

**Chapter 2: The chalk depositional system in the Central Graben – concepts of regional geology, sedimentology and stratigraphy**



## 1. Introduction

Before the science of geology was established, the chalk was already noted by the Romans for the prosperity that grapevines reached on chalk soil during their conquering campaign across Europe throughout the 1st century BC. Chalk acts as a sponge, storing excess water for dry periods and with the ability to store heat. In the seventeenth century, such properties were also noted by the celebrated wine maker and monk, Pierre Perignon – or Dom Perignon – the “father of champagne”, making the fortune of the French Champagne and Chablis wine yards. Romans commonly referred to chalk as *Creta*, from which the name of the stratigraphic system *Cretaceous* is derived. The white colour of chalk is possibly the source for the Latin name of England *Albion* (Latin *alba* = white) after the Roman Empire first set foot on English soil at the White Cliffs of Dover (Ewans, 2002).

The Cretaceous is a special and complex period in the Earth’s history characterized by episodes of rapid climatic change, abrupt oceanic anoxic events and significant variations in the carbon cycle, as well as intensification of volcanic activity with development of large igneous provinces (Weissert *et al.*, 1998; Jenkyns, 1994; Wissler *et al.*, 2003; Bodin *et al.*, 2006; Jarvis *et al.*, 2006). In large parts of NW Europe, the sedimentary succession deposited during the Late Cretaceous–Early Paleocene (100–61 Ma) largely consists of thick and conspicuous intervals of white chalk sediments. Deposition of chalk during these 40 million years of Earth’s history followed a significant rise in sea-level at the beginning of the Late Cretaceous (Hancock, 1975; Hancock & Kauffman, 1979). This transgression was probably one the most important during the Phanerozoic and has been commonly attributed to rapid ocean crust formation and an increase in the volume of oceanic ridges (Orth *et al.*, 1993; Kerr, 1998) caused by intensification of mantle magma upwelling to the Earth’s surface (Larson, 1991).

A second peculiarity of the Late Cretaceous is the equally warm climatic conditions over the continents associated with higher tropical and polar temperatures and lower latitudinal temperature gradients than present day (Wilson & Norris, 2001; Wilson *et al.*, 2002). The warmer climate was one of the consequences of the volcanic release into the atmosphere of large quantities of greenhouse gases such as CO<sub>2</sub> and CH<sub>4</sub>. It is possible that warm climate conditions were also reinforced by the higher content of atmospheric water vapour in response to the higher temperatures (Huber *et al.*, 2002). The increase in the volume of oceanic ridges, the prolonged periods of ice-free conditions at the poles as well as increased sea-surface temperatures, led to a significant sea-level rise with inundation of the continental shelves and establishment of epeiric seas in regions previously dominated by shallow marine or paralic environments. Consequently, the shelf break front observed today at the inshore-open ocean boundary did not exist during the Upper Cretaceous and oligotrophic surface water conditions, previously confined to the open oceans, were now established in the newly formed epicontinental seas.

The high sea-level in connection with the warm, but arid, climate and the low topographic relief of the peneplaned NW European land masses reduced the influx of erosional detritus into the oceans. The ensuing clarity of the surface waters and their warm temperature, associated with oligotrophic nutrient levels and normal salinity, were optimal conditions for the flourishing of the coccolithophorid

algae. The widespread sedimentation of the calcareous skeletal remains of these algae, the coccoliths, formed thick layers of calcitic chalk ooze on the sea floor (Håkansson *et al.*, 1974; Hancock, 1975).

At the end of the Cretaceous, several marine and terrestrial biota disappeared, including dinosaurs, ammonites, inoceramid and rudist bivalves and many nanno- and microfossil groups (Smit, 1990; Keller, 1989, 2001). This large-scale extinction has been attributed to a meteorite impact (Alvarez *et al.*, 1980; Smit & Hertogen, 1980), based on the increased level of cosmogenic Ir and the presence of spherulitic glass ejecta in the K-T sequences in NE Mexico, Texas and at the Chicxulub impact crater. However, in recent years researchers have disputed this thesis and confirmed that the Chicxulub impact predates the K-T boundary by ~300 kyr, suggesting that the concomitant effects of volcanism, meteorite bolide impacts, and extreme climate variations before and throughout the K-T boundary were the main cause of the mass extinction at the end of the Cretaceous (Keller, 2008).

In the Central Graben area, sedimentation of chalk continued throughout the Danian, although the end-Cretaceous extinction resulted in the collapse of Cretaceous phytoplankton systems and replacement of the Maastrichtian coccolithophorid by different assemblages (Kennedy, 1987a). At the end of the Danian, following the compressive pulses of the Alpine orogenesis, the landmasses surrounding the chalk epeiric sea were gradually uplifted with subsequent increase in terrigenous detrital influx into the oceans. This led to deposition of siliciclastic successions over broad areas, terminating the proliferation of coccolithophorid algae and the associated sedimentation of chalk (Ziegler, 1990). In the North Sea, the total thickness of the Chalk Group may exceed 2000 m (Fig. 1), while in the study area within the Norwegian Central Graben it reaches a maximum thickness greater than 1500 m. Present-day burial depth is variable, but may reach more than 3000 m in the centre of the basin (Fig. 2) (Ziegler, 1990).

Sedimentation of chalk primarily occurred through pelagic settlement of coccolithophorid tests in marine conditions (Håkansson *et al.*, 1974; Hancock, 1975). In the North Sea, deposition of chalk was accompanied by rise of ridges, domes and anticlines due to halokinetic movements and tectonic inversion of the pre-Cretaceous extensional faults (Cartwright, 1989). Seismic activity increased the instability of sediments and triggered downslope mass movements of previously deposited chalks through slides, slumps, debris flows and turbidity currents (Watts *et al.*, 1980; Hatton, 1986; Kennedy, 1987a, b; Van der Molen *et al.*, 2005). The physiography of the chalk sea floor was intensively sculpted by bottom currents that were periodically intensified during the Late Cretaceous. Bottom currents created important topographic features and thickness variations through formation of channels, ridges, moats and drifts (Lykke-Andersen & Surlyk, 2004; Esmerode *et al.*, 2007, 2008; Surlyk & Lykke-Andersen, 2007; Surlyk *et al.*, 2008; Esmerode & Surlyk, 2009).

Geological interest in chalk has increased since the 19<sup>th</sup> century, but it was the discoveries of the Kraka and Ekofisk hydrocarbon fields in the North Sea during the 1960s that drove the focus of geologists towards this sedimentary rock. In the following years, chalk sediments proved to be prolific reservoirs in a series of oilfields. These included the Eldfisk, Tor, Albuskjell, Hod, Valhall, Tommeliten, West Ekofisk and Edda fields in the southern Norwegian offshore sector, the Gorm, Dan, Tyra and Skjold fields in the Danish sector and the Joanne and Banff fields in the UK sector, while the Hanze Field is

the only hydrocarbon accumulation in chalk so far discovered in the Dutch sector (Fig. 3) (Megson, 1992; Oakman & Partington, 1998; Bramwell *et al.*, 1999; Megson & Hardman, 2001; Hofmann, 2002; Surlyk *et al.*, 2003).

Among the chalk hydrocarbon fields in the North Sea, those present in the Norwegian sector show the highest porosity and permeability with allochthonous facies forming the main reservoir intervals. In the Norwegian offshore sector, proven reserves originally in place were estimated to be around  $2300 \times 10^6 \text{ Sm}^3$  of oil and  $550 \times 10^9 \text{ Sm}^3$  of gas (www.npd.no). In the Danish fields, rhythmically bedded pelagic chalks also form hydrocarbon reservoirs (Scholle *et al.*, 1998; Damholt & Surlyk, 2004), although these show lower porosity and permeability compared to coeval successions in the Norwegian sector.

## 2. Geological setting

The study area is located within the Central Graben, an intracratonic basin that represents the southern branch of the North Sea triple rift system. The other NE and NW branches of this rift are the Viking Graben and the Moray Firth Basin, respectively. The Central Graben extends from the northern termination of the Dutch Central Graben to the western end of Rinkobing-Fyn High in Danish waters. It then continues toward the northwest entering into the UK sector of the North Sea where it intersects the southern part of the Viking Graben (Fig. 3).

In Norwegian waters, the Central Graben occurs along listric normal faults directed NNW–SSE which are locally cut by transverse faults with W–E to WSW–ENE directions. Most of the graben is made up of intra-basinal lows and highs (Fig. 4) and its present-day structural configuration is the result of several tectonic phases that can be dated back to the Palaeozoic (Fig. 5) (Ziegler, 1990; Gowers *et al.*, 1993; Knott *et al.*, 1993).

