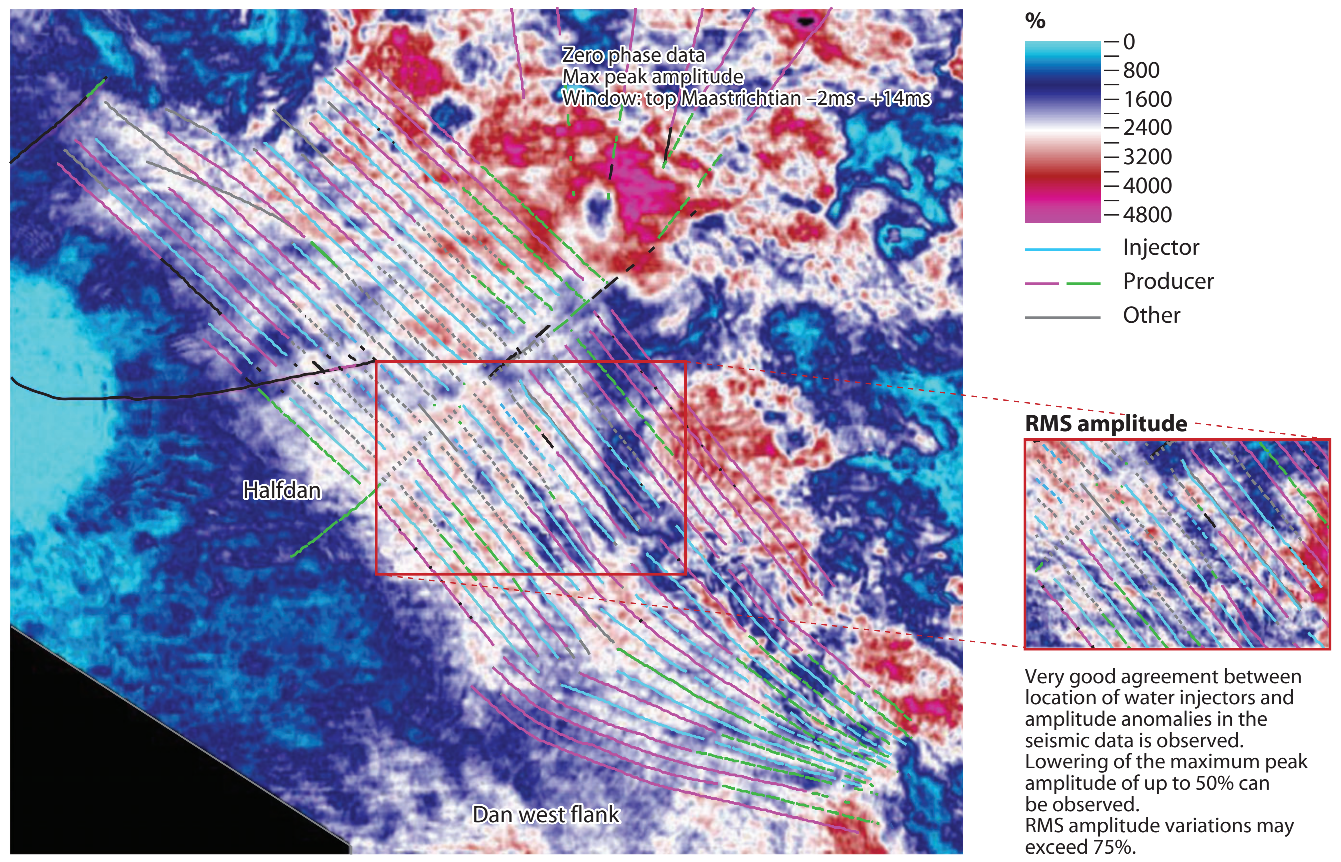


Figure 11.33 Seismic response to water injection (root mean square (RMS) amplitude).

Table 11.2 Properties of the Halfdan oilfield.

