Cretaceous Oceanic Anoxic Event 2 in eastern England: further palynological and geochemical data from Melton Ross

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Abstract: The lowermost 1.45 m of the Welton Chalk Formation, including the regional sedimentary record of Oceanic Anoxic Event 2 (OAE-2), has been sampled at Melton Ross Quarry in eastern England, UK. The section is investigated for organic geochemistry and stable isotopes for the first time, while a detailed palynological study follows previously published preliminary results. It comprises a condensed interval that spans the Cenomanian–Turonian Stage boundary. A locally preserved, lower ‘anomalous’ succession (Beds I–VII) and a ‘Central Limestone’ (Bed A) are shown to correlate respectively with the pre-Plenus sequence and Plenus Bed at Misburg and Wunstorf in the Lower Saxony Basin (LSB), NW Germany. They are overlain by a succession of variagated marls (Bed B to Bed H), including the Black Band (Beds C–E), that can be correlated across eastern England. Based on a carbon isotope (δ13C) profile and dinoflagellate cyst and acritarch bio-event correlation, Beds B–H appear to be a highly attenuated post-Plenus equivalent of the LSB succession, including part of the ‘Fish Slate’. The δ13C profile shows possible ‘precursor’/‘build-up’ events in the lower succession at Melton Ross, with the main OAE-2 δ13C excursion occurring in the Central Limestone and overlying Beds B–H. The darker coloured marls from the Black Band and Bed G contain 1.43–3.47% total organic carbon (TOC), hydrogen index values of 78–203 mg HC/g TOC and oxygen index values of 15–26 mg CO2/g TOC, indicating type III and type II–III organic matter, of mixed terrigenous and marine algal sources. The corresponding palynological assemblages are dominated by marine dinoflagellate cysts, comprising mainly gonyaulacoid taxa, with subordinate terrigenous miospores, mainly gymnosperm bisaccate pollen, consistent with a distal marine setting. The interbedded lighter-coloured marls contain less than 0.4% TOC and lower proportions of miospores and peridinioid dinoflagellate cysts compared with the darker layers. This is suggestive of moderately raised levels of productivity during deposition of the darker layers, possibly related to greater nutrient availability from land-derived sources. The occurrence of the peridinioid taxa Eurydinium saxoniense and Bosedinia spp., together with higher proportions of prasinophyte phycomata in the darker layers, may also point to stimulation of organic-walled phytoplankton productivity by reduced nitrogen chemo-species encroaching the photic zone, possibly by expansion of an oxygen-minimum zone. Exceptionally high concentrations of palynomorphs (in the tens of thousands to lower hundreds of thousands per gramme range) in the darker layers at Melton Ross and eight other eastern England localities is consistent with increased quality of seafloor preservation in a low oxygen environment, coupled with a high degree of stratigraphic condensation. Two new dinoflagellate cyst species are described from Melton Ross, Canningiopinopsis? lindseyensis sp. nov. and Trithyridinium maculatum sp. nov., along with two taxa described in open nomenclature.

Supplementary material: One pdf file, with detailed sample positions and descriptions, tables of supporting information (also available in Excel format), quarry photographs and a palynological distribution chart, is available at https://doi.org/10.6084/m9.figshare.c.4987205

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The Late Cretaceous Epoch was characterized by sustained warm climate, resulting in high eustatic sea-levels (Miller et al. 2005; Haq and Huber 2016; Hay et al. 2018). Numerous epicontinental seaways became established, submerging large areas of Western Europe (e.g. Gale 1995) and the Western Interior of North America (Kauffman and Caldwell 1993). Major global perturbations in the carbon cycle occurred, termed Oceanic Anoxic Events, the most prominent spanning the Cenomanian–Turonian Stage boundary (CTB, 93.9 Ma; International Commission on Stratigraphy 2019), named OAE-2 (Schlanger and Jenkyns 1976; Schlanger et al. 1987) and lasting for up to c. 900 kyr (Eldrett et al. 2015a; Gangl et al. 2019). This interval was marked by a globally recognized positive carbon isotope (813C) excursion, reflecting the widespread sequestration of 812C-enriched organic matter in marine sediments under global anoxic conditions (Jenkyns 2010 and references therein). However, the deposition of dark-coloured, organic-rich, fine-grained sediments (‘black shales’) varied both temporally and spatially, being modulated and ultimately dependent on local and regional processes (basin restriction, water stratification, bottom-currents, sediment input) in addition to global phenomena (Large Igneous Province activity, sea-level change, orbital forcing; e.g. Ernst and Youbi 2017; Clarkson et al. 2018; Minisini et al. 2018). A global mass extinction/turndown bio-event occurred around the CTB and is probably associated with OAE-2 (Raup and Sepkoski 1982, 1984; Kauffman 1984; Milne et al. 1985; Hart 2005, 2019).

In the North Sea Basin and adjacent outcrop areas in NW Germany and eastern England, UK, deposition of dark coloured mudstones interbedded with light coloured mudstones...
and limestones commenced during the Late Cenomanian, and continued into the Early Turonian in some areas, including Wunstorf and Misburg in the Lower Saxony Basin, NW Germany (Fig. 1). At these locations, the CTB succession is 12–30 m thick (Ernst et al. 1983; Hilbrecht 1986). The dark mudstone layers form a distinctive interval in predominantly whitish limestone chalk successions. In eastern England (Fig. 2), a highly condensed (relative to NW Germany) ‘Black Band’ has developed in Yorkshire (0.3–0.7 m thick at Flixton, East Knapton, Bishop Wilton and Market Weighton; Jeans et al. 1991; Dodsworth 1996) and northern Lincolnshire (0.1–0.2 m thick at South Ferriby, Elsham, Melton Ross, Bigby and Caistor). Jenkyns (1985) envisaged a pelagic shelf depositional setting for the Black Band and adjacent strata in a relatively shallow (several hundred metres) epicontinental sea. The Black Band wedges out to the south of Louth (Hart et al. 1991). It appears to represent the ‘feather-edge’ of OAE-2, which dies out when traced towards a palaeohigh that appears to have been located in the region of the Wash (Hart et al. 1991). Schlanger et al. (1987) suggested that it may have been deposited near the upper limit of an oxygen-minimum zone that lapped on to the continental shelf.

The Cenomanian and Turonian stages probably record the highest sea-levels within the Cretaceous Period, with maximum eustatic sea-levels being reached during the Early Turonian (Haq et al. 1988; Sahagian et al. 1996; Miller et al. 2003, 2005). They were probably the time of minimal continental relief during the Phanerozoic and the time of minimal detrital sediment delivery to the ocean (Hay et al. 2018). The CTB interval was also a time of extreme warmth with mid-latitude sea-surface temperature possibly exceeding 35°C (Huber et al. 2002; Voigt et al. 2004; Forster et al. 2007; Robinson et al. 2019). This warmth was likely associated with a more stratified water column, resulting in poor atmosphere–ocean gas exchange, oxygen-deficient photic zone waters, including euxinic conditions in some regions, and enhanced organic matter preservation at the sediment–water interface (Sinninghe Damsté and Köster 1998; Monteiro et al. 2012). Rhythmic bedding in CTB successions has been linked to orbital-forcing (e.g. Arthur et al. 1986; Eldrett et al. 2015b; Boulila et al. 2020), which in some regions, including NW Germany, may have caused an intensified hydrological cycle during warmer/wetter periods (van Helmond 2015; Charbonnier et al. 2018; Gharaie and Kalanat 2018).

The early stages of OAE-2 were characterized by a relatively extended interval (c. 150–200 kyr; Clarkson et al. 2018) of c. 3–5°C cooling of sea surface temperatures in the

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Fig. 1. Turonian palaeogeographic reconstruction with the main Northern Hemisphere site locations discussed in the text: brown shaded area, landmass; light blue, palaeoshelf; CLIP, Caribbean Large Igneous Province; HALIP, High-Arctic Large Igneous Province. Modified from Eldrett et al. (2014, 2017) and Du Vivier et al. (2015).
proto-Atlantic and the European shelf (Forster et al. 2007; Pearce et al. 2009; Sinninghe Damsté et al. 2010; Jarvis et al. 2011; van Helmond et al. 2014, 2015). Boreal realm fauna, including the belemnite Praeactinocamax plenus (Blainville), migrated southwards across Europe (Jeffries 1962, 1963; Gale and Christensen 1996; Marcinowski et al. 1996; Košták et al. 2004). The interval was termed the ‘Plenus Cold Event’ (PCE) by Gale and Christensen (1996) and probably marks a shift towards improved oxygenation of bottom waters during OAE-2 (Forster et al. 2007; Eldrett et al. 2014, 2017; van Helmond et al. 2014), possibly on a global scale (Clarkson et al. 2018; O’Connor et al. 2020), and relatively drier climates (Heimhofer et al. 2018; Gharai and Kalanat 2018). Across regions of Europe that record organic-rich sedimentation during OAE-2, including NW Germany and SE France (e.g. Pont d’Issole, Fig. 1), ‘black shale’ deposition temporarily gave way in some areas to more oxygenated sediments during the PCE, and a resumption of deposition of relatively thick biogenic pelagic limestones (Wiese et al. 2009; Jarvis et al. 2011; Grosheny et al. 2017; Jenkyns et al. 2017). In NW Germany, a conspicuous tripartite limestone bed, called the Plenus Bed or Plenus Bank because it yields the eponymous belemnite, is developed at this level (Ernst et al. 1984; Hilbrecht and Dahmer 1994), separating underlying pre-Plenus and overlying post-Plenus successions of dark coloured mudstones interbedded with light coloured mudstones and limestones.

Wood and Mortimore (1995) and Wood et al. (1997) reported the CTB succession from Melton Ross Quarry in Lincolnshire (Fig. 2). Temporary deep excavations in the late

**Fig. 2.** The Cretaceous outcrop in eastern England and the location of sections discussed in the text (modified from Hart 2019). The Black Band wedges out to the south of Louth (line A–A). It appears to represent the ‘feather-edge’ of OAE-2 when traced towards a palaeohigh located in the region of the Wash. Between lines A–A and B–B the chalks and nodular chalks of latest Cenomanian–earliest Turonian age are dull red in colour, while south of line B–B the chalks at the same stratigraphic level are pale green/grey in colour.
90s (Fig. 3) exposed a section hitherto unreported in eastern England, presumed eroded by unconformity elsewhere, that appears to correlate with the lowermost 2–3 m of the NW Germany CTB succession, including the pre-Plenus interval and the Plenus Bed. These deposits are overlain by the regionally correlative Bed B, Black Band and Beds F–H (Fig. 4).

In a recent review of the Black Band, Hart (2019) discussed advances in foraminiferal knowledge a quarter of a century after previous publications on their distribution across the CTB in eastern England (Hart and Bigg 1981; Hart et al. 1993). The present paper follows up on this work, reviewing advances in palaeoenvironmental knowledge and biostratigraphic dating using organic-walled phytoplankton since this was last discussed for eastern England in the publications of Dodsworth (1996, 2000) and Wood et al. (1997). Previous analysis of palynological recovery from the Black Band at South Ferriby and Flixton (Hart et al. 1993; Dodsworth 1996) has revealed exceptionally high concentrations of dinoflagellate cysts, in the thousands to lower hundreds of thousands per gramme range. Hart and Koutsoukos (2015) recommended further investigation of whether the abundance of dinoflagellate cysts is a function of their increased productivity under eutrophic conditions, increased quality of seafloor preservation in a low oxygen environment, or a normal level of organic productivity accentuated by a loss of biogenic carbonate sediment (including planktonic foraminifera and calcareous nanofossils) during the CTB mass extinction (Jarvis et al. 1988a; Lamolda et al. 1994; Paul and Mitchell 1994; Hart 1996). This paper provides the first detailed palynological data from Melton Ross and attempts to discriminate the relative contribution of these factors. A taxonomic review of new and problematic dinoflagellate cysts is included. Other marine palynomorphs are documented, including prasinophyte phycomata and acanchnomorph acritarchs, along with land-derived (terrigenous) pollen and spores. For regional comparison, summary palynological data are published for eight correlative eastern England localities. From south to north, these are Caistor, Bigby, South Ferriby, Market Weighton, Bishop Wilton, East Knapton, Flixton and Speeton (Fig. 2). Total organic carbon, Rock-Eval pyrolysis and stable isotope data are presented for the first time from Melton Ross. The integrated bio- and chemostratigraphy of the section is assessed in an inter-regional context.

In this paper, chronostratigraphic substages and their corresponding ages are treated as formal units, using initial upper case letters, consistent with the usual practice, although not all have been ratified. The base of the Cenomanian Stage and Lower Cenomanian Substage have a ratified Global boundary Stratotype Section and Point (GSSP) at Mont Risou in southern France (Kennedy et al. 2004). The base of the Turonian Stage and Lower Turonian Substage have a ratified GSSP near Pueblo, Colorado (Kennedy et al. 2005). The base of a Middle Cenomanian Substage has a proposed GSSP in Sussex, southern England.
(Tröger et al. 1996) and that of a Middle Turonian Substage is proposed near Pueblo (Bengston et al. 1996; Kennedy et al. 2000; Dodsworth and Eldrett 2019). Tröger et al. (1996) recommended southern France as a suitable region for an Upper Cenomanian Substage GSSP, though they did not propose a site. Replacement of Acanthoceras ammonites by the genus Calycoceras (Hancock 1991), which can also be correlated using carbon isotopes (Kennedy and Gale 2006), provides a possible datum for the base of an Upper Cenomanian Substage.

Lithostratigraphy

The Late Cretaceous Epoch in England is represented by the bio-micritic, white limestones of the Chalk Group (Fig. 2), which Wood and Smith (1978) grouped into three major faunal and depositional provinces: a Northern Province (investigated here), which links eastern England to the north of the Wash with contemporaneous North Sea and NW German successions; a Southern Province linking southern England and northern France (Anglo-Paris Basin); and a Transitional Province in the Chilterns and East Anglia. In eastern England, chalk bedding is developed on a decimetre scale with more clay-rich marls forming thin (<5 cm) interbeds (e.g. Hancock 1976; Jeans 1980). From Louth northward in Lincolnshire and Yorkshire, the lowermost c. 0.5 to 1.5 m of the Welton Chalk Formation (Wood and Smith 1978) is characterized by an atypically thick succession of variegated marls, including the organic-rich Black Band (Fig. 4).

