# Isotopic and palynological evidence for a new Early Jurassic environmental perturbation

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## ARTICLE INFO

## ABSTRACT

The Early Jurassic Epoch was a predominantly greenhouse phase of Earth history, but a comprehensive understanding of its climate dynamics is hampered by a lack of high resolution multi-proxy environmental records. Here we report a geologically brief (approximately several hundred thousand years) negative carbon isotope excursion (CIE) of 2–3‰ in both marine and terrestrial materials, recognised for the first time for the Late Sinemurian Substage (Early Jurassic, ~194 Ma) of eastern England. The Late Sinemurian carbon isotope excursion, which is termed the S-CIE herein, is accompanied by peaks in the abundance of the pollen grain *Classopollis classoides* and the dinoflagellate cyst *Liasidium variabile*. *Classopollis classoides* was thermophilic and is a reliable proxy for hot/warm climatic conditions. *Liasidium variabile* is interpreted as thermophilic and eutrophic using multivariate statistics, its fluorescence properties being similar to living heterotrophic dinoflagellate cysts, and its association with *C. classoides*. Moreover, the morphological and ecological similarities of *L. variabile* to the Cenozoic genus *Apectodinium* are noteworthy. The co-occurrence of the acmes of *C. classoides* and *L. variabile* with a negative CIE is interpreted here as having wide geographical significance due to the marine and terrestrial carbon isotope signals being precisely in phase within an open marine setting. This is consistent with an oceanic–atmospheric injection of isotopically-light carbon, coupled with global warming and increased marginal marine nutrient supply, possibly the result of increased precipitation due to an enhanced hydrological cycle or a seasonally-stratified water column. A probable sea level rise of at least regional extent has been identified at the *L. variabile* event in other records, which supports this putative phase of global warming. All these features are common to the Paleocene/Eocene thermal maximum (PETM, ~56 Ma), and there are also similarities with the Early Toarcian oceanic anoxic event (T-OAE, ~182 Ma).

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## 1. Introduction

The Jurassic Period was a mainly greenhouse phase of Earth history, characterised by global palaeotemperatures that were 5–10 °C higher than modern levels based on palaeo-modelling results (Chandler et al., 1992; Rees et al., 1999; Sellwood and Valdes, 2008). During the Early Jurassic the supercontinent Pangaea fragmented into Laurasia and Gondwana, creating major seaways including the proto Central Atlantic Ocean and marine connections between western Tethys and the north polar region (Fig. 1). Sea levels rose steadily throughout the Early Jurassic (Haq et al., 1987, 1988) and much of Europe and surrounding areas in the northern hemisphere were covered by warm, relatively shallow and stable epicontinental seas (summary in Coward et al., 2003). Sedimentation was dominated by rhythmic cycles of fossiliferous marine shelfal muds with subordinate carbonates and

sands within large fault-bounded basins that were created by crustal extension. Northwest Europe probably lay around 10° south of its present latitude (Bradshaw et al., 1992). British Lower Jurassic stratigraphy has been summarised by Cope et al. (1980) and Simms et al. (2004).

The Early Toarcian oceanic anoxic event (T-OAE, ~182 Ma according to the timescale of Gradstein et al., 2012) was a short-lived major phase of global warming during the late Early Jurassic, and is marked by a significant negative carbon isotope ($\delta^{13}$C) excursion (CIE) (Jenkyns, 2010). The T-OAE and other large CIEs have been interpreted as being due to the release of isotopically-light, largely biogenic methane from sub-seafloor accumulations of methane clathrate (Hesselbo et al., 2000; Kemp et al., 2005; Cohen et al., 2007). Similar, but smaller scale, events have also been recognised in the Early Jurassic at the Pliensbachian–Toarcian boundary (~183 Ma; Hesselbo et al., 2007; Bodin et al., 2010; Littler et al., 2010), and at the Sinemurian–Pliensbachian boundary (~191 Ma; Korte and Hesselbo, 2011).

In this study of a succession from eastern England, we report geochemical and palynological data which are consistent with another
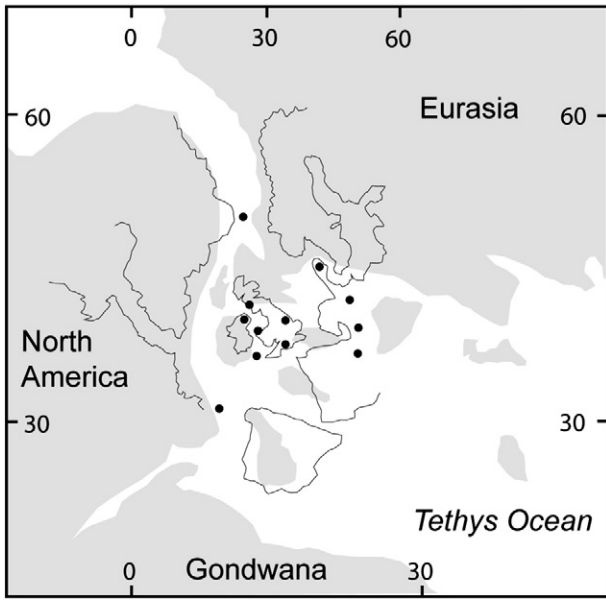
**Fig. 1.** A palaeogeographical map of eastern North America and northwestern Eurasia for the Late Sinemurian (adapted from Smith et al., 1994) with the locations of records of *Liasidium variabile* indicated by the 12 filled circles. The locations were taken from Brittain et al. (2010). Continental areas are indicated in grey, and the black lines depict selected modern coastlines.

Early Jurassic warming event in the Late Sinemurian (~194 Ma). This evidence comprises a significant negative CIE (termed herein the S-CIE) which is temporally coincident with abundance acmes of the terrestrially-derived pollen grain *Classopollis classoides* and the marine dinoflagellate cyst *Liasidium variabile* (Figs. 2, 3). Both these taxa were thermophilic, and *L. variabile* has a narrow stratigraphical range.

The Mesozoic gymnosperm pollen *Classopollis* was produced by plants belonging to the family Cheirolepidiaceae. This genus is interpreted as being thermophilic and xerophytic and hence a proxy for hot/warm palaeotemperatures (e.g. Pocock and Jansonius, 1961; Srivastava, 1976). *Classopollis*-producing plants have been interpreted to have preferred both coastal and upland slope habitats (Batten, 1975; Filatoff, 1975).

The distinctive dinoflagellate cyst *L. variabile* is an index fossil for the Late Sinemurian of the northern hemisphere and is the oldest known peridinialean species (Brittain et al., 2010). In an interval characterised by extremely low dinoflagellate cyst diversities, *L. variabile* suddenly appeared, became abundant and abruptly disappeared. The range of *L. variabile* is typically entirely within the *Oxynoticeras oxynotum* ammonite Biozone, and this taxon has also been interpreted as a warm water species using multivariate statistics (Riding and Hubbard, 1999). Because of some similarities to the apparently thermophilic dinoflagellate cyst genus *Apectodinium* Costa and Downie, 1976, *L. variabile* has been suggested to have tracked warmer water polewards (Feist-Burkhardt, 2009).

These three linked phenomena are interpreted here as being consistent with a major biogeochemical perturbation associated with a brief phase of significant ocean/atmosphere carbon injection and global warming ~8 Ma prior to the T-OAE.

## 2. Material and methods

The Copper Hill Borehole was an exploratory stratigraphical borehole drilled by the British Geological Survey (BGS) during 1991 in Copper Hill Quarry, near Ancaster, Lincolnshire, eastern England; the national grid reference is SK 9787 4265 (Fig. 4; Berridge et al., 1999, fig. 19). The Lower and Middle Jurassic (Sinemurian to Bajocian) palynology of the Copper Hill Borehole was described by Riding (1992). The palynology samples were prepared using the standard acid digestion method including mild oxidation with cold nitric acid (Wood et al., 1996), and the rock samples, prepared organic residues, and microscope slides are all housed
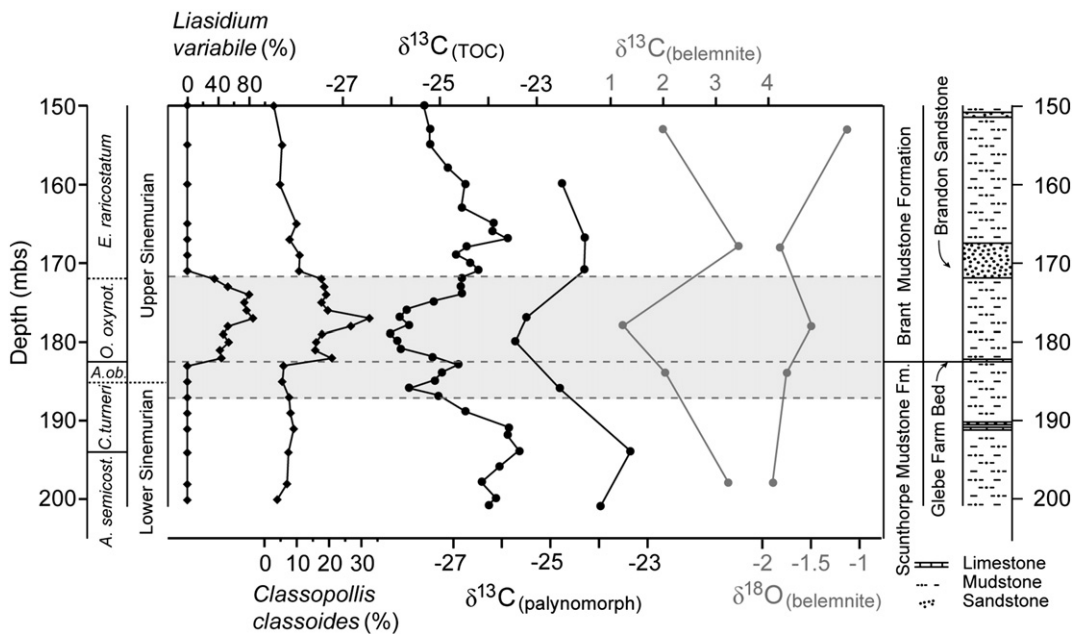