Reservoir	Maastrichtian to Danian chalk
Lithology	Chalk
Depth to top (m)	2050
Maximum column height (m)	15
Net reservoir thickness (m)	50
Porosity (%)	25-37
Permeability (mD)	0.5-4
Fluid type	Oil, gas and condensate
Initial pressure (bar)	288.8 at 2081.8 m
Temperature (°C)	72.2
Source rock	Kimmeridgian-Berriasian mudstones (Farsund Formation)
Seal	Tertiary mudstones

4.3 Tyra gasfield, offshore Denmark

The Tyra gasfield is located offshore of Denmark, approximately 230 km west-north-west of Esbjerg. The field was discovered in 1968 and came on stream in 1984. It acts as a swing producer to maintain reservoir pressure to optimise production of the fluid hydrocarbons. The field was initially developed as a gasfield due to poor production tests of the oil rim during appraisal of the field. However, a better understanding of the fluid distribution and the advent of horizontal-drilling technology accelerated the drilling of oil-rim wells during the 1990s. The best oil wells are in areas where gas production is impaired by tight chalk directly above the gas-oil contact, and where the oil rim is thickest.

The Tyra gasfield, with an underlying oil rim, produces from Danian and Maastrichtian chalk trapped in a low-relief structure sealed by Upper Paleocene shales (Nykjaer, 1994). The structure, which has vertical relief of about 76 m (**Figure 11.34**) and an area of 90 km^2, is part of the north-west–south-east-trending