Hopson (2005) gave a comprehensive review of the historical naming of the lowermost part of the Welton Chalk Formation, along with evidence for its stratigraphic position, based on chemostratigraphy (stable carbon isotopes: Schlanger et al. 1987; Hart et al. 1991; Gale et al. 1993; Wood and Mortimore 1995; and subsequently, Clarkson et al. 2018) and macrofossils (e.g. Jefferies 1963; Whitham 1991; Gaunt et al. 1992; Wood et al. 1997). Hopson (2005) proposed using the term Plenus Marl Member for marl and limestone beds between the base of the Welton Chalk Formation and the base of the Black Band, on the grounds of their probable correlation with the Plenus Marls of the Anglo-Paris Basin. He also proposed using the term Black Band Member for marls and limestones between the base of the Black Band and the base of the Buckton Member (Mitchell 2000), on the grounds of these deposits probably being stratigraphically higher than the Plenus Marls, equivalent to part of the Melbourn Rock Member/Ballard Cliff Member in southern England.

However, subsequent workers have retained one unit, the Fixton Member of Jeans (1980), for both intervals (Hart 2019; Mitchell 2019). In Table 1, the subdivision of the Fixton Member into two sub-members (Brett et al. 2018) is suggested, based on the definitions of Hopson (2005).

In the Anglo-Paris Basin, Jefferies (1963) labelled the eight correlative beds of a ‘standard’ Plenus Marls succession as Beds 1–8 in ascending order. Any strata that could not be correlated with these beds were labelled locally with lower case Roman numerals. Jefferies (1963) could not trace Beds 1–8 farther north than Marham in Norfolk (Fig. 2), though subsequent work (Voigt et al. 2006) tentatively identified the uppermost units (Bed 7 and Bed 8) at Heacham and Barret Ringstead, near Hunstanton. Jefferies (1963) investigated one locality from Lincolnshire (South Ferriby) and another from Yorkshire (Speeton), with the Black Band assigned to ‘bed i’ at Speeton and ‘bed iii’ at

Table 1. Lowermost part of the Welton Chalk Formation: a comparison of bed nomenclature. Sample positions are indicated

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South Ferriby. Hart et al. (1991, 1993) applied Jeffreys’ (1963) informal bed names at South Ferriby. Subsequently, individual beds within the Flixton Member have been assigned different names by different authors during research published in the 1990s (Table 1). Most of these schemes use ascending numbers comparable to the Anglo-Paris Basin succession whilst not intending to imply lithostratigraphic correlation with southern England beds of the same number. To avoid confusion, Dodsworth (1996) alternatively proposed using letters rather than numbers for the eastern England succession, and erected a regional lithostratigraphic scheme of comparable resolution to previous schemes.

In hydrocarbon wells from the central North Sea Basin, e.g. well 47/10-1 (Fig. 2), the distinctive black, grey and green lithologies described from the CTB interval (Rhys 1974; Burnhill and Ramsay 1981) are comparable to those from the lowermost part of the Welton Chalk Formation of onshore eastern England, and contain comparable foraminiferal assemblages (Burnhill and Ramsay 1981; Crittenden et al. 1991) and palynological assemblages (Marshall and Batten 1988; Dodsworth 1996). The term ‘Plenus Marl Formation’ was initially applied to the corresponding high gamma log-response unit in UK sector North Sea wells by Batten (1988; Dodsworth 1996). The term was initially applied to the corresponding high gamma log-response unit in UK sector North Sea wells by Weise (1997, 1997, fig. 4; Fig. 3), which contain a thicker lower succession than Site 1, and categorized the localized deposits into seven units, Beds I–VII. They expressed uncertainty about how Beds III–VI correlate with Site 1. At the time of writing (2019), sites 1–4 are buried c. 10 m below the restored surface of the quarry.

In addition, Wood et al. (1997) collected their own set of 27 samples from Site 2 (labelled CJW-1 to -26) and presented rare earth element data, analysis of clay mineralogy and preliminary palynological comments from those samples. They reported shale-normalized rare earth element profiles to be sub-horizontal with no Europium anomaly, suggesting to them a detrital rather than volcanogenic provenance for the CTB succession (Wood et al. 1997, fig. 5). They further reported the clay mineralogy to comprise a mixture of illite, kaolinite and smectite at the base of the succession, but becoming predominantly smectite upwards with the loss of kaolinite in the regionally correlative upper succession (Wood et al. 1997, fig. 4). They interpreted this change in mineralogy as probably suggesting either deepening or increasing distance from shorelines through the CTB interval (Wood et al. 1997, p. 339).

Wood et al. (1997, p. 342, fig. 3) recovered Praeactinocamax plenus from Site 1 at Melton Ross, in silty lithologies immediately above the Central Limestone (Fig. 4). They considered the latter to correlate with the Plenus Bed in NW Germany. Wood and Mortimore (1995) and Wood et al. (1997) did not publish logs of the chalk section above the variegated marls (Bed B to Bed H) at Melton Ross, i.e. from the upper part of the Flixton Member. This section is logged in the present study (Fig. 4) and reveals a thickness of chalk (0.65 m) between the top of the variegated marls and ‘Adrian’s pair of marls’ (of Mortimore 2014), immediately below the base of the Buckton Member, comparable to that at nearby South Ferriby and Caistor (Hart et al. 1991, fig. 2; Mortimore 2014, fig. 4.23a; Hart 2019, fig. 4).

**Previous palynological work**

There have been several previous palynological investigations of the lowermost part of the Welton Chalk Formation. R.J. Davey, in Hart and Bigg (1981), reported dinoflagellate cysts to be the dominant palynomorphs at Elsham (Fig. 2), with subordinate bisaccate pollen and very rare spores. Hart and Bigg (1981) suggested that marine algae were probably the main source of the abundant organic matter in the Black Band. They also noted that adjacent marls and chalks at Elsham were palynologically barren, probably a function of unfavourable lithologies for the preservation of organic-walled microfossils. A species list and distribution chart were not provided. Marshall (1983, p. 172–174) and Marshall and Batten (1988) undertook detailed analysis of six South Ferriby samples. Cyclonephelium compactum–membraniphorum was reported to be the dominant dinoflagellate cyst in the Black Band, with sparse recovery of palynomorphs from marls immediately below and above (Beds B and F), and palynologically barren samples from adjacent chalks. Microplankton compose 76% of an assemblage from the lower part of the Black Band (Bed C) and 89% from the upper part (Bed E). The Cenomanian marker taxon Litosphaeridium siphoniforum was not recorded in any of the samples. Duane (1992, p. 277–292; in Hart et al. 1993)
provided palynological data for four productive samples from the east wall at South Ferriby quarry (Beds C to G); adjacent chalks and marls from c. 75 cm above and 65 cm below were palynologically barren. *Cyclonephelium compactum–membraniphorum*, along with *Eurydinum saxoniense*, are dominant in the dinoflagellate cyst assemblages of the productive samples, while terebrigenous bisaccate pollen and spores compose <5% of the total palynological assemblages. Exceptionally high concentrations of dinoflagellate cysts (thousands per gramme), relative to Middle Cenomanian chalks and marls of southern England (hundreds per gramme; Paul et al. 1994), were reported. Duane (1992; in Hart et al. 1993) did not record *Litossaeridium siphoniphorum* in any of the South Ferriby samples.

Dodsworth (1996) published palynological data from three samples of the Black Band from the south wall at South Ferriby quarry (recovery from adjacent marl samples was sparse), six samples from the Black Band at Flixton (splits of the same samples analysed for geochemistry by Jeans et al. 1991), two samples from Bed G at Market Weighton, and commented on the palynology of the Black Band at Caistor, Bigby, Market Weighton, Bishop Wilton, East Knapton and Spieeton. As with previous investigations, *Litossaeridium siphoniphorum* was not recorded in any of the samples. However, taxa whose last occurrences Marshall and Batten (1988) calibrated approximately to an influx of the latest Cenomanian zonal ammonite *Neocardioceras juddii* Barrois and Guerne in NW Germany (Hilbrecht 1986), namely *Adnatosphaeridium tutulosum* (in Bed C at Flixton) and *Carpodinium obliquicostatum* (in Bed C at South Ferriby, and Beds C and E at Flixton), were recorded and illustrated. Batten, in Wood et al. (1997), studied two samples from the Black Band at Melton Ross (one each from Bed C and Bed E) and confirmed an absence of *L. siphoniphorum*. However, analysis of five samples from local Beds II–VI at Site 2 revealed the first records of this taxon in the Northern Province, where it is common in occurrence. Dodsworth (2000, fig. 12), provided preliminary marker taxa distribution data from the samples documented more fully here (Fig. 4), corroborating Batten’s record of common *L. siphoniphorum* in local Beds I–III at Melton Ross (Site 1), and indicating the presence of *A. tutulosum* and *C. obliquicostatum* higher in the succession (Site 4), from Beds C–F and Bed C, respectively.

**Material and methods**

Twenty-seven channelled (composite) rock samples (MR97 series) were collected from the top of the Ferriby Chalk Formation and lowermost 1.45 m of the Welton Chalk Formation in Melton Ross Quarry by one of us (PD) on 24 February 1997 and are documented in the present paper. Eleven of our samples (MR97-23 to -14) were collected from Site 1 of Wood and Mortimore (1995). Sites 2 and 3 and the lower part of Site 4 of Wood et al. (1997) were submerged below the quarry’s water table in February 1997 and were unavailable for sampling. The upper part of Site 4 was accessible and 16 of our samples (MR97-13 to -1) were collected from the relatively freshly-exposed, regionally correlative beds there. The thickness of channelled samples varies from 3 mm to 230 mm, depending on the thickness of rock layers (Fig. 4; see the Supplementary Material for details of the samples). The samples were prepared and analysed for organic geochemistry, palynology and carbonate stable isotopes. In the following text, the sample series prefix is not written out in full for each sample; thus, sample MR97-1 is shortened to sample -1.

**Organic geochemistry**

All samples were analysed for total organic carbon (TOC). Samples with >0.25% TOC (-1 to -11.5, and -22) were also analysed for Rock-Eval pyrolysis. Analyses were undertaken in the laboratory of Applied Petroleum Technology (APT) AS, Oslo, Norway. All procedures followed Weiss et al. (2000). For TOC, a Leco SC-632 instrument was used. Diluted hydrochloric acid (HCl) was added to the crushed rock samples to remove carbonate. The samples were then introduced into the Leco combustion oven, and the amount of carbon in the sample was measured as carbon dioxide (CO₂) by an IR-detector. For Rock-Eval pyrolysis, a HAWK instrument was used. Jet-Rock 1 was run as every tenth sample and checked against the acceptable range given in Weiss et al. (2000). The temperature programme was a five minutes purge before pyrolysis: 300°C (three minutes) plus 25°C per minute until 650°C was reached.

**Palynological processing**

Five grammes (dark coloured lithologies) or ten grammes (light coloured lithologies) of crushed, dried material from samples -1 to -22 were dissolved in hydrochloric acid (35% HCl) and hydrofluoric acid (40% HF) in order to remove carbonate and silicate minerals, respectively. Preparations were sieved with a 10 µm mesh. Kerogen slides were prepared at this stage. Palynomorphs (illustrated in Figs 5–7), brown and black wood fragments (vitrinite and inertinite phytoclasts) and clumps of granular amorphous organic matter (AOM) tend to occur in comparable proportions in the >10 µm kerogen from the <1% TOC samples at Melton Ross (Table 2; Fig. 8). To improve remaining residues for the counting of palynomorphs, many of these preparations were cleaned by treatment with a ‘nitric wash’, i.e. one minute of oxidation with nitric acid (70% HNO₃), or a rinse with potassium permanganate (KMnO₄), and in not selectively destroying gonyaulacoid dinoflagellate cysts (cf. Dodsworth 1995, 2004a, fig. 4; Dodsworth and Eldrett 2019). All oxidized preparations were stained with Safranin O solution (red stain). Full laboratory processing records are available in Supplementary Table A.

Where palynological recovery permitted, approximately equal portions of quantified organic residues from each sample were strewn over four 22 × 22 mm cover slips, dried
Fig. 5. Dinoflagellate cysts and miospores from Melton Ross. All specimens were photographed at ×400. Sample/slide numbers, specimen England Finder co-ordinates, and BGS specimen registration numbers (MPK prefix) are given. Specimens 1–16 are dinoflagellate cysts; 17–19 are gymnosperm pollen; 20–21 are angiosperm (Normapolles) pollen. The 50 µm scale bar applies to all specimens.

and mounted on to microscope slides using Norland Optical Adhesive 61. The proportion of organic residue strewn on each cover slip (e.g. 10% of that derived from a 5 g sample) was used to calculate the equivalent mass of original dried rock material represented (e.g. 5 g × 0.10 = 0.50 g represented on the cover slip).

To obtain an estimate of the number of palynomorphs on coverslips, the number in a 1/44 traverse of each was counted and multiplied by 44. To give an estimate of 'absolute abundance', i.e. the concentration of palynomorphs in each sample (counts per gramme, cpg), the mean number of palynomorphs per coverslip was divided by the approximate mass of dried rock material represented on each coverslip. Calculations for each sample are given in Supplementary Tables B–E. Relative abundances were estimated by counting the first three hundred...
palynomorphs identified (0.3% = 1 specimen; 0.7% = 2 specimens; 1% = 3 specimens, etc.). The remainder of the first coverslip and, where applicable, the three additional coverslips, were subsequently scanned for additional rare taxa. Relative abundances are described as ‘rare’ (outside the count), ‘frequent’ (0.3–0.7%), ‘common’ (1–9.7%) or ‘abundant’ (10% +). Separate counts of 100 kerogen particles (AOM, phytoclasts and palynomorphs) were made from the unoxidized kerogen slides. Further samples from Caistor, Bigby, South Ferriby, Market Weighton, Bishop Wilton, East Knapton, Flixton and Speeton were processed and analysed for palynology only, using the methods described above, with preliminary summary results presented herein.