**Fig. 2.** The Copper Hill Borehole between ~200 m and 150 m, with the Sinemurian ammonite biozones (left) and the lithostratigraphy (right). The left hand curve illustrates the range and acme of *Liasidium variabile* in the samples studied; the data are expressed as percentages of the entire marine palynofloras. To the immediate right of this are the percentages of *Classopollis classoides* expressed as percentages of the terrestrially-derived palynomorphs. The remaining four curves depict the stable isotope records (black: organic carbon, grey: carbonate). The light grey shading indicates the extent of the S-CIE. The dashed line in the centre of the light grey shading indicates the range base of *L. variabile* which is coincident with the second $\delta^{13}C$ pulse. The right hand column is a lithological summary from the unpublished core description of N.G. Berridge, together with the principal lithostratigraphical units.

**Fig. 3.** A low magnification view of the palynomorph residue from 177 m in the Copper Hill Borehole illustrating the prominence of *Liasidium variabile* and *Classopollis classoides* in this sample. The scale bar represents 50 μm.

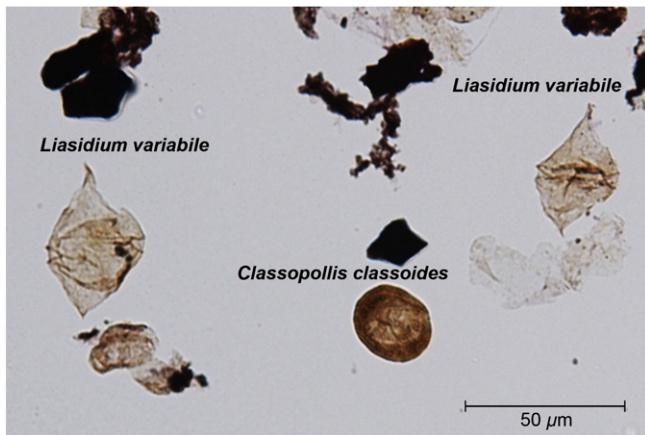in the collections of the British Geological Survey, Keyworth, Nottingham NG12 5GG, United Kingdom.

For the carbon isotope analysis on the bulk and palynomorph concentrates, the samples were homogenised and treated with 5% hydrochloric acid to remove any calcite. $^{13}C/^{12}C$ analyses were

performed by combustion in a Costech Elemental Analyser (EA) on-line to a VG TripleTrap and Optima dual-inlet mass spectrometer, with $\delta^{13}C$ values calculated to the VPDB scale using within-run laboratory standards calibrated against NBS18, NBS-19 and NBS-22. Replicate analysis of well-mixed samples indicated a precision of $\pm$<0.1‰ (1 SD). For the belemnite carbon and oxygen isotope analysis, approximately 50 μg of carbonate was used for isotope analysis using a GV IsoPrime mass spectrometer plus Multiprep device. The isotope values ($\delta^{13}C$, $\delta^{18}O$) are reported as per mille (‰) deviations of the isotopic ratios ($^{13}C/^{12}C$, $^{18}O/^{16}O$) calculated to the VPDB scale using a within-run laboratory standard calibrated against NBS standards. Analytical reproducibility of the standard calcite (KCM) is <0.1‰ for $\delta^{13}C$ and $\delta^{18}O$.

The interval studied in the Copper Hill Borehole is relatively fossiliferous and the standard ammonite biozones and biosubzones have been determined (Tables 1–3). The ammonite biostratigraphy, however, is of limited utility in terms of identifying the duration of relatively short-lived palaeoclimatic events such as the S-CIE. The S-CIE is not currently well-constrained chronostratigraphically because there are no accurate estimates for the absolute age ranges of the three Late Sinemurian ammonite biozones as suitable cyclic sedimentary successions have not been studied. A cyclostratigraphical calibration of the long-term decline in seawater $^{87}Sr/^{86}Sr$ values throughout the Early Jurassic gives an estimate for ammonite biozone duration of ~1 Ma each (Weedon and Jenkyns, 1999; Gradstein et al., 2012). This means that the S-CIE may have lasted a few hundred
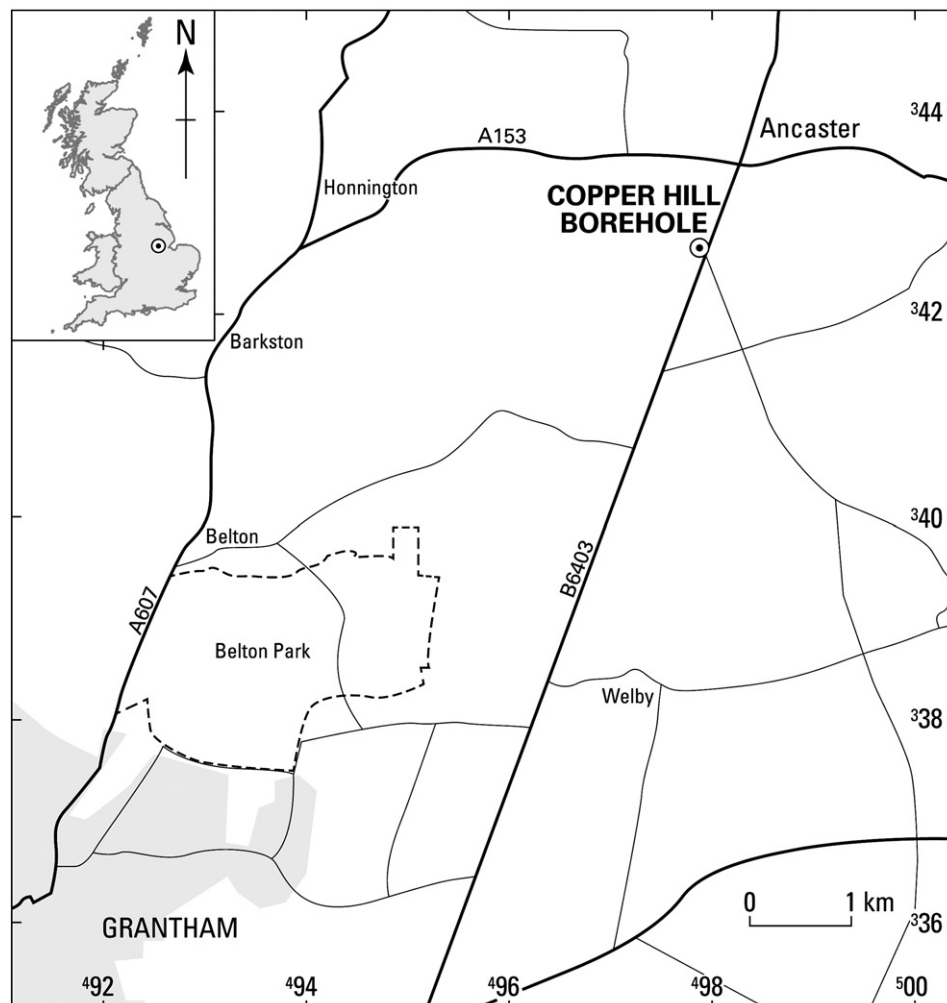


**Fig. 4.** The location of the Copper Hill Borehole, near Ancaster, Lincolnshire, eastern England, United Kingdom.

**Table 1**

The quantitative stratigraphical distribution of palynomorphs between 200 m and 150 m in the Copper Hill Borehole. The numbers in the cells are percentages. An 'X' denotes that the respective palynomorph has been recorded outside the count. An ellipsis (....) denotes the absence of the respective palynomorph. In the ammonite biozone/biosubzone column, the index taxa have been abbreviated. A.s. = *Arnioceras semicostatum*, E.r. = *Euagassiceras resupinatum*, C. tur. = *Caenisites turneri*, As.o? = *Asteroceras obtusum?*, Ox.ox. = *Oxynoticeras oxynotum*, Ox.si. = *Oxynoticeras simpsoni*, E. rar. = *Echioceras raricostatum*, E.rari. = *Echioceras raricostatoides* and L.m. = *Leptechioceras macdonnelli*.