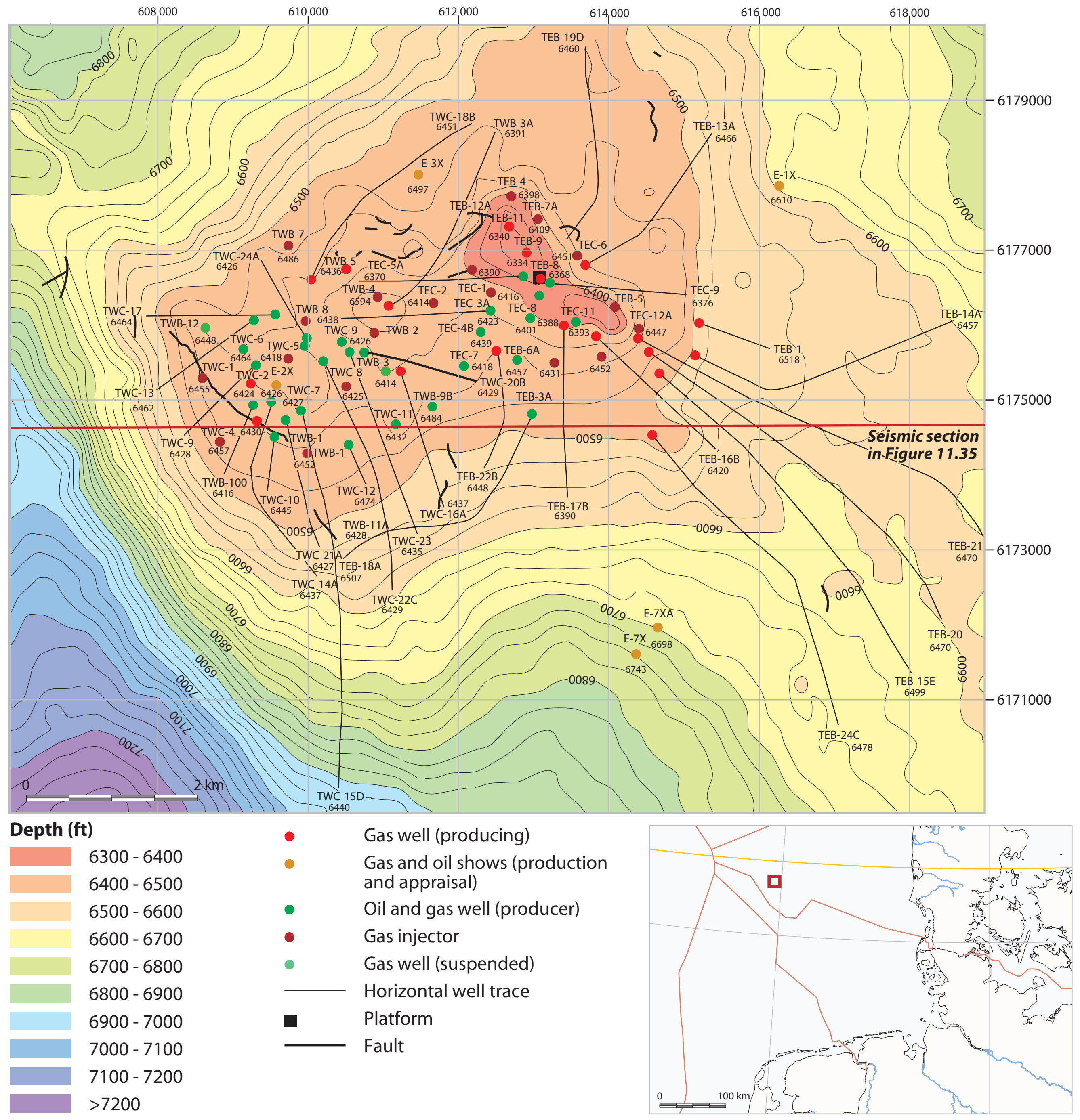


Figure 11.34 Depth-structure map to the top of the Chalk Group in the Tyra gasfield showing well trajectories.

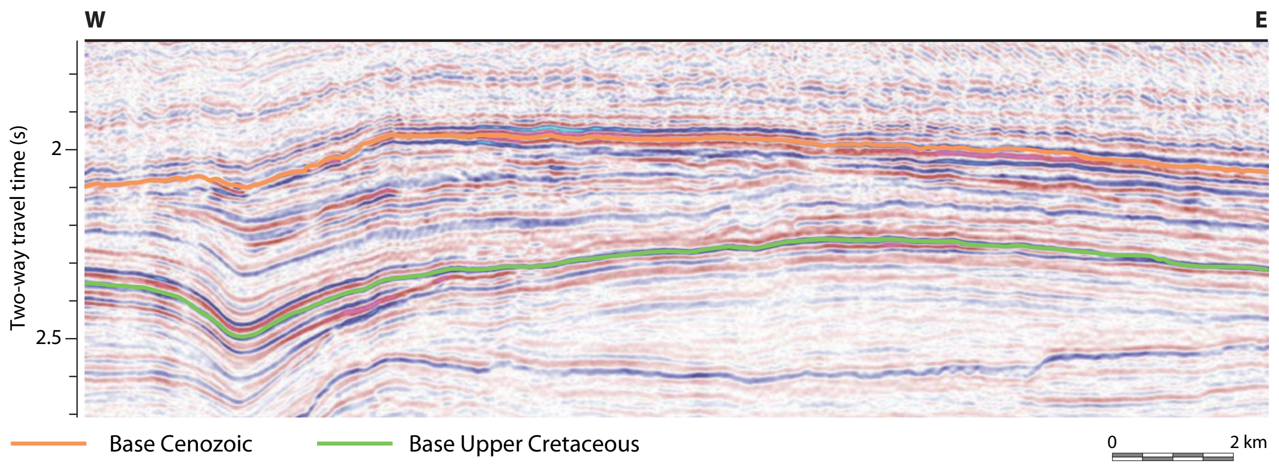


Figure 11.35 Seismic section across the Tyra field. Note the amplitude brightening and phase reversal at the base of the Cenozoic. See Figure 11.34 for location.

Tyra-Igor inversion ridge located in the Danish Central Graben. **Figure 11.35** illustrates the gentle low-relief Tyra structure with seismic amplitude brightening and phase reversal at the top of the Chalk Group on the crest caused by the high-porosity gas-bearing reservoir. The fluid distribution is complex, with tilted contacts towards the south-east controlled by the regional hydrodynamic trend (**Figure 11.36**). The oil zone is divided vertically into an active and an immobile residual zone. In the active zone, oil saturations are increasing with height above a FWL determined by capillary forces. In the residual zone, oil-rim saturations are independent of height as seen in well TEC-8 (**Figure 11.37**). The redistribution of early trapped oil is believed to be caused by shrinkage and leakage of the gas cap during burial and a late structural tilt.