Standard palaeoenvironmental parameters have been calculated for the Melton Ross section, including: (i) the ratio between terrestrial (T) and marine (M) palynomorphs (T/M ratio) as a proxy for terrestrial input, (ii) the ratio between peridinioid and gonyaulacoid dinoflagellate cysts (P/cysts and G/cysts, the P/G ratio) as a proxy of nutrient input and (iii) the species richness, i.e. the number of dinoflagellate cyst taxa recorded as a proxy of their diversity. Our detailed discussion of palaeoenvironmental parameters is provided in Eldrett et al. (2017); see also McLachlan et al. (2018) for a recent review of the P/G ratio.

All Melton Ross palynological slides are curated in the MPA and MPK collections of the British Geological Survey, Keyworth, Nottingham, UK (slide numbers MPA 70686 to 70778 and MPA 71640 to 71642; type and figured specimen numbers MPK 14662 to 14716). For the relationship between MPA numbers and the original sample numbers used in this paper, see Supplementary Table F. A full range chart of palynological data is available in the Supplementary Material. Full author names and synonyms of dinoflagellate cysts, prasinophyte phycomata and acritarchs can be found in Fensome et al. (2019).

Stable isotopes
All samples were analysed for stable carbon and oxygen isotopes on carbonates (Supplementary Table G). Analyses were undertaken in the laboratory of the Department of Earth Sciences, Oxford University, UK. Oxygen and carbon isotope analytical methods were adapted from those described in Day and Henderson (2011, section 2.7). All oxygen isotope measurements were performed on a Delta V Advantage isotope mass spectrometer fitted with a Gas Bench II. The Gas Bench II device converted the carbonates to carbon dioxide (CO2) with 100% phosphoric acid (H3PO4) at 72°C (McCrea 1950). The relative $\delta^{18}O$ of carbonate are expressed in per mil ($\delta$‰) relative to Vienna Pee Dee Belemnite (VPDB) on a normalized scale such that the $\delta^{18}O$ of NBS-19 is -2.2‰ and the $\delta^{18}O$ of NBS-18 is -23.01‰. The relative $\delta^{13}C$ values ($\delta^{13}C$) of carbonate are expressed in per mil relative to VPDB on a normalized scale such that the $\delta^{13}C$ of NBS-19 is 1.95‰ and the $\delta^{13}C$ of NBS-
18 is -5.01‰. External error (0.07 and 0.09 for $\delta^{13}C$ and $\delta^{18}O$, respectively) is calculated from repeat measurements of Oxford University’s in-house standard NOCZ. It is assumed that the phosphoric acid–carbonate fractionation is the same for NBS-19 and Oxford University’s calcite samples (Coplen 1996). For carbonates and waters, results are expressed on the same normalized scale such that $\delta^{18}O$ of SLAP2 reference water is -55.5‰.

Results

Organic geochemistry

The two chalk samples (top Ferriby Chalk Formation, sample -23, 0.04% TOC; Central Limestone, sample -14, 0.10% TOC) contain the lowest organic carbon values in the sampled section. The ‘anomalous’ lower succession marls (local Beds I–VII; samples -22 to -15) range from 0.11% to
0.16% TOC, with the exception of the darker coloured marl layer (local Bed II, sample -21), which has 0.28% TOC. In the ‘standard’ upper succession, the lighter coloured lithologies of Bed B (sample -13, 0.12% TOC; sample -12, 0.20% TOC), Bed D (sample -9, 0.36% TOC) and Bed F (samples -3 to -5, 0.12% to 0.34% TOC) contain the lowest values while the dark grey marl samples contain the highest values: Bed C (samples -11.5, -11 and -10, 1.43% to 2.2% TOC), Bed E (samples -8.5 and -8, 2.74% and 2.18% TOC, respectively) and Bed G (sample -2, 3.47% TOC). The hydrogen index (HI) in the dark grey marl samples ranges from 78 (sample -11) to 203 mg HC/g TOC (sample -2) while the oxygen index (OI) ranges from 15 (sample -11.5) to 26 mg CO2/g TOC (sample -8) in the same samples (Supplementary Table II). On a modified Van Krevelen diagram (Fig. 9), these samples plot within the Type III organic matter (mainly terrestrially derived) field, with samples -11.5 and -2 plotting towards Type III–II organic matter (mixed terrigenous and marine type). The lighter coloured lithologies between the dark grey marls give HI

Fig. 9. Van Krevelen-type diagram for the Melton Ross samples analysed for Rock-Eval pyrolysis. Blue-filled circles are data for the >1 wt% TOC samples. Open circles are data for the <1 wt% TOC samples; the latter are unreliable, due to probable elevation by occluded CO2 within carbonate of inorganic origin, in addition to that derived from early diagenesis of organic matter.
values of 40 to 78 mg HC/g TOC and relatively high OI values (27 to 165 mg CO$_2$/g TOC), as does sample -22 from local Bed II ($\text{HI} = 39$, $\text{OI} = 53$; Supplementary Table H). However, with regard to characterizing organic matter type, the OI data may be unreliable in these low TOC, carbonate-rich samples, due to probable elevation by occluded CO$_2$ within carbonate of inorganic origin, in addition to that derived from early diagenesis of organic matter (P. Barnard pers. comm. 2019).

**Palynology**

In the lower succession, local Bed I (sample -22) and Bed ?III (sample -20), respectively, yielded an estimated 53 and 95 palynomorph counts per gramme (cpg) while Bed II (sample -21) has a much higher concentration of palynomorphs, 12 408 cpg (Table 2). Beds ?IV to VII (samples -19 to -17) yielded between 5 and 88 cpg, dominated by one taxon, Dinoflagellate? type D of Ioannides (1986) (Table 3; Figs 6.13, 6.14). The topmost part of the lower succession (samples -15 and -16) and overlying Central Limestone (sample -14) are palynologically barren or contain up to two dinoflagellate cysts. The dinoflagellate cyst assemblage from local Bed II (sample -21) is diversified with 77 taxa present (Table 2). The adjacent samples (-22 and -20) from local Beds I and ?III, along with the samples from Bed B (-12 and -13) contain fewer taxa (32–33) but this may mainly reflect a much smaller number of specimens inspected from these lower recovery samples. Relatively high diversity dinoflagellate cyst assemblages are noted from the darker lithologies of the Black Band (68–79 taxa) with slightly fewer taxa in the lighter coloured inter-beds (Bed C, sample -10.5, and Bed D; 65 and 63 taxa, respectively) and overlying Bed F (53–69 taxa). Bed G yielded 62 taxa, fewer than from comparable dark lithologies in the Black Band.

The lower succession has the highest P/G ratio in the sampled section (average 0.61), with P-cyst taxa
Palaeohystrichophora infusorioides (Fig. 5.13) and Subtilisphaera pontis-mariae (Figs 5.14, 5.15), respectively comprising an average 19.4% and 23.0% of total palynological assemblages from samples -22 to -20 (Table 4). In the upper succession, P-cysts are subordinate to G-cysts in all samples, though they occur in higher relative and absolute abundances in the more organic-rich (>1% TOC), dark grey marl samples (average P/G ratio of 0.33) than the lighter coloured lithologies of Beds B–D (average P/G ratio, 0.04) and upper Bed E to Bed F (average P/G ratio, 0.26; Table 2).

Trithyrodinium suspectum (Fig. 5.12) and Ginginodinium? sp. A of Prauss (2006, 2012a) (Figs 6.3, 6.7) occur in most of the productive samples from both lower and upper successions, respectively comprising 0.3–7% and 0.3–1.7% of total palynological assemblages. Palaeohystrichophora infusorioides and S. pontis-mariae also occur consistently in the upper succession. With the exception of sample -6, the former is relatively rare (1–5%) in the Black Band and more common (7.7–11.3%) in Bed F while the latter is more common in the >1% TOC samples of the Black Band (10.7–18.7%). Bosedia cf. sp. 1 of Prauss (2012b) (Figs 7.5, 7.9, 7.10) is also relatively common in the >1% TOC samples (1.3–6%). Eurydinium saxoniense (Fig. 5.6) occurs consistently in Beds C and D, with its first confirmed occurrence in sample -11.5 at the base of the former, and is relatively common (1.7–8%) in Beds E, F and G (Table 4).

The G-cyst taxon Spiniferites spp. is prominent in most productive samples throughout the section, 5.4–28.3% of total palynological assemblages, while Pterodinium spp. comprises 1–5% of assemblages, with an acme (10%) in sample -3 from Bed F (Table 3). Litosphaeridium siphoniphorum (Figs 5.1, 8.12) is restricted to the lower succession, where it is common (1–2.7%). Cyclonephelium compactum–membraniphorum (Figs 5.5, 6.4) is rare in the lower succession but is consistently abundant throughout the upper succession, from basal Bed B to Bed H (average 21.5%). An isolated specimen of Oligosphaeridium toton (Figs 5.9, 5.10, 8.9) was recorded from sample -17 but the taxon is only consistently recorded from Bed B to Bed F and has an acme in upper Bed B (40.3%; Table 3). Sepispinula? huguoniotii (Fig. 5.16) is common to abundant from Bed B to lower Bed E (samples -13 to -8.5).

There is a relative increase in prasinophyte phycomata, mainly the genera Leiosphaeridia, Pterospermella and Tasmanites, in the darker intervals at Melton Ross (0.7–1% in Bed C; 1.3–6% in Bed E; 6.3% in Bed G). Acanthomorph acritarchs are also present in higher percentages in the darker >1% TOC lithologies (average 3.9% of the total palynological assemblage) than in the <1% TOC samples (average 0.6%). They are mainly represented by Veryhachium spp., apart from Bed G (sample -2), in which Micrhystridium is prominent (11%; Table 2).

Terrigenous palynomorphs are present in higher relative and absolute abundances in the more organic-rich (>1% TOC) samples (average T/M ratio of 0.103 and concentration of 12 555 cpg) than in adjacent lithologies with lower (<1%) TOC values (Table 2; average T/M ratio, 0.023; average concentration, 524 cpg; Supplementary Table E). Terrigenous palynomorphs are mainly composed of gymnosperm taxa (4.7–13.7% of total palynological assemblages...
from >1% TOC samples), predominantly bisaccate pollen, including Ruguhibesiculites rugosus (Fig. 5.17) and Alisporites microsaccus (Fig. 5.18), along with subordinate Classopolis spp. (Fig. 5.19) and Inapertiropollenites hiatus. Peridiphyte spores, including Deltoidospora spp. and Gleichenioidites spp. are persistent and rare to common but do not exceed 3% of the total palynological assemblages. Normapolles angiosperm pollen, Atlantopollis microreticulatus (Fig. 5.20) and Complexiopollis spp. (Fig. 5.21) occur in most productive samples but are rare and compose less than 1% of the palynological assemblages. With the probable exception of an occurrence of the dinoflagellate cyst Sindiridinium borealis in sample -11 (reported Albian–Early Cenomanian stratigraphic range; Nøhr-Hansen et al. 2018), palynomorphs reworked from older formations were not recorded in the Melton Ross section.

**Stable isotopes**

The sample (−23) from the top of the Ferriby Chalk Formation yielded a δ13C value of 2.49‰. In the lower succession, there are three ‘peaks’ in the δ13C data above 3‰ (samples -22 to -21, -19, and -17 to -15), separated by two ‘troughs’ below 3‰ (samples -20, -18 to -18.5). Maximum values for the section are from the Central Limestone (Bed A, sample -14, 4.25‰) and Bed B (sample -13, 4.21‰; sample -12, 4.17‰). Values show an overall decline through the Black Band, from 3.91‰ (sample -11.5) to 3.52‰ (sample -6). Above a trough in Bed F (3.34% in sample -4), there is a subsequent peak in Bed G (3.77‰ in sample -2). The highest sample analysed (-1) from Bed H yielded a value of 3.14‰ (Supplementary Table G).

The δ18O data fluctuate in the −3.58 to −5.12‰ range in the lower succession. Minimum negative values for the section occur in the Central Limestone (−3.03‰ in sample -14; Fig. 4) and overlying basal Bed B (−3.25‰ in sample -13). Upper Bed B and Beds C to H all yielded δ18O values more negative than −4.5‰, with three notable peaks in the samples with >1% TOC (lower Bed C, lower Bed E and Bed G; −5.35‰ to −6.46‰).

**Stratigraphy**

**Micropalaeontology**

A study of foraminifera at Melton Ross has not yet been undertaken. The Cenomanian planktonic marker taxon Rotaliopora cushmani (Morrow) has been recorded from the Ferriby Chalk Formation in eastern England, at South Ferriby and Elsham, but has to date not been recorded from above the erosion surface at the base of the Welton Chalk Formation (Hart et al. 1993). In most southern England sections, it ranges up into the lower part of the Plenus Marls, e.g. Bed 3 at Dover (Jarvis et al. 1988a) and Bed 4 at Eastbourne (Jarvis et al. 2006). At Misburg, R. cushmani occurs in Cenomanian strata and has its last occurrence (LO) in the pre-Plenus ‘black shale’ succession (Hilbrecht 1986). Future inspection of our samples from Melton Ross will test for its presence in localized Beds I–VII.

The regionally correlative Beds B–H in eastern England are dominated by a Hedbergella/Whiteinella assemblage, with an increase in small buliminids and simple agglutinated foraminifera in the dark coloured mudstones (Hart and Bigg 1981; Hart et al. 1993; Dodsworth 1996, fig. 6). The assemblage in the mudstones adjacent to the Black Band is comparable to those of the upper part of the Plenus Marls in southern England (Beds 4–8). This led Hart and Bigg (1981) to suggest that the Black Band may be the lateral equivalent of Bed 6, a relatively clay-rich unit. On grounds of event stratigraphy, i.e. the most argillaceous levels within the CTB successions of the Northern and Southern Provinces, Jeans et al. (1991) also suggested a correlation of eastern England Black Band, Beds C and E, with southern England Beds 4 and 6, respectively. However, a correlation of the Black Band with a level in the Plenus Marls is not supported by the palynological and stable isotopes data presented in this study (see below).