Prolonged extension occurred during the Permian–Triassic with coeval sedimentation of thick successions of Zechstein salt. During the Triassic, tectonic extension continued mainly with a regional WNW–ESE direction, while the Middle Jurassic was characterized by pre-rift doming, which developed into the Mid-Cimmerian Unconformity. This doming represents the early stages of an important extensional phase that culminated during the Late Jurassic. These rifting phases caused uplift and tilting of fault blocks on the rift shoulders favouring erosion by shallow marine processes during the Early Cretaceous. As a result, the regional Base Cretaceous Unconformity developed over the entire area of the Central Graben (Gowers *et al.*, 1993). Opening of the Atlantic Ocean during the Early Cretaceous changed completely the regional stress regime causing the cessation of the rifting. The horizontal stress became compressive mostly with an E–W direction. This compression resulted in local shortening and inversion of the Triassic–Jurassic faults, with transpressive movements along the NNW–SSE oriented faults.

Throughout the Late Cretaceous and Palaeogene, the Central Graben underwent post-rift thermal subsidence periodically punctuated by NNE–SSW compressional tectonic pulses with variable

intensity. During deposition of the Chalk Group, four tectonic phases of increased intensity occurred (Vejbæk & Andersen, 2002):

- (1) Latest Santonian (Sub- Hercynian tectonic phase);
- (2) Mid Campanian (Sub- Hercynian tectonic phase);
- (3) Late Maastrichtian (Sub- Hercynian tectonic phase);
- (4) Late Paleocene (Laramide tectonic phase).

Inversion of major basement faults generated anticlinal fold structures such as the Lindesnes Ridge (Fig.4) (Cartwright, 1989; Farmer & Barkved, 1999). The ductility and intrusive behaviour of the Zechstein salt along major basement faults enhanced the inversion movements, but also created a wide array of halokinetic structures such as diapirs, domes, salt walls and salt withdrawal basins (Knott *et al.*, 1993; Oakman & Partington, 1998). During progressive uplift, the crestal areas of the inversion zones and halokinetic structures were subject to gravitational collapse with formation of local horsts and grabens (Farmer & Barkved, 1999).

The thermal subsidence of the North Sea Basin associated with the eustatic sea-level rise led to the progressive overstepping of its margins and, by the beginning of the Early Cretaceous, the sea-level exceeded the present-day levels by ~100–300 m (Fig. 6) (Haq *et al.*, 1987; Ziegler, 1990). During this period, inundation of the land masses allowed the widespread occurrence of oligotrophic oceanic conditions with consequent deposition of chalk (Fig. 7) (Ziegler, 1990).

Climate throughout the Cretaceous is generally considered to have been warm with lower tropical to polar temperature gradients than at present-day. Estimated tropical sea-surface temperatures varied between 32° and 34° C, while in polar regions temperatures of 10° to 18° C are thought to have occurred (Hay *et al.*, 2008). Significant climatic variations have been inferred during the late Cretaceous with the Cenomanian–Campanian being the warmest, which coincided with the peak of the sea-level transgression. Sea water temperatures declined strongly during the Maastrichtian (Huber *et al.*, 2002), which was also characterized by marked temperature fluctuations (Li & Keller, 1999).

Global palaeoenvironmental changes during the Cretaceous accompanied the sea-level transgression, leading to widespread oceanic anoxic events (OAEs; e.g. Valanginian, Hauterivian, Barremian–Aptian, Aptian/Albian, Albian–Cenomanian, Cenomanian–Turonian, and Coniacian). The the sea-level transgression is commonly attributed to rapid sea floor spreading and an increase in the volumes of mid-oceanic ridges (Schlanger *et al.*, 1981; Larson, 1991; Wignall, 2001).

The concomitant release of volcanic CO<sub>2</sub> led to more intensive terrestrial weathering and nutrient input into the oceans, as well as lowering of the atmospheric oxygen. These factors, associated with increased water-column stratification, led to eutrophication of ocean surface waters and an increased flux of organic matter to the seafloor. Bacterial decomposition of organic matter depleted the dissolved oxygen, causing anoxic environments and sedimentation of black shale (Schlanger *et al.*, 1987; Wignall, 1994).

### 3. Stratigraphy

The first formal lithostratigraphic nomenclature for the central and northern North Sea was published by Deegan & Scull (1977). The Cretaceous and Tertiary stratigraphy for the Norwegian North Sea was later revised by Isaksen & Tonstad (1989). Lieberkind *et al.* (1982) published an informal nomenclature for the Chalk in the Danish Central Graben in 1982. The JCR Chalk Monograph compiled by Andersen (1995) summarizes the North Sea chalk lithostratigraphy. The stratigraphic nomenclature for chalk formations in the Central Graben used in this study follows the approach of Bailey *et al.* (1999) as summarized in Figure 8. These authors reviewed the North Sea chalk lithostratigraphy and integrated internal and informal chalk subdivisions used by the operating companies. This subdivision comprises the Hydra, Blodøsk, Narve, Thud, Magne, Tor and Ekofisk formations.

### 4. Chalk constituents

The chalk primarily consists of the minute skeletal remains of coccolithophorid algae. These calcareous nanofossils are composed of individual calcareous plates (coccoliths) that form clay to silt-sized spherical bodies called coccospheres (Fig. 9). Complete coccospheres are sporadically present within the chalk but the majority are broken up into single coccolith plates or laths (Håkansson *et al.*, 1974; Hancock, 1975; Scholle, 1977).

The size of the coccosphere and its components give to the chalk a fine mudstone to wackestone micro-texture and coarser textures, such as packstone and grainstone, are restricted to sporadic and thin intervals. Secondary calcareous components in the North Sea chalks are calcispheres, foraminifers, macrofossil mollusc debris, bryozoan, brachiopod debris and ostracods, while the non-carbonate biogenic fraction commonly consists of radiolarians, diatoms and sponge spicules (Kennedy, 1985). The low-Mg content of the calcite that forms the coccoliths makes the chalk less prone to diagenetic changes compared to carbonate of mixed aragonitic and high-Mg composition (cf. Scholle, 1977).

Chalk also contains other secondary components, for example, the non-biogenic terrigenous fraction is largely represented by clay minerals and detrital quartz and although clay content is generally low, in certain intervals this material may reach up to 20% wt of the bulk sediment (Lindgreen *et al.*, 2002). It is generally believed that clay material in the chalk was transported as erosional detritus by wind or by river plumes, although volcanic ash may also have contributed (Fabricius, 2007; Lindgreen *et al.*, 2008).

The chalk also contains a siliceous fraction in the form of flint, which was derived from the siliceous tests and skeletons from radiolarians, diatoms and sponge spicules (Håkansson *et al.*, 1974; Kennedy, 1985). Flint is composed of cryptocrystalline quartz and it is the result of dissolution, reprecipitation and subsequent diagenetic transformation of the initial biogenic opal-A silica. Besides flint and detrital grains within clay-rich intervals, quartz is present as euhedral particles or particle clusters of nanometre size dispersed in the chalk matrix (Lindgreen *et al.*, 2010; Madsen, 2010). This kind of quartz occurs within apparently homogenous chalk intervals similar to other non-silicified chalk beds. In addition, pyrite and sulphate minerals may be also present in concretions as the result of

microbial action (Fabricius, 2007). Other minor components present in the chalk as a product of burial diagenesis are zeolites, barite, celestite and feldspar (Fabricius & Borre, 2007).

## 5. Chalk sedimentology and depositional model

The primary depositional mechanism of the chalk was from pelagic rain of coccoliths in a marine setting at depths up to few hundred metres. Due to their minute size, coccoliths would not reach the sea floor by settlement alone. It is therefore more probable that the pelagic rain (Fig. 10A) consisted of faecal pellets that originated from plankton feeder organisms or algae mucose filaments bounding planktonic pellets known as “marine snow” (Hancock, 1975; Honjo & Roman, 1978; Scholle, 1983; Steinmetz, 1994; Damholt & Surlyk, 2004). This pelagic rain deposited on the sea floor a highly water-saturated calcareous ooze with porosity ranging from 70% to 80% (Scholle, 1977) which is analogous to modern deep-sea nanoplankton ooze.

After deposition, the calcareous ooze was subject to progressive compaction and bioturbation by the action of benthic organisms. The slow rate of sedimentation of chalk facilitates intensive bioturbation, enhancing dewatering and early compaction (Surlyk *et al.*, 2003). As a result, most of the autochthonous North Sea chalks are bioturbated and only remnants of the primary sedimentary structures are visible (Ekdale & Bromley, 1983; Kennedy, 1980). In chalk sediments, a wide range of trace fossils can be observed. These include shallow obliterated tiers passing downward to tiers characterized by *Planolites*, *Thalassinoides*, *Taenidium*, *Zoophycos* and *Chondrites* (Fig. 11) (Ekdale & Bromley, 1983, 1991).