The stratigraphic development of the chalk reservoirs is similar in several fields in the southern Danish Central Graben. The Maastrichtian Tor Formation chalk of pelagic origin exhibits a regionally correlatable characteristic cyclical pattern in the 1 to 3 m scale of high- and low-porosity interval. The Danian chalk of the Ekofisk Formation is divided into an upper unit of relatively clean chalk and a lower unit characterised by impure argillaceous and siliceous chalk with lower hydrocarbon saturations. The reservoir consists of Danian and Maastrichtian chalks with porosities of up to 45%. Matrix permeabilities vary from 0.1 to 5 mD (**Table 11.3**). Natural fractures have significantly enhanced local effective permeabilities. The high porosities are partly an effect of overpressure and early invasion of hydrocarbons generated in the underlying Upper Jurassic Farsund Formation, which has prevented compaction and diagenetic alterations.

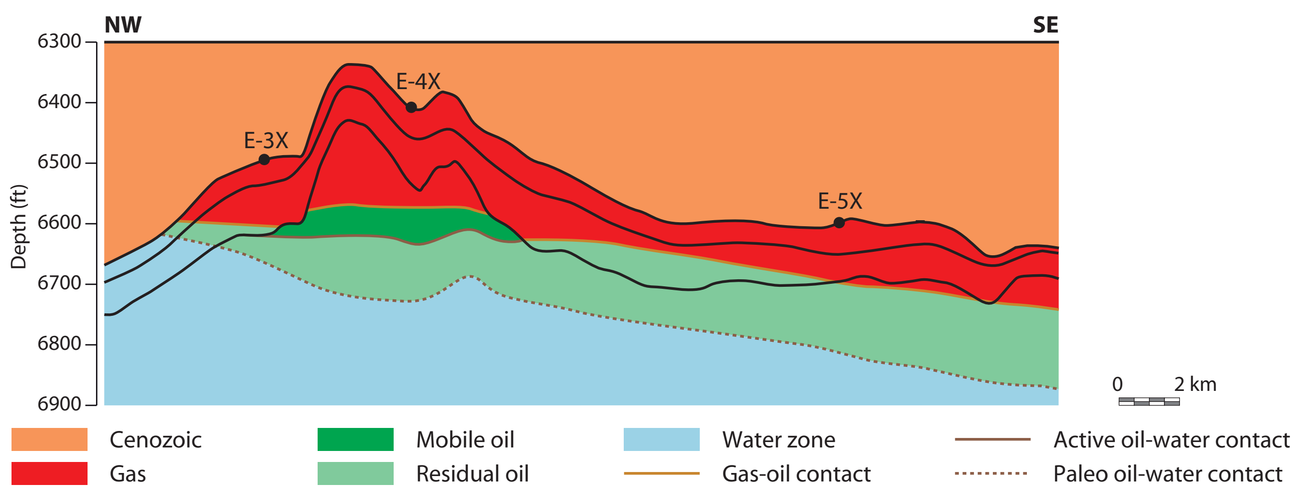


Figure 11.36 Schematic profile of the Tyra field showing fluid distribution. See Figure 11.34 for locations of wells.