Hart and Bigg (1981) found the first occurrence (FO) of Helvetoglobotruncana praecheveltica (Trujillo) in Bed H at Elsham. This taxon is an early morphotype of Helvetoglobotruncana helvetica (Bolli), the diagnostic zonal marker for the Early to Middle Turonian in Tethyan areas. The FO of H. helvetica was tentatively recorded from a marl seam at South Ferriby (Hart and Leary 1989) that probably correlates with the marl seam above the Turnus Bed at Melton Ross (Fig. 4). However, its FO is based on the (highly subjective) evolutionary boundary between praecheveltica and helvetica (Hart and Leary 1989). The rarity of H. helvetica at high latitudes such as Lincolnshire also renders its FO impracticable as a confident time-diagnostic event in the Northern Province (Hart 2019). In NW Germany, Hilbrecht (1986) reported the FO of H. helvetica within the Fish Shale. At Dover, southern England, its FO is difficult to locate precisely in the nodular chalk lithology of the Ballard Cliff Member (above the Plenus Marls), as this is difficult to process for calcareous microfossils (Hart and Leary 1989). At Eastbourne, the FO of H. helvetica is picked between Mead Marls 3 and 4 of the Ballard Cliff Member (Jarvis et al. 2006). In Figure 4, the planktonic foraminiferal zonation is extrapolated to Melton Ross from Elsham and South Ferriby.

**Palynology**

Several dinoflagellate cysts have widespread range bases and tops in the Cenomanian and Turonian stages, and have been used to zone their substages (Clarke and Verdier 1967; Burgess 1971; Foucher 1981; Williams 1977; Williams et al. 2004). Regional zonations have recently been revised for Central and Northern Europe (Olde et al. 2015a) and adapted for the Western Interior of the USA (Dodsworth and Eldrett 2019).

The LO of consistent and common Litospaeridiidium siphonophorum in the lower succession at Melton Ross (sample -20) correlates with its LO in the pre-Plenus sequence of NW Germany, at Wunstorff outcrop (Marshall and Batten 1988) and core (van Helmond et al. 2015) and Misburg outcrop (Marshall and Batten 1988). In southern England, the LO of common L. siphonophorum occurs in Plenus Marls Bed 6 at Lulworth and Eastbourne (Fig. 1; Dodsworth 2000; Pearce et al. 2009) while isolated specimens have been recorded from Bed 7 and Bed 8 at Lulworth (Dodsworth 2000). Isolated specimens have also been sporadically recorded from Turonian and Coniacian deposits.
at some localities in southern England, France, NW Germany (Clarke and Verdier 1967; Foucher 1982, 1983; Marshall and Batten 1988) and Pueblo, Colorado (Dodsworth 2000). These isolated Turonian and younger occurrences are probably reworked specimens. Previous reports of the LO of consistent L. siphoniphorum in Lower Turonian deposits (Williams and Bujak 1985; Costa and Davey 1992) probably derive from former assignment of the Plenus Marls in southern England to this substage (e.g. Jefferies 1962, 1963); the Plenus Marls have subsequently been confidently reassigned to a Late Cenomanian age (e.g. Jarvis et al. 1988a; Gale et al. 1993). A Late Cenomanian LO of consistent L. siphoniphorum has been reported worldwide, including other locations in southern England (Davey 1969; Hart et al. 1987; Jarvis et al. 1988b), the Witch Ground Graben, Central North Sea (Harker et al. 1987), France (Foucher 1979, 1980; Courtilat et al. 1991), northern Spain (Mao and Lamolda 1999), Crimea (Dodsworth 2004a), Poland (Dodsworth 2004b), eastern USA (Aurisano 1989), the Western Interior of the USA (Courtilat 1993; Dodsworth 2000, 2016; Eldrett et al. 2015a; Dodsworth and Eldrett 2019), Ocean Drilling Project holes including Demerara Rise (Fig. 1; Leg 207, Site 1260) and Kerguelen Plateau (Leg 183, 2019), Ocean Drilling Project holes including other locations in southern England (Davey 1969; Hart et al. 1987; Jarvis et al. 1988b), the Witch Ground Graben, Central North Sea (Harker et al. 1987), France (Foucher 1979, 1980; Courtilat et al. 1991), northern Spain (Mao and Lamolda 1999), Crimea (Dodsworth 2004a), Poland (Dodsworth 2004b), eastern USA (Aurisano 1989), the Western Interior of the USA (Courtilat 1993; Dodsworth 2000, 2016; Eldrett et al. 2015a; Dodsworth and Eldrett 2019), Ocean Drilling Project holes including Demerara Rise (Fig. 1; Leg 207, Site 1260) and Kerguelen Plateau (Leg 183, Site 1138; Eldrett et al. 2017), Australia (Morgan 1980; McMinn 1988) and New Zealand (Hasegawa et al. 2013; Schiøler and Crampton 2014).

At Melton Ross, the LO of Pterodinium crassimuratum (Fig. 5.7) occurs in sample -21, close to the LO of consistent/common L. siphoniphorum in sample -20. Davey and Williams (1966) and Clarke and Verdier (1967) recorded it from beds of Middle and Late Cenomanian age in southern England, its LO coinciding with that of L. siphoniphorum within the Plenus Marls on the Isle of Wight and at a slightly higher level at Eastbourne, in Plenus Marls Bed 8 (Pearce et al. 2009). At Pueblo, Colorado, the LO of P. crassimuratum also occurs in the Upper Cenomanian (Dodsworth and Eldrett 2019).

The FO of abundant Cyclonephelium compactum–membraniphorum in lower Bed B (sample -13), immediately above the Central Limestone at Melton Ross, correlates with the base of its abundant occurrence immediately above the Plenus Bed in NW Germany, at Wunstorf outcrop (Marshall and Batten 1988) and core (van Helmond et al. 2015), and Misburg outcrop (Marshall and Batten 1988). The event is less clearly defined in southern England, but can be tentatively picked near the top of the Plenus Marls, in Bed 7 at Eastbourne (Pearce et al. 2009) and Bed 8 at Lulworth (Dodsworth 2000). Cyclonephelium membraniphorum is rare or absent in coeval Tethyan sections farther south, in Spain (Lamolda and Mao 1999; Peyrot 2011; Peyrot et al. 2011, 2012) and Morocco (Prauss 2012a, b).

At Melton Ross, an influx of Oligospheriaidium totum from Bed B to lower Bed C can be correlated with an influx of O. totum from the same levels at Caistor (samples CL-11 to CL-9; 1–3%) and from upper Bed B at Market Weighton (samples MW-20 to MW-18; 3–6%; see Supplementary Table 1 for sample positions). Duane (in Hart et al. 1993) recorded its highest relative abundance (3.4%) in the lower part of the Black Band (their sample SFE-18) at South Ferriby (the underlying Bed B samples SFE-19 and SFE-20 at South Ferriby were palynologically barren). In NW Germany, Marshall and Batten (1988) only recorded O. totum from above the Plenus Bed at Wunstorf (their sample 38) and at Misburg (their samples 27, 22 and 20). At Misburg, an influx of O. totum was indicated, with its base in sample 22, collected from a bed that contains the only occurrence of the latest Cenomanian zonal ammonite Neocardioceras juddii in the section. An acme of O. totum occurs in overlying sample 20. In central Poland, O. totum occurs in sample Pul-17 of the Pulawy borehole (Fig. 1), within the δ13C excursion that spans the CTB (Peryt and Wyrwicka 1993), one sample above the FO of abundant C. compactum–membraniphorum (sample Pul-16). This, in turn, is stratigraphically higher than the LO of consistent/common L. siphoniphorum (sample Pul-15). Thus, the three bio-events also occur in the same relative order over a condensed (<1 m thick) interval in that section (Dodsworth 2004b).

The LO of Adnatosphaeridium tutulosum (Fig. 5.4) occurs in Bed F at Melton Ross (sample -5). It was recorded from Bed C at Flixton (Dodsworth 1996). At other European and North America locations, the LO of A. tutulosum also occurs within the uppermost Cenomanian, above the LO of consistent/common L. siphoniphorum and the FO of abundant C. compactum–membraniphorum: in NW Germany, above the Plenus Bed at Wunstorf and Misburg outcrops (Marshall and Batten 1988); in southern England, at Hooken Cliffs, Devon (Jarvis et al. 1988b), Eastbourne (Plenus Marls Bed 8; Pearce et al. 2009) and 0.5 m above the Plenus Marls at Lulworth (Dodsworth 2000); and in southern France (cf. Courtilat et al. 1991; Jarvis et al. 2011) and the Western Interior of the USA (Dodsworth 2000; Harris and Tocher 2003; Dodsworth and Eldrett 2019). The LO of A. tutulosum is at the same level as the LO of L. siphoniphorum in northern France (Foucher 1983) and Crimea (Dodsworth 2004a).

The LO of Carpodinium obliquicosatum (Fig. 5.3) occurs in Bed C (sample -11.5) at Melton Ross. It was recorded from Bed E at South Feribby and Flixton (Dodsworth 1996). In NW Germany, Marshall and Batten (1988) reported it above the Plenus Bed in the Wunstorf and Misburg outcrops. In southern England, it occurs 0.5 m above the Plenus Marls at Lulworth (Dodsworth 2000). Although rare and sporadic in occurrence, its range top in the Western Interior is reported to coincide approximately with that of A. tutulosum at Pueblo, Colorado (Dodsworth 2000).

Microdinium setosum (Fig. 6.6) occurs fairly consistently in the Black Band at Melton Ross, with its LO in Bed E (sample -6). In southern England, it occurs throughout the Cenomanian (Clarke and Verdier 1967) and is sporadic in the Plenus Marls with an LO in Bed 2 at Eastbourne (Jarvis et al. 2011) and Bed 7 at Lulworth (Dodsworth 2000). However, it occurs sporadically as high as Middle Turonian in the North Sea Basin (Costa and Davey 1992).

At Melton Ross, Adnatosphaeridium? chonetum (Fig. 5.8) was recorded from local Bed II (sample -21), with an isolated, questionable specimen also present in Bed E (sample -6). It has previously been recorded from Bed C at Flixton (Dodsworth 1996). The occurrence of A.? chonetum in Cenomanian deposits was noted by Davey (1969) in Northern Europe and by Cookson and Eisenack (1962) and Backhouse (2006) in Australia. In the Western Interior of the USA, it is common in the Upper Cenomanian Substage, with
an uppermost Cenomanian LO (Harris and Tocher 2003; Dodsworth and Eldrett 2019). In the Shetland Group, offshore Norway, it reappears in Coniacian and Lower Santonian deposits (PD, personal observation).

The FOs of Canningia glomerata, Heterosphaeridium difficile, Florentinia buspina, F.? torulosa and Senoniasphaera turonica are intra-Lower Turonian biostratigraphic marker events in Europe (Davey and Verdier 1976; Foucher 1980, 1981; Tocher and Jarvis 1987; Jarvis et al. 1985a; Costa and Davey 1992; FitzPatrick 1995; Pearce et al. 2003, 2009, 2011). These taxa have not been recorded at Melton Ross or in any of the other eastern England localities/samples indicated in Supplementary Table I. Thus, Beds C–G are probably stratigraphically lower than their range bases.

In terms of dinoflagellate cyst zonation, the palynologically productive samples from the lower succession at Melton Ross are assigned to the Litosphaeridium siphoniphorum Interval Zone of Olde et al. (2015) and Dodsworth and Eldrett (2019). The productive part of the upper succession (Bed B to Bed G) is assigned to the Cyclonephelium membraniphorum Zone of these authors, with Bed B to lowermost Bed F belonging to the Adnatosphaeridium tutulosum Subzone of Dodsworth and Eldrett (2019; Fig. 4).

At Wunstorf, Prauss (2006) reported acanthomorph acritarchs. Higher relative abundances of Veryhachium (c. 1–2%) occur in darker lithologies from the pre-Plenus and lower post-Plenus succession, relative to lighter coloured inter-beds. There is a large influx of Micrhystridium in the upper part of the Fish Shale (c. 15–70%). Following Wall (1965) and Downie et al. (1971), Prauss (2006) attributed the Veryhachium occurrences to a distal offshore, hydrodynamically quiet environment, and the prominent acritarch peak dominated by Micrhystridium within the upper part of the Fish Shale to the influence of a relatively near-shore turbulent water environment. An up-section change from common Veryhachium (Beds C to E) to abundant Micrhystridium (Bed G) at Melton Ross may correlate with that reported from Wunstorf.

The consistent presence of the bisaccate pollen Rugubivesiculites rugosus in the upper succession at Melton Ross suggests that it is not a local, but a regional phenomenon. This pollen species has been previously reported from the Lower to Middle Turonian interval in the foreland areas of the North Sea basin (Fröhlich 1980; Fröhlich and Ohrner 1985; Fröhlich and Schmitz 1988; Fröhlich and Ohrner 1989; Fröhlich and Schmitz 1989; Wehrl and Neumüller 1990; Neumüller 1995; Neumüller et al. 2003; Neumüller and Back 2004; Wehrl et al. 2006).

