Abstract

The generally arid Late Triassic climate was interrupted by a wet phase during the mid Carnian termed Carnian Pluvial Episode (CPE). Quantitative palynological data from the Mercia Mudstone Group in the Wessex Basin (UK), reveals vegetation changes and palaeoclimate trends. Palynostratigraphy and bulk organic carbon isotope data enable
correlation to other Carnian successions. The palynostratigraphy indicates that the
Dunscombe Mudstone is Julian and the lowest part of the overlying Branscombe Mudstone
Formation is Tuvalian. The Aulisporites acme characterizing the CPE in Tethyan successions
and the Germanic Basin, is missing in the UK. The quantitative palynological record suggests
the predominance of xerophyte floral elements with a few horizons of increased hygrophytes.
A humidity signal is not seen due to the dry climate in central Pangea. Secondly, the signal
might be masked by the overrepresentation of xerophyte regional pollen and the
predominance of xerophyte hinterland flora. The bias towards regional pollen rain is enhanced
by the potential increase in continental runoff related to seasonally humid conditions and
differences in pollen production rates and transport mechanisms. The vegetation of British
CPE successions suggests a more complex climate history during the Carnian indicating that
the CPE is not recognized by the same changes everywhere.

**Supplementary material:** Detailed lithological log of the Strangman’s Cove (Devon)
outcrop with the description of the MMG lithostratigraphical units, description of the
laboratory techniques, seven photoplates with selected spores and pollen grains, a list of all
identified palynomorphs, Excel sheets with the palynological and palynofacies counts, bulk
organic carbon isotope ratios and TOC values of the Strangman’s Cove outcrop are available
at https://doi.org/xxxx
During the Late Triassic the continental interior of Pangea was characterized by predominantly arid climates in low-mid latitudes with strong seasonality and a monsoonal regime (Kutzbach & Gallimore 1989; Parrish 1993; Sellwood & Valdes 2006; Preto et al. 2010). In central Pangea (Preto et al. 2010), the depositional setting of the Mid- to Late Triassic Mercia Mudstone Group (MMG) in the Wessex Basin (SW England, UK) was marked by playa lake deposits with red beds and local evaporites similar to the Keuper Group of the Southern Permian (or Germanic) Basin (Bachman et al. 2010; Hounslow et al. 2012). During the Carnian, the depositional style of the Mercia Mudstone Group changed significantly, marked by short-lived lacustrine interval and more pronounced fluvial influence (Simms & Ruffell 1989, 1990; Porter & Gallois 2008; Hounslow & Ruffell 2006; Ruffell et al. 2016). In the Wessex Basin, this transition is manifested by the lithological change from gypsiferous red mudstone to a green-grey mudstone with locally sandy beds: termed the Dunscombe Mudstone Formation (Porter & Gallois 2008; Ruffell et al. 2016). This facies change is also expressed in the Keuper Group in the Germanic Basin, where red playa lake deposits were temporarily interrupted by sandy fluvial channel and overbank deposits of the Schilfsandstein (Stuttgart Formation) during the late Julian (e.g., Bachmann et al. 2010; Kozur & Bachmann 2010; Shukla et al. 2010). The lithological shift seen at the base of the Dunscombe Mudstone Formation and in the Schilfsandstein may be coincident with a climate change towards more humid conditions known as the Carnian Pluvial Episode (CPE, Ruffell et al. 2016), which in other localities, appears to have commenced in the early Carnian, close to the early to late Julian boundary (Dal Corso et al. 2015; Mueller et al. 2016a, b). The global nature of the CPE has been debated, but evidence from successions in Europe (e.g., Schlager & Schöllnberger 1974; Simms & Ruffell 1989, 1990), the Middle East (Bialik et al. 2013), Iberia and eastern North America (Arche & López-Gómez 2014) and also Asia (Hornung et al. 2007 a, b; Nakada et al. 2014; Sun et al. 2016), suggest that the CPE wet
conditions had global extent (Ogg 2015; Ruffell et al. 2016). Previously Visscher et al. (1994) had rejected the presence of a wetter climatic phase during the Carnian based on palynological evidence from the Schilfsandstein explaining the facies change by the establishment of a large river system in an overall dry floodplain, but with locally wet environments near the river banks; with the present-day Nile Valley as an analogue. Despite differing local interpretations, major environmental change is evident from the switching in lithology occurring in both continental and marine Carnian successions. The CPE was probably accompanied by sea level changes, global warming (Trotter et al. 2015; Sun et al. 2016) increased continental weathering (Rostási et al. 2011), demise of carbonate platforms (Keim et al. 2006; Breda et al. 2009; Lukeneder et al. 2012; Arche & López-Gómez 2014) and deepening of the carbonate compensation depth in the oceans (Rigo et al. 2007; Lukeneder et al. 2012; Nakada et al. 2014). At low palaeolatitudes enhanced terrigenous input appears to have lasted from the late Julian (Julian 2) to the early Tuvalian (Roghi et al. 2010, Rostási et al. 2011). The CPE is characterized by wet-dry cycles and multiple humid pulses; before the climate returned to persistent aridity in the late Carnian or Norian (Preto et al. 2010; Lukeneder et al. 2012; Bialik et al. 2013; Mueller et al. 2016b; López-Gómez et al. 2017). At high latitudes (in the Boreal Realm, e.g Svalbard) higher temperatures, increased humidity and the local development of coals represent the equivalent of the CPE (Mueller et al. 2016a).

In the marine realm the onset of the CPE in the mid Julian is associated with a negative carbon isotope excursion in the marine realm suggesting the injection of a significant amount of $^{13}$C-depleted CO$_2$ into the atmosphere (Dal Corso et al. 2012, 2015). The first evidence of a negative carbon isotope excursion in the CPE from a terrestrial realm has been provided by the Wiscombe Park-1 Borehole succession in the MMG from the Wessex Basin (Miller et al. 2017). There, the initial carbon isotope excursion (labelled IIE) in both total organic carbon
and plant leaf waxes (Miller et al. 2017), is followed by four other negative C-isotope excursions. Miller et al. (2017) recognised ca. 413 ka eccentricity cycles from the Dunscombe Mudstone Formation suggesting the establishment of a duration for the isotope excursions, lasting for ca. 1.09 Ma; comparable to previous estimates of 0.8–1.2 Ma from marine units in China (Zhang et al. 2015). The release of isotopically lighter CO$_2$ into the atmosphere caused the intensification of Pangean monsoon activity and is the most likely responsible for the increase in rainfall (Parrish 1993). The origin of $^{13}$C-depleted CO$_2$ and the carbon isotope excursion may be linked to enhanced volcanic activity and associated feedbacks (warming, dissociation of methane clathrates, reduction in marine primary productivity) (Simms et al. 1995; Hornung et al. 2007a, b). The emplacement of the Wrangellia Large Igneous Province basalts is considered the most likely trigger of the CPE (Furin et al. 2006, Dal Corso et al. 2012) although Greene et al. (2010) and Xu et al. (2014) showed that the Wrangellia eruptions started earlier than the Carnian. There is evidence for regionally widespread contemporaneous exhalative, acidic volcanic activity from the Anatolian Terrane in Greece and Turkey (Huglu-Pindos Series: Moix et al. [2008, 2013]), Apennines (Furin et al. 2006) and Iberia (e.g., Arche & López-Gómez 2014), but the extent of this volcanic activity is less voluminous compared to the Wrangellia LIP.