| Substage | Ammonite biozone/subbiozone | Depth (m) | POLLEN | | | | | | SPORES | | | | | | | | | | | | | | | | | | | | | AQUATIC PALYNOMORPHS | | | | | | Number of specimens counted |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | *Araucariacites australis* | bisaccate pollen (undifferentiated) | *Cerebropollenites macroverrucosus* | *Chasmatosporites* spp. | *Classopollis classoides* | *Perinopollenites elatoides* | *Aurtitulinasporites scanicus* | *Baculatisporites/Osmundacidites* spp. | *Cibotiumspora juriensis* | *Concavissimisporites verrucosus* | *Coronatispora valdensis* | *Cyathidites* spp. | *Dictyophyllidites* spp. | *Duplexisporites* sp. | *Gleicheniidites senonicus* | *Ischyosporites variegatus* | *Kraeuselisporites reissingeri* | *Leptolepidites rotundus* | *Neoraistrickia gristhorpensis* | *Retitriletes austroclavatidites* | spores (undifferentiated) | *Sestrosporites pseudoalveolatus* | *Stereisporites* spp. | *Todisporites* spp. | *Uvaesporites argenteaeformis* | reworked Carboniferous spores | *Botryococcus* | foraminiferal test linings | *Liasidium variabile* | *Micrhystridium* spp. | *Tasmanites* spp. | *Veryhachium* sp. | |
| Upper Sinemurian | E. rar./L.m. | 150 | ... | 15.5 | X | ... | 2.9 | 72.1 | ... | ... | ... | ... | ... | 1.9 | X | ... | ... | ... | ... | ... | ... | 0.6 | 2.2 | ... | ... | ... | X | ... | ... | 1.3 | ... | 3.5 | X | ... | 315 |
| | E. rar./L.m. | 155 | X | 24.3 | 1 | 1.7 | 5.4 | 58.6 | ... | 0.7 | ... | ... | X | 2 | ... | ... | X | ... | ... | ... | ... | 0.3 | 1.7 | ... | ... | 0.3 | ... | X | ... | 0.3 | ... | 3.7 | X | ... | 297 |
| | E. rar./L.m. | 160 | ... | 24.8 | 0.3 | 0.7 | 4.7 | 59.4 | ... | X | ... | X | ... | 1 | X | ... | ... | ... | ... | ... | X | 2 | 1.7 | ... | ... | ... | X | ... | ... | 0.7 | ... | 4.7 | X | ... | 298 |
| | E. rar./L.m. | 165 | ... | 28.9 | 1.4 | 0.4 | 9.5 | 50.3 | ... | X | ... | X | ... | 2.1 | X | ... | ... | X | ... | ... | ... | 1.4 | 0.7 | ... | X | ... | X | X | ... | 0.4 | ... | 4.9 | ... | ... | 284 |
| | E.rar./E.rari. | 167 | ... | 29.1 | 0.3 | ... | 7.6 | 57.3 | ... | 0.3 | ... | ... | ... | 0.3 | ... | ... | ... | ... | ... | ... | ... | 1.2 | ... | ... | ... | ... | X | ... | X | ... | ... | 3.9 | X | ... | 330 |
| | E.rar./E.rari. | 169 | ... | 18.8 | 1.3 | ... | 10.5 | 58.6 | ... | X | ... | ... | ... | 0.6 | 0.3 | ?X | ... | ... | ... | X | ... | 1 | 4.5 | ... | ... | ... | X | ... | X | 0.9 | ... | 3.5 | ?X | ... | 314 |
| | E.rar./E.rari. | 171 | X | 17.1 | 0.9 | X | 10.1 | 59.8 | ... | ... | ... | ... | ... | 1.9 | ... | ... | ... | ... | ... | ... | ... | 0.9 | 2.6 | ... | ... | ... | X | ... | X | 1.3 | ... | 5.4 | ... | ... | 316 |
| | E.rar./E.rari. | 172 | X | 18.8 | X | 2.3 | 15.6 | 45.1 | ... | 0.3 | ... | ... | X | 2.3 | ... | ... | X | X | ... | ... | ... | 0.3 | 3.3 | ... | X | ... | X | ... | X | ... | 4.2 | 6.5 | 1.3 | ... | 308 |
| | Ox.ox/Ox.ox. | 173 | ... | 19.4 | ... | 2 | 15.5 | 42.1 | ... | X | ... | X | X | 1.6 | ... | ... | ... | ... | ?X | ... | ... | 0.3 | 2.6 | ... | ... | ... | X | X | ... | ... | 8.6 | 7.9 | ... | ... | 304 |
| | Ox.ox/Ox.ox. | 174 | ... | 16.6 | 1.1 | 2.1 | 15.5 | 37.8 | ... | 1.1 | ... | ... | X | 2.8 | X | ... | ... | ... | ... | ... | X | 1.8 | 2.1 | ... | ... | ... | X | ... | X | ... | 15.2 | 3.9 | X | ... | 283 |
| | Ox.ox/Ox.ox. | 175 | ... | 23.2 | 0.3 | 1.7 | 15 | 35.8 | ... | X | ... | ... | X | 2.4 | ... | ... | ... | ... | X | ... | ... | 2.1 | 3.8 | ... | ... | ... | 0.3 | ... | X | ... | 11.3 | 3.8 | 0.3 | ... | 293 |
| | Ox.ox/Ox.ox. | 176 | X | 15.7 | 2.1 | 1.7 | 15 | 35.2 | X | 0.7 | ... | ... | ... | 2.4 | ... | 0.4 | ... | ... | X | ... | X | 1.4 | 1.7 | X | ... | X | ... | ... | X | X | 18.1 | 4.9 | 0.7 | ... | 287 |
| | Ox.ox/Ox.ox. | 177 | X | 17.4 | 0.3 | 1.3 | 24.3 | 26.6 | ... | ... | ... | ... | ... | 2 | ... | ... | ... | ... | X | ... | X | 0.6 | 2.3 | ... | ... | X | ... | X | X | X | 21.3 | 3.9 | ... | ... | 305 |
| | Ox.ox/Ox.si. | 178 | X | 20.2 | 0.7 | 0.3 | 23 | 35.9 | ... | 0.7 | ... | ... | ... | 2.4 | ... | ... | ... | ... | X | ... | X | 1.1 | 1.7 | ... | ... | ... | X | X | X | ... | 7.3 | 5.6 | 1.1 | ... | 287 |
| | Ox.ox/Ox.si. | 179 | ... | 31.6 | 1.4 | 1.8 | 16.2 | 34.2 | ... | X | ... | ... | ... | 2.2 | ... | X | ... | X | ... | ... | X | 0.4 | 2.9 | ... | ... | ... | X | ... | X | ... | 4.3 | 4.7 | 0.3 | ... | 278 |
| | Ox.ox/Ox.si. | 180 | ... | 20.8 | 1 | 3.2 | 13.1 | 39.5 | ... | X | ... | ... | X | 1.6 | ... | ... | ... | X | ... | ... | X | X | 2.2 | ... | ... | ... | ... | ... | ... | ... | 9.9 | 8.7 | ... | ... | 312 |
| | Ox.ox/Ox.si. | 181 | ... | 25.3 | X | 1 | 14.5 | 43.4 | ... | X | ... | ... | ... | 1 | ... | ... | ... | X | ... | ... | X | 0.3 | 6.1 | ... | ... | ... | X | ... | X | 0.3 | 3.4 | 4 | 0.7 | ... | 297 |
| | Ox.ox/Ox.si. | 182 | ... | 21.9 | 0.9 | 0.3 | 19.2 | 44.7 | ... | 0.6 | ... | X | ... | 1.5 | X | ... | X | X | ... | ... | X | 0.3 | 2.4 | ... | ... | X | ... | X | X | ... | 3.6 | 4.3 | 0.3 | ... | 329 |
| | As.o? | 183 | ... | 27.5 | 0.6 | 3.6 | 5.4 | 44 | ... | 0.9 | ... | ... | ... | 3.9 | 0.3 | ... | ... | ... | ... | ... | ... | 0.6 | 2.1 | ... | ... | ... | 0.3 | ... | ... | X | ... | 10.8 | ... | ... | 334 |
| | As.o? | 185 | X | 22.8 | X | 5.5 | 4.3 | 37.2 | ... | X | ... | ... | ... | 3.4 | ... | ... | ... | ... | ... | ... | X | 0.6 | 2.5 | ... | ... | ... | X | ... | X | ... | ... | 22.5 | 1.2 | X | 325 |
| Lower Sinemurian | ? | 187 | X | 20.8 | X | 3.7 | 6.4 | 45 | ... | ... | ?X | ... | X | 4.4 | 0.7 | X | ... | ... | ... | ... | ... | X | 1.3 | X | ... | ?X | 0.3 | ... | 0.3 | ... | ... | 17.1 | ... | ... | 298 |
| | C.tur. | 189 | X | 22.2 | 1.3 | 2.2 | 6.9 | 45 | ... | 0.3 | ... | ... | ... | 3.4 | 0.3 | ... | ... | ... | ... | 0.3 | ... | 0.6 | 1.6 | ... | ... | ... | ... | ... | ... | ... | ... | 15.3 | 0.6 | ... | 320 |
| | C.tur. | 191 | ... | 22.3 | 0.3 | 1.9 | 8.4 | 52.8 | ... | X | ... | ... | ... | 3.6 | X | ... | ... | ... | ... | ... | ... | 0.3 | 1.9 | ... | ... | X | ... | ... | ... | 0.7 | ... | 7.5 | 0.3 | ... | 309 |
| | A.s./E.r. | 194 | . | 17.7 | X | 1.8 | 6.9 | 58.5 | ... | 0.7 | ... | ... | X | 1.4 | 0.4 | ... | ... | ... | X | ... | ... | 0.4 | 0.4 | 2.5 | ... | ... | ?X | ... | 0.7 | X | ... | 8.6 | ... | ... | 277 |
| | A.s./E.r. | 198 | ... | 19.7 | 0.7 | 1.7 | 6.5 | 58.8 | ... | X | ... | X | X | 1.4 | X | ... | ... | ... | X | X | ... | X | 0.7 | 2 | ... | X | ... | ... | X | 0.3 | ... | 8.2 | ... | ... | 294 |
| | A.s./E.r. | 200 | ... | 24.5 | ... | ... | 3.8 | 61.8 | ... | ... | ... | ... | ... | 0.3 | X | ... | ... | ... | ... | ... | ... | ... | 0.7 | 1 | ... | ... | ... | ?X | ... | ... | ... | ... | 7.9 | ... | ... | 293 |