Fig. 10. Lithological succession in the Bridge Creek Limestone Member of the Greenhorn Formation, on the north side of the Pueblo Reservoir State Recreation area, west of Pueblo, Colorado, USA. This section contains the Global boundary Stratotype Section and Point (GSSP) for the base of the Turonian Stage (38° 16′ 56″ N, 104° 43′ 39″ W) and the proposed GSSP for the base of the Middle Turonian Substage (Bengtson et al. 1996; Kennedy et al. 2000, 2005; Dodsworth and Eldrett 2019). (a) Metres above/depth below the base of Bridge Creek Member; (b)chronostratigraphy; (c) lithostratigraphy; (d) lithology; (e)δ13Corg, data from Bowman and Bralower (2005). Peaks ‘a’, ‘b’ and ‘c’ in the OAE-2 excursion follow Jarvis et al. (2006); (f) summary of proposed regional correlation with sections from NW Europe (this paper); (g) Pueblo ammonite zones after Kennedy and Cobban (1991); Cobban (1993) and Kennedy et al. (2000). The Watinoceras devonense, Pseudaspidoceras flexuosum and Vascoceras birchbyi units are often treated as subzones of a Watinoceras coloradoense Zone or Watinoceras spp. Zone in published literature; (h) Pueblo inoceramid bivalve zonation by Walaszczuk, and Cobban in Kennedy et al. (2000); (i) Pueblo planktonic foraminiferal zones are from Caron et al. (2005) and Keller and Pardo (2004); (j) dinoflagellate cyst zones and subzones are from Dodsworth and Eldrett (2019).
Melton Ross is noteworthy. It has recently been reported throughout the Upper Cenomanian–Middle Turonian section exposed at Pueblo, Colorado (Dodsworth and Eldrett 2019) and in Turonian cores from Texas (e.g. Iona-1; Eldrett et al. 2017). In the Greenland area, Nohr-Hansen (2012, appendix 1), Fensome et al. (2016) and Nohr-Hansen et al. (2016) indicated that Rugulibivesiculites spp./R. rugosus has a consistent LO within the Turonian.

Calcareaous nanofossils
A study of calcareaous nanofossils at Melton Ross has not yet been undertaken, but Bralower (1988) reported their ranges from the CTB interval at other quarries, including the nearby sections at South Ferriby and Elsham (Fig. 2). The LOs of the coccoliths Azopodorhabdus albicans (Black), Helena chiaistia Worsley and Rhagodicus asper Stradner occur in Bed H at South Ferriby. At Elsham, the LO of R. asper also occurs in Bed H but the LOs of H. chiaistia and A. albicans occur, respectively, 20 cm and 15 cm lower, within undifferentiated variegated marls. At Misburg, NW Germany, all three events occur within Lower Turonian deposits above the Fish Shale (cf. Bralower 1988, fig. 21, and Hilbrecht and Hoefs 1986, fig. 2), with the LOs of H. chiaistia and A. albicans in a ‘triple band’ of stratigraphically higher, dark coloured mudstones, and the LO of R. asper c. 5 m above the ‘triple band’. At Dover, southern England, the three events occur stratigraphically lower, within Upper Cenomanian deposits, i.e. the highest 0.55 m of the Plenus Marls that is 4.45 m thick at the sampled location. At Pueblo, Colorado, the three events occur over a c. 4.5 m interval that spans the CTB, in the stratigraphic order of LO A. albicans, LO H. chiaistia and LO R. asper (Fig. 10). Bralower (1988) suggested that slow sedimentation rates, a probable hiatus between Bed H and the Turnus Bed, and subsequent bioturbation may have ‘smeared’ these events at South Ferriby and Elsham, while at Dover they occur in proximity to a lithological change to the hard, condensed limestone of the Ballard Cliff Member that overlies the Plenus Marls.

Stable isotopes
Before discussing the possible stratigraphic significance of the measured changes in the δ13C stable isotope record, diagenetic or local lithological influences have to be excluded. Diagenetic alteration is commonly observed in fine-grained carbonate sedimentary rocks. Deep burial cementation and recrystallization can result in the addition of isotopically depleted calcite to the bulk carbonate pool, shifting the bulk δ18O record towards lower values (e.g. Jarvis et al. 2011, 2015). However, during burial diagenesis, carbon isotope values are less prone to diagenetic alteration than oxygen isotope values as the carbon isotope system is rock-dominated and δ13C is subject to a much smaller temperature-controlled fractionation (Marshall 1992). A cross-plot of δ13C and δ18O allows the elucidation of the depositional and post-depositional controls on stable isotope values, with a strong co-variance between δ13C and δ18O indicating potential diagenetic influence (Fig. 11).

In general, the stable isotope results for the Melton Ross section show poor co-variance (R2 = 0.0076), indicating limited diagenetic overprint. However, the samples from Beds I–VII and Beds A and B that contain low TOC (<0.2%) do show a relatively strong co-variance (R2 = 0.84), indicating potential dissolution and re-precipitation of isotopically light 13C cements through interaction with meteoric pore fluids during burial in this interval. Conversely, the δ13C and δ18O co-variance remains poor in samples with low TOC from Beds C–H. It is uncertain why the lower part of the succession may be preferentially affected by diagenesis. The range of δ13C and δ18O values from Melton Ross appear consistent with bulk stable isotope values from coeval European Cenomanian–Turonian sections that record a 2–5% positive carbon isotope excursion (CIE) associated with OAE-2 (Fig. 11). The δ13C profile from Beds A–H at Melton Ross broadly correlates with that assigned to OAE-2 at South Ferriby (Schlanger et al. 1987; Hart et al. 1991; Clarkson et al. 2018), which also shows highest δ13C values in Beds A and B and an overall decreasing trend in values through Beds C to H, with a relatively minor, terminal peak around Bed G/H. The South Ferriby studies sampled metres of chalk below the Ballard Cliff Member that overlies the Plenus Marls.
in Southern Province in England, our new record occurs in the overlying part of the Plenus Marls (Beds 4 relatively thick limestone unit. Melton Ross, Site 1 (Wood and Mortimore 1995; Clarkson et al. 2018), NW Germany, background OAE-2 occur within the Plenus Bed. The new record of peak δ13C values in the Central Limestone at Melton Ross supports its correlation with the Plenus Bed, as initially predicted by Wood and Mortimore (1995). The magnitude and top of the δ13C excursion vary across NW Germany (Hilbrecht et al. 1992), but it shows an overall decreasing trend through the Fish Shale at Misburg and Wunstorf, possibly correlating with that seen through Beds C to H at Melton Ross and South Ferriby.

In the Southern Province of England, a ‘build-up phase’ of overall increasing δ13C values through Beds 1–2 of the Plenus Marl precedes a lower peak (‘a’) around Bed 3, a relatively thick limestone unit. Praeactinocamax planus occurs in the overlying part of the Plenus Marl (Beds 4–8), particularly in Bed 4 (Jefferies 1962, 1963). Above a trough in δ13C values through Beds 4–8, a second and maximum peak (‘b’) occurs close to the boundary between the Plenus Marl and overlying Ballard Cliff Member. Above peak ‘b’, δ13C values remain high but there is an overall decline in values across the CTB, through the Ballard Cliff Member, including some prominent marl seams (Mead Marl 1–6), below a final OAE-2 peak ‘c’ (Jarvis et al. 2006).

In terms of correlation between the Northern Province and Southern Province in England, our new δ13C data from Melton Ross could be interpreted as follows. The three (>3%) peaks of δ13C in Beds 1–7 may be ‘precursor’/build-up events to OAE-2, correlating with southern England Beds 1–2; the base of the >4% interval in Bed A (Central Limestone) equals peak ‘a’ of Jarvis et al. (2006; southern England Bed 3); lower Bed B may correlate with part of Beds 4–8 in southern England, supported by the record of P. planus at Melton Ross, Site 1 (Wood et al. 1997; Figure 4); the top of the >4% interval in upper Bed B equals peak ‘b’ of Jarvis et al. (2006); the trough in Beds C–F and peak in Bed G may correlate with the trough through the Ballard Cliff Member and peak ‘c’ in southern England (Fig. 10). A comparable correlation has been made between the δ13C profile from Beds A–H at South Ferriby and that of the Southern Province (Wood and Mortimore 1995; Clarkson et al. 2018).

Notwithstanding concerns about diagenetic alteration of the eastern England Chalk Group oxygen isotope signal discussed above, the δ18O curve from Melton Ross (Fig. 4) is comparable to those from other parts of Europe (cf. Jarvis et al. 2011, fig. 8). The minimum values in the Central Limestone and overlying lower Bed B may correlate with the PCE. The higher values from Beds C–H could be considered consistent with subsequent warmning around the CTB. The peaks in δ18O values from the most organic-rich layers could reflect temperature maxima and/or increased run-off during their deposition (see below).

Discussion

Organic geochemistry

In an overview of OAE-2 in North European shelf settings, Jenkyns (1985) reported TOC values mainly within the range of 1–3% and HI from 150–300 mg HC/g TOC (mixed marine and terrigenous sources) in Cenomanian–Turonian organic-rich deposits, including eastern England (South Ferriby, 0.79–3.15% TOC) and NW Germany (1.2–2.8% TOC; Schlanger et al. 1987). The new HI data from the >1% TOC dark mudstones at Melton Ross plot around the lower end of this range (78–203 mg HC/g TOC; Fig. 9), consistent with a relatively large terrigenous component. Jenkyns (1985) noted that there appears to be a general trend of an increase in preservation of organic carbon as the depositional environment deepened from stable, relatively shallow shelf to rifted graben and deeper continental margin. Higher TOC values have since been reported from the Black Band farther north in eastern England, from Speeton (13% TOC, Farrimond et al. 1990) and Flixton (4.5–10.2% TOC, Jeans et al. 1991) at the edge of the Cleveland Basin (Fig. 2). Herbin et al. (1986) reported 30% TOC from the Black Band Bed in the Central Graben of the North Sea, including large quantities of Type III organic matter, consistent with a substantial terrigenous component. In palynological assemblages from Melton Ross and other onshore Black Band localities, marine organic-walled phytoplankton dominate over terrigenous pollen and spores (Supplementary Table I), although at Flixton, the latter compose approximately one half of the assemblages from Bed C.

The Cyclonephelium compactum–membraniphorum issue

The southward incursion of abundant Cyclonephelium compactum–membraniphorum into Central Europe has been attributed to a stressed marine environment associated with oceanic anoxia (Marshall and Batten 1988; Courtinat et al. 1991; Hart et al. 1991). However, its initial migration has more recently been correlated with the PCE, including sites outside Europe. The latter include a proto-Atlantic coastal setting in New Jersey (Bass River; van Helmond et al. 2014; Fig. 1) and the Western Interior Seaway of the USA (e.g. Portland-1 core, central Colorado; Iona-1 core, SW Texas; Eldrett et al. 2014, 2017; van Helmond et al. 2016; Fig. 1). At Melton Ross, its basal abundant occurrence in sample -13 is at the same level as records of Praeactinocamax planus (Fig. 4) and it is consistently abundant up to the top of the sampled section (sample -1). While the initial southward migration may have been linked to the PCE, its subsequent abundance in latest Cenomanian and Early Turonian deposits occurred during a time of extreme warmth. Its post-PCE prosperity during the OAE-2 interval may be related to the relative tolerance of areoligeracean dinoflagellate cysts in general to stressed marine environments, where their abundant occurrence is often associated with nearshore environments/falls in relative sea-level (Brinkhuis and Zachariasse 1988; Harker et al. 1990, p. 202–204; Li and Habib 1996; Olde et al. 2015b). In this case, the environmental stress is possibly associated with oxygen-deficient conditions in the water column and photic zone (Marshall and Batten 1988) and/or increased runoff (van Helmond et al. 2015).

Samples with prominent Dinoflagellate? type D of Ioannides (1986)

Dinoflagellate? type D occurs consistently throughout the sampled section at Melton Ross, and in dominant relative
abundance (48–100% of palynological assemblages) in the poorly preserved material from samples -20 to -17, from the local Beds III?–VII. However, inspection of the absolute abundance data (Table 3) indicates that its occurrences in these beds are in the single units to tens per gramme range, less than its hundreds per gramme concentration estimates from beds with richer recovery. It is suggested here that its dominance in Beds III?–VII is due to its preferential preservation, probably related to a relatively thick and resilient wall structure. This interpretation is supported by the data of Batten (in Wood et al. 1997), who reported diversified dinoflagellate cyst assemblages with *L. siphoniphorum* up to local Bed VI at Melton Ross, Site 2, i.e. in correlative strata a few tens of metres away (Fig. 3).

Weathering of exposed quarry sections probably destroys palynomorphs and other organic matter over a short time, of months to a few years, with only central levels within the Black Band being unaffected at some other sampled localities, e.g. South Ferriby and Caistor. Weathered samples contain lower concentrations of poorly preserved palynomorphs, mainly those with relatively thick walls: Dinoflagellate? type D, *Kalyptea* spp., prasinophyte phycomata and indeterminate peridinioid endocysts. Samples worst affected are marked with an asterisk in Supplementary Table I.

**Terrigenous input**

In eastern England, the higher T/M ratio in relatively organic-rich (>1% TOC) lithologies documented here from Melton Ross (Table 2) has previously been noted in Black Band samples from Flixton (Dodsworth 1996, fig. 9; T/M ratio 0.27–0.54 in 4.5–10.2% TOC samples, and 0–0.04 in 0.1–0.4% TOC samples). Analyses from other Yorkshire sections (Fig. 2) appear to confirm this pattern: East Knapton (T/M ratio average 0.11 in dark lithologies and 0.02 in light lithologies), Market Weighton (T/M ratio average 0.08 in dark lithologies and 0.01 in light lithologies) and, to a lesser extent, Bishop Wilton (T/M ratio average 0.06 in dark lithologies and 0.04 in light lithologies; Supplementary Table I). This trend cannot be discerned at other Lincolnshire localities (South Ferriby, Bigby and Caistor), where Bed B and Bed F/G samples are heavily weathered or barren and Bed D is relatively argillaceous and tentatively picked on slightly lighter-coloured marls within the Black Band.

At Wunstorf, Prauss (2006) calculated an average T/M ratio of c. 0.2 for the CTB succession. Prauss (2006) and van Helmond et al. (2015) also reported elevated relative and absolute abundances of spores and pollen in dark coloured mudstones compared with interbedded organic-poor lithologies, particularly within the Fish Shale (T/M ratio c. 0.2–0.46, in c. 1–3% TOC rocks). Van Helmond et al. (2015) interpreted this pattern as indicative of an intensified hydrological cycle during OAE-2, i.e. increased evaporation, precipitation and run-off during ‘black shale’ deposition, associated with global warming after the PCE (see also Heimhofer et al. 2018). Reworking of mud to the basin (e.g. Jenkyns 1980) and/or the formation of swamps due to drowning of land masses (cf. Ioannides et al. 1976) during phases of CTB transgression could also have contributed to the increase in spores and pollen (Dodsworth 2000).