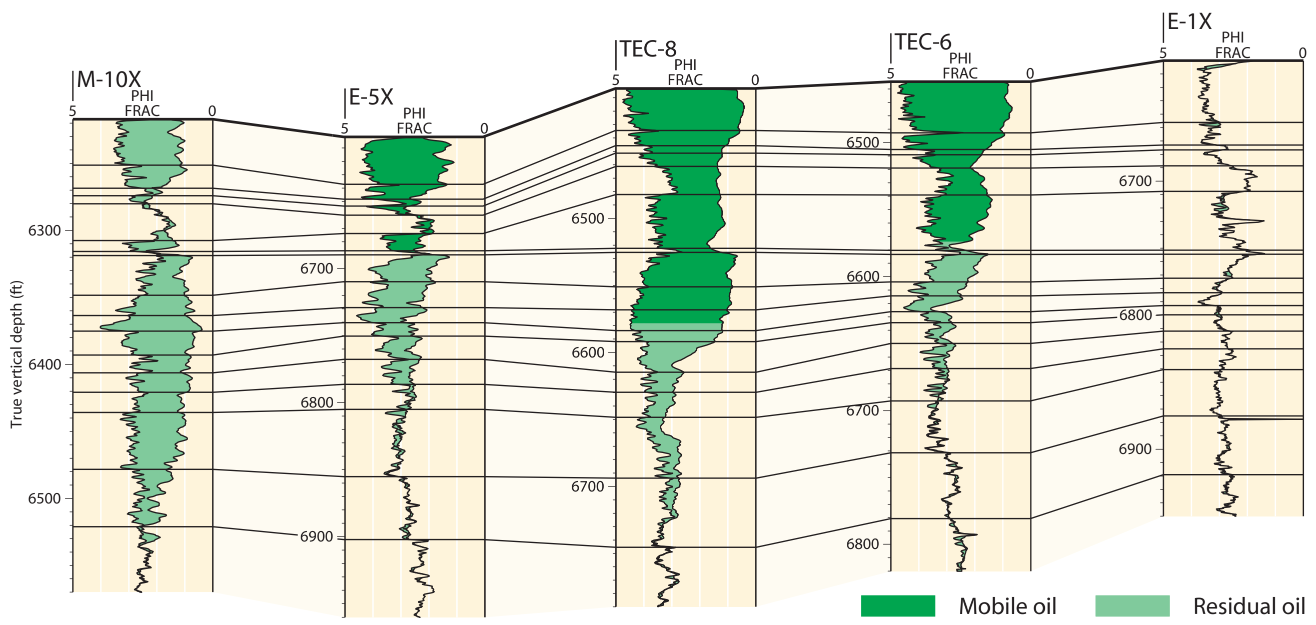


Figure 11.37 Tyra-Dan correlation panel flattened on the top of the Maastrichtian hardground showing reservoir zonation. The distance between Tyra and the M-10X well on the flank of the Dan field is about 30 km. Note the oil saturation profile in TEC-8 indicating an active oil zone with upward-increasing saturations and a residual zone with saturations independent of height. See Figure 11.34 for locations of wells.

Table 11.3 Properties of the Tyra field.

Reservoir	Maastrichtian to Danian chalk
Lithology	Chalk
Depth to top (m)	2163
GWC/GOC.OWC (m)	2975
Maximum column height (m)	27
Porosity (%)	20-45
Permeability (mD)	0.1-5
Fluid type	Oil and gas
Initial pressure (bar)	293.8 at 1996.4
Temperature (°C)	71.1
Source rock	Kimmeridgian-Berriasian mudstones (Farsund Formation)
Seal	Tertiary shales

4.4 *Valdemar oilfield, offshore Denmark*

The Valdemar oilfield is located offshore of Denmark, approximately 250 km west-north-west of Esbjerg. It is the only oilfield in the North Sea currently producing from Lower Cretaceous chalk, marly chalks, and marlstones of late Hauterivian to early Aptian age. A secondary hydrocarbon-bearing reservoir is found in overlying Upper Cretaceous to Danian chalks. The field was discovered in 1977 by the Bo-1 well and has been developed by long horizontal wells with artificial sand-filled fractures, and came on stream in 1993. The recovery mechanism is natural depletion. Although the in-place volumes are in the order of 115 bcm, the recovery is rather low, with ultimate oil reserves of 12.8 bcm (DEA, 2008), primarily due to low permeability.

The accumulation is situated on an elongate north–south-orientated anticline with two structural culminations separated by a saddle at the Lower Cretaceous reservoir level, formed as a result of Late Cretaceous and Paleogene inversion movements (**Figures 11.38** & **11.39**).