**Table 2**

Raw data on carbon and oxygen isotopes and the percentages of *Liasidium variabile* and *Classopollis classoides* between 200 m and 150 m in the Copper Hill Borehole. The geochemical data comprise $\delta^{13}C_{org}$ measured on the bulk organic carbon for 37 samples, $\delta^{13}C_{bel}$ and $\delta^{18}O_{bel}$ measured from five belemnite specimens and $\delta^{13}C_{palynomorph\ concentrate}$ measured on the organic residues from eight samples. The figures for *L. variabile* and *C. classoides* are percentages of the overall palynoflora. In the ammonite stratigraphy column, the ammonite biozones and biosubzones have been abbreviated; A.s. = *Arnioceras semicostatum*, E.r. = *Euagassiceras resupinatum*, C. tur. = *Caenisites turneri*, ?As.o. = ?*Asteroceras* obtusum, Ox.ox. = *Oxynoticeras oxynotum*, Ox.si. = *Oxynoticeras simpsoni*, E. rar. = *Echioceras raricostatum*, E. rari. = *Echioceras raricostatoides* and L.m. = *Leptechioceras macdonnelli*.

| Substage | Ammonite biozones | Depth (m) | Bulk organic C | Belemnites | | Paly. conc. | L. variabile (%) | C. classoides (%) |
|---|---|---|---|---|---|---|---|---|
| | | | $\delta^{13}C_{org}$ | $\delta^{13}C_{bel}$ | $\delta^{18}O_{bel}$ | $\delta^{13}C_{paly.conc.}$ | | |
| Upper Sinemurian | E.rar./L.m | 150 | −25.3 | | | | … | 2.9 |
| | | 153 | −25.2 | +2.00 | −1.13 | | | |
| | | 155 | −25.2 | | | | … | 5.4 |
| | | 158 | −24.8 | | | | | |
| | | 160 | −24.5 | | | −24.8 | … | 4.7 |
| | | 163 | −24.5 | | | | | |
| | | 165 | −23.9 | | | | … | 9.5 |
| | | 166 | −23.9 | | | | | |
| | E.r./E.r. | 167 | −23.6 | | | −24.3 | … | 7.6 |
| | | 168 | −24.4 | +3.42 | −1.81 | | | |
| | | 169 | −24.7 | | | | … | 10.5 |
| | | 170 | −24.4 | | | | | |
| | | 171 | −24.2 | | | −24.3 | … | 10.1 |
| | | 172 | −24.5 | | | | 4.2 | 15.6 |
| | Ox.ox./Ox.ox. | 173 | −24.6 | | | | 8.6 | 15.5 |
| | | 174 | −24.5 | | | | 15.2 | 15.5 |
| | | 175 | −25.1 | | | | 11.3 | 15 |
| | | 176 | −25.7 | | | | 18.1 | 15 |
| | | 177 | −25.8 | | | −25.5 | 21.3 | 24.3 |
| | Ox.ox./Ox.si. | 178 | −25.6 | +1.24 | −1.50 | | 7.3 | 23 |
| | | 179 | −26.0 | | | | 4.3 | 16.2 |
| | | 180 | −25.9 | | | −25.7 | 9.9 | 13.1 |
| | | 181 | −25.8 | | | | 3.4 | 14.5 |
| | | 182 | −25.1 | | | | 3.6 | 19.2 |
| | ?As.o. | 183 | −24.6 | | | | … | 5.4 |
| | | 184 | −24.9 | +2.05 | −1.75 | | | |
| | | 185 | −25.1 | | | | … | 4.3 |
| Lower Sinemurian | ? | 186 | −25.6 | | | −24.8 | | |
| | | 187 | −25.0 | | | | … | 6.4 |
| | C. tur. | 189 | −24.5 | | | | … | 6.9 |
| | | 191 | −23.6 | | | | … | 8.4 |
| | | 192 | −23.6 | | | | | |
| | A.s./E.r. | 194 | −23.4 | | | −23.4 | … | 6.9 |
| | | 196 | −23.8 | | | | | |
| | | 198 | −24.1 | +3.24 | −1.89 | | … | 6.5 |
| | | 200 | −23.8 | | | | … | 3.8 |
| | | 201 | −24.0 | | | −24.0 | | |

thousand years, which is comparable to the Paleocene/Eocene thermal maximum (PETM, ~56 Ma). Other potentially analogous CIEs are at the T-OAE, which has a poorly constrained duration of ~200–900 ka (Suan et al., 2008; Kemp et al., 2011), and at the Selli Event in the Early Aptian (~120 Ma). The Selli Event, or OAE 1a, was reported to have lasted $1.11 \pm 0.11$ Ma by Malinverno et al. (2010).

## 3. Palynomorph and kerogen assemblages

The palynofloras in 26 samples between 200 m and 150 m from the Sinemurian of the Copper Hill Borehole are illustrated in Table 1. The succession studied represents an open marine palaeoenvironment within a geographically extensive seaway (Coward et al., 2003). All the samples yielded abundant and well-preserved palynofloras of relatively low taxonomic diversity. The floras are overwhelmingly dominated by gymnospermous pollen grains, particularly bisaccate forms including *Vitreisporites pallidus*, *Classopollis classoides* and *Perinopollenites elatoides*. *Classopollis classoides* is a long-ranging species and is present throughout the entire Sinemurian to Bajocian succession drilled by the Copper Hill Borehole (Riding, 1992). The acme of this species is at 177 m, at which level it comprises 24.3% of the entire palynoflora and 32.5% of the terrestrial palynomorphs (Fig. 2; Tables 1, 2). This maximum is coincident with the lowest percentage (26.6%) of *P. elatoides* (Table 1). Other pollen present comprise *Araucariacites australis*, *Cerebropollenites macroverrucosus*

and *Chasmatosporites* spp. Pteridophyte spores are significantly subordinate to gymnospermous pollen, and are dominated by simple, smooth forms attributable to the genus *Cyathidites*. Other spores encountered include *Baculatisporites*/*Osmundacidites* spp., *Concavissimisporites verrucosus*, *Coronatispora valdensis*, *Dictyophyllidites* spp., *Duplexisporites* sp., *Gleicheniidites senonicus*, *Ischyosporites variegatus*, *Kraeuselisporites reissingeri*, *Neoraistrickia gristhorpensis*, *Retitrilites austroclavatidites*, *Sestrosporites pseudoalveolatus*, *Todisporites* spp. and *Uvaesporites argenteaeformis*. Rare reworked Carboniferous spores were recognised; these include *Densosporites* spp., *Lycospora pusilla* and *Tripartites vetustus*. The freshwater/brackish alga *Botryococcus* is present in low proportions (Table 1).