Chalk sediments may show evidence of primary bedding at a decimetre to metre scale, usually visible as variation in oil staining, grey tone or alternation of more bioturbated to laminated beds. The bedding can result from numerous mechanisms, for instance primary pelagic lamination possibly emphasized by sea floor diagenesis, variation in the influx of silicates, current winnowing of unconsolidated chalk or alternating bioturbation and pelagic sedimentation punctuated by intermittent deposition of distal low density turbidity currents (Kennedy, 1987a, b; Scholle *et al.*, 1998; Damholt & Surlyk, 2004; Fabricius, 2007).

Cyclic deposits of marl-limestone couplets referred to as periodites frequently occur within pelagic chalk successions (Kennedy, 1987a, b). Within these rhythmical successions, clay-rich intervals usually show well-developed planar parallel laminations while the limestones are highly bioturbated, though remnants of primary laminations can be present. In chalk succession, other cyclicity typically occur as variations in the degree of cementation (nodular chalk and hardgrounds), colour changes, variations in the carbon and oxygen isotopic ratios, flint bedding distribution and variations of petrophysical properties, e.g. gamma ray, porosity and magnetic susceptibility (Hancock, 1975; R.O.C.C. Group, 1986; Hart, 1987; Ditchfield & Marshall, 1989; Gale, 1989; Zijlstra, 1994, 1995; Niebuhr & Prokoph, 1997; Molenaar & Zijlstra, 1997; Scholle *et al.* 1998; Gale *et al.*, 1999; Stage 1999, 2001a, b; Niebuhr *et al.*, 2001; Damholt & Surlyk, 2004). Periodites and other cyclicity present in chalk are interpreted to result from variations in the calcareous nannoplankton productivity, changes



in the terrigenous input or variations in the redox conditions at sea-floor as well as modification of the carbonate dissolution rates. These factors may be linked to several causes, for instance sea-level fluctuations, changes in organic productivity, ocean currents and upwelling or climatic fluctuation ascribed as Milankovitch cycles (Barron *et al.*, 1985; R.O.C.C. Group, 1986; Gale, 1989; Scholle, 1998; Damholt & Surlyk, 2004).

The pelagic rain of coccoliths accumulated thick piles of sediment on the sea-floor. The calcium carbonate (CaCO<sub>3</sub>) particles making up the thixotropic chalk ooze would have had no unbalanced electric charges or platy interlock, hence little or no cohesion existed. In submarine environments, small inclinations around 1–2 degrees are sufficient for gravity-flows to occur (Lewis, 1971) and even mild vibrations or ground motions of the substrate would result in remobilization and downslope movement of the incohesive chalk material. Triggering mechanisms of such movements could be earthquakes, storms or rapid release of clathrates from the sediment column (Kennedy, 1985, 1987a, b; Bramwell *et al.*, 1999).

In the Central Graben, while coccoliths were being deposited ubiquitously, tectonic seismicity increased sediment instability over the inversion areas, leading to gravity-driven resedimentation and emplacement of allochthonous material into the basins (Fig. 10B). Kennedy (1987a) concluded that this was the main cause of the great thickness of the Chalk Group over the basinal areas (2000 m) compared to the flanks (200 m) of the Central Graben. Numerous types of mass-flow deposit can be observed in chalk from cores, well-logs and seismic data (Fig. 10A) (Perch-Nielsen *et al.*, 1979; Kennedy, 1985, 1987a, b; Watts *et al.*, 1980; Hatton, 1986; Bramwell *et al.*, 1999; Skirius *et al.*, 1999; Van der Molen *et al.*, 2005).

Based on the degree of deformation, which reflects the initial state of consolidation and distance from source area, gravity-flow deposits can be categorized from proximal to distal: (1) slides and creeps, formed by rotated and slightly deformed blocks with parallel dips; (2) slumps, consisting of plastically deformed chinks; (3) debris flows, usually comprising chalk clasts floating in a homogeneous matrix; (4) mudflows, which are usually characterized by massive and homogeneous deposits and are considered as debris flow where all the clasts have been deposited or disrupted during movement; (5) turbidites, comprising planar-to cross-laminated packstone often showing normal grading. The majority of the redeposited chinks are not burrowed with the exception of thin gravity-flows or the uppermost interval of the resedimented unit which was in contact with the sea-floor. An extensive description of the different depositional facies in the chalk can be found in Kennedy (1985, 1987a, b), Bromley & Ekdale (1987), Crabtree *et al.* (1996), Bramwell *et al.* (1999) and Røgen *et al.* (1999).

Bottom currents played a significant role in the deposition of the chalk, as recently described from the offshore chalk successions in the North Sea and in the Danish Basin (Lykke-Andersen & Surlyk, 2004; Esmerode *et al.*, 2007, 2008; Surlyk & Lykke-Andersen, 2007; Surlyk *et al.*, 2008; Esmerode & Surlyk, 2009). These currents created a series of important erosional and depositional features on the chalk sea floor which have been described as moats, drifts and channels (Fig. 10B). Mass movements were also triggered by alongslope current erosion, which decreased the stability of the sediments on the slope favouring failures and mass movements (Esmerode *et al.*, 2008). Channel features in chalk

successions have also been described from onshore France (Quine & Bosence, 1991) and in the offshore sectors of the UK (Evans & Hopson, 2000; Evans *et al.*, 2003), Denmark (Back *et al.*, 2011) and the Netherlands (Van der Molen, 2004). These channels have been otherwise interpreted to result from submarine erosion by tidal or oceanic currents during lowstand periods (Quine & Bosence, 1991) induced by tectonic uplift (Evans & Hopson, 2000; Evans *et al.*, 2003) or to represent conduits for submarine gravity-flows (Back *et al.*, 2011).

## 6. Diagenesis

Chalk diagenesis in general is well documented by several studies that have addressed the depositional and post-depositional history of this sediment. These include Mapstone (1975), Scholle (1977), Clayton (1984), Taylor & Lapre (1987), Herrington *et al.* (1991), Maliva & Dickson (1992), Brasher & Vagle (1996), Scholle *et al.* (1998), Lindgreen *et al.* (2002, 2008, 2010), Fabricius (2007), Fabricius & Borre (2007) Fabricius *et al.* (2007, 2008), Madsen (2010), Madsen & Stemmerick (2010).

Intense current action or intervals of low coccolithophore productivity are likely to reduce the net sedimentation rate of coccoliths, enhancing early sea floor cementation. Figure 12 summarizes the progressive stages of hardground formation according to Kennedy & Garrison (1975). Reduced rate or breaks in sedimentation led to the formation of an initial omission surface characterized by intensive *Thalassinoides* tiers. Afterwards, nodular chalks develop between the burrows as a product of cementation. Eventually, the nodules may aggregate forming a cemented bed and, if followed by erosion, the nodules may be reworked to form an intraformational conglomerate. If erosion does not occur and sedimentation restart, *Thalassinoides* burrows may be filled with the newly deposited chalk (Surlyk *et al.*, 2003).

Hardgrounds represent extensive periods of reduced deposition and sea floor exposure characterized by lithification and bioerosional processes. This allows long-term water circulation in the sediments with consequent precipitation of calcite cement in the pore spaces. Hardgrounds are also characterized by cement filling of pre-existing *Thalassinoides* burrows, coating of the sediment surface by phosphatic and glauconitic mineralizations, *Entobia* borings and an irregular upper surface due to submarine erosion (Bromley, 1975, Kennedy & Garrison, 1975, Lasseur *et al.*, 2009).

Early diagenesis, through de-watering and compaction, takes place from the time of deposition until pore water in the sediment ceases to be exchanged with seawater, which commonly occurs at about 1 kilometre of burial where porosity is ~40% (Taylor & Lapre, 1987). The presence of clay minerals strongly influences the reservoir quality of chalk sediments. A high clay content prevents contact between grains and development of intergranular cement during early diagenesis. This decreases the degree of sediment consolidation, resulting in greater mechanical and chemical compaction during subsequent burial (Surlyk *et al.*, 2003).

A second diagenetic effect relates to the complex area of silica dissolution and reprecipitation in the form of flint nodules. Flint is composed of cryptocrystalline quartz and represents the results of relatively early diagenetic phenomena that occur at the redox boundary within the first tens of metres

in the sediment column (Clayton, 1984; Madsen & Stemmerick, 2010). Dissolution of biogenic opal-A silica and subsequent reprecipitation of flint is the main mechanism of flint development in chalk. The source of biogenic silica has been commonly attributed to the remnants of radiolarian, diatoms and sponge spicules. Some silica is preferentially precipitated in burrows, notably *Thalassinoides*, while other replaces very fine-grained chalk.