In eastern England, gymnosperm bisaccate pollen are invariably the dominant terrigenous palynomorph group in the Black Band and Bed G (Dodsworth 1996; this paper). The productive samples from the lower succession at Melton Ross (samples -22 to -20; Table 2) contain few terrigenous palynomorphs (T/M ratio 0.01 or less), which are also mainly bisaccate pollen. At Wunstorf, however, saccate pollen constitute less than one half of the terrigenous palynomorphs in most samples from the CTB succession, with trilete spores and non-saccate pollen often composing most of the assemblage (Prauss 2006; van Helmond et al. 2015). This contrast between the regions probably reflects greater proximity to sources of detrital input in the Lower Saxony Basin, as reflected by a much thicker succession there. Prauss (1993) noted that trilete spores show higher relative abundances than saccate pollen in marine deposits with greater fluvial input because the latter are more readily transported by wind (e.g. Mudie and McCarthy 1994). Different hydrodynamic properties, i.e. the relative buoyancy of saccate pollen, is probably also a factor in their differential sorting with distance from fluvial-deltaic sources (cf. Muller 1959). In the Cenomanian deposits of Texas, spores form much larger proportions of terrigenous palynomorph assemblages in the proximal East Texas Basin than in the relatively distal Maverick Basin (Dodsworth 2016).

**Pollen provincialism**

Herngreen and Chlonova (1981); Herngreen et al. (1996) and Traverse (2007) described Late Cretaceous palynofloral provinces. Costa and Davey (1992) reported the southerly limit of triprojectate pollen (the *Aquitapollenites* Province) to correlate approximately with the northerly siliciclastice Shetland Group – southerly calcareous Chalk Group transition in Northern Europe, around modern latitude 59° N. The rare but consistent occurrences of Normapolles pollen from the CTB succession at Melton Ross complement previous records of their presence in the Black Band at Flixton, South Ferriby and North Sea well 47/10-1 (Fig. 2; Dodsworth 1996). These may be the most northerly published Cenomanian–Turonian records in Europe (cf. Peyrot et al. 2008, fig. 4) and indicate assignment of the eastern England Chalk Group to the Normapolles Province. Normapolles pollen are a more common component of palynological assemblages from coeval deposits farther south in Europe (e.g. Bulgaria, Pavlishina and Minev 1998; Romania, Ion et al. 2004; SE France, Heimhofer et al. 2018; Spain, Peyrot et al. 2008, 2011). In SE France, Heimhofer et al. (2018) suggested that the PCE may have fostered a first spread of Normapolles-type angiosperms. Climatic conditions would have been cooler and less humid during the PCE, possibly resulting in an open, savannah-type vegetation community with increased abundances of Normapolles-producing angiosperms.

**Phytoplankton productivity and preservation**

The P-cyst *Palacoehystrichophora infusoriaoides* and the G-cyst *Spiniferites ramosus* constitute the main components of European dinoflagellate cyst assemblages from Late Cretaceous offshore/deeper water environments and upwelling zones (Pearce et al. 2003; Olde et al. 2015b). The P-cyst
genera *Palaeohystrichophora* and *Subtilisphaera* have been suggested as heterotrophic taxa. High numbers may reflect increased nutrient availability from upwelling in the Atlantic Ocean off the NW African (*Prauss 2012b*) and European margins, and subsequent transport of nutrients via a proto-Gulf Stream to the Anglo-Paris Basin (*Prince et al. 1999, 2008; Pearce et al. 2009*). *Subtilisphaera pontis-mariae* and *P. infusorioides* mainly compose the high P/G ratio in Melton Ross local Beds I–III (average 0.61). A source for enhanced nutrient supply during this interval is uncertain. The low T/M ratio is inconsistent with increased run-off. Upwelling as a nutrient source requires a thermal difference and seafloor topographical variation but there is no evidence for these in eastern England.

Phytoplankton productivity during OAE-2 ‘black shale’ deposition has previously been discussed from Wunstorf. *Linnert et al.* (2010) studied assemblages of calcareous nannofossils, showing a shift from a generally oligotrophic ecosystem during deposition of lighter coloured beds to more mesotrophic or even eutrophic conditions during deposition of dark coloured mudstones. *Van Helmond et al.* (2014, 2015) suggested that an acceleration of the hydrological cycle during the warmer intervals of OAE-2 may have played a key role in supplying nutrients offshore and enhancing stratification, thus contributing to the development of ocean anoxia. Increased precipitation, run-off (reflected by a higher T/M ratio) and associated nutrients may have contributed to productivity of organic-walled phytoplankton during deposition of the dark coloured layers in eastern England.

Within the Black Band and Bed G at Melton Ross, common *Eurydinium saxoniense* and *Bosedinia* cf. sp. 1 of *Prauss (2012b)* and rare *Bosedinia laevigata*, make a greater contribution to the P/G ratio in the >1% TOC samples (Table 4). *Eurydinium saxoniense* was initially described in higher relative abundances in dark coloured mudstone layers from Wunstorf and Misburg by *Marshall and Batten* (1988), who associated it with a stressed marine environment with a low level of oxygen extending high up the water column.

High relative abundances of *Bosedinia* spp. in Late Cretaceous marine settings have been interpreted as reflecting reduced salinity surface water and enhanced density stratification (at Tarfaya; *Prauss 2012b*, fig. 5), and/or increased nutrient (nitrate/nitrite) availability from oxygen-deficient waters encroaching into the photic zone (in the Western Interior of the USA and Demerara Rise; *Dodsworth 2016; Eldrett et al. 2017*). In the Black Band and Bed G, higher numbers of *E. saxoniense* and *Bosedinia* spp. occur in the absence of fresh/barkish-water algae and may record some stimulation of dinoflagellate productivity by increased supply of reduced nitrogen chemo-species in photic-zone waters (cf. *Prauss 2007*). This is supported by the relative increase in prasinophyte phycomata in the darker layers at Melton Ross, Flixton (5.7–9.3% in Bed C, 14% in Bed E; *Dodsworth 1996*) and Wunstorf (up to 15%; *Prauss 2006*). Prasinophyte productivity may be mainly related to productivity stimulated by introduction of nitrogen/ammonium-enriched waters of the denitrification zone into the photic zone (*Prauss 2006, 2007*). However, the average P/G ratio at Melton Ross (0.33) in the >1% TOC samples is much lower than that discussed above for *Bosedinia*-dominated organic-rich lithologies from, for example, the Cenomanian of the Western Interior (c. 0.75–0.85), and may therefore, reflect a relatively moderate increase in peridinioid productivity. Likewise, prasinophyte increases are moderate relative to uppermost Cenomanian organic-rich mudstones in Aksudere, Crimea (Fig. 1; up to 7.2% TOC; *Naidin 1993*), where they typically constitute over half the palynomorph assemblage (*Dodsworth 2004a*). Regardless of the relative contribution of organic-walled phytoplankton productivity to OAE-2, the higher proportions of *E. saxoniense, Bosedinia* and prasinophytes in the >1% TOC lithologies at Melton Ross and Flixton are consistent with increased quality of seafloor preservation in a low O2 environment. They may indicate periodic hydrographically restricted and eutrophic conditions (cf. *Eldrett et al. 2017*).

**Palynomorph concentration**

Most palynological studies of Cenomanian and Turonian deposits report relative abundances only. Records of specimen numbers per microscope slide from Dover, Eastbourne and the Isle of Wight, southern England (*Jarvis et al. 1988a; Fitzpatrick 1996*) have given a semi-quantitative indication of recovery. Until the last ten years, there have been few reports of fully quantitative concentration, counts per gramme (cpg) data. *Duane* (in *Paul et al.* 1994), using a volumetric method, reported dinoflagellate cysts in the hundreds (mainly 100–800 cpg) range from Middle Cenomanian chalks and marls in the Dover–Folkestone Warren area, southern England. *Pearce et al.* (2003), using a method involving adding a known number of *Lycopodium* spores to samples (*Stockmarr 1971; Mertens et al. 2009*), reported dinoflagellate cysts in the tens to 600 cpg range from Turonian chalks and marls at Banterwick Barn, southern England.

Much higher concentrations of palynomorphs (mainly dinoflagellate cysts) were reported from Upper Cenomanian deposits in southern England, at Eastbourne and Lulworth. *Pearce et al.* (2009), using the *Lycopodium* method, reported c. 50 000–75 000 cpg from six marl and chalk samples over a 10 m interval below the Plenus Marls at Eastbourne. *Dodsworth* (2000), using this study’s volumetric method, reported nearly 38 000 cpg from a marl sample one metre below the Plenus Marls at Lulworth. *Pearce et al.* (2009) noted that assemblages from below the Plenus Marls at both localities are dominated by the P-cyst *Palaeohystrichophora infusorioiides*. They attributed its high concentrations to upwelling in the Atlantic Ocean off the European and NW African margins, with associated nutrients being transported to the southern England area by dominantly southwesterly surface winds and ocean surface currents. Both *Dodsworth* (2000) and *Pearce et al.* (2009) documented overall decreasing concentrations of palynomorphs through the Plenus Marls (to just over 100 cpg in Bed 8 at Lulworth) and barren or impoverished recovery in the overlying Ballard Cliff Member. *Jarvis et al.* (1988a); *FitzPatrick* (1996) and *Pearce et al.* (2003) also reported a palynologically barren or impoverished interval above the Plenus Marls at other southern England localities, with better palynological recovery from Turonian chalks and marls above the Ballard Cliff Member and Holywell Member.
Melton Ross palynology and geochemistry

Dodsworth (2000) attributed poor palynological recovery from the upper part of the Plenus Marls and the Ballard Cliff Member to upward coarsening through the succession (Hancock 1989; Jeans et al. 1991; Lamolda et al. 1994) and the associated reduced palynomorph preservation potential of lithologies with a relatively high coarse silt and sand grade component. Pearce et al. (2009) acknowledged this but, following Jarvis et al. (1988a); FitzPatrick (1996) and Lamolda and Mao (1999), suggested that a primary decline in phytoplankton productivity across the CTB in the Anglo-Paris Basin may have been an additional cause of the low recovery. The Eastbourne and Lulworth localities record a marked decrease in the relative and absolute abundance of P. infusorioides above Bed 4 of the Plenus Marls, consistent with a decrease in nutrient supply (Pearce et al. 2009). An overall decline in the nannofossil fertility index also occurs between Bed 4 and Ballard Cliff Member (Gale et al. 2000).

Palynomorph concentration data are becoming more frequently available from areas outside southern England. From the CTB succession at Bass River, New Jersey, van Helmond et al. (2014), using the Lycopodium method, reported dinoflagellate cysts in the 2000–30 000 cpg range, and terrigenous spores and pollen in the 400–11 000 cpg range, from silty clay to clayey silt lithologies (0.61–1.51% TOC) that were deposited in an inner neritic setting. From the CTB succession at Bass River, New Jersey, van Helmond et al. (2014), using the Lycopodium method, reported dinoflagellate cysts mainly in the 1000–15 000 cpg range and terrigenous spores and pollen in the 2500–5000 cpg range, from samples with 0.5–3% TOC. Using this study’s volumetric method, comparable values for organic-walled phytoplankton were obtained from the CTB succession at Aksudere, Crimea, 3520–20 400 cpg, from samples with 0.3–7.2% TOC (Dodsworth 2004a). Comparable values have also been obtained from relatively unweathered dark coloured mudstones of the Black Band at South Ferriby and Caistor; average phytoplankton is in the range of 10 000–15 000 cpg with pollen and spores at 200–900 cpg (Supplementary Table 1).

Significantly higher concentrations of palynomorphs have been recorded from the dark coloured mudstones of the Black Band and Bed G elsewhere in Lincolshire, at Melton Ross (average phytoplankton c. 84 000 cpg; pollen and spores c. 10 000 cpg; Table 2) and Bigby (average phytoplankton c. 60 000 cpg; pollen and spores c. 3500 cpg). These values may be more representative than those from South Ferriby and Caistor, given that samples from the latter sections are apparently more affected by quarry wall weathering. However, the relatively expanded sections at Melton Ross and Bigby could have been deposited in local topographic troughs in the seafloor (Gaunt et al. 1992; Wood and Mortimore 1995; Wiese et al. 2009), in which organic matter preservation may have been enhanced (cf. Jenkyns 1985). Farther north in Yorkshire, average phytoplankton concentrations from relatively unweathered dark coloured mudstones of the Black Band and Bed G are c. 40 000–50 000 cpg at Market Weighton, Bishop Wilton and East Knaptown, with pollen and spores at c. 3500–5000 cpg, i.e. values relatively comparable to Melton Ross and Bigby. The highest concentrations of palynomorphs in the region are from sections on the margins of the Cleveland Basin (e.g. Jeans et al. 1991; Mitchell 2000) at Flixton and Speeton, with average phytoplankton concentrations of c. 95 000–100 000 cpg, and pollen and spores c. 10 000 cpg at Speeton and c. 80 000 cpg at Flixton (Supplementary Table 1). These regional data appear to be consistent with the trend of an increase in preservation of organic matter as the depositional environment deepened from relatively shallow shelf to deeper continental margin (Jenkyns 1985). Unlike the samples below the Plenus Marls at Eastbourne and Lulworth, which also contain high concentrations of dinoflagellate cysts (c. 38 000–75 000 cpg; see above), P. infusorioides is a relatively minor component of assemblages from dark coloured mudstones in eastern England (Table 4), suggesting that nutrients originating from upwelling might not have been the cause of high concentrations of dinoflagellate cysts there.

The fact that the Wunstorf and Misburg sections are at least ten times as thick as the correlative eastern England sections, yet show a succession of dinoflagellate cyst and acritarch bio-events comparable to those at Melton Ross, is consistent with a high degree of stratigraphic condensation in eastern England. The greater palynomorph concentrations documented from the dark coloured mudstones in eastern England may partly reflect this, along with increased quality of seafloor preservation (in a low O\textsubscript{2} environment), particularly in potential local topographic troughs in the seafloor and in deeper areas of the continental shelf. A loss of biogenic carbonate sediment during the CTB mass extinction interval, and a reduction of siliciclastic input during maximum flooding associated with the base of the Black Band (eastern England; Wood et al. 1997) and the correlative base of the Fish Shale (NW Germany; Ernst et al. 1984), may also have contributed to stratigraphic condensation and enhanced palynomorph concentration.