Marine palynomorphs are present in consistently lower proportions than terrestrially-derived taxa. *Liasidium variabile* is the only dinoflagellate cyst recognised and was recorded between 182 m and 172 m, within the Upper Sinemurian part of the Brant Mudstone Formation, where it is largely confined to the *O. oxynotum* ammonite Biozone (Fig. 2; Tables 1–3). The acme of *L. variabile* is at 177 m, at which level it comprises 21.3% of the entire palynoflora (Fig. 2; Tables 1, 2). The marine acanthomorph acritarch genus *Michrystridium* is present throughout in significant proportions. Other marine palynomorphs are present in low numbers; these are foraminiferal test linings, *Tasmanites* spp. and *Veryhachium* sp. (Table 1). The palynomorph species recognised in this study are listed, with author citations, in Appendix 1. The palynology of the entire Lower and

**Table 3**
The percentages of the four main kerogen macerals (i.e. black wood, other plant tissues, palynomorphs and amorphous organic material—AOM) between 200 m and 150 m in the Copper Hill Borehole. The percentages of terrestrially-derived and indigenous marine palynomorphs are also depicted. Note the consistent dominance of palynomorphs and terrestrially-derived palynomorphs respectively. In the stratigraphy column, the ammonite biozones and biosubzones have been abbreviated; *A.s.* = *Arnioceras semicostatum*, *E.r.* = *Euagassiceras resupinatum*, *C. tur.* = *Caenisites turneri*, *?As.o.* = *?Asteroceras obtusum*, *Ox.ox.* = *Oxynoticeras oxynotum*, *Ox.si.* = *Oxynoticeras simpsoni*, *E. rar.* = *Echioceras raricostatum*, *E. rari.* = *Echioceras raricostatoides* and *L.m.* = *Leptechioceras macdonnelli*.

| Stratigraphy | | Depth (m) | Percentages of the principal kerogen macerals | | | | Palynomorphs | |
|---|---|---|---|---|---|---|---|---|
| | | | Black wood (%) | Plant tissues (%) | Palynomorphs (%) | AOM (%) | % terrestrial | % marine |
| Upper Sinemurian | *E. rar./L. m.* | 150 | 18 | 15 | 57 | 10 | 95.2 | 4.8 |
| | | 155 | 8 | 17 | 70 | 5 | 96 | 4 |
| | | 160 | 18 | 18 | 60 | 4 | 94.6 | 5.4 |
| | | 165 | 12 | 12 | 70 | 6 | 94.7 | 5.3 |
| | *E.rar./E.rari.* | 167 | 18 | 15 | 55 | 12 | 96.1 | 3.9 |
| | | 169 | 22 | 12 | 60 | 6 | 95.6 | 4.4 |
| | | 171 | 20 | 8 | 67 | 5 | 93.3 | 6.7 |
| | | 172 | 17 | 12 | 68 | 3 | 88 | 12 |
| | *Ox.ox./Ox.ox.* | 173 | 8 | 12 | 77 | 3 | 83.5 | 16.5 |
| | | 174 | 12 | 16 | 70 | 2 | 80.9 | 19.1 |
| | | 175 | 12 | 8 | 73 | 7 | 84.6 | 15.4 |
| | | 176 | 12 | 12 | 70 | 6 | 76.3 | 23.7 |
| | | 177 | 12 | 15 | 68 | 5 | 74.8 | 25.2 |
| | *Ox.ox./Ox.si.* | 178 | 8 | 7 | 75 | 10 | 86 | 14 |
| | | 179 | 15 | 10 | 43 | 32 | 90.7 | 9.3 |
| | | 180 | 10 | 12 | 63 | 15 | 81.4 | 18.6 |
| | | 181 | 18 | 14 | 50 | 18 | 91.6 | 8.4 |
| | | 182 | 15 | 12 | 65 | 8 | 91.8 | 8.2 |
| | *?As.o.* | 183 | 17 | 15 | 47 | 21 | 89.2 | 10.8 |
| | | 185 | 15 | 13 | 64 | 8 | 76.3 | 23.7 |
| Lower Sinemurian | *?* | 187 | 12 | 13 | 70 | 5 | 82.6 | 17.4 |
| | *C.tur.* | 189 | 23 | 18 | 44 | 15 | 84.1 | 15.9 |
| | | 191 | 22 | 25 | 42 | 11 | 91.5 | 8.5 |
| | *A.s./E.r.* | 194 | 12 | 30 | 53 | 5 | 91.4 | 8.6 |
| | | 198 | 18 | 25 | 40 | 17 | 91.5 | 8.5 |
| | | 200 | 23 | 32 | 37 | 8 | 92.1 | 7.9 |

Middle Jurassic (Sinemurian–Bajocian) succession cored by the Copper Hill Borehole was documented by Riding (1992).

The percentages of the main kerogen macerals are given in Table 3. The organic residues are generally dominated by palynomorphs. Black wood and various plant tissues are subordinate to palynomorphs. Amorphous organic material (AOM) is largely relatively rare, although it is most abundant at 179 m where it attained 32% (Table 3).

## 4. Carbon and oxygen isotope geochemistry

Carbon isotope values were obtained from bulk organic matter, palynomorph concentrates and belemnite fragments between 201 m and 150 m in the Copper Hill Borehole; these geochemical data are given in Table 2. This interval includes the acme of *Classopollis classoides* and is within and adjacent to the range of *L. variabile* (Table 2). The materials analysed are both terrestrial and marine and thus allow a test as to whether any carbon isotope perturbation affected the global carbon cycle. Specifically, the palynomorph concentrates are dominated by terrestrial materials and the belemnite fragments represent the marine realm. A consistent negative CIE of 2–3‰ occurs in the Copper Hill Borehole (Fig. 2). The CIE as defined by the bulk sample $\delta^{13}C_{TOC}$ is broadly symmetrical, and is recorded between 187 m and 176 m. There are two separate peaks; an older one at 186 m and a younger, larger one at 179 m (Fig. 2). The negative CIE is also recorded in the $\delta^{13}C$ of palynomorph concentrates, which are overwhelmingly dominated by terrestrial material such as wood fragments, pollen and spores, and hence in part reflect isotopic changes in atmospheric $CO_2$, in addition to a potentially minor unknown contribution from changing vegetation types. The $\delta^{13}C$ values of the only five belemnites (nekto-benthic marine molluscs) recovered in this interval also show a negative CIE. The peak excursion of all three materials is temporally coincident with the acmes of *Classopollis classoides* and *L. variabile* (Fig. 2).

In contrast to the carbon isotope record, the $\delta^{18}O_{belemnite}$ values are relatively uniform (Fig. 2; Table 2) (mean $\delta^{18}O = -1.6 \pm 0.3‰$). Although the negative CIE and the coincident *Classopollis classoides* and *L. variabile* peaks are inferred here to represent a period of global warming, the lack of a clear warming signal from belemnites may represent the effect of sea level rise due to thermal expansion of sea-water, and hence locally deeper cooler water, as has recently also been suggested for the Sinemurian–Pliensbachian boundary (Korte and Hesselbo, 2011). Alternatively, water column stratification and/ or local upwelling of cool intermediate waters may have increased. It is not possible to use $TEX_{86}$ or MBT/CBT as temperature proxies because the succession studied here is too thermally mature. Indeed, most Triassic and Jurassic successions have been too thermally altered for these methods, which have been used successfully on Early Cretaceous to modern material (Kuypers et al., 2001; Schouten et al., 2002).

## 5. Palaeoecology

### 5.1. The pollen genus Classopollis

The gymnospermous pollen genus *Classopollis*, also known illegitimately as *Corollina* (see Traverse, 2004; McNeill et al., 2006, appendix 3 F), is present throughout much of the Mesozoic. It ranges from the Norian to the Turonian (Srivastava, 1976, 1978; Morbey, 1978; Helby et al., 1987). *Classopollis* pollen was produced by plants belonging to the extinct gymnosperm conifer family Cheirolepidiaceae (Francis, 1983), which were superficially similar to the extant family Cupressaceae. *Classopollis* is considered to be a reliable proxy for hot/ warm palaeoclimates (Pocock and Jansonius, 1961; Pocock, 1972; Srivastava, 1976; Vakhrameev, 1978, 1981, 1987, 1991; Volkheimer et al., 2008). The abundance of *Classopollis* in the Jurassic and Cretaceous of Russia decreased markedly northwards, i.e. towards higher

and cooler palaeolatitudes (Vakhrameev, 1981). A similar distribution pattern was observed in North America by Pocock (1972).

There has been considerable debate as to the precise palaeoecological preference of *Classopollis*-producing plants. The genus has been reported in high proportions from near-coastal settings such as sandy bars or coastal islands with well-drained soils (Batten, 1975; Srivastava, 1976; Alvin, 1982; Abbink, 1998). This distribution led Batten and MacLennan (1984) to propose that the family Cheirolepidiaceae were possibly salt marsh shrubs or trees resembling modern mangroves. However Filatoff (1975, fig. 10) and Vakhrameev and Doludenko (1977) suggested that the Cheirolepidiaceae preferred upland slope habitats, and that the abundant *Classopollis* pollen preserved in nearshore palaeoenvironments was therefore allochthonous. Furthermore, several authors, such as Vakhrameev (1970), considered the Cheirolepidiaceae to be drought-resistant, thermophilous, xerophytic shrubs and trees which resembled modern juniper bushes.

The occurrence of abundant *Classopollis* during a global warming event is somewhat inconsistent with this genus being a proxy for dry (semiarid to arid) conditions as suggested by, for example, Pocock and Jansonius (1961), Pocock (1972 and Vakhrameev 1991). This is because global warming events are normally associated with an enhanced hydrological cycle, and *Classopollis* has an extremely wide geographical distribution. Hence the conclusions of Batten and MacLennan (1984) that the parent plants of *Classopollis* were salt marsh or mangrove shrubs/trees are supported herein. However, it is possible that the Cheirolepidiaceae lived in a wide variety of habitats and that they could also survive in semiarid to arid conditions.