The Valdemar field reservoir occurs within the upper levels of the Lower Cretaceous Cromer Knoll Group. The lower part is located in the approximately 50 m-thick upper Hauterivian to middle Barremian Tuxen Formation consisting of a heterogeneous layered succession of deep-water chalk and marly chalk interbedded with marlstones. The formation consists of two discrete chalk-dominated units separated by up to 2 m-thick organic-rich marlstones, the Munk Marl Bed. The upper part is included in the upper

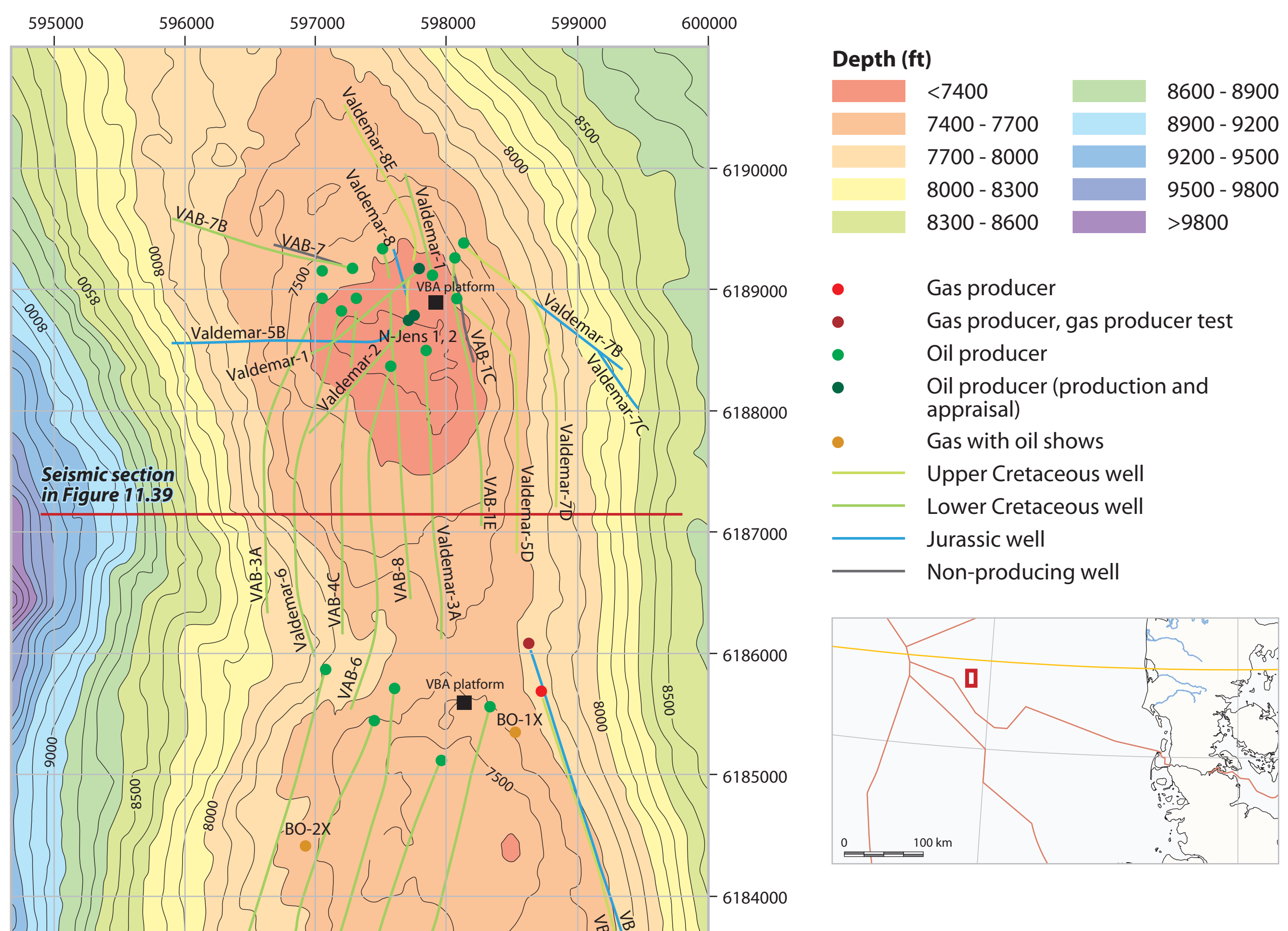


Figure 11.38 Depth-structure map of the top of the Sola Formation showing well trajectories. The Valdemar field is located in a N–S elongated, structurally inverted anticline with two culminations separated by a saddle.