Below 1000 m of burial, the effects of chemical compaction and pressure solution become the predominant porosity and permeability reducing processes. Pressure solution strongly affects the reservoir properties, leading to a rapid decline in porosity from 30–50% to about 20–30% at burial depths of 1000–2000 m. The main diagenetic features produced during this stage are pressure solution parallel laminations (Ekdale & Bromley, 1988) and lenticular chalk consisting of small lenses of pure chalk enveloped by clay-rich solution seams (Garrison & Kennedy, 1977). At deeper burial, solution seams change to stylolites, which become the predominant expression of pressure solution (Fig. 13). This occurs in high porosity chalk, as well in chalk with porosity less than 25% (Scholle, 1977). Stylolites generally have an amplitude of few millimetres with the insoluble residue at the stylolite dentate surface formed by various clay minerals, pyrite and dolomite (Dons *et al.*, 1995).

Original porosity of chalk ooze is estimated at ~70–80%, however porosity values up to 50% in chalks at burial depths greater than 3000 m were considered unusual. Early dewatering of the coccolith ooze followed by mechanical and chemical compaction during progressive burial normally would have reduced the original porosity to less than 15% (Scholle, 1977; D'Heur, 1984). Preservation of high porosities at more than 3000 m of burial results from several factors (Fig. 14): (1) creation of a rigid grain-to-grain framework due to early cementation (Mapstone, 1975); (2) overpressure due to rapid subsidence (Scholle, 1977); and (3) early hydrocarbon charging which inhibit or even stop chemical compaction (Scholle, 1977; D'Heur, 1984).

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Figures

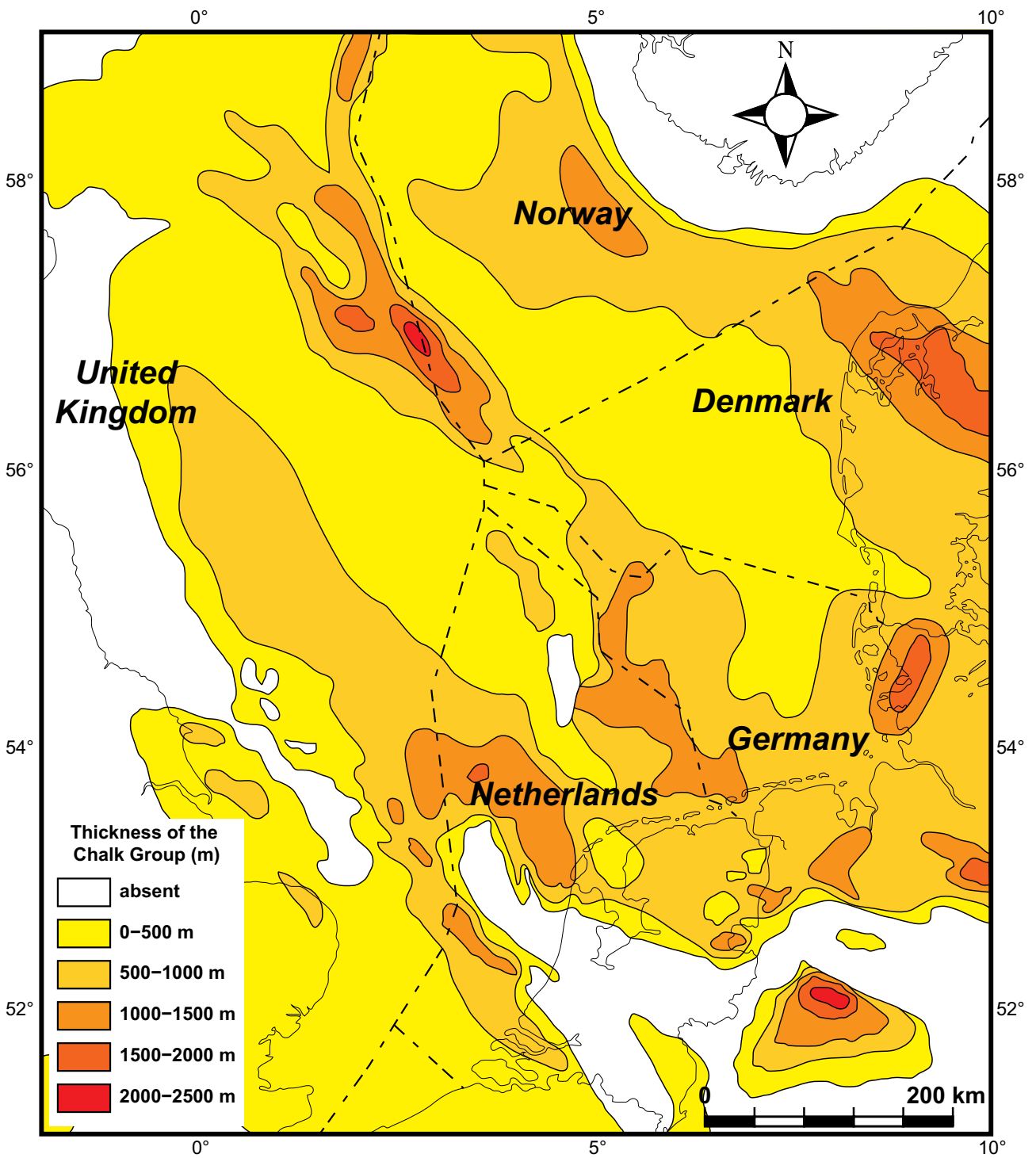


Fig.1. Thickness of the Chalk Group in the central and southern North Sea (Ziegler, 1990).

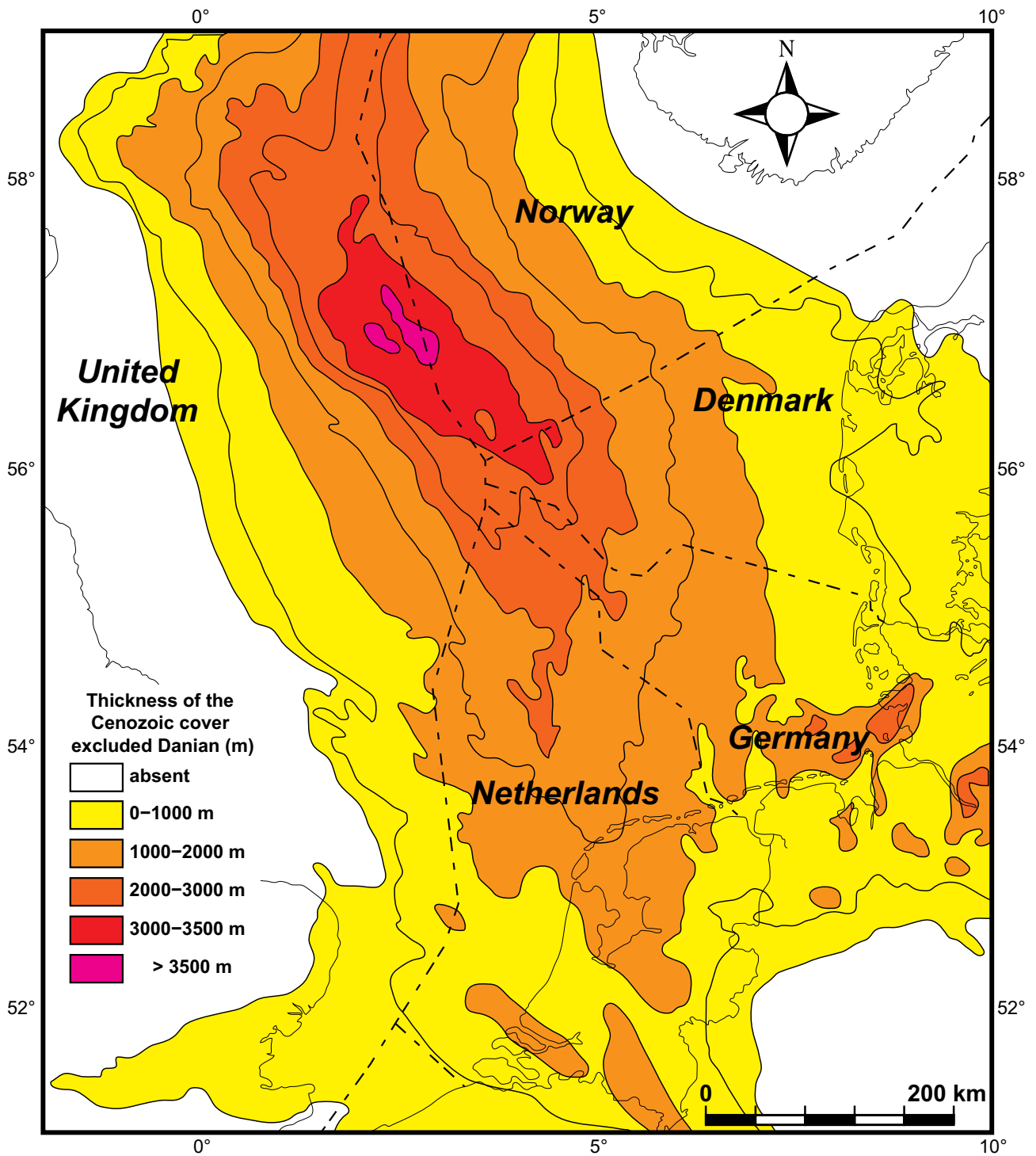


Fig.2. Thickness of the post Chalk Group succession in the central and southern North Sea (Ziegler, 1990).