### 5.2. Other pollen and spores

The remaining pollen and spore floras are typical of Sinemurian terrestrial palynofloras from Europe (e.g. Srivastava, 1987). The non-*Classopollis* gymnosperm pollen is dominated by coniferous plants from the orders Pinales and Cycadales. For example *Perinopollenites elatoides* belongs to the Family Cupressaceae (cypresses). These groups are generally widely distributed, and hence are not especially palaeoecologically significant. However, *Chasmatosporites* is a possible representative of the Order Cycadales (see Pocock and Jansonius, 1969), and these forms are typical of subtropical and tropical regions today. The majority of the spores are from ferns or club mosses. These groups have extensive geographical extents today. However, *Ischyosporites variegatus* belongs to the chiefly tropical Family Schizaeaceae (see Couper, 1958). Therefore, the other pollen and spores support the warm climate signal provided by *Classopollis classoides*, and are entirely consistent with an equable, subtropical setting.

### 5.3. The dinoflagellate cyst L. variabile

The dinoflagellate cyst *L. variabile* was originally described from the Upper Sinemurian strata of southwest Germany by Drugg (1978). Despite relatively little being known concerning the tabulation style of this species, the accepted view is that *L. variabile* is the oldest known representative of the order Peridiniales (Bujak and Davies, 1983, p. 56; Evitt, 1985, p. 176, 177; Fensome et al., 1993, p. 121). Its morphology was thoroughly reviewed by Below (1987, p. 128), who noted that the tabulation of this species is not fully developed and that the size and shape of the archaeopyle differ from most other peridinialean taxa in being unusually large. Furthermore the anterior margin of the archaeopyle is geniculate and it lies near to the apex, encroaching on the apical horn (Below, 1987, fig. 68). Riding and Hubbard (1999, p. 27) interpreted *L. variabile* as a warm water species based on evidence derived from the study of a comprehensive database of Jurassic dinoflagellate cysts using principal component analysis.

*Liasidium variabile* is an excellent marker for the Upper Sinemurian of the northern hemisphere (Riding, 1987; Poulsen and Riding, 2003; Van

de Schootbrugge et al., 2005; Brittain et al., 2010). This species is especially prominent in, and characteristic of, the *O. oxynotum* ammonite Biozone (~192–191 Ma) of northwest Europe, and is confined to northwest Europe and offshore eastern Canada (Williams et al., 1990) and has not been reliably reported elsewhere (Fig. 1). Dinoflagellates have a planktonic habit, some taxa also having a benthic resting cyst phase (Taylor, 1987). Thus the apparent geographical restriction of *L. variabile* to northwest Europe and offshore eastern Canada is deemed to reflect the lack of suitable studies of the Lower Jurassic from Africa, the Middle East, South America and southern Asia, coupled with the dominance of non-marine facies in the Early Jurassic of regions such as, for example, Russia and Australia (Riding et al., 1999, 2010).

Lower Jurassic dinoflagellate cyst assemblages have low species richness, especially in the Hettangian to Lower Pliensbachian (Bujak and Williams, 1979; Woollam and Riding, 1983; Fensome et al., 1996, 1999; MacRae et al., 1996). Upper Sinemurian successions yielding *L. variabile* are frequently underlain and overlain by strata that are entirely devoid of other dinoflagellate cysts (e.g. Riding, 1987, fig. 3). Van de Schootbrugge et al. (2005, p. 87–88) discussed the highly unusual sudden appearance, abundance and disappearance of *L. variabile* in Europe and eastern Canada, and explained this stratigraphically isolated acme as being due to major palaeoceanographical change, possibly linked to the opening and flooding of the Hispanic Corridor. These authors suggested that *L. variabile* migrated into Europe and eastern Canada from western Panthalassa through the Hispanic Corridor into the western Tethys in response to changes in ocean circulation during a sea-level highstand. This contention, however, is somewhat speculative because there are no records of *L. variabile* from Central America and the eastern Pacific area (Brittain et al., 2010, p. 72). It is equally possible, however, that *L. variabile* migrated northwestwards into Europe from western Tethys. This putative northerly migration of *L. variabile* may be an indication that this species tracked warmer water northwards. This contention is consistent with the interpretation of *L. variabile* as a thermophilic species by Riding and Hubbard (1999), and certain similarities between *L. variabile* and the peridinioid genus *Apectodinium*.

Feist-Burkhardt (2009) suggested that the acme of *L. variabile* in the Late Sinemurian is similar to the abundant occurrence of *Apectodinium* at the PETM, and also demonstrated that some specimens of *L. variabile* in unoxidised material exhibit dark brown-pigmentation with low autofluorescence intensities (Fig. 5). No pigmented forms were found in the Copper Hill Borehole material, but these samples were all oxidised using nitric acid during the preparation procedure which may have had a bleaching effect. Extant pigmented dinoflagellate cysts, such as produced by *Archaeperidinium*, *Congruentidium* and *Protoperidinium*, with little or no autofluorescence are all heterotrophic (Head, 1994, 1996; Brenner and Biebow, 2001). These characteristics led Feist-Burkhardt (2009) to conclude that *L. variabile* may have been produced by a heterotrophic and thermophilic dinoflagellate by analogy with extant floras, supposedly like the (consistently non-pigmented) genus *Apectodinium*. However, the peridinioid affinity of *L. variabile* is not prima facie evidence of this species being indicative of elevated sea surface temperatures; as with *Apectodinium*, palaeotemperature is one of several parameters which controlled its distribution.

There are, however, possible alternatives to *L. variabile* being a warm-water species. For example, a rise in sea level may have caused salinity changes that led to the influx of *L. variabile* during the Late Sinemurian. Alternatively, upwelling cold oceanic waters may have introduced an influx of nutrients which caused the acme of *L. variabile* at this time. The low-resolution but apparently invariant levels of $\delta^{18}O$ throughout the succession studied (Fig. 2; Korte and Hesselbo, 2011) would be consistent with this hypothesis. Despite these caveats, multivariate statistical evidence based on principal component analysis of the thermophilic nature of this species developed by Riding and Hubbard (1999), together with the co-occurrence of

abundant *Classopollis classoides* and the negative CIE, makes a compelling case for *L. variabile* being thermophilic. Hence this independent evidence for thermopyly in *L. variabile*, and the fact that the evidence for heterotrophy in *L. variabile* is sporadic pigmentation, means that the analogy to *Apectodinium* is not as strong as suggested by Feist-Burkhardt (2009).

### 5.4. The dinoflagellate cyst genus Apectodinium

The comparison of *L. variabile* to the Paleogene genus *Apectodinium* by Feist-Burkhardt (2009) mentioned above in Section 5.3 is an important one in the context of this contribution. Both these peridinialean taxa have relatively transient acmes, and *Apectodinium* has been considered to be both heterotrophic and thermophilic (e.g. Crouch et al., 2003b). It should however be emphasised that, other than a shared peridinalean affinity and therefore a probable heterotrophic feeding strategy, similarities between these taxa are relatively few. This is especially the case in terms of morphology. *Apectodinium* has a subpentagonal outline, exhibits a quadra-style archaeopyle, typically bears five horns, is cornucavate and is consistently unpigmented (Costa and Downie, 1976, 1979; Harland, 1979). By contrast, *L. variabile* is biconical, normally with two horns, acavate, has a large, heptagonal anterior intercalary (type I) archaeopyle and is occasionally pigmented (Drugg, 1978; Bucefalo Palliani and Riding, 2000, Figs. 5, 6; Feist-Burkhardt, 2009). The similarities between *L. variabile* and *Apectodinium* are mostly autoecological, i.e. are related to their similar geological settings. Both these taxa probably exploited rising sea levels and warming climates that would have caused the widespread development of water-column stratification on the continental shelves. These conditions may have caused algal blooms, thereby providing a food source for the heterotrophic motile dinoflagellate stages.

The palaeoecological preference of *Apectodinium* has recently been extensively debated. It is known as a neritic genus that thrived in estuarine settings (Crouch et al., 2003a; Sluijs et al., 2005). At low-mid latitudes, *Apectodinium* typically comprises >40% of the dinoflagellate cyst associations during the PETM (e.g. Crouch et al., 2003b; Sluijs et al., 2007; Kender et al., 2012). By contrast, *Apectodinium* is less common (~20%) in the Arctic, where low salinity dinoflagellate cysts

increase (Sluijs et al., 2006; Harding et al., 2011). Several studies have concluded that, due to the low latitude preference of *Apectodinium*, it must have tracked warm water northwards during the PETM (e.g. Crouch et al., 2003a). However, Canonical Correspondence Analysis (CCA) data from the New Jersey Shelf indicates that *Apectodinium* has only a weak correlation with palaeotemperature based on $TEX_{86}$ (Sluijs and Brinkhuis, 2009, p. 1791). Furthermore, Sluijs et al. (2009) found that *Apectodinium* is generally absent throughout another hyperthermal event, the Eocene thermal maximum 2 (ETM2). Recently, Sluijs et al. (2011, fig. 4) reported the main peak in abundance of *Apectodinium* occurred well before the PETM at ODP Site 1172 on the East Tasman Plateau. The principal acme of *Apectodinium* at this locality attained >60% of the assemblage at ~0.8 m below the thermal peak of the PETM. Following a decrease to around 2% there was a later peak of *Apectodinium*, attaining ~30% of the assemblage, within the PETM (Sluijs et al., 2011, fig. 3). The fact that the older, more significant, acme of *Apectodinium* on the East Tasman Plateau is associated with normal sea surface temperatures (~26 °C) clearly indicates that this genus responded to variables other than simply temperature. For example, abundant *Apectodinium* has been interpreted as being indicative of elevated nutrient levels, water stratification and low salinities by Crouch et al. (2003b), Sluijs et al. (2008) and Kender et al. (2012). This is also consistent with a warming world and elevated fluvial nutrient runoff from an associated enhanced hydrological cycle with greater precipitation in mid-high latitudes (Pagani et al., 2006). In conclusion, *Apectodinium* is normally abundant during the PETM, but temperature is only one of several factors which controlled its abundance.