Changes in plant communities are good proxies for terrestrial climate, therefore palaeobotany and palynology have been widely utilized in understanding the climate change during the Late Triassic (e.g., Reitz 1985; Visscher et al. 1994; Roghi 2004, Pott et al. 2008; Roghi et al. 2010; Bonis & Kürschner 2012; Mueller et al. 2016a, b). Palynology of sediments from the CPE typically show a shift towards hygrophyte vegetation with increased abundance of ferns, equisetalans and cycadaleans (Roghi 2004; Hochuli & Vigran 2010; Roghi et al. 2010; Mueller et al. 2016a, b). Particularly distinctive is the widely distributed Aulisporites astigmosus assemblage typical of the late Julian in the western Tethys (Roghi et al. 2010).
Aulisporites astigmosus is a hygrophyte vegetation element and its widespread distribution during the Carnian is a crucial argument for the global scale of a wet phase (Roghi et al. 2010). However, the palaeoclimatic significance of the Aulisporites acme during the Carnian is controversial (Visscher et al. 1994) and some recent studies indicate that it may be diachronous (Mueller et al. 2016a).

Here we provide quantitative palynological data from four terrestrial successions of the Dunscombe Mudstone Formation from the UK and interpret the vegetation changes and palaeoclimatic trends. The palynological data are further integrated with organic carbon isotope stratigraphy and are compared to other Carnian successions in Europe, in order to evaluate regional differences.

**Geological setting**

Thick (up to 1 km, usually ca. 450 m) packages of fluvial-lacustrine sediments of the Mercia Mudstone Group accumulated in SW England during the Mid-Late Triassic, in fault-bounded basins that were formed during the syn-rift phase of crustal extension as a consequence of Pangean rifting and thermal relaxation (Ruffell & Shelton 1999; Howard et al. 2008; McKie & Williams 2009; Hounslow et al. 2012; Fig. 1). In the Wessex Basin, the Mercia Mudstone Group comprises ca. 450 m of predominantly red mudstones and local evaporites that indicate deposition in a low-relief sabkha environment in a hot desert (Gallois & Porter 2006; Hounslow & Ruffell 2006; Hounslow et al. 2012; Fig. 1). The MMG in central and southern England has been nationally rationalised into the Sidmouth Mudstone Formation, Arden Sandstone Formation, Branscombe Mudstone Formation and the Blue Anchor Formation (Howard et al. 2008 (Fig. 1, detailed description of each unit can be found in the Supplementary Data). In the Wessex Basin, Dunscombe Mudstone Formation distinguishes a variation of the mid parts of the MMG with predominantly green, grey to purple mudstone unit between the red mudstones of the under-and overlying Sidmouth and Branscombe
Mudstone formations (Porter & Gallois 2008). In the coastal sections in Devon, the Arden Sandstone Fm of Howard et al. (2008) occupies only the mid 24 m of the Dunscombe Mudstone Fm (Warrington 2004), so we prefer to use the formation designations defined used by Gallois (2001), Gallois & Porter (2006) and Porter & Gallois (2008), which better represent the local lithological change. The Dunscombe Mudstone Formation consists of a 25 to 43 m-thick laterally variable succession of green, purple and grey laminated mudstones interbedded in the lower part with thin (typically, 5 cm to 30 cm) calcareous siltstone/sandstone beds. Breccia beds, caused by gypsum/anhydrite and halite collapse are also present (Gallois 2003; Gallois & Porter 2006; Porter & Gallois 2008). Palaeosols, bioturbated horizons, hardgrounds and lag deposits are evidence of condensation and minor hiatuses through the DMF (Gallois & Porter 2006). The DMF represents a fluvial–lacustrine succession with shallow freshwater lakes fed by shallow distributary channels in low-relief topography (Gallois & Porter 2006; Porter & Gallois 2008). On the Devon coast in the lower part of the formation, a lenticular unit of calcareous siltstones and fine-grained sandstones, (the Lincombe Member) is geographically restricted, but evidence of an oxygenated freshwater lake (Gallois & Porter 2006).

The palynomorph assemblages of the Dunscombe Mudstone Fm and Arden Sandstone formations were reconnaissance sampled by Clarke (1965), Warrington (1967, 1970, 1971), Fisher (1972), Warrington (1974, 1984), Warrington & Williams (1984), Warrington (1997) and Kousis (2015). The previous studies of the Arden Sandstone Fm have suggested a late Carnian (Tuvalian) age (e.g., Warrington et al. 1980; Barclay et al. 1997). The only work that has systematically sampled the DMF in the Devon coastal area using the logs of Jeans (1978) is that of Fisher (1985) who recognized an older and a younger Carnian assemblage. In contrast, based on data from the nearby Wiscombe Park-1 Borehole, Miller et al. (2017) assigned a Julian age to the DMF. Kozur (in Gallois & Porter 2006) found the conchostracan...
Laxitexella multireticulata within the Lincombe Member in the coastal sections which is indicative of the early Carnian L. mucroreticulata Zone (Kozur & Weems 2010). This species is common in the early Carnian ‘Estheria Beds’ (in the upper Grabfeld Formation in Germany), but is also reported from the Schilfsandstein in Germany. Hounslow et al. (2002) suggested that the Ladinian-Carnian boundary may be in the upper part of the Sidmouth Mudstone Formation and the Carnian-Norian boundary near the boundary between the DMF and overlying Branscombe Mudstone Formation, however, this was based on hitherto unpublished magnetostratigraphic data. The DMF crops out along coastal cliffs between Higher Dunscombe Cliff and Strangman’s Cove and has a wide distribution in the sub-surface of the Wessex Basin (Gallois & Porter 2006; Porter & Gallois 2008) (Fig. 1). The DMF was cored in two boreholes Wiscombe Park 1 and 2 (WP-1 and 2) about 5 km north of the coastal outcrop (Porter & Gallois 2008) (Fig. 1). Further north in Somerset, several lenticular sandstone units of the Arden Sandstone Fm can be found around Taunton, North Curry and Sutton Mallet areas (Fig. 1), similar to the Lincombe Member of the Dunscombe Mudstone Fm. However, the stratigraphic correlation of these arenaceous units remains uncertain (Gallois 2003; Gallois & Porter 2006). On lithostratigraphical and palaeoenvironmental grounds, Howard et al. (2008) argued that the sandstone bodies in Somerset are contemporaneous with the Lincombe Member, but Gallois (2001) suggested they might occur in a higher stratigraphic position within the DMF compared to the Lincombe Member.