Palynomorphs are recovered from most levels within the Chalk Group of the Southern Province (Anglo-Paris Basin; see examples above) and the Transitional Province, e.g. the Trunch borehole (Fig. 2) that has palynological recovery throughout the Upper Cretaceous section apart from a Middle Cenomanian to Middle Turonian barren interval (Pearce 2010, 2018; Oldé et al. 2015a; Pearce et al. 2020). All chalk samples analysed to date from the Cenomanian and Lower Turonian of the Northern Province, below and above the Black Band, and from the Central Limestone at Melton Ross (this study), are palynologically barren. This is also the case for chalk samples from the CTB interval in NW Germany (Marshall and Batten 1988; Prauss 2006; van Helmond et al. 2015). In the North Sea Basin, poor or patchy palynological recovery from the Chalk Group has led to palynology not being routinely used in stratigraphic studies of hydrocarbon wells (with exceptions such as Maastrichtian chalk in the Dan Field, Danish sector; Schieler and Wilson 1993), whereas it is a primary stratigraphic discipline in the coeval northerly, siliciclastic Shetland Group, which has consistent rich palynological recovery (Costa and Davey 1992). In England, a different diagenetic history of Cenomanian and Turonian chalks in the Northern Province, which are hardened relative to those of the Transitional and Southern Provinces (Jeans 1980; Jeans et al. 2014), probably contributed to post-depositional loss of palynomorphs.
**Correlation with Norwegian sector well 35/6-2 S**

Ditch cuttings samples from the Grosso exploration well 35/6-2 S in the Norwegian sector, North Viking Graben (Shetland Group facies; Figs 1 and 12) were analysed for palynology by one of us (PD) in 2009. The well has a relatively thick (37 m) Blodøks Formation gamma log profile that probably correlates with the more expanded outcrop lithological successions discussed in this paper.

A sharp rise in gamma log values marks the base of the Blodøks Formation at 2657 m (log depth). The LO of consistent/common *Litosphaeridium siphoniphorum*, and rare *Pterodinium crassimuratum* at 2655 m, indicate the probable presence of the pre-Plenus succession. A fall in gamma log values from 2650.4 m to 2648.6 m may correlate with the Plenus Bed in NW Germany and Central Limestone/Bed A in eastern England. This is supported by the FO of *Cyclonephelium compactum–membraniphorum*, *Eurydinium saxoniense* and *Alterbidinium daveyi* in the next up-hole sample from 2646 m, in an interval of rising gamma log values from 2648.6 m to 2641.7 m (likely equivalent of eastern England Bed B). The remainder of the Blodøks Formation from 2641.7 m to 2620.0 m contains two gamma log plateaux, separated by a lower gamma interval, probably correlative with the relatively clay- and organic-rich Fish Shale in NW Germany and Beds C to H (eastern England) in eastern England. This interpretation is supported by the LO of *Carpodinium obliquicostatum* from a cuttings sample near the top of the formation at 2616 m. Common *C. compactum–membraniphorum*, *E. saxoniense* and *A. daveyi* persist higher than the log pick for the top of the Blodøks Formation. The highest common occurrence of *C. compactum–membraniphorum* is often used as a proxy for the top of the Lower Turonian (Dodsworth and Eldrett 2019), occurring in this well at 2466 m. *Rugubivesiculites rugosus* occurs sporadically in eight samples in the Turonian interval, from 2256 m, but is more consistent in Cenomanian deposits from 2646 m and below.

**Conclusions**

A sequence of dinoflagellate cyst and acritarch bio-events supports the proposed correlation of the Melton Ross CTB succession with sections in NW Germany. The top of consistent/common *Litosphaeridium siphoniphorum* occurs in the lower succession at Melton Ross and the pre-Plenus beds at Misburg and Wunstorf; the base of abundant *Cyclonephelium compactum–membraniphorum* occurs immediately above the Central Limestone at Melton Ross, in lower Bed B, and immediately above the Plenus Bed at Misburg and Wunstorf; a regional influx of *Oligosphaeridium totum*, with an acme in upper Bed B at Melton Ross, occurs at the same level as a record of the latest Cenomanian zonal ammonite *Neocardioceras juddii* at Misburg; rare specimens of the dinoflagellate cysts *Adnatosphaeridium tutulosum* and *Carpodinium obliquocystatum* have their range tops in or just above the Black Band in eastern England, and within the Fish Shale at Misburg and Wunstorf. An influx of the acanthomorph acritarch *Micrhystridium* spp. in Bed G at Melton Ross may correlate with that recorded from the upper parts of the Fish Shale at Wunstorf.

Correlation between eastern and southern England is less straightforward. Carbon isotope data from Melton Ross and South Ferriby, and the distribution of dinoflagellate cyst bio-events, support the interpretation that the Black Band is stratigraphically higher than the Plenus Marls. The restriction of *Praeactinocamax plenus* (belemnite) to the Plenus Marls and *Neocardioceras juddii* (ammonite) to the overlying Ballard Cliff Member are consistent with this. The correlation points to separate depositional histories and marked differences in siliciclastic input between the Northern and Southern Provinces during post-Plenus Marls times. However, the foraminiferal and calcareous nanofossil data can alternatively be interpreted in terms of the Black Band being equivalent to the upper part of the Plenus Marls. Poor preservation and/or recovery of all three microfossil
groups in the Ballard Cliff Member hampers confident assessment of correlative bio-events around the CTB in southern England.

Palynotological assemblages from the >1% TOC samples at Melton Ross contain higher relative abundances of the P-cysts *Eurydinium saxoniense* and *Bosedinia* spp., along with prasinophyte phycocysts, consistent with periodic hydrographically restricted and eutrophic conditions during deposition of the darker mudstone layers. Exceptionally high concentrations of palynotomorphs suggest enhanced preservation in such a low O2 environment at the sediment–water interface coupled with a high degree of stratigraphic condensation. This is in agreement with the view of Hart and Leary (1989), who suggested that the Black Band was deposited during ‘sluggish’ oceanic conditions that would have been ideal for the concentration and preservation of organic-rich sediments, with or without high surface-water productivity.

**Systematic palaeontology**

Division DINOFLAGELLATA (Bütschli, 1885) Fensome *et al.*, 1993

Class DINOPHYCEAE Pascher, 1914

Order PERIDINIALES Haeckel, 1894

Suborder PERIDININEAE (Autonym)

Family PERIDINIACEAE Ehrenberg, 1831

Genus *Alterbidinium* Lentin and Williams, 1985

Type species: *Alterbidinium* ‘recticorne’  (Vozzhennikova, 1967) Harker *et al.*, 1990

*Alterbidinium daveyi* basionym nov.

**Derivation of name.** In honour of the palynologist Roger J. Davey.

**Designation of holotype.** Davey (1970, plate 1, fig. 3). Location, International Yarbo Borehole no. 17, SE Saskatchewan, Canada (coordinates supplied by Davey 1969, p. 115, fig. 8, area: ‘east of Regina at Lsd. 1, Sec. 24, Twp. 20, Rg. 33, W1st Meridian’). Sample depth 254.5 m (835 ft) below Kelly Bushing, Second White Specks Shale, Colorado Group. The slide is curated at the Natural History Museum, London (slide/specimen reference number V.51979).

**Description.** See Davey (1970, p. 338).

**Discussion.** Davey (1970, pl. 1, figs 3, 4, p. 338) described specimens from Cenomanian deposits in Saskatchewan that he assigned to *Deflandrea* (now *Subtilisphaera*) *pirnaensis*. He pointed out differences with the type material of *S. pirnaensis*, as described by Alberti (1959) from Turonian deposits in Germany, including a smaller size, length 46 (62.7) 87 µm (compared with 80–106 µm) and width, 34 (45.5) 63 µm (compared with 58–64 µm), and presence of an archaeological in many Saskatchewan specimens but absence of one in the type specimens. It is also noted here that the holotype of *S. pirnaensis* (Alberti 1959, pl. 8, fig. 1) has possible pre- and postcircular tabulation that is absent from the Saskatchewan specimens. Stover and Evitt (1978) proposed a new species, *Alterbia daveyi*, based on the specimens illustrated in Davey (1970) but did not designate one of the specimens as a holotype. The name was therefore not validly published (Fensome *et al.* 1999). Subsequently, the genus *Alterbia* has been considered an illegitimate name and its species have been transferred to the genus *Alterbidinium* (Fensome *et al.* 2019). *Alterbidinium* ‘daveyi’ has become widely accepted as a separate but informal species. The designation herein of one of the specimens illustrated in Davey (1970) as a holotype, gives it formal status.

*Alterbidinium daveyi* and *Eurydinium saxoniense* have comparable morphology, and both possess a single intercalary (2a) plate archaeopyle that is steno- to iso-deltaform. *Alterbidinium daveyi* may be slightly larger (cf. length, 52–66 µm, and width, 37–54 µm for *E. saxoniense*; Marshall and Batten 1988) and appears to have a more strongly developed apical horn and cingulum than *E. saxoniense*, but intergradations may occur. The cingulum in *A. daveyi* is marked by low ridges that sometimes possess pustules distally and is occasionally crossed by low ridges delimiting plate boundaries (Davey 1970, p. 338); the latter features are not reported in *E. saxoniense*. The holotype of *A. daveyi* appears to show some anterior dorsal intercalary tabulation (2a and 3a plate boundaries). Dorsal tabulation is usually restricted to the archaeopyle in *E. saxoniense* (Marshall and Batten 1988), though our illustrated specimen (Fig. 5.6) may possess sutures around both 2a and 3a intercalary plates.

The eastern England specimens inspected in this paper and previous studies have been assigned to *E. saxoniense*, although the additional presence of *A. daveyi* cannot be ruled out. The dinoflagellate cyst distribution charts of Marshall and Batten (1988) indicate an absence of *A. daveyi* and the common to abundant occurrence of *E. saxoniense* from the (post-Plenus) CTB interval in NW Germany. Conversely, in coeval deposits from North America, *A. daveyi* (sometimes recorded as *S. pirnaensis*) is prominent (Bloch *et al.* 1999, fig. 23; Dodsworth 2000, 2016; Harris and Tocher 2003; Dodsworth and Eldrett 2019). In the Shetland Group of Northern Europe, both *A. daveyi* and *E. saxoniense* are common, e.g. in Norwegian well 35/6-2 S, where confident differentiation between the species can be problematic for some specimens.


Type species: *Bosedinia granulata* (He and Qian, 1979) He, 1984

*Bosedinia laevigata* (Jiabo, 1978, ex He and Qian, 1979) He, 1984

Figures 7.11, 7.12, 7.13, 7.17, 7.21

**Description.** Small to intermediate-sized, sub-spheroidal shaped, smooth and thin-walled proximate autocysts with an ornament of c. 10–50 sparsely distributed, non-tubar, solid verrucae, each of which is c. 1–3 µm in diameter. Tabulation is indicated by an archaeopyle only. A flap-like operculum, probably involving fused apical and anterior intercalary plates, is often attached. Inclusions (omphali) are present in some specimens.

**Dimensions.** Diameter (central body, 12 measured specimens): 35.7 (43.5) 56.0 µm.
Discussion. The type specimens of *Bosedinia laevigata* (He et al. 2009, pl. 17, figs 11–15), from lacustrine Neogene deposits in China, are of comparable size (35–60 µm diameter) to the specimens measured from Melton Ross, and contain a comparable number, size and distribution of small verrucae ornament. *Bosedinia laevigata* has been recorded from marine Upper Cretaceous deposits at Tarfaya, Morocco (Prauss 2012c, 2015) and the Abu Gharadig Basin in Egypt (Ahmed et al. 2020; Figure 1), but not previously in Europe.

**Stratigraphic range/occurrence.** At Melton Ross, Bed C (sample -11.5) to Bed E (sample -8).

*Bosedinia* cf. sp. 1 of Prauss (2012b)

Figures 7.5, 7.9, 7.10

Discussion. The abundant occurrence of mainly enclosed spheres, containing omphali, in organic-rich Cenomanian–Coniacian deposits at Tarfaya, has been discussed by Prauss (2012a, b, c, 2015). Prauss (2012b) provided evidence for their assignment to the genus *Bosedinia* from occasional specimens that have an archaeopyle/operculum. Eldrett et al. (2017) documented the abundant occurrence of comparable specimens at other proto-Atlantic sites (Demerara Rise, DSDP Sites 1260 and 1261; Fig. 1) and within the Western Interior Seaway of the USA (SW Texas and central Colorado). Specimens from Melton Ross constitute the first record of their presence in Europe. However, there are issues with consistent identification. While the specimen illustrated in Figure 7.9 shows a rare example of discernible anterior horns, the fully enclosed specimen in Figure 7.10 could alternatively be interpreted as an endocyst of *Subtilisphaera pontis-mariae*. (cf. the specimens in Figs 5.14, 5.15). The two overlapping specimens in Figure 7.5, resemble *Eyrea nebulosa*, which also possesses omphali (cf. plate 11, in Cookson and Eisenack 1971), but may be distinguished by the presence of a kalyptra, when present. In the Melton Ross data, specimens encountered are mainly single enclosed spheres with omphali and have been assigned to *Bosedinia* cf. sp. 1/peridinioid endocysts’ in the supplementary distribution chart.