Several authors have stated that the genus *Apectodinium* was probably heterotrophic and fed on other plankton and/or organic detritus (e.g. Brinkhuis et al., 1992, 1994; Crouch et al., 2003b; Sluijs et al., 2005). This interpretation of trophic mode is based on the morphological similarity of *Apectodinium* with modern cysts of the unequivocally heterotrophic genus *Protoperidinium* (subfamily Protoperidinioideae). The vast majority of living protoperidiniacean dinoflagellates (including the diplopsalioideans) produces pigmented cysts, thus indicating that pigmentation is connected to heterotrophy in this group of dinoflagellates (Head, 1994, 1996; Rochon et al., 1999; Brenner and Biebow, 2001). However, *Apectodinium* belongs to the subfamily Wetzelielloideae (see Fensome et al., 1993). *Apectodinium* and cysts of *Protoperidinium* are hence both members of the suborder Peridiniineae, but there are substantial familial and subfamilial differences which are based on significant differences in tabulation style. The subfamily Wetzelielloideae has, for example, a quadra-style 2a plate, a stable epicystal tabulation and a fully subdivided cingulum. In contrast, cysts of the subfamily Protoperidinioideae are characterised by a highly variable cinctioid or bipesioid epicystal tabulation and a largely undivided cingulum (Fensome et al., 1993). The subpentagonal outline of *Apectodinium* is not a primary characteristic of the subfamily Protoperidinioideae (see Fensome et al., 1993, fig. 142). Hence, the evidence for heterotrophy in *Apectodinium*, based on similarities to cysts of *Protoperidinium* at the suborder level, appears to be purely circumstantial.

Several studies from marginal marine environments suggest an elevation in levels of marine nutrients in shelf seas, possibly in response to an intensification of increased weathering, terrestrial runoff and/or water column stratification during the PETM (e.g. Crouch et al., 2003b; Sluijs et al., 2006, 2007; Sluijs and Brinkhuis, 2009; Harding et al., 2011; Kender et al., 2012). It is highly likely therefore that the abundance of *Apectodinium* at or around the PETM is an indirect response to higher global palaeotemperatures and/or eutrophication.

## 6. Similarities of the S-CIE to other environmental change events

The temporal coincidence of the highly restricted thermophilic/neritic *L. variabile* and the thermophilic *Classopollis* acmes in the
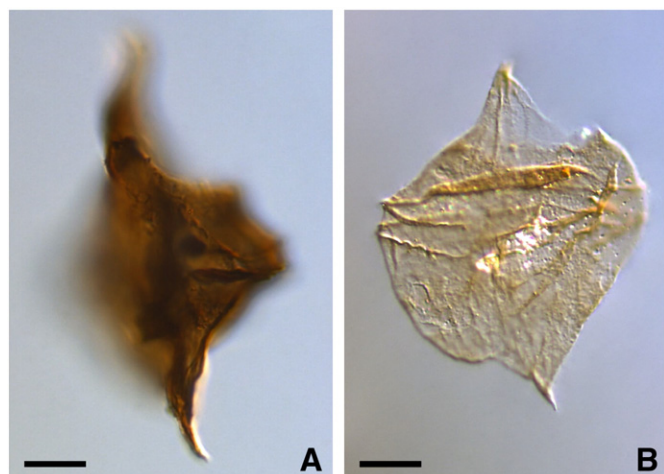


**Fig. 5.** Transmitted light photomicrographs of *Liasidium variabile* from the Upper Sinemurian Obtususton Formation (*Oxynoticeras oxynotum* ammonite Biozone) of Aselfingen, southwest Germany. A—a specimen of *L. variabile* in unoxidised material which has dark brown-pigmentation and a low autofluorescence intensity. Slide SF902-oc-1, England Finder reference L59/3. B—an unpigmented specimen with high autofluorescence intensity, similar to the other palynomorphs in this sample. Sample SF901, slide NHM/10/03-10, England Finder reference O35/1-2. Both specimens are from Feist-Burkhardt (2009, figs. 5H and 5A respectively) and the scale bars represent 10 μm.
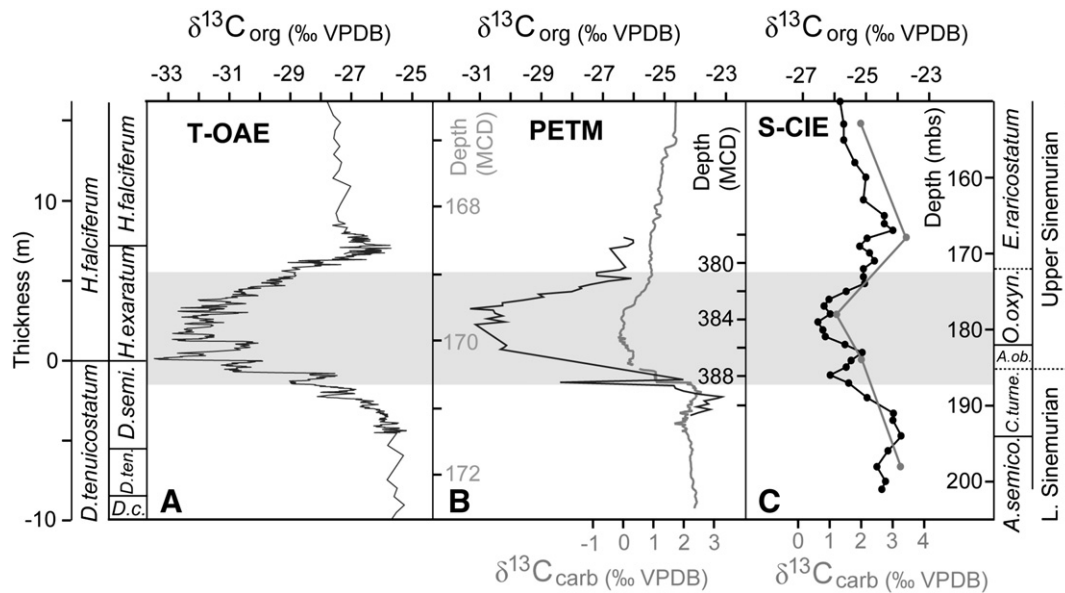
**Fig. 6.** Comparisons of the $\delta^{13}C_{org}$ profiles (black lines) of the S-CIE with the T-OAE and the PETM, and comparison of the $\delta^{13}C_{carb}$ profiles (grey lines) of the S-CIE with the PETM. A—carbon isotope record from the T-OAE of the North Yorkshire coast, United Kingdom (summarised in Cohen et al., 2007). B—carbon isotope records from the PETM of the Arctic Ocean (black), and the Southern Ocean (grey) (summarised in Cohen et al., 2007). C—carbon isotope records from the S-CIE of Lincolnshire, United Kingdom (this study). The light grey shading indicates the extent of the three CIEs.

Copper Hill Borehole, with a marked negative CIE (Fig. 2), suggests that the inferred environmental changes were associated with a perturbation to the carbon cycle. This may have been global in nature due to isotope excursions in both marine and terrestrial materials. Furthermore, the palynological records appear to be at least regional in extent. Van de Schootbrugge et al. (2005, table 3) also recorded coincident abundances of *Classopollis* spp. and *L. variabile* in the Late Sinemurian strata of the Mochras Borehole, northwest Wales. Our new records are similar in some respects to other Mesozoic and Cenozoic hyperthermal events such as the T-OAE and the PETM (Cohen et al., 2007), which exhibit negative CIEs coeval with global warming and major changes in marine palynomorph assemblages. Furthermore, the S-CIE association with the peridiniacean dinoflagellate cyst *L. variabile* exhibits a striking similarity to the PETM. The latter hyperthermal event, occurring ~56 Ma ago, is characterised by a negative CIE and abundant levels of the peridiniacean genus *Apectodinium* (e.g. Crouch et al., 2003b). During the PETM, the proportion of isotopically-light carbon in the atmosphere and oceans rapidly increased, hypothetically due to the catastrophic destabilisation of methane clathrate, magmatic heating of organic-rich sediments or direct volcanic outgassing (e.g. Dickens, 2011). The PETM is a particularly well-documented hyperthermal event ~56 Ma ago, that lasted approximately 120–220 ka (Röhl et al., 2007; Murphy et al., 2010). The duration of the S-CIE may have been a few hundred thousand years, which is therefore comparable to the PETM and the T-OAE (see above, Section 2).