**Methods**

**Palynology**

Palynological samples were taken from four locations: the Strangman’s Cove outcrop, the Wiscombe Park-1 Borehole section in Devon and Sutton Mallet and Lipe Hill outcrops in Somerset. Detailed description of the sample locations can be found in the Supplementary Data. A detailed lithological log for the WP-1 and Strangman’s Cove sections is provided in
the Supplementary Data (Fig. S1) with the exact position of the palynological samples. A total of 104 samples were processed for palynological analysis from the studied sections and cores. The palynological preparation follows standard procedures Wood et al. (1996) and as described in Kuerschner et al. (2007). Details of the preparation technique can be found in the Supplementary Data. The palynological slides and organic residues are stored at the Department of Geosciences, University of Oslo, Norway. In each sample, ca. 300 terrestrial palynomorphs (spores and pollen) were determined after scanning 2-4 slides (quantitative analysis). After encountering at least 300 terrestrial taxa all remaining slides were scanned for rare taxa (qualitative analysis). Lycopodium, undetermined palynomorphs and aquatic palynomorphs were counted concomitantly but excluded from the palynomorph sum. Identification of the taxa is based on the works of Klaus (1960), Clarke (1965), Schulz (1967), Scheuring (1970, 1978), Fisher (1972), Roghi (2004), Planderová (1972, 1980), Van der Eem (1983), Hochuli & Frank (2000), Mehdi et al. (2009), Fijalkowska-Mader et al. (2015), Paterson et al. (2016). Spore coloration index (SCI) values follow those of Batten (2002). Relative palynomorph abundances were calculated and plotted using the Tilia/TiliaGraph computer program (Grimm, 1991–2001). Palynomorph assemblages at Strangman’s Cove and Lipe Hill were distinguished by stratigraphically constrained cluster analysis using CONISS within Tilia (Grimm 1987). For plotting this diagram, the counted abundance data of all identified taxa were used, but unidentified forms and aquatics were excluded from the cluster analysis. For palynofacies analysis different types of sedimentary organic matter (SOM) particles were distinguished in the samples. Approximately 300 SOM particles were counted in each sample. The subdivision of the different groups and terminology follows Oboh-Ikuenobe & de Villiers (2003). The complete palynofacies and palynomorph raw data set is provided in the Supplementary Data (Table S3-S9).
Palaeoecological affinity of palynomorphs and palynomorph source areas

The classification of dispersed palynomorphs as hygrophyte and xerophytes (Table S1, Supplementary Data) is based on the concept of Visscher & Van der Zwan (1981), taking into consideration the known or supposed ecological preferences of the parent plant. It can be regarded as the first approximation of a climatic signal, yet it should be applied with caution as the exact botanical affinity and ecological needs of many Mesozoic dispersed spore and pollen are uncertain. All spores identified in this study are assigned to the hygrophyte group.

The elements of Alisporites spp. are considered to be transitional elements in the sense of Visscher & Van der Zwan (1981), but some workers attribute them to a hygrophyte affinity (e.g., Roghi et al. 2010; Whiteside et al. 2015, Mueller et al. 2016a). All other bisaccate pollen, monosaccate pollen and the members of the Circumpolles group are assigned to the xerophyte group (e.g., Hochuli & Vigran 2010; Roghi et al. 2010; Mueller et al. 2016a, b).

Cycadopites sp. and Aulisporites astigmosus are generally assigned to hygrophyte pollen (Roghi 2004; Roghi et al. 2010).

Unlike plant megafossils which represent predominantly local vegetation, the palynological assemblages record plant communities of different habitats, as well as local and regional vegetation types (Jacobson & Bradshaw 1981; Demko et al. 1998; Kustatscher et al. 2012). Elements of the local pollen rain originate within a distance of ca. 20 m from the sampling site in the sense of Jacobson & Bradshaw (1981) (Table S1). Extra local elements grow between 20 m and several hundred metres of the sampling site and regional pollen derives from plants at greater distance (Jacobson & Bradshaw 1981). To assign the dispersed palynomorphs to local or regional elements, the habitat of the parent plant and the transport mechanisms characteristic for palynomorph types have to be considered. These environmental parameters form the basis of the sporomorph ecogroup model (SEG) method of Abbink et al. (2004) for assigning the pollen and spores to various habitats. The SEG is an ecological
model which groups dispersed palynomorphs into different habitats based on the known or presumed parent plants (Abbink et al. 2004). The original SEG method was established for Jurassic–Cretaceous assemblages, but several workers have applied the method to Triassic palynomorph assemblages (Ruckwied et al. 2008; Götz et al. 2009, 2011; Kustatscher et al. 2012; Paterson et al. 2016). In fully terrestrial settings four SEGs (habitats) are defined: river SEG, dry and wet lowland SEG and upland/hinterland SEG. The upland/hinterland SEG includes upland communities growing on higher terrains, well above groundwater level and is never submerged by water (Abbink et al. 2004). The river SEG reflects riverbank communities that are periodically submerged and subject to erosion, the dry lowland SEG reflects floodplain vegetation that can be occasionally submerged, wet lowland SEG represents marshes and swamps and the hinterland SEG reflects plant communities on well-drained terrains, above groundwater table (Abbink et al. 2004). Due to proximity of habitats to permanently wet environments (river, lake or marsh), the palynomorphs in the river and lowland SEG consists of the local and extra local vegetation elements. The hinterland SEG consists of the elements of the regional pollen flora that can be transported over long-distance (Olivera et al. 2015) (Table S1).

Organic carbon isotope analysis and TOC

A total of 36 samples from the Strangman’s Cove outcrop were selected for $\delta^{13}$C$_{TOC}$ analyses. Details of the analytical technique and the data set (Table S10) can be found in the Supplementary Data.

Results

Palynological assemblages

Only 36 out of 104 processed samples provided palynomorph assemblages. The rest of the samples were barren or the scanned slides contained less than 10 specimens. The
Palynomorphs are generally moderately to well preserved, the wall colour varies between pale yellow to golden brown, their SCI index ranges from 2 to 7 depending on wall-thickness variations between the grains (Batten 2002). A total of 81 spore and pollen taxa and 5 aquatic palynomorphs are distinguished. The most important palynomorphs are illustrated in Fig. 4. A list of all identified taxa with full author citation and seven additional photoplates can be found in the Supplementary Data.

**Sidmouth Mudstone Formation**

Palynological assemblages from the Sidmouth Mudstone Formation are only recorded in core WP-1 from depths 111.98 m and 109.11 m (Fig. 2). The assemblages are characterized by the predominance of bisaccate pollen grains with a trilete mark (*Triadispora* group) (up to 45% of the total abundance) along with Circumpolles (up to 22% of the total abundance). The species *Triadispora obscura*, *T. plicata* and *T. aurea* are predominant and the stratigraphically useful *T. verrucata* is present at 111.98 m. Among the Circumpolles group, various species of the genus *Partitisporites* (e.g., *P. novimundanus*, *P. scurrilis*) and *Praecirculina granifer* are common as well as *Duplicisporites granulatus*. *Camerosporites secatus* is present but rare in both samples from the Wiscombe Park Borehole. Alete bisaccates are less common, with only *Alisporites* spp. and *Ovalipollis ovalis* reaching more than 10% of the total spore-pollen flora. Monosaccate pollen grains are represented only by *Vallasporites ignacii*. *Aulisporites astigmosus* and *Cycadopites* sp. are observed in both samples, but they are minor components of the assemblage. During quantitative analysis only a few spore taxa are documented; among them *Aratrisporites* spp. and *Calamospora tener* are more common (2-3% of the total abundance). Other spore taxa were recorded only during the qualitative analysis and then only represented by one or two specimens (e.g., *Thomsonisporis toralis*, *Verrucosisporites morulae*, *Krauselisporites* sp.).
Dunscombe Mudstone Formation