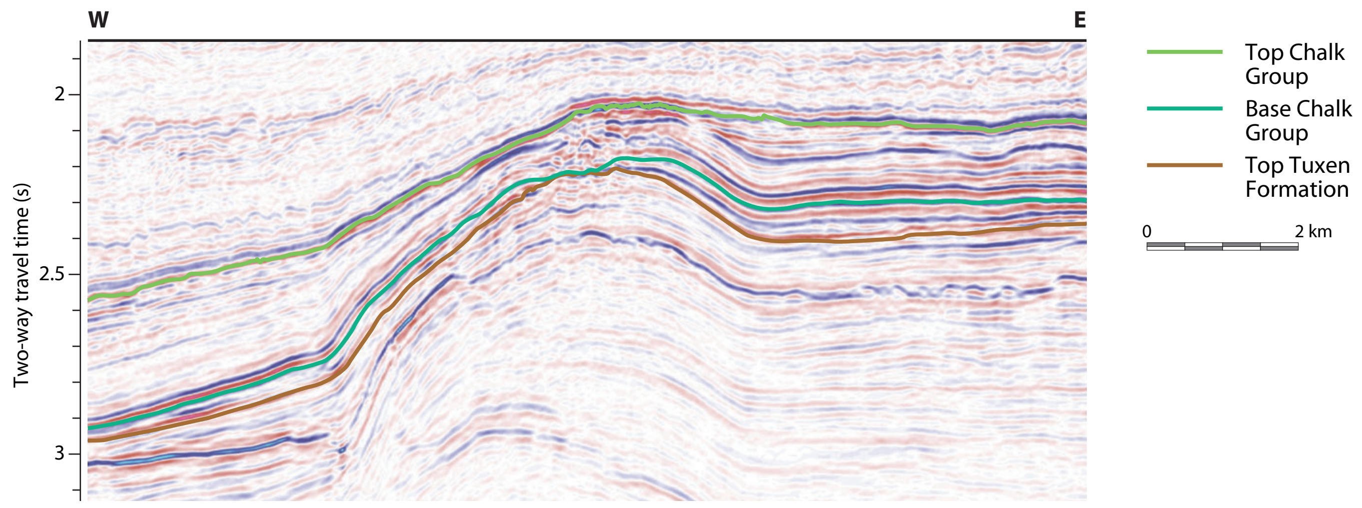


Figure 11.39 Seismic section across the Valdemar field. See Figure 11.38 for location.

Barremian to lower Albian Sola Formation (approximately 30 m thick), with a similar mixed lithology, although with a higher proportion of low-permeable marlstone. It is capped by a 5 m-thick claystone, the 'Albian Shale', which forms part of the top seal. A detailed reservoir zonation consisting of 14 reservoir units has been established based on integration of core and petrophysical data (**Figure 11.40**). The argillaceous chalks of the reservoir are highly faulted and fractured, overpressured and undercompacted. Due to the lithological heterogeneity, internal stratigraphic barriers are common and result in compartmentalisation and therefore a complex distribution of hydrocarbons. The chalk (6-10% insoluble residue (IR)) and marly chalk reservoirs (10-30% IR) are characterised by porosities between 25 to 40% and matrix permeabilities of 0.1 to 0.4 mD (**Table 11.4**). The porosities of the chalks correlate negatively with the IR (particularly clay) content. For a given porosity, the matrix permeability of the Valdemar field is a factor of ten lower than that of an equivalent Maastrichtian chalk.

Table 11.4 Properties of the Valdemar field.