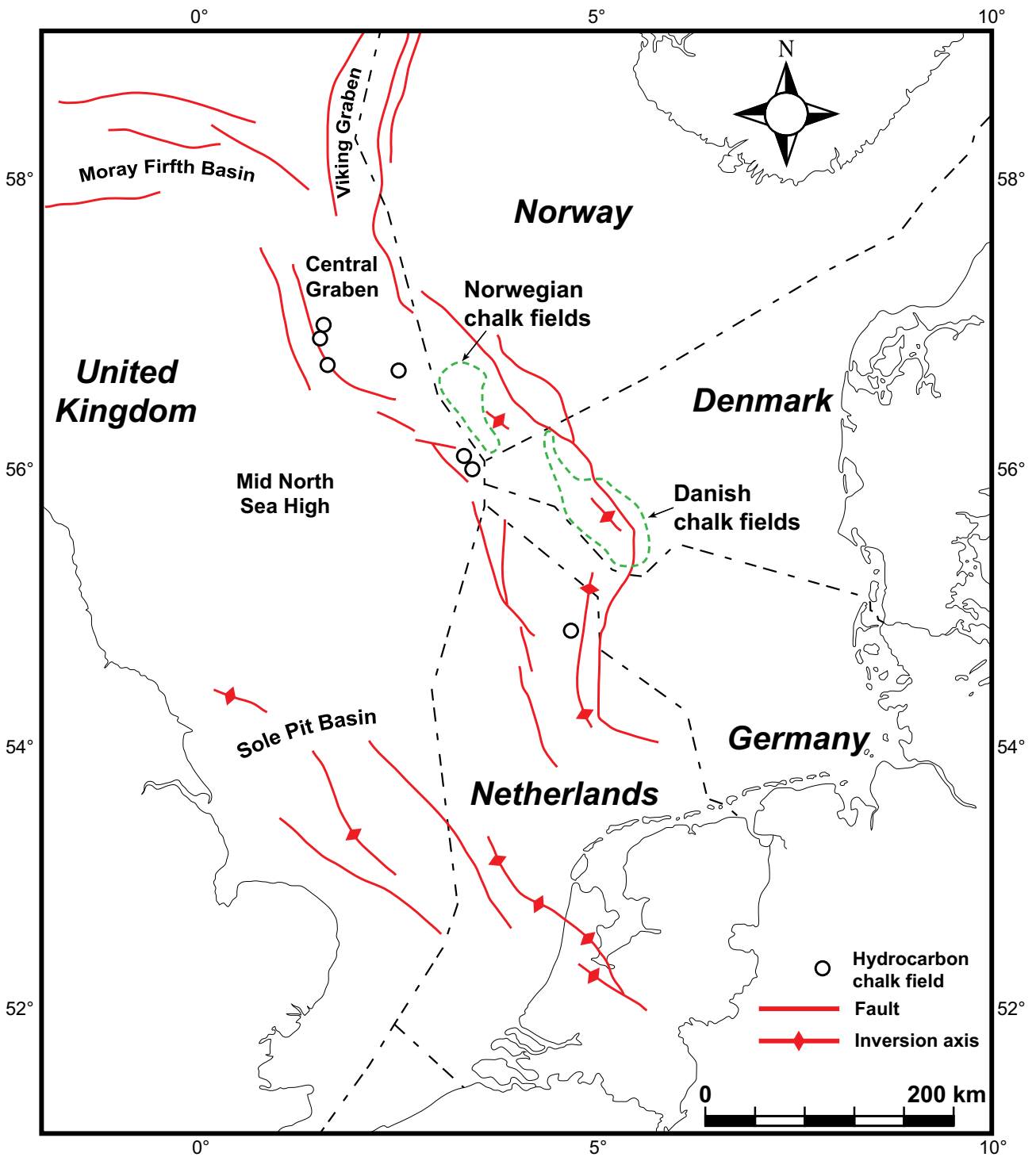


Fig.3. Cretaceous tectonic elements and major structural lineaments of the central and southern North Sea with indicated the major hydrocarbon provinces in the chalk (Ziegler, 1990; Megson & Hardman, 2001).

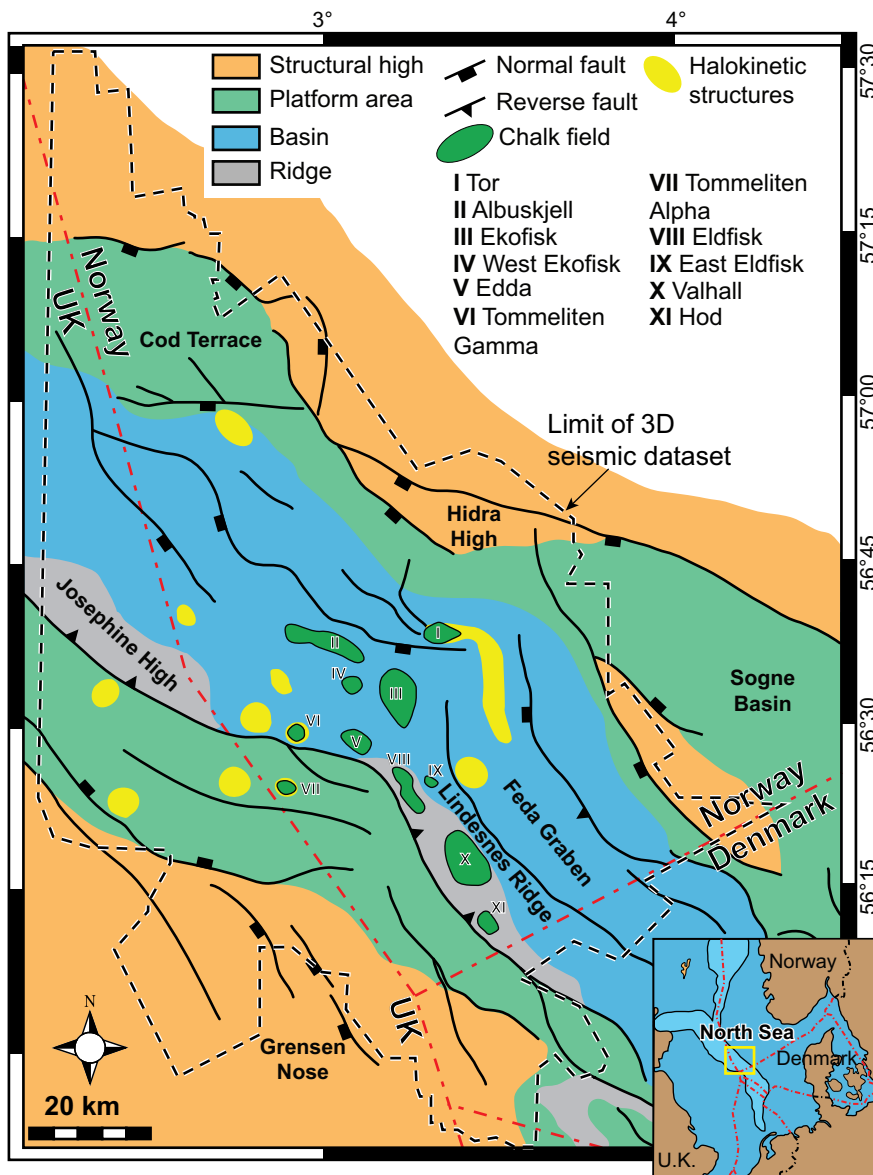


Fig.4. Structural map of the Southern Norwegian Central Graben illustrating the dominant structural and halokinetic features during Late Cretaceous time and the relative position of the chalk fields (after Ziegler, 1990; Gowers *et al.*, 1993; Knott *et al.*, 1993; Bailey *et al.*, 1999). Position of the area relative to the North Sea is shown in the inset map in the bottom right corner.

System - Series	Stress state	Orientation	Salt activity	W-E structural sketch	Structural style
Miocene	Compression		Major diapirism		Rapid subsidence
Oligocene					Subsidence
Eocene					
Paleocene	Compression		Involved in transpressive movements - Diapirism		Transpressional en-echelon folds
Upper Cretaceous					
Lower Cretaceous	Thermal subsidence		Moved into faults		Basin fill
	Compression				
Upper Jurassic	Strong Extension		Withdrawal basins along transverse faults		
Middle Jurassic	Extension				
Lower Jurassic	Extension		Salt diapirs and pillows		Fault development guided by pre-Zechstein lineaments
Triassic					
Permian	Extension		Deposition		Reactivation of basement faults

Fig. 5. Tectono-stratigraphic chart summarizing the main tectonic and salt activity from the Permian to the Miocene (after Ziegler, 1990; Gowers *et al.*, 1993; Knott *et al.*, 1993; Bramwell *et al.*, 1999).