The palynological assemblages from the Dunscombe Mudstone Formation recorded in the Strangman’s Cove section and in the WP-1 core (Fig. 2) contain predominantly bisaccate pollen grains and members of the Circumpolles group. *Triadispora* spp. are very frequent throughout the DMF, especially the species *T. obscura*, *T. plicata* and *T. sulcata*. Alete bisaccate pollen grains are represented mainly by *Ovalipollis* spp., *Alisporites* spp., *Pityosporites* sp. and *Klausipollenites gouldii*. The abundance of *Minutosaccus crenulatus*, *Ellipsovelatisporites plicatus* and *Microcachrydites doubingeri* occasionally increases. Striate bisaccate pollen grains (*Lunatisporites acutus*) are scarce. Among monosaccate pollen grains only *Enzonalasporites vigens* is common. *Patinasporites densus* is recorded first in sample WE017 (within the Lincombe Member) in the Strangman’s Cove section and at the top of the WP-1 core, at 56.51m (Fig. 2), but its abundance is low. *Camerosporites secatus*, *Duplicisporites granulatus* and *Praecirculina granifer* are the most abundant members of the Circumpolles group. *Aulisporites astigmosus* is virtually absent in the DMF except for one questionable specimen at 71.41m in core WP-1. The colonial chlorococcalean algae *Plaesiodictyon mosellanum* dominate the palynological samples from the Lincombe Member at Strangman’s Cove (Fig. 2). Cluster analysis helped to distinguish two assemblages (assemblage I and II) at the Strangman’s Cove. Assemblage I and II differ mainly in the increase of *C. secatus* among the Circumpolles and the presence of *Patinaporites densus* in Assemblage II. The boundary between these two assemblages is placed between samples WE017 and WE109 (Fig. 2, Supplementary Data).

Branscombe Mudstone Formation

In the Branscombe Mudstone Formation, only one palynological assemblage (Assemblage III) is encountered from the upper part of the Strangman’s Cove section, in sample WE203 (Fig. 2). This sample is characterized by an increase in the abundance of spores (up to 30% of the
total abundance in WE203) compared to the DMF *Calamospora tener, Porcellispora longdonensis, Todisporites major* and *T. rotundiformis* are the most frequent taxa. Among the bisaccate pollen grains, the abundance of genus *Triadispora* decreases in favour of alete bisaccates e.g., *Alisporites* spp., *Klausipollenites gouldii* and *Ovalipollis* spp. The abundance of monossaccate pollen (*Patinasporites densus, Enzonalasporites vigens, E. manifestus, Pseudoenzonalasporites summus, Vallasporites ignacii*) is higher compared to the DMF and SMF. *Camerosporites secatus* dominates the Circumpolles group. *Partitisporites quadruplices* and cf. *Partitisporites tenebrous* are also recorded in this assemblage, while *P. maljawkinae, P. novimundanus* and *P. scurrilis* are absent. The enigmatic sporomorph *Brodispora striata* is also more frequent here compared to its single occurrence in the DMF.

**Somerset Sandstone units**

From the Somerset localities (Fig. 1), samples from the Sutton Mallet and Lipe Hill sections yielded palynomorphs, but the samples at Knapp Quarry were barren (Fig. 3). At Lipe Hill two assemblages are distinguished by cluster analysis (Assemblage LH I and II) (Fig. 3). Ten samples from the Lipe Hill section in Somerset provided well-preserved assemblages (Fig. 4). The palynomorph abundance in sample LH-1 is very low, and the proportion of the terrestrial palynomorphs decreases significantly in samples LH-5 and LH-6 as these are especially rich in algae, predominantly *P. mosellanum* (Fig. 3). The terrestrial palynomorphs are characterized by the predominance of the *Triadispora* group (up to 60%), mainly *Triadispora obscura* and *T. plicata*. The Circumpolles group is also common; the proportion of *D. granulatus* is somewhat elevated compared to *C. secatus*. Two palynomorph assemblages are distinguished by cluster analysis, with sample LH-1 as a third group that differs from the other two assemblages by a low total palynomorph count. The two other assemblages differ primarily in the proportion of *Ovalipollis* spp. (LH I 20-30%, LH II ca. 10%) to *Triadispora* spp (LH I ca. 40%, LH II up to 60%). Spores are present in all samples but their diversity and
abundance is very low (<5%). In samples LH-1–LH-2, LH-5–LH-6 a few acanthomorph acritarchs (*Micrhystrydium* sp.) are observed. Some of these acritarchs are probably reworked as suggested by a darker wall colour or a damaged vesicle. However, some specimens are well-preserved and show no signs of reworking.

In the Sutton Mallet section (Figs 1 and 3) the palynomorph assemblages predominantly contain bisaccate pollen grains (Fig. 3). *Triadispora* spp. are the most frequent (*ca. 50-60%). Alete bisaccates are less common and are represented mainly by *Ovalipollis ovalis, O. lunzensis, Alisporites grauvogeli* and *A. grandis*. Spores are more abundant in this locality than in the coastal section and the WP-1 core. *Arattrisporites* spp. is especially frequent; it reaches around 10% of the total abundance in each sample. The only striate bisaccate pollen encountered is *Lunatisporites acutus*. Among the Circumpolles, *Camerosporites secatus* is predominant, while other genera occur only in minor proportions. *Ricciisporites tuberculatus* is recorded in sample SM2 at Sutton Mallet. *Plaesiodictyon mosellanum* is frequent also in the samples from this locality. *Botryococcus* is present but less abundant than *P. mosellanum*. In sample SM4 a few acanthomorph acritarchs (*Micrhystrydium* sp.) are observed. These specimens are well-preserved and show no ambiguous signs of reworking.

**Organic carbon isotopes**

Bulk organic carbon isotope ratios (δ^{13}C_{TOC}) from WP-1 core (Miller *et al.* 2017) and samples from the Strangman’s Cove section allow stratigraphical correlation of these data to be assessed (Fig. 6). The Strangman’s Cove Section outcrop contains several negative isotope excursions: the older part of the section shows larger variations but a gradual upwards shift to more negative values, which culminates in the negative isotope excursion in sample WE015. A shift to more positive values occurs near the top of the DMF (Fig. 6).
Palynostratigraphy

Palynology provides a powerful tool in the correlation of Late Triassic marine and non-marine successions; however, many assemblages are not dated independently with other fossils (e.g., ammonoids, conodonts) or other geochronological methods (e.g., radiometric dating) (Cirilli 2010). Important taxa often lack well-calibrated stratigraphical ranges and these are often diachronous in different regions (e.g., between Germanic Basin and Alpine realm) (Cirilli 2010).