The negative CIEs at the T-OAE, OAE 1a and the PETM are similar in form to that of the S-CIE (Fig. 6; Gröcke et al., 1999; Mort et al., 2007; Kuhnt et al., 2011). The 2–3‰ magnitude of the S-CIE is also similar to that of carbonate records of OAE 1a (e.g. Mort et al., 2007; Jenkyns, 2010). However, the magnitude of the S-CIE is smaller than that of the organic PETM (~5‰; summary in McInerney and Wing, 2011) and many T-OAE records (6‰; e.g. Hesselbo et al., 2007; Hermoso et al., 2009), suggesting either a smaller volume of injected carbon and/or a heavier isotopic value. The pulsed onset of the S-CIE is similar to those of the T-OAE, OAE 1a, and PETM (Fig. 6). However the S-CIE is of a much longer duration (~1 Ma) at the Early–Late Sinemurian transition in the Copper Hill Borehole, being perhaps therefore closer in character to the excursion at the Pliensbachian–

Toarcian boundary followed by the T-OAE (Hesselbo et al., 2007). Specifically, the $\delta^{13}C$ (bulk) record of the S-CIE shows an initial isolated spike at 186 m in the upper *Caenisites turneri* and *Asteroceras obtusum* ammonite biozones, followed by a partial $\delta^{13}C$ recovery towards pre-S-CIE levels, before the main part of the S-CIE at 179 m (Fig. 2). It is possible that the initial onset of the S-CIE at 190 m represents an early phase of warming that was temporarily reversed, either by a reduction in carbon injection and/or by negative feedbacks. It is noteworthy that the range base of *L. variabile* and the onset of the *Classopollis classoides* acme are coincident with the largest (second) negative pulse at ~180 m (Fig. 2), which we therefore infer to represent the largest environmental shift. In the Copper Hill Borehole, the onset of this main phase of the S-CIE coincides with the Glebe Farm Bed, an erosional horizon containing a concentration of bored carbonate nodules (Fig. 2; Brandon et al., 1990) that also represents the base of the *O. oxynotum* ammonite Biozone. Thus the record reported here is interrupted by a hiatus and is unlikely to represent the full extent of this palaeoenvironmental perturbation.

The aforementioned Toarcian, Aptian and Paleogene CIE events have been hypothesised to be due to injections of isotopically-light carbon into the ocean–atmosphere system (e.g. Dickens et al., 1997; Hesselbo et al., 2000), and this is likely also to be the case for the S-CIE. Another possible cause of a negative carbon excursion could be a contracting biosphere due to cooling, but this would be evident as major changes in biomes and very extensive glaciation for which there is no evidence at this particular time (Sellwood and Valdes, 2008; Korte and Hesselbo, 2011). The coincident and transient nature of the range and acme of *L. variabile*, the acme of *Classopollis classoides* and the S-CIE is highly suggestive of a global warming event associated with carbon injection event.

There is as yet no known major magmatic/volcanic event, or previously reported indication of significant global warmth, during the Late Sinemurian. Furthermore, there are no known planktonic and/or vegetational changes in the Late Sinemurian that could have significantly affected the bulk carbon isotope signal. However, there is some evidence for a sea-level rise of at least regional scale over this interval (Haq et al., 1988), which is consistent with a period of global warmth. The upper part of the *A. obtusum* and the lower part of the *O. oxynotum* ammonite biozones represent an interval of shale

deposition associated with relative sea-level rise in the Cleveland Basin, northern England (Hesselbo and Jenkyns, 1998; Van Buchem and Knox, 1998). Similarly, the lithology of Late Sinemurian age strata on Skye, in the Hebrides Basin, northwest Scotland, closely approaches black (i.e. organic-rich) shale in the *O. oxynotum* ammonite Biozone (Hesselbo et al., 1998) and Brittain et al. (2010) reported *L. variabile* from coeval strata on Raasay in the Inner Hebrides. As well as indicating relative sea-level rise in this example, organic-rich shale is typical of deposition in stratified marine waters during periods of global warming and high nutrient levels such as the T-OAE, OAE 1a and PETM. The Late Sinemurian is represented by hiatus concretions and an omission surface on the Dorset coast in the Wessex Basin, southern England (Coe and Hesselbo, 2000), interpreted by Haq et al. (1988) and Hesselbo and Palmer (1992) to have resulted from bioerosion during a period of regional sediment starvation caused by sea-level rise (contra Hallam, 1999). The Glebe Farm Bed in the Copper Hill Borehole (Fig. 2), overlain by mudstone, grading upwards to sandstone, probably represents the same condensation phenomenon brought about by sea-level rise and deepening (upper *A. obtusum* and lower *O. oxynotum* ammonite biozones), followed by progradation and shallowing (upper *O. oxynotum* ammonite Biozone) (Hesselbo, 2008).

## 7. Conclusions

This study demonstrates that a significant abundance of the thermophilic pollen *Classopollis classoides* and the range and acme of the thermophilic/eutrophic dinoflagellate cyst *L. variabile* are linked with a marked coincident negative CIE (herein termed the S-CIE) in the Late Sinemurian *O. oxynotum* ammonite Biozone (~194 Ma) of eastern England (Fig. 2). The S-CIE reflects a perturbation to the whole exchangeable carbon reservoir because terrestrially-derived organic material (the palynomorph concentrate) and marine carbonate exhibit the same excursion (Fig. 2). The relatively brief record of *L. variabile* together with the transient acme of *Classopollis classoides* and the negative S-CIE is highly suggestive of a major climate-warming-driven biogeochemical perturbation, such as a carbon injection event, at ~194 Ma, and indicates that Jurassic climates were relatively susceptible to significant and transient change. The level of the S-CIE can be confidently delineated by the range and acme of *L. variabile*. The S-CIE shares many characteristics with the global negative CIE event at the PETM, a phase of abnormally high global temperatures coincident with a peak in the apparently thermophilic/eutrophic dinoflagellate cyst genus *Apectodinium*. The S-CIE also shares similarities to negative carbon isotope excursions at the T-OAE and OAE 1a, which are other major phases of global warming.

The S-CIE has so far only been recorded from a single succession in eastern England. However, the magnitude and remarkable synchronicity of the geochemical and palynological data from the Copper Hill Borehole indicate that the S-CIE probably represents a significant environmental change event. Eastern England is known to have been within an extensive, open seaway during the Early Jurassic (Coward et al., 2003), and hence may be representative of a large part of the Earth system at this time. It is considered that the marine and terrestrial geochemical and palynological signals documented here are not attributable to local factors, nor are they a result of background noise or oscillations in the carbon cycle. However it is fully acknowledged that more sections need to be studied to test our hypothesis. These should include more complete successions and those from localities outside Europe in order to verify this potentially important, apparently global, palaeoenvironmental change event.

## Acknowledgements

## Appendix 1

This appendix lists all valid palynomorph taxa below generic level which are mentioned in this contribution with full author citations. The palynomorphs are listed alphabetically within three groups.

**Dinoflagellate cyst**:

*Liasidium variabile* Drugg, 1978

**Gymnospermous pollen**:

*Araucariacites australis* Cookson, 1947
*Cerebropollenites macroverrucosus* (Thiergart, 1949) Schulz, 1967
*Classopollis classoides* (Pflug, 1953) Pocock and Jansonius, 1961
*Perinopollenites elatoides* Couper, 1958
*Vitreisporites pallidus* (Reissinger, 1950) Nilsson, 1958

**Pteridophyte spores**:

*Auritulinasporites scanicus* Nilsson, 1958
*Cibotiumspora juriensis* (Balme, 1957) Filatoff, 1975
*Concavissimisporites verrucosus* Delcourt and Sprumont, 1955
*Coronatispora valdensis* (Couper, 1958) Dettmann, 1963
*Cyathidites australis* Couper, 1953
*Cyathidites minor* Couper, 1953
*Gleicheniidites senonicus* Ross, 1949
*Ischyosporites variegatus* (Couper, 1958) Schulz, 1967
*Kraeuselisporites reissingeri* (Harris, 1957) Morbey, 1975
*Leptolepidites rotundus* Tralau, 1968
*Lycospora pusilla* (Ibrahim, 1932) Schopf et al., 1944
*Neoraistrickia gristhorpensis* (Couper, 1958) Tralau, 1968
*Osmundacidites wellmanii* Couper, 1958
*Retitriletes austroclavatidites* (Cookson, 1953) Döring et al., 1963
*Sestrosporites pseudoalveolatus* (Couper, 1958) Dettmann, 1963
*Todisporites major* Couper, 1958
*Todisporites minor* Couper, 1958
*Tripartites vetustus* Schemel, 1950
*Uvaesporites argenteaeformis* (Bolkhovitina, 1953) Schulz, 1967

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