Reservoir	Upper Aptian to upper Hauterivian marly chalk	Danian chalk
Lithology	Chalk, limestone and argillaceous chalk	Chalk, limestone and argillaceous chalk
Depth to top (m)	2270	2025
GWC/GOC.OWC (m)	Oil down to 2720 m	
Maximum column height (m)	100	30
Porosity (%)	25-40	30-40
Permeability (mD)	0.1-0.4	2-5
Fluid type	Oil and gas	Oil and gas
Initial pressure (bar)	349.2 at 2255.5 m	303.5 at 2049.8 m
Source rock	Kimmeridgian to Berriasian mudstones (Farsund Formation)	Kimmeridgian to Berriasian mudstones (Farsund Formation)
Seal	Lower Cretaceous shales	Lower Cretaceous shales

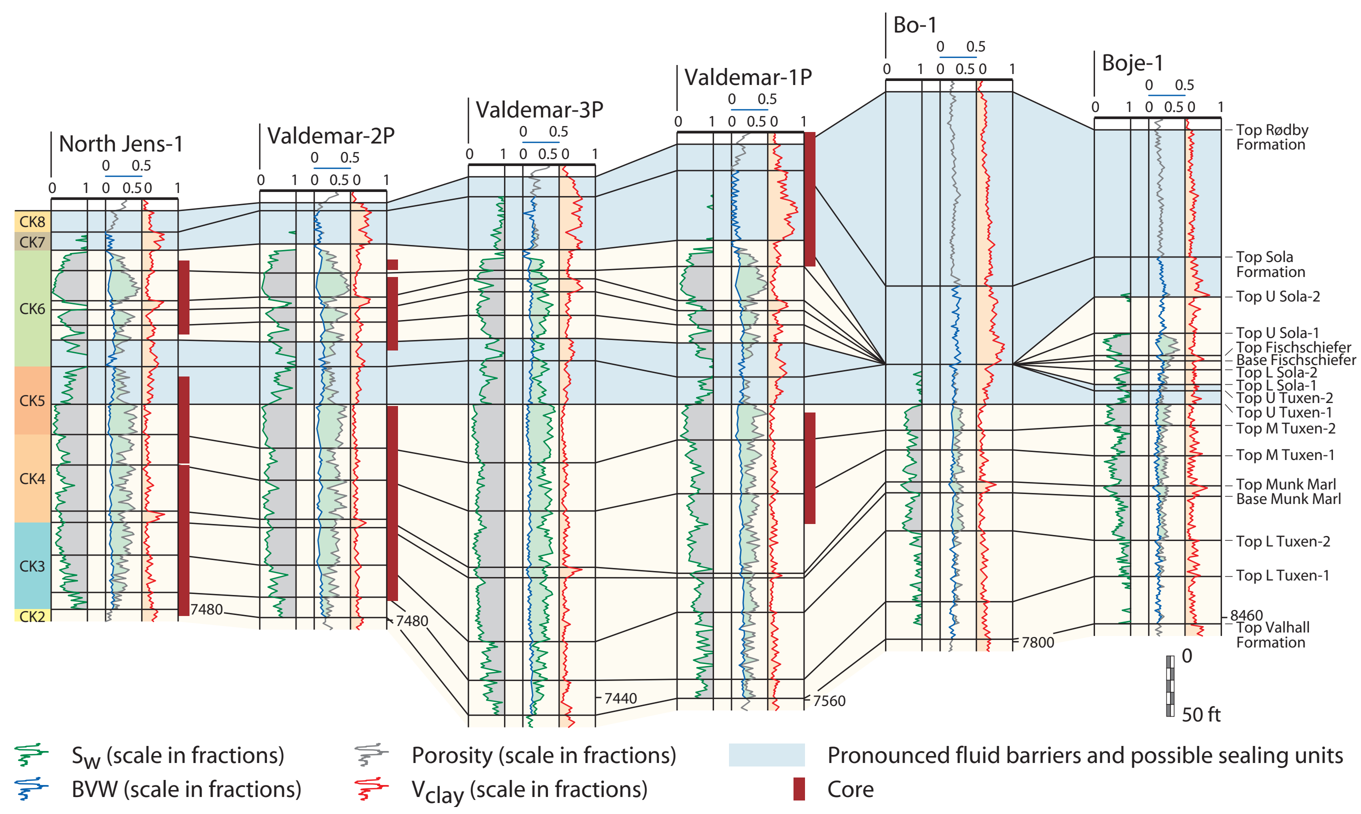


Figure 11.40 Correlation of reservoir zones in vertical wells across the Valdemar area. Note the thickness variation and lateral variation in porosity, clay content and saturation (from Jakobsen et al., 2004).