In the Southern Permian Basins (mainly Germany, Poland and southern North Sea), early Julian assemblages are characterized by Triadispora verrucata, Camerosporites secatus, Patinasporites densus and Vallasporites ignacii (e.g., Scheuring 1970, 1978; Orłowska-Zwolińska 1983, 1985; Reitz 1985; Heunisch 1999; Kürschner & Herngreen 2010; Fijałkowska-Mader et al. 2015) with an increase in Aulisporites astigmosus in the younger part of the Julian. The Tuvalian is characterized by the appearance of Ricciisporites tuberculatus and Classopollis spp. (Kürschner & Herngreen 2010). The common occurrence of Granuloperculatipollis rudis with Classopollis zwolinskae and Chasmatosporites spp. marks the beginning of the assumed Norian Granuloperculatipollis rudis Zone in the Germanic basins (e.g., in the Arnstadt Formation) (Kürschner & Herngreen 2010). However, late Carnian and Norian palynological records with independent marine biostratigraphical control or other geochronological age constrains are still lacking from the Germanic Triassic basins (Cirilli 2010; Kürschner & Herngreen 2010). In the GSSP section for the Carnian, Prati di Stuores/Stuores Wiesen, Vallasporites ignacii, Patinasporites densus, Aulisporites astigmosus and Camerosporites secatus all have their first occurrences in the lowermost part of the Daxatina canadensis Subzone of the Trachyceras Zone (Fig. 5) in the lowermost Carnian (Mietto et al. 2012).
The palynological assemblages in Devon and Somerset yielded typical Carnian palynological assemblages with *Camerosporites secatus*, various *Partitisporites* and *Duplicisporites* species, *Enzonalasporites* spp., *Vallasporites ignacii* and *Patinasporites densus*. However, distinct Julian palynomorph assemblages e.g., the *Triadispora verrucata* assemblage and the *Aulisporites astigmosus* acme reported from the Schilfsandstein (Visscher *et al.* 1994) and marine series in the Alps (Roghi *et al.* 2010), are not seen in the British succession. The common taxa in the uppermost part of the Sidmouth Mudstone Fm (Fig. 2), *Partitisporites* spp., *Camerosporites secatus* and *Praecirculina granifer*, are generally characteristic in Carnian palynological assemblages in the Germanic realm (from the e.g., Lower Gipskeuper, Schilfsandstein and Upper Gipskeuper). Few specimens of *Triadispora verrucata* are recorded in the studied part of the Sidmouth Mudstone Fm, which in contrast is a common species in the early Julian assemblages from the Lower Gipskeuper in the Germanic Basins (e.g., Kürschner & Herngreen 2010; Fijałkowska-Mader *et al.* 2015) (Fig. 5). This species is frequent in the *Triadispora verrucata* subzone of the *Camerosporites secatus* Zone in NW and central Europe (Kürschner & Herngreen 2010), and in the *verrucata* Subzone of the *longdonensis* Zone in the Polish Keuper (Orłowska-Zwolińska 1983, 1985; Fijałkowska-Mader *et al.* 2015), and in the GTr 12-13 zones of Heunisch (1999) (Fig. 5). The assemblages with abundant *T. verrucata* are usually found in the upper part of the Lower Gipskeuper in the Germanic Basin (Kürschner & Herngreen 2010; Fijałkowska-Mader *et al.* 2015), but the complete range of *T. verrucata* has a longer duration, first appearing in the Grenzdolomit (topmost Erfurt Formation, late Ladinian-earliest Carnian, [e.g., Szulc 2000]) and disappearing around the Carnian/Norian boundary in Germany (Heunisch 1999; Kürschner & Herngreen 2010). The palynological assemblages from the ‘Mudstone I’ unit of Fisher (1985), which underlies the Dunscombe Mudstone Formation is similar to our assemblage of the Sidmouth Mudstone Fm in the WP-1 core that contains *Camerosporites*.
secatus, Duplicisporites granulatus and Ovalipollis ovalis but lacks Enzonalasporites vigens. Fisher (1985) did not identify Triadispora verrucata in any of his samples from ‘Mudstone I’. He assigned this assemblage to the latest Ladinian (Longobardian)-early Carnian (Cordevolian) based on correlation to the assemblages from the Meridenkalken (Scheuring 1978). The preliminary magnetostratigraphy of the MMG (Hounslow et al. 2002) suggests that the Ladinian/Carnian boundary may be placed in the upper part of the SMF. However, characteristic Ladinian taxa from the Germanic Basins such as Heliosaccus dimorphus or Echinitosporites liiacoides (Kürschner & Herngreen 2010) are absent in ‘Mudstone I’ of Fisher (1985) and in the upper part of the SMF in our study. Hence, the co-occurrence of Triadispora verrucata with other characteristic Carnian taxa e.g., C. secatus, Partitisporites spp., V. ignacii, A. astigmosus within the palynoflora from the SMF suggest an early Julian age (Orłowska-Zwolińska 1983, 1985; Heunisch 1999; Kürschner & Herngreen 2010; Fijalkowska-Mader et al. 2015). Therefore the Ladinian/Carnian boundary is located lower than the investigated interval according to the palynological results. Aulisporites astigmosus is a characteristic component of Julian assemblages in the western Tethys (e.g., Roghi et al. 2010), as well as in the Schilfsandstein (Visscher et al. 1994). An acme of the species is recorded in several late Julian successions in the Dolomites, Julian Alps and Northern Calcareous Alps (Roghi 2004; Roghi et al. 2010). Mueller et al. (2016b) found an increase in A. astigmosus only in the Rheingraben Shales of mid Julian age. In the British successions, the species has been recorded only in the Sidmouth Mudstone Formation represented by a few specimens and one poorly preserved specimen was found in the DMF (WP-1, sample 71.41 m, Fig. 6). The Aulisporites acme is often associated with the presence of other Cycadophyte related pollen grains such as Cycadopites spp. and lycopsid spores of the Aratrisporites group (Roghi et al. 2010). A palynoflora with similar composition to that is absent in the British successions. In the WP-1 core Cycadopites is extremely rare, and in the
Strangman’s Cove coastal outcrop only a few specimens of *Cycadopites* sp. are recorded in the interval from sample WE017 in the Lincombe Member, to sample WE214 in the upper part of the DMF (Figs 2 and 6). This interval might reflect an expression of the *Aulisporites* assemblage however, the British successions appear to completely lack an acme of *Aulisporites astigmosus* as seen in the western Tethys or in the Schilfsandstein. The previous study of Fisher (1985) distinguished two palynological assemblages in the DMF in the South Devon Coast: an older and younger palynomorph assemblage with the boundary between the two placed within the Lincombe Member (Fisher 1985; Gallois & Porter 2006). The assemblages were informally termed the “Dunscombe cycle” and “Weston cycle”. These “cycles” most likely correspond to the Assemblages I and II from the Strangman’s Cove coastal outcrop respectively. The composition of the “Dunscombe cycle” is similar to the Assemblage I from the Strangman’s Cove (Fig. 2), with a high volume of bisaccate pollen grains, *Praecirculina granifer, Enzonalsporits vigens, Ovalipollis ovalis, Ellipsovelatisporites plicatus* and *Triadispora* spp. Fisher (1985) correlated the older “Dunscombe cycle” to the palynological assemblages of the Gipskeuper and Zones D-E of Scheuring (1970) from the Keuper in Switzerland based on the restricted occurrence of *Protodiploxypinus gracilis* and *Triadispora plicata* in this assemblage and suggested an early Carnian (“Cordevolian” age, now considered early Julian). The “Weston cycle” of Fisher (1985) has a similar palynological composition to Assemblage II in the upper part of the DMF in our study, characterized by an increase in *Camerosporites secatus* and the restricted occurrence of *Patinasporites densus*. Fisher (1985) correlated the “Weston cycle” to the upper part of the Gipskeuper and the Schilfsandstein (Zones F-G of Scheuring 1970) and suggested a Julian-Tuvalian age. The local first occurrence (FO) of *Patinasporites densus* is recorded in the upper part of the WP-1 (at 56.51 m). In the Strangmans’s Cove section, the species is present from ca. 20 m above the base of the section (sample WE017, Fig. 2). *Patinasporites*
densus and Partitisporites maljawkinae are characteristic for the densus–maljawkinae phase of Van der Eem (1983) (Fig. 5) characteristic for the late Julian in the Dolomites. As the last local occurrence of Partitisporites novimundanus is recorded in the upper part of the DMF at Strangman’s Cove (Figs 3 and 6), the uppermost studied part of the DMF is most likely still Julian according to the recent palynological zonation for NW and central Europe (Kürschner & Herngreen 2010). Characteristic Tuvalian taxa e.g., Ricciisporites or Classopollis spp. are absent in the studied part of the DMF.