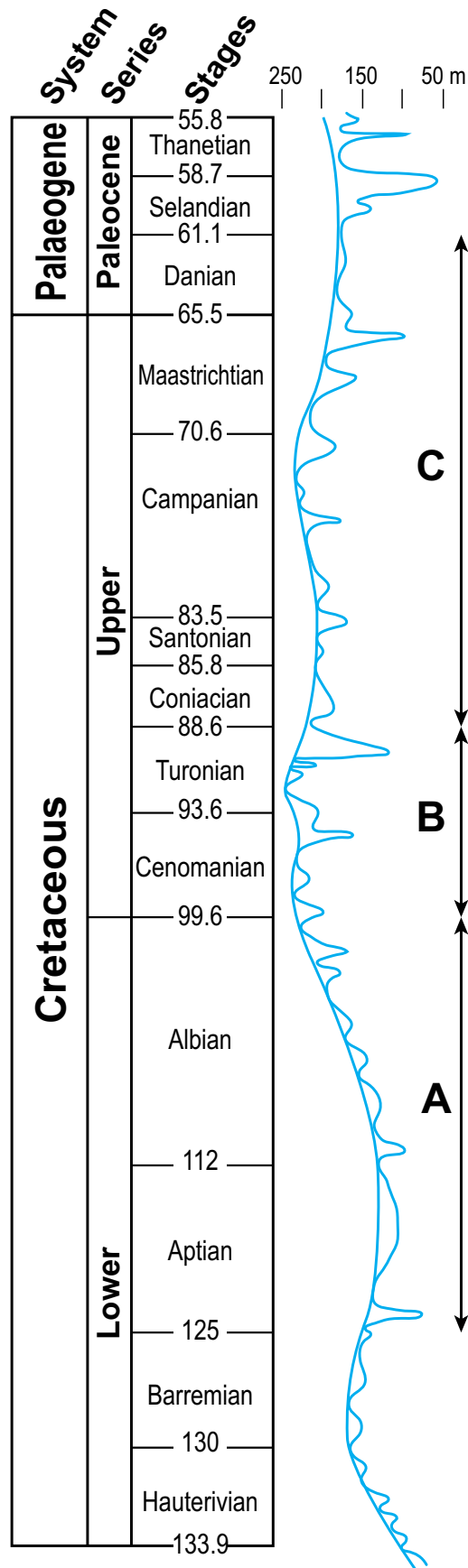


Fig.6. Lower Cretaceous to Paleocene long-term and short-term eustatic sea-level curve (after Ziegler 1990; Hardenbol *et al.*, 1998). Letters referred to intervals displayed in the palaeogeographic maps in Figure 7.



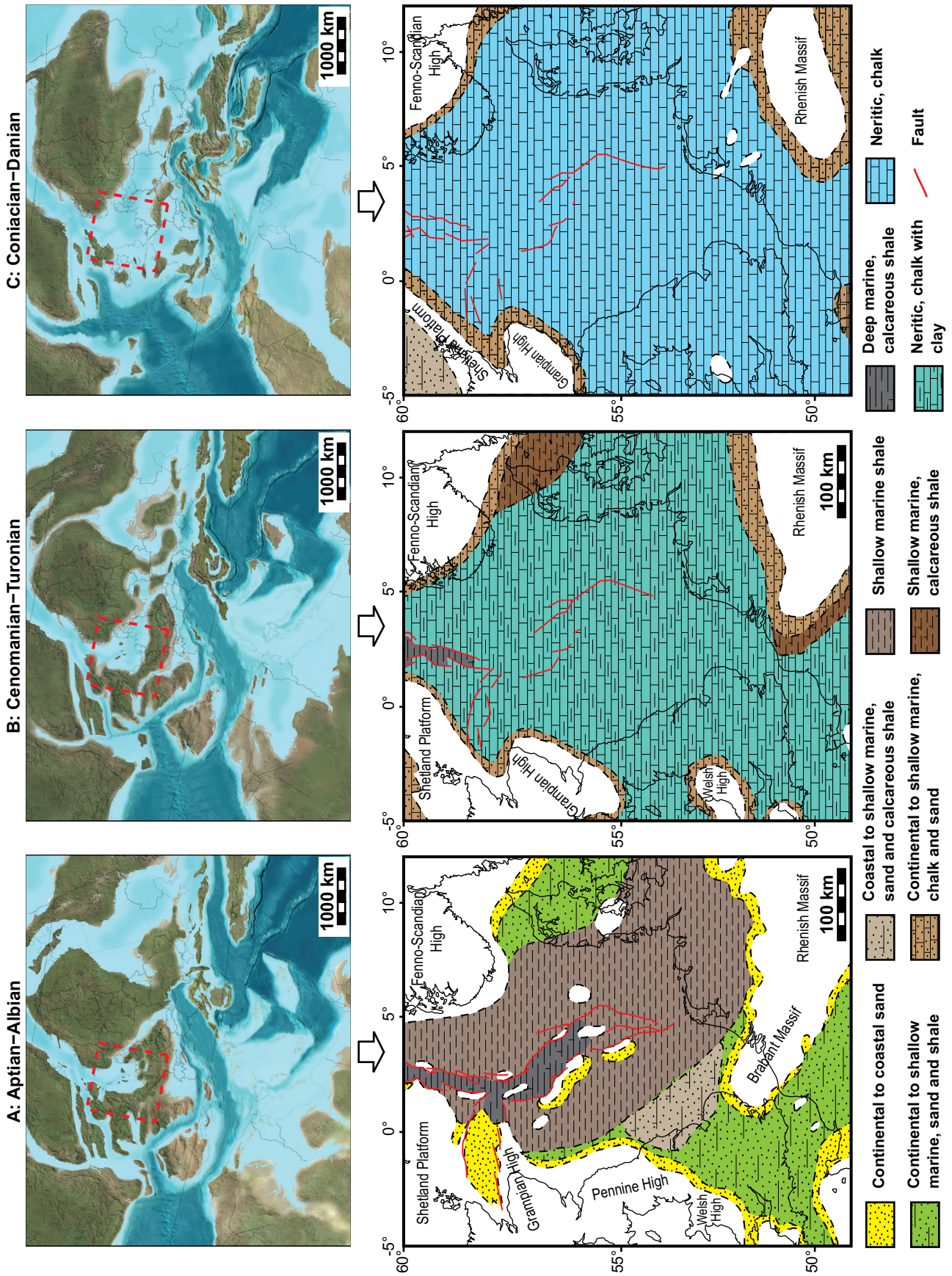


Fig. 7. NW European palaeogeography (A) Aptian–Albian; (B) Cenomanian–Turonian; and (C) Coniacian–Danian. (after Ziegler, 1990). Palaeogeographic maps of NW Europe in the upper row were produced by Ron Blakey, Colorado Plateau Geosystems.



System	Stages	LITHOSTRATIGRAPHIC UNITS				
		Offshore Norway (Isachsen & Tonstad, 1989)		Offshore Denmark (Bailey <i>et al.</i> , 1999)	Offshore UK (Lieberkind <i>et al.</i> , 1982) (Johnson & Lott, 1993)	
Paleo- cene	61.1 Ma. Danian	Chalk Group	Ekofisk Fm.	Ekofisk Fm.	Chalk 6 Unit	Ekofisk Fm.
	65.5 Maastrichtian		Tor Fm.	Tor Fm.	Chalk 5 Unit	Tor Fm.
	70.6 Campanian		Hod Fm.	Magne Fm.	Chalk 4 Unit	Mackerel Fm.
	83.5 Santonian				Thud Fm.	
	85.8 Coniacian			Narve Fm.	Chalk 2 Unit	
	88.6 Turonian				Turonian shale	
	93.6 Cenomanian		Blodøks Fm.	Blodøks Fm.	Plenus Marl Mb.	
	99.6		Hidra Fm.	Hidra Fm.	Chalk 1 Unit	Hidra Fm.

Fig.8. Lithostratigraphic subdivisions of the Chalk Group in the offshore sector of Norway, Denmark and UK.