**Stratigraphic range/occurrence.** At Melton Ross, local Bed II (sample -21) to Bed H (sample -1).

Genus *Ginginodinium* Cookson and Eisenack, 1960, emend. Lentin and Williams, 1976

Type species: *Ginginodinium spinulosum* Cookson and Eisenack, 1960

*Ginginodinium?* sp. A of Prauss 2006, 2012a

Figures 6.3, 6.7

Discussion. In Cenomanian and Turonian deposits from NW Germany (Wunstorf) and Morocco (Tarfaya), Prauss (2006, 2012a) differentiated specimens that are morphologically close to *Trithyrodinium suspectum* but with a more complex archaeopyle, tentatively assigning them to the genus *Ginginodinium*. The archaeopyle in *Ginginodinium?* sp. A involves from one intercalary plate (2a) to three intercalary plates (1a–3a), often attached, and sutures are usually present between precingular plates that are attached at the singular border (3I3Pa). Prauss (2006) noted that *Ginginodinium?* sp. A probably includes specimens informally described as ‘Cyst form A’ by Marshall and Batten (1988). The type species, *Ginginodinium spinulosum* (Cookson and Eisenack 1960, p. 7, pl. 2, fig. 9), and other accepted species, sometimes exhibit accessory sutures on dorsal precingular plates, in addition to a three plate intercalary archaeopyle (Stover and Eivitt 1978). In the Melton Ross material, consistent differentiation of this taxon from *T. suspectum* was problematic.

**Stratigraphic range/occurrence.** At Melton Ross, local Bed I (sample -22) to Bed H (sample -1).

Genus *Subtilisphaera* Jain and Millepied, 1973, emend. Lentin and Williams, 1976

Type species: *Subtilisphaera senegalensis* Jain and Millepied, 1973

*Subtilisphaera pontis-mariae* (Deflandre, 1936) Lentin and Williams, 1976

Figures 5.14, 5.15

Description. Small to intermediate-sized, elongate to ovoidal-shaped, smooth-walled, bicavate to circumcavate peridinioid dinoflagellate cysts with pericoels developed into a sub-conical apical horn and one similar antapical horn that is positioned asymmetrically. The epicyst is larger than the hypocyst. The cingulum is delimited by two low ridges. An archaeopyle has not been observed. The endocyst is sub-spherical and slightly thicker walled than the periphram. The horns are approximately 1/5 to 1/4 of the diameter of the endocyst.

**Dimensions** (seven measured specimens). Length 47.3 (52.7) 57.0 µm; width 32.6 (36.0) 42.2 µm.

Discussion. The *Subtilisphaera* specimens encountered in the present study have comparable morphology, with some variation in the degree of cavation and the length of horns. They are recorded here as *Subtilisphaera pontis-mariae*. The holotype (Deflandre 1936, pl. 2, fig. 7) is closely comparable to the Melton Ross specimen illustrated in Figure 5.14. Some subsequently illustrated specimens (e.g. Davey 1970, pl.1, figs 10–11; Dodsworth 2016, pl. 1, fig. 7) differ slightly in having longer horns, c. 1/2 to 1/3 the diameter of the endocyst. While Dodsworth (1996, 2000) used *Subtilisphaera* spp. for eastern and southern England specimens from the CTB interval, other workers in eastern England (Marshall and Batten 1988; Hart et al. 1993), southern England (FitzPatrick 1995; Pearce et al. 2009) and northern France (Foucher 1979) have considered the range of morphological variation of *Subtilisphaera* specimens to fall within acceptable intra-specific limits for *S. pontis-mariae* in material of this age. However, Foucher (1980) assigned the latest Cenomanian specimens to *S. cf. pontis-mariae* in the Boulonnais region of northern France.

**Stratigraphic range/occurrence.** At Melton Ross, local Bed I (sample -22) to Bed H (sample -1).


Type species: *Trithyrodinium evitti* Drugg, 1967
Trithyrodinium maculatum sp. nov.

Figures 7.14, 7.15, 7.18, 7.19

Derivation of name. From maculate, meaning spotted or stained, with reference to the distinctive surface markings.


Diagnosis. A species of Trithyrodinium possessing an ornament of ring-shaped indentations on the endophagm and periphragm.

Description. Small to intermediate-sized, spheroidal to ovoidal-shaped, circumcavate, peridinioid dinoflagellate cysts. The endophagm (c. 1 µm thick) and periphragm (c. 0.5 µm thick) are smooth to finely-granular, both possessing regular ornament elements that are circular to sub-circular, ring-shaped indentations (c. 3–6 µm diameter), which are at least half the thickness of the surrounding wall areas. Cavation is c. 2–3 µm wide at the lateral margins and up to 10 µm in the antapical region. An archaeopyle involving three anterior intercalary plates, type 3I(1–6) is formed in the endocyst, with operculum plates attached or detached. The 2a intercalary plate is isodeeltaform, hexa-type (Fig. 7.19; cf. text-figure 6 in Bujak and Davies 1983). The periarchaeopyle type has not been confirmed. There is no evidence of tabulation, other than that indicated by the archaeopyle.

Dimensions (length × width of the endocyst). Holotype, 39.6 × 42.5 µm. Other specimens (5 measured): length 38.0 (45.6) 55.3 µm, width 39.5 (45.6) 55.3 µm.

Discussion. Only six specimens of this taxon have been recorded to date. However, its distinctive morphology makes it easy to identify and warrants the erection of a formal species. The endocyst archaeopyle type is consistent with the genus Trithyrodinium but an outer wall layer (periphragm) has only been observed on two specimens, including the holotype. The circular indentations on the cyst walls are ring-shaped, as opposed to the whole area within the circle being thinned. The ring shapes are reminiscent of those attributed to impressions made by coccoliths on the surface of specimens of an alga (Campenia sp.) by Prauss (2012b, figs 10A, 11H). However, the T. maculatum specimens occur in assemblages of abundant dinoflagellate cysts, with other taxa not exhibiting comparable markings. The markings are a key identifying feature but could be a preservation artefact. This type of surface ornament has not previously been reported on other species of P-cysts, including those belonging to the genus Trithyrodinium. In the G-cyst species Apteodinium maculatum, comparable ring-shaped thinning surrounds small thickened circular areas (Eisenack and Cookson 1960).

Stratigraphic range/occurrence. At Melton Ross, Bed C (sample -11) to Bed G (sample -2).

Order GONYAULACALES Taylor, 1980
Suborder GONYAULACINAE (Autonym)
Family GONYAULACEAE Lindemann, 1928
Genus Dissiliodinium Drugg, 1978
Type species: Dissiliodinium globulus Drugg, 1978

?Dissiliodinium globulus Drugg, 1978

Figures 7.4, 7.8

Description. Intermediate-sized, spheroidal-shaped, smooth and thin-walled proximate autocysts lacking ornament. Tabulation is indicated by sutures between precingular plates on the epicyst. Archaeopyle formation involves one or more precingular plate. There is no evidence of tabulation on the hypocyst.

Dimensions (four specimens measured). Diameter: 57.0 (60.7) 65.0 µm.

Discussion. The observed features of the four available specimens allow tentative assignment to Dissiliodinium globulus. The taxon has been documented from the Upper Jurassic of Central Europe (Drugg 1978) and Northern Europe (e.g. Bailey et al. 1997) and is present in higher proportions within the Lower Cretaceous of SW Morocco (Below 1981). Prauss (2012b, fig 9a–i) reported specimens from around the CTB at Tarfaya.

Stratigraphic range/occurrence. At Melton Ross, Bed E (sample -8 only).

Type species: Leptodinium subtile Klement, 1960

Leptodinium? aff. delicatum (this paper)

Figure 6.12

Description. Intermediate-sized, smooth and thin-walled autocysts with sub-polygonal shape and septae (entire; c. 1.5 to 3.5 µm high) that probably correspond to tabulation. The cingulum appears to have strongly offset ventral ends. An archaeopyle has not been observed on the two specimens recorded.
Dimensions. Diameter: illustrated specimen, 61.8 µm; second specimen, 60.8 µm.

Discussion. In *Leptodinium? delicatum*, the wall is very thin (less than 0.5 µm thick) and only attains a thickness of 0.5 µm when forming plate boundaries (Davey 1969, p. 123). The Melton Ross specimens resemble *L.? delicatum* but differ in possessing sutural septae. The detailed plate formula is not yet determined. The probable offset sulcus is consistent with the genus *Leptodinium* though the presence of septae is more compatible with the genus *Impagidinium* (Stover and Evitt 1978).

Stratigraphic range/occurrence. At Melton Ross, Bed C (samples -11 and -11.5).

Genus *Oligosphaeridium* Davey and Williams, 1966, emend. Davey, 1982
Type species: *Oligosphaeridium complex* (White, 1842)
Davey and Williams, 1966

*Oligosphaeridium totum* Brideaux, 1971

Figures 5.9, 5.10, 8.9

Discussion. The published taxon *Oligosphaeridium totum* is considered here to accommodate a skolochorate dinoflagellate cyst from NW Germany and the North Sea that was informally assigned by Marshall and Batten (1988) to *Litosphaeridium* sp. A. The latter name was adopted in subsequent studies of the CTB interval in eastern England (Hart et al. 1993; Dodsworth 1996), southern England (FitzPatrick 1995) and Poland (Dodsworth 2004b). The processes of *O. totum* are tubular, sometimes slightly narrower medially, with apices usually flaring, buccinate and possessing an entire margin (Brideaux 1971). Two subspecies are differentiated, based mainly on process length relative to central body, *O. totum* subsp. *totum* (relatively long processes, approximately two-thirds to one central body diameter in length; comparable to most specimens encountered in this study, e.g. Figs 5.9, 8.9) and *O. totum* subsp. *minus* (relatively short processes, approximately one half of the central body diameter; cf. Fig. 5.10). The Melton Ross specimens have the same process formula as those described by Brideaux (1971) for *O. totum* and by Marshall and Batten (1988) for *Litosphaeridium* sp. A; 4', 6'l, 0c, 5'l1, 1p, 1'l1, l's. This is a process formula common to all species of *Oligosphaeridium* and some species of *Litosphaeridium* (Lucas-Clark 1984). *Litosphaeridium* differs from *Oligosphaeridium* in having dome-shaped processes that are typically not expanded distally (Stover and Evitt 1978). The type material of *O. totum* has a slightly thicker, more scabrate central body wall than the smooth walls observed on Melton Ross specimens and those described and illustrated from NW Germany (Marshall and Batten 1988, p. 92, pl. I, fig. 9; Marshall 1983, pl. 13, figs 16, 18). However, Singh (1971) reported a smooth cyst wall on *Oligosphaeridium diastema*, a junior synonym of *O. totum* (as agreed by Singh and Brideaux; Brideaux and McIntyre 1975, p. 29). Some variation in thickness and ornament of the central body may therefore be attributed to intra-specific variation.

Stratigraphic range/occurrence. At Melton Ross, local Bed VII (sample -17) to Bed F (sample -3).

Type species: *Pterodinium aliferum* Eisenack, 1958

*Pterodinium crassimuratum* (Davey and Williams, 1966)
Thurow et al., 1988

Figure 5.7

Discussion. Kjellström (1973) and Pavlishina (1990) considered *Pterodinium? pterotum* to be a senior synonym of *Pterodinium crassimuratum* and *Pterodinium cingulatum* subsp. *polygonale*. In Dodsworth and Eldrett (2019, plate 1, fig. 4), specimens encountered were assigned on the basis of this proposed synonymy to *P.? pterotum*. However, subsequent inspection of the holotype photographs and original descriptions for the three taxa indicates that our specimens, both from the USA and Europe, are closest to *P. crassimuratum* and *P. cingulatum polygonale*, having thickened intra-plate areas and ‘a linear depressed area on each side of the ledges [septae] thereby separating a more elevated central plate area from the ledges’ (Clarke and Verdier 1967, p. 47). We agree with Clarke et al. (1968, p. 181) that the latter subspecies is a junior synonym of *P. crassimuratum*. However, Cookson and Eisenack (1958, p. 50) make no reference to comparable thickened intra-plate areas in *P.? pterotum*, and none are visible on their holotype photograph (plate 11, fig. 7). We therefore conclude that *P.? pterotum* is not a senior synonym of *P. crassimuratum*.

Family AREOLIGERACEAE Evitt, 1963
Genus *Apteia* Eisenack, 1958
Type species: *Apteia polymorpha* Eisenack, 1958

Fensome et al., 1994

Figures 6.1, 6.2

Description. Intermediate-sized, lenticular-shaped cysts with rounded or flattened apical and antapical areas, lacking well-developed horns. The autophragm is smooth to lightly pitted and is covered with a coarse reticulum. Rod-like structures with expanded to bifurcating tips support the reticulum muri and are c. 6 – 10 µm long. The muri are fibrous or finely fenestrate. The lacuna between the muri vary in size, from c. 5 – 13 µm across. A slight indentation in the reticulum in the antapical region on one specimen (Fig. 6.1) may represent a gap between antapical horns. Tabulation is indicated by an apical archaeopyle only.

Dimensions. (five measured specimens, length × width without operculum). Length 61.0 (66.9) 73.0 µm; width 83.8 (86.3) 89.0 µm.

Discussion. The Melton Ross specimens conform to the type material of *A.? spongireticulata*, illustrated from NW Germany by Prössl (1990). *Canningia macroreticulata* differs in having smaller lacuna between its muri of c. 2 – 7 µm and has shorter (c. 2 µm high), entire muri without fibrous/fenestrate structure (Lebedeva, in Ilyina et al. 1994). The Melton Ross specimens may be the first published record of *A.? spongireticulata* outside Germany.
Schematophora shape and lacks any tabulation in the cingulum region and Helby 1987, fig. 4) but its septae are coarsely perforate. The shape, presence of a sulcal notch and a probable small postcingular 1III plate (Fig. 6.8) indicate an areoligeracean affinity (cf. Evitt 1985, text-figs 10.2, 10.6). Although the type species and most other specimens lack an outer layer. Comparable specimens from Poland were assigned to Eyrea sp. nov. by Dodsworth (2004b). In poorly preserved material from eastern England, it can be difficult to distinguish specimens of Dinoflagellate? type D from degraded and/or broken specimens of Sentusudinium ringnesiorum (cf. Fig. 7.20) and the granular endocysts of Trithyriodinium spectum and Ginginodinium? sp. A.

**Stratigraphic range/occurrence.** At Melton Ross, local Bed 1 to Bed H. A broad Late Cretaceous range is suggested by the records of Ioannides (1986) and Cookson and Eisenack (1971).