The assemblage from the Branscombe Mudstone Fm is characterized by the presence of Porcellispora longdonensis, Calamospora tener, Partitisporites quadruplices and Brodispora striata is more common compared to the single occurrence of the species in the DMF (Strangman`s Cove, sample WE111). Brodispora striata is a characteristic element of the Arden Sandstone Formation of the English Midlands (Clarke 1965) (e.g., Warrington 1970, Warrington et al. 1980; Fisher 1985; Barclay et al. 1997). This formation has been assigned to the Tuvalian substage based on the palynological assemblage with Duplicisporites spp., Camerosporites secatus, Haberkornia gudati, Vallasporites ignacii, Ovalipollis pseudoalatus, Ellipsovelatisporites plicatus, Ricciisporites umbonatus, Patinasporites densus and Brodispora striata (Clarke 1965; Warrington 1970a; Warrington et al. 1980; Barclay et al. 1997) although many of these taxa are not confined to the Tuvalian (Cirilli 2010; Kürschner & Herngreen 2010). In the Alpine Realm, B. striata has been recorded from the Opponitz Formation in the Lunz am See area in Austria (Dunay & Fisher 1978) dated to the uppermost Carnian Tropites subbulatus ammonoid Zone suggesting Tuvalian age (Fig. 5). The species Partitisporites quadruplices characterizes the Tuvalian in the western Tethys (Visscher & Krystyn 1978; Cirilli & Eshet 1991). Brodispora striata and P. quadruplices are found together in the Carnitza Formation and Travenanzes Formation in the Dolomites (Roghi 2004), which based on ammonoids is Tuvalian in age (De Zanche et al. 2000). According to
the national rationalisation of the Mercia Mudstone Group lithostratigraphy by Howard et al. (2008), the DMF includes the equivalent of the Arden Sandstone Formation of the English Midlands there considered to be Tuvalian in age. Our palynological data and the results of Fisher (1985) from the DMF are inconsistent with a Tuvalian age assignment. The studied part of the DMF on the South Devon Coast being Julian in age, with Tuvalian assemblages confined to the overlying Branscombe Mudstone Formation in the upper part of the Strangman’s Cove section. The Julian/Tuvalian boundary might be placed in the uppermost part of the DMF or the lowermost part of the Branscombe Mudstone which has not been investigated in this study.

**Correlation of the sandstone units from Somerset and the English Midlands**

Based on lithostratigraphy and log correlations Howard et al. (2008) suggested that the North Curry Sandstones from Somerset are equivalent to the Lincombe Member of the Devon coast and also the Arden Sandstone Formation in the English Midlands. **Brodispora striata** and **Ricciisporites** are frequent elements of the Arden Sandstone Formation (Warrington et al. 1980). **Brodispora** is common only in the Branscombe Mudstone Formation on the Devon coast and **Ricciisporites** was only recorded at Sutton Mallett, in Somerset. The palynological assemblages at Lipe Hill are comparable with Assemblage I at Strangman’s Cove and the “Dunscombe cycle” of Fisher (1985) from the coast, based on the high amount of **Triadispora** spp. especially **T. obscura**, **Ovalipollis spp.**, **Praecirculina granifer**, **Duplicisporites granulatus**. The assemblages at Lipe Hill suggests Julian (probably late Julian) age but it cannot be defined more precisely based on the palynological assemblages only.

The assemblage at the Sutton Mallet section is younger, based on the presence of **Ricciisporites tuberculatus** in sample SM 2, indicating the Tuvalian. In the Alpine realm this species appears first in the Travenanzes Formation in the Dolomites and in the Carnitza
Formation in the Julian Alps in the Tropites dilleri Zone of the Tuvalian (Fig. 5) (Roghi 2004; Roghi et al. 2010).

Chemostratigraphy and correlation

In marine successions, $\Delta^{13}C_{\text{CARB}}$ shows an overall rise of 3‰ in the earliest Carnian ($\Delta^{13}C_{\text{CARB}}, +1 \%o–+3 \%o$ compared to V-PDB) with a small negative excursion near the Ladinian-Carnian boundary (-1 ‰) (Korte et al. 2005) and a sharp negative excursion at the early-late Julian boundary both in carbonate carbon and organic carbon (Korte et al. 2005; Muttoni et al. 2014; Mueller et al. 2016a, b).

The correlation of the negative carbon isotope excursion at the Ladinian-Carnian boundary and the IIE excursion in the WP-1 core is unlikely since palynological evidence does not suggest a Ladinian age for the lowermost part of the DMF or topmost part of the Sidmouth Mudstone Fm. The correlation of the IIE excursion in the WP-1 core to the early-late Julian boundary excursion seems to be more plausible and it can be matched to the Julian 1/2 zonal boundary excursion in the Alpine areas (Dal Corso et al. 2012, 2015). The lithostratigraphical and palynological correlation between the Strangmans’ Cove section and WP-1 core (Gallois & Porter 2006) is supported by $\Delta^{13}C_{\text{TOC}}$ wiggle matching in Fig. 6. The third isotope excursion (CIE-3) in WP-1 can be matched with the excursion in sample WE015 in the coastal outcrop. This excursion is located at the base of the Lincombe Member, just below calcareous marker bed C (Gallois 2007; Gallois & Porter 2008) at Strangman’s Cove (Fig. 6). The correlation of this interval between the core and the outcrop is supported by the local first appearance date of *Patinasporites densus*. CIE-4 is located in both sections at calcareous marker bed G (Fig. 6). CIE-5 is just above bed I (Fig. 6) at Strangman’s Cove which correlates to the interval between 51 m and 52 m in WP-1 (Gallois & Porter 2006) with the fifth negative carbon isotope excursion (Fig. 6). CIE-2 is located at calcareous marker bed A.
and the IIE is few metres above the base of the DMF in both sections (Fig. 6). However, the latter two excursions cannot be precisely matched.