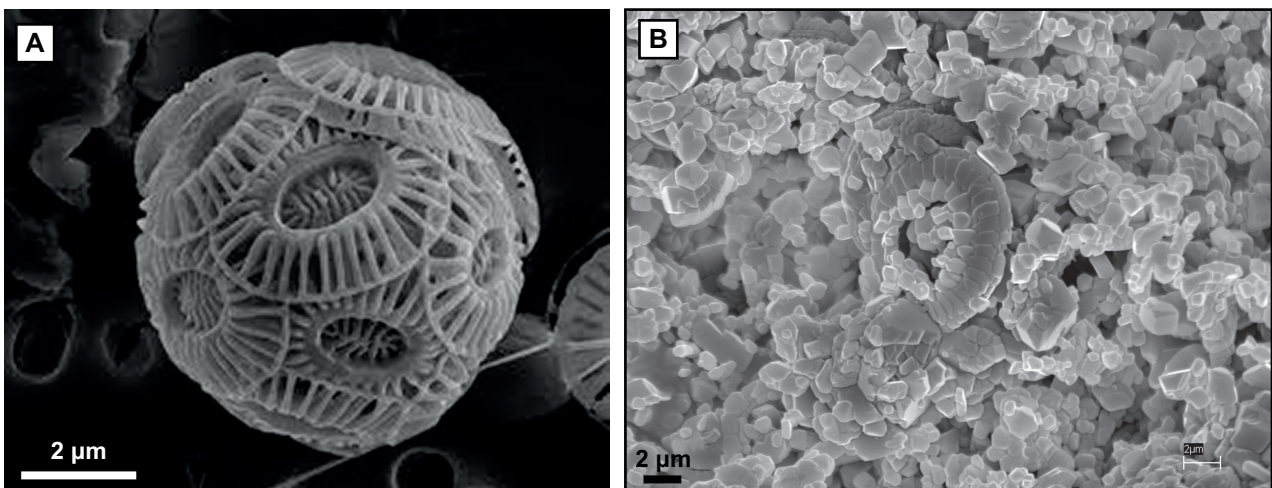
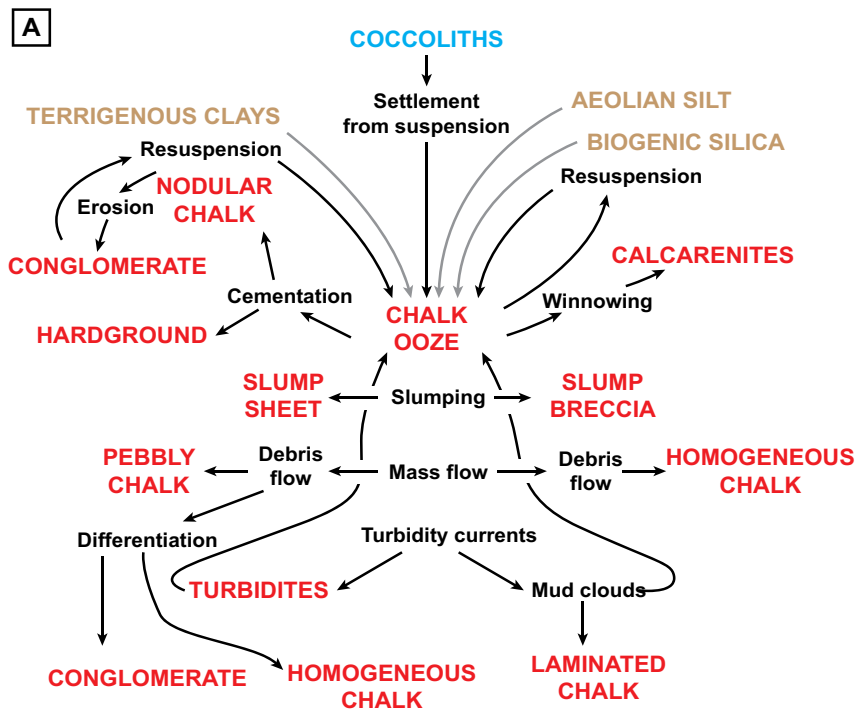
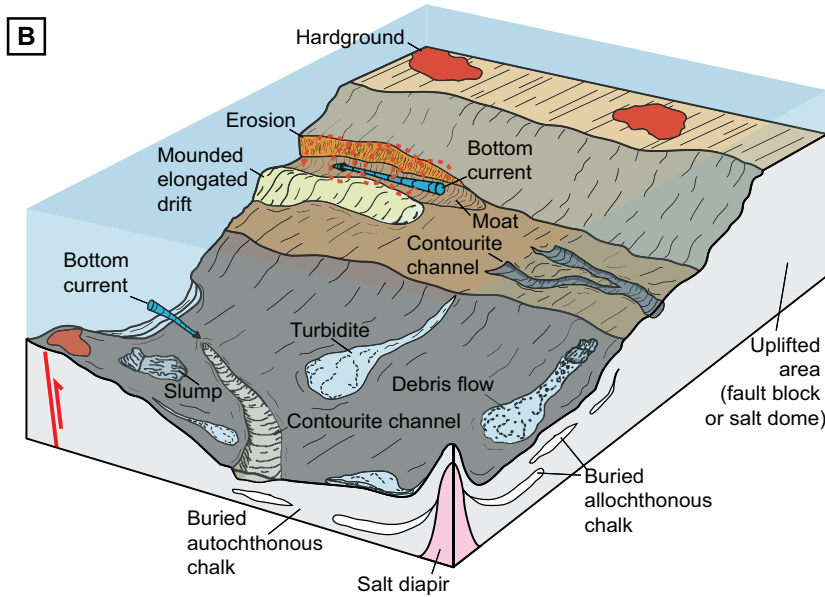


Fig.9. (A) Coccolithophore species *Emiliana huxleyi* (from Tyrrel & Merico, 2004). (B) Scanning-electron micrographs at x10000 of rock chip surface from the Ekofisk Formation in the Ekofisk Field.



**LEGEND**

- COCCOLITHS** - carbonate input
- TERRIGENOUS CLAYS** - non-carbonate input
- Debris flow** - depositional process
- PEBBLY CHALK** - facies



**Fig.10. (A) Overview of the depositional processes and resulting facies types (after Kennedy, 1987). (B) Depositional model of chalk in the Central Graben (modified from Taylor & Lapré, 1987; Surlyk & Lykke-Andersen 2007).**

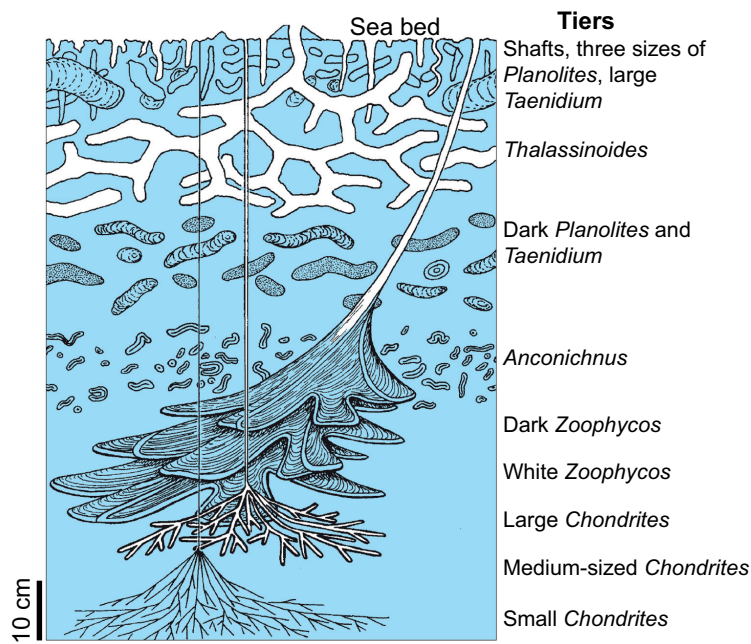


Fig.11. Trace fossil tiering in Maastrichtian chalk. The diagram illustrates the types of trace fossil found at and beneath the palaeo-sea bed in Maastrichtian chalk. Although this example is derived from onshore outcrop in Denmark, analogous features are observed offshore (from Surlyk *et al.*, 2003).