As the bulk organic matter analysed for its δ¹³C composition comprises not only the particulate organic matter and palynomorphs identified here but includes the whole sedimentary organic matter (kerogen), it is not straightforward to directly compare the palynological/palynofacies data with the bulk carbon isotope data. The bulk δ¹³C_TOC values are more negative in the Strangman’s Cove outcrop than in the Wiscombe Park core. These differences may be due to variations in the composition of the organic matter, indeed TOC values, although generally low, do differ between the core and the outcrop (Fig. 6).

Unfortunately, Rock-Eval data which may help to quantify the composition of the total sedimentary organic matter are not available.

Multiple carbon source contributes to and thus complicate the interpretation of the bulk organic δ¹³C signal. Fluctuations in bulk organic δ¹³C can be associated with both changes in lithology and changes in environmental conditions (Pancost et al. 1999). Major factors controlling bulk organic δ¹³C values include variations in the composition of the organic matter (i.e. aquatic vs. terrestrial) and the carbon isotopic composition of atmospheric CO₂ during inorganic carbon uptake by plants (e.g. Diefendorf & Freimuth, 2017). Compound specific carbon isotope analysis of plant leaf waxes (n-alkanes) can help resolve the multiple isotopic signals that contribute to the bulk organic carbon δ¹³C signal. Miller et al. (2017) suggest that the negative carbon isotope excursions recorded simultaneously in both WP-1 δ¹³C_TOC and in plant leaf waxes (n-alkanes) represent an injection of light carbon into the atmosphere, resulting in significant global perturbations of the carbon cycle. The good correlation between WP-1 and Strangman’s Cove δ¹³C_TOC may suggest that changes in δ¹³C_TOC at Strangman’s Cove are also likely a result of changes in the carbon isotopic composition of atmospheric CO₂ and not fully related to changes in organic matter source.
Nevertheless, climate change often results in vegetation shifts. The amount of carbon isotope fractionation, is heavily controlled by the plant’s photosynthetic pathway, which varies depending on plant type (e.g. Diefendorf & Freimuth, 2017; Collister et al. 1994). The negative carbon isotope excursions at Strangman’s Cove (CIE-3, -4 and -5) coincide with higher bisaccate pollen and spore abundances (Supplementary Data), although there is no statistically significant correlation with any of the individual palynomorph groups ($R^2$ below 0.5 for each palynomorph group, see Supplementary Data, Fig. S2). Therefore, we suggest that the observed shifts in bulk carbon isotopes may also be the result of changes in vegetation as well as shifts in the C isotopic composition of atmospheric CO$_2$. Moreover, precipitation amount (humidity) is a strong factor governing plant carbon isotopic fractionation. Except for in extremely wet environments, the $\delta^{13}C$ of C$_3$ plants tends to decrease with increasing rainfall (e.g. Kohn 2010). Thus, the increased humidity experienced during the CPE may have also contributed to the observed negative C isotope excursions.

Climatic and palaeoenvironmental implications

In the Southern Permian Basin and western Tethys marine sections, hygrophyte pteridophytic spores and pollen grains with hygrophyte affinity (e.g., Aulisporites astigmosus) arose in abundance during the Julian (Roghi et al. 2010) (Table S2). However, the pollen record across the Dunscombe Mudstone Fm shows the predominance of xerophyte-related land vegetation during this interval (Figs. 2–4, Figs 7–8, Table S1. Despite the shift in climate, evident from the lithological change into the base of the Dunscombe Mudstone Fm: Porter & Gallois, 2008), a clear humid climate signal is missing from the palynological record. There are two different ways to interpret the lack of a clear humid climate signal. Firstly, during the Late Triassic NW Europe was at ca. 30°N in the continental interior of Pangaea within the dry climate belt with low annual precipitation (Fig. 1), likely falling during the Northern Hemisphere-summer period when the main rainfall zone with the ITCZ migrated to the north
Hence, in the Wessex Basin, the lack of an expected increase in hygrophyte floral elements e.g., ferns or the Bennettitalean parent plant of *Aulisporites* *(Williamsonianthus keuperianus)* (Kräusel & Saarschmidt 1966; Balme 1995) might be related to this strongly seasonal precipitation, which during the dry season may have prohibited the development of a permanent extensive wetland vegetation and the preservation of spores and pollen grains. Also, the regional palaeogeography and topography of the studied area such as subdued upland surrounding a low-relief interior basin (Talbot et al. 1994) may have caused isolation from other basins and the development of a locally more arid climate than in other region of the Germanic Triassic, even during the CPE. *Aulisporites astigmosus* seems to be extremely rare in the British Carnian palynological assemblages, as it has not been identified in most palynological studies (e.g., Warrington 1970, 1971; Fisher 1972; Warrington 1974, 1984; Warrington & Williams 1984; Fisher 1985; Warrington 1997, 2004). Even though the species is present in some coeval sections of the UK (pers. comm. with G. Warrington 2017), the parent plants might have formed only minor part of the vegetation. Although the acme of *A. astigmosus* is recognized in different depositional settings and at different palaeolatitudes there are locations across Europe, like the UK, where it is absent in the Carnian palynological assemblages. The *A. astigmosus* acme has not been recorded in the Carnian Manuel Formation from Spain (Arche & López-Gómez 2014) (Table S2) and Lindström et al. (2017) do not record any well expressed hygrophyte palynological assemblages from the mid– to late Carnian assemblages of the Danish Basin which may suggest that these areas were probably too dry even during the CPE. Secondly, if the parent plants of *Aulisporites* and other humidity indicators (ferns) were present in the Wessex Basin, the balance between local and regional vegetation elements in the palynological record could have masked a real humid signal. Pollen and spores come from...
both local and regional sources, and catchments of different sizes can represent vegetation at
different spatial scales (Berglund 1973; Jacobson & Bradshaw 1981; Hicks & Hyvärinen
1999). In modern lacustrine settings, the contribution of regional pollen increases with larger
lake size or catchment area (Jacobson & Bradshaw 1981), which can lead to the
underrepresentation of local hygrophyte vegetation growing on the shoreline. This principle,
when applied to the Late Triassic vegetation elements of the UK, suggests the regional pollen
rain is probably associated with the xerophyte pollen group originating from conifers and
upland seed ferns (monosaccates, bisaccates, Circumpolles), while the local and extra local
elements were likely the hygrophyte lowland and riparian elements including *Aulisporites*
*astigmosus* (Table S1), fern- and lycopsid spores.

Such differences between the contribution of local and regional pollen rain has already been
discussed by Visscher et al. (1994) in the case of the Schilfsandstein. They considered the*Aulisporites*-acme as only a local signal and rejected the possibility of a European-wide
humid climatic phase, as the regional pollen was likely characterized by xerophyte pollen
during the whole Julian. In contrast to the mainly lacustrine DMF, the Schilfsandstein in
Germany is considered a predominantly fluvial-deltaic deposit (Shukla et al. 2010; Franz et
al. 2014). The local wet spots on the floodplain probably collected more material from the
local and extra local pollen rain resulting in hygrophyte palynological assemblages enriched
in *Aulisporites astigmosus* and spores.