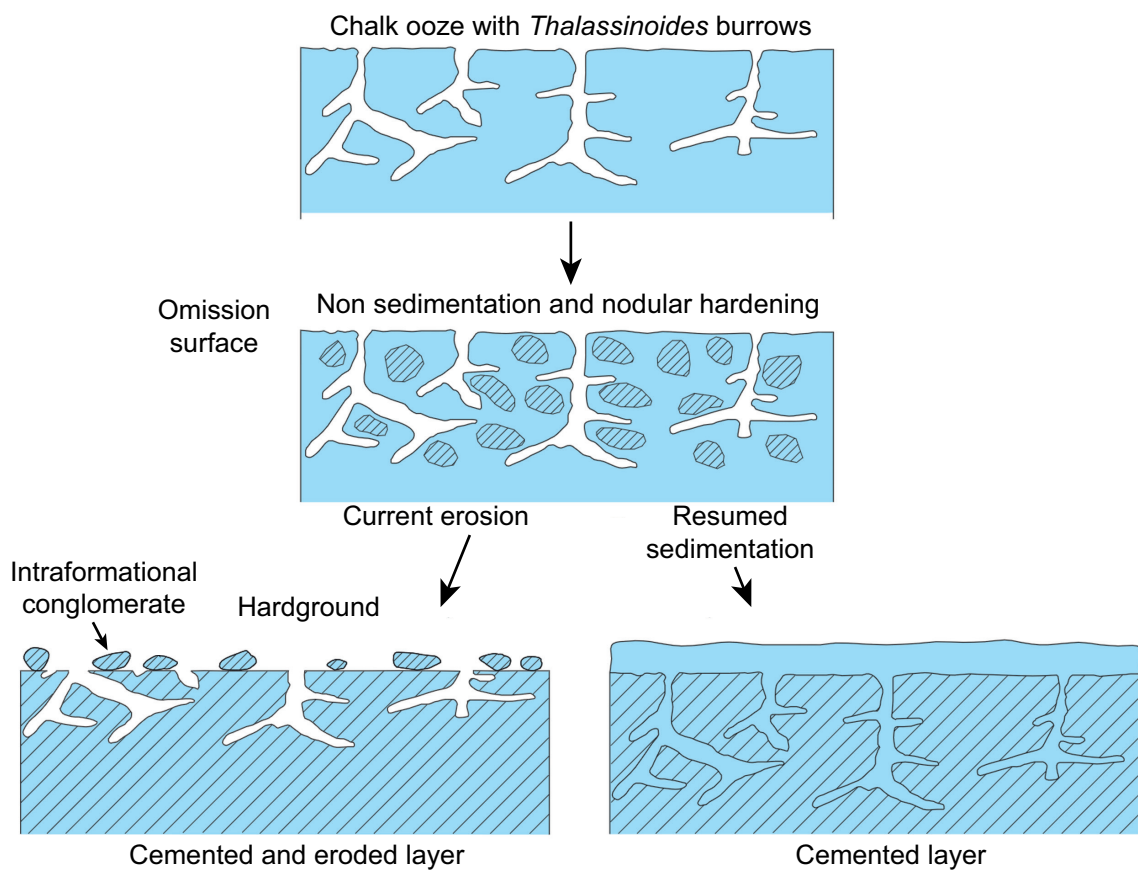
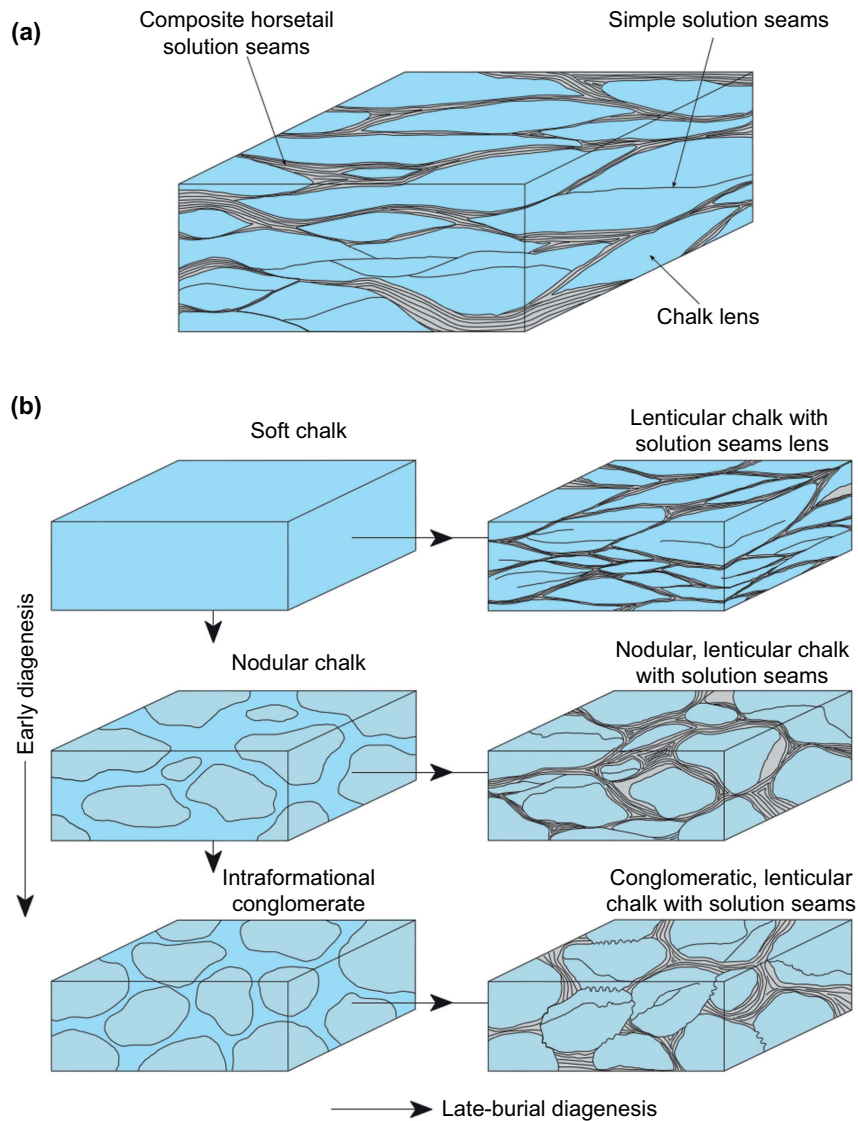


Fig. 12. Diagram illustrating the progressive development of nodular chalk and hardgrounds (from Surlyk *et al.*, 2003).



**Fig. 13. (a) Schematic diagram of lenticular chalk illustrating the terminology used. (b) Early diagenesis of soft chalk may lead to the formation of nodular chalk while late-burial diagenesis form lenticular chalk with solution seams. During late-burial diagenesis of nodular chalk, solution seams concentrate in uncemented areas in between the nodules. Reworking of nodular chalk may produce intraformational conglomerate that under burial diagenesis develop solutions seams separating the intraclasts but not the nodules. Pressure solutions occur where the intraclasts are in contact (from Surlyk *et al.*, 2003).**

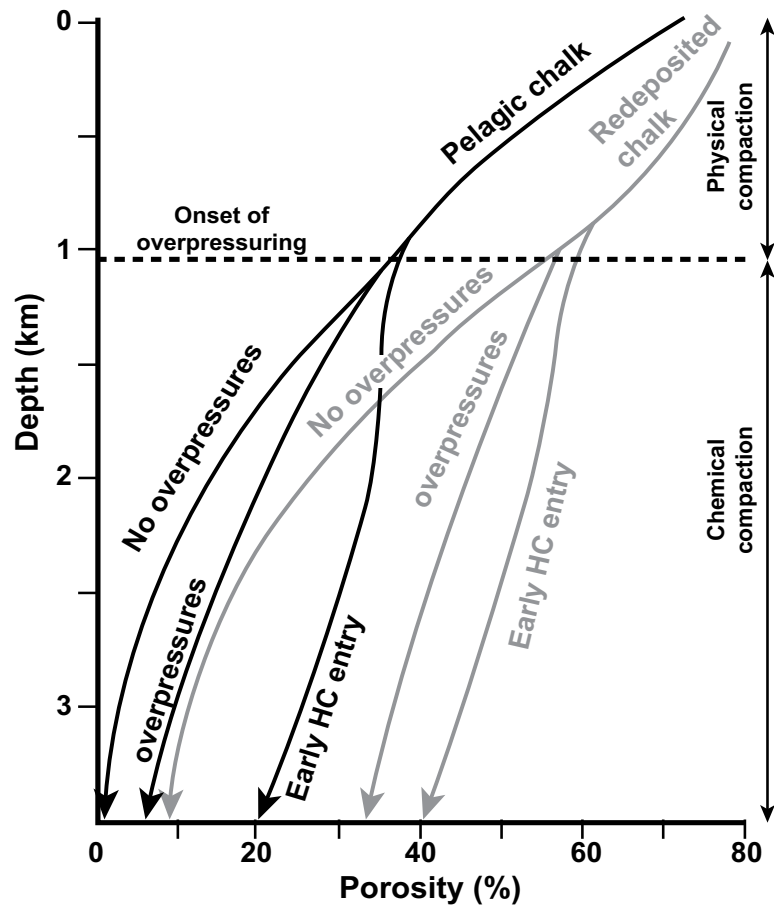


Fig. 14. Diagenetic pathways of Norwegian Central Graben chalk during progressive burial. Differences in porosity between pelagic and redeposited chalk by gravity flows is accentuated during mechanical compaction. Chemical compaction below 1 km of burial varies depending on the timing of overpressure and hydrocarbon charging (after Brasher & Vagle, 1996).