The Schilfsandstein depositional system probably provided a more favourable habitat for the
parent plant of *Aulisporites* (Bennettitalean) which grew in warm and often in deltaic settings

The Lincombe Member in Devon represents a geographically restricted freshwater lake
(Porter & Gallois 2008). If this lake was of considerable size (>1km), according to the pollen
catchment model of Jacobson & Bradshaw (1981) the size of the lake might explain the
predominance of the regional pollen rain of xerophyte vegetation and the lack of local and extra local pollen rain including spores and pollen from hygrophyte vegetation (e.g., *Aulisporites*) (Fig. 7). Within the Sidmouth Mudstone and Branscombe Mudstone formations, the samples with high spore abundance come from cm-thick greenish-grey intervals within between thicker red mudstone units (Fig. 2). These intervals likely represent spatially restricted locally wet environments with higher representation of extra local and local pollen and spores (Fig. 7). The sandstone units in Somerset contain more spores suggesting a more local source for the palynomorphs, but the *Aulisporites* acme is still absent. The taphonomic bias between local and regional pollen is further exacerbated by the higher pollen production rates of wind-blown pollen from the hinterland (e.g., conifers) compared to insect-pollinated palynomorph types such as Cycadales pollen, *Cycadopites* sp., *A. astigmosus* (e.g., Fægri & van der Pijl 1966).

The relatively wetter periods during the CPE should have supported the expansion of the hygrophyte vegetation living proximal to a lake. However, the proposed enhanced seasonal runoff associated with the CPE could have transported more regional pollen types to the lacustrine depositional setting of the DMF, leading to the underrepresentation of the more proximal local and extra local vegetation elements e.g., ferns or Bennettitales. A similar scenario was suggested by Bonis et al. (2010) from the Triassic/Jurassic boundary palynological assemblages of St. Audrie`s Bay (Somerset, UK) where the intensification of seasonal monsoons was connected to the increase in hinterland Cheirolepidiaceae pollen and lower spore abundance.

Climate change reflected by the hinterland vegetation (regional pollen) might provide new criteria for tracking climate change during the CPE. However, interpreting the ecological signal of the regional hinterland pollen types is difficult due to the uncertainty in parent plants and in the uncertain assignment of dispersed pollen grains into the hygrophyte or xerophyte
groups. *Alisporites* species were most likely produced by seed ferns and although the parent plant is uncertain, with some placing them in transitional and/or hygrophyte groups (e.g., Visscher & Van der Zwan, 1981; Whiteside *et al.* 2015; Lindström *et al.* 2016; Mueller *et al.* 2016a, b). The increase in the total *Alisporites* and total spore abundance in the lower part of the DMF in the Strangman's Cove section coincides with a peak in freshwater algae (*P. mosellum*) in the Lincombe Member (Fig. 8). This increase in algae suggests expansion of the lacustrine facies (i.e. the Lincombe Member), hence can be linked to a climatic interval with a relatively wetter season (Fig. 8), which is also supported by the increase in abundance of *Alisporites*.

**Acritarchs in Somerset**

Marine microplankton are known from some other Late Triassic mudstone units in the Cheshire Basin and the Midlands (Earp & Taylor 1986; Wilson & Evans 1990; Warrington & Ivimey-Cook 1992; Barclay *et al.* 1997) indicating possible marine influence, but with limited additional evidence for a marine incursions in the MMG depositional basins. The existence of halites in the MMG has always been contentious, with purely marine, sabkha and wind-blown origins considered (Hounslow & Ruffell 2006). Fauna from the Arden Sandstone Formation in the Worcester Graben includes possible marine bivalves *Nucula?*, *Thracia?* and *Pholadomya?* and shark teeth (Old *et al.* 1991), indicating possible marine connection. However, the frequent occurrence of colonial freshwater algae (*P. mosellum*) in the Lincombe Member (South Devon) and in Somerset suggests freshwater conditions during the deposition of this lithological unit. The acritarchs are simple acanthomorph acritarchs (*Michrystridium*) with no significant stratigraphical value and could easily be reworked Palaeozoic forms. The acritarchs could be reworked and transported to the lakes during periodic flash floods. The source of the reworked palynomorphs is likely Carboniferous or Devonian sediments 50-80 km to the west and north-west (Porter & Gallois, 2008). In
addition as Hounslow & Ruffell (2006) conclude, the marine microplankton may also have arrived in far-travelled aeolian dust, in part sourced from marine aerosols, as occurs in modern desert settings (Glennie & Evans 1976).

Conclusions

The Carnian Pluvial Episode during late Julian to early Tuvalian in the western Tethys is considered to be the one of the most pronounced climate change during the Triassic associated with higher precipitation rates and carbon cycle perturbation. The shift to relatively more humid conditions is manifested in the expansion of hygrophyte vegetation in many locations worldwide (Roghi et al. 2010). The palynological assemblages from four localities in the Wessex Basin, SW UK have been shown to track vegetation changes and palaeoclimate trends during the Carnian from a fluvio-lacustrine environment and in addition provide a palynostratigraphical framework for correlation.

The Dunscombe Mudstone Formation in South Devon and the Lipe Hill succession from Somerset contain a Julian (early Carnian) palynoflora, while the assemblage from the overlying Branscombe Mudstone Formation is likely Tuvalian (late Carnian) based on the co-occurrence of Brodispora striata and Partitisporites quadruplices. The presence of Ricciisporites tuberculatus also indicates a Tuvalian age for the Sutton Mallett section in Somerset. Based on the palynological assemblages and the chemostratigraphy, the stratigraphical range of the Dunscombe Mudstone Formation can be extended down into the Julian in contrast to previous Tuvalian age assignments of the Arden Sandstone Fm (Clarke 1965; Warrington 1970; Warrington et al. 1980; Barclay et al. 1997). A humidity signal associated with the CPE is not seen in the Wessex Basin successions where quantitative palynology suggests the dominance of xerophyte floral elements through the Dunscombe Mudstone Formation with only a few horizons of increased hygrophyte flora. The British
successions appear to lack an acme of *Aulisporites astigmosus* which has been linked to the shift to wetter climate as seen in the western Tethys or in the Germanic Basin. Firstly, the prevailing dry climatic conditions in the inner part of Pangea and the strong seasonality in precipitation might explain the lack of suitable permanent habitat for the hygrophyte vegetation. Secondly, the lack of a clear humid signal is likely caused by overrepresentation of the regional pollen rain in the lacustrine units leading to the predominance of xerophyte hinterland floral elements. The bias towards regional pollen rain is further enhanced by the higher pollen production rate of hinterland elements—mainly the conifers. In addition, the potential increase in continental runoff related to more humid conditions might have further increased the proportion of the regional hinterland floral elements in the palynological assemblages. Changes in the regional flora might provide a tool for recording climate change.

A slight shift to wetter climate is inferred from increased *Aulisporites* abundance in the hinterland flora that coincides with increase of fresh water algae suggesting the expansion of local lacustrine environments during a relatively wetter period.

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Figure captions

Fig. 1. (a) Map of the UK and Ireland showing the position of the study area, (b) palaeogeography during the Late Triassic after Golonka (2007), the asterisk marks the position of the study area, (c) Mercia Mudstone lithostratigraphy of south-west England from Howard et al. (2008), names shown in italics are abandoned in the revised nomenclature, (d, e) outline geology maps and location of the four studied outcrops and the Wiscombe Park-1 Borehole, modified after Gallois & Porter (2006) and Porter & Gallois, (2008).

Fig. 2. Lithostratigraphy, sample positions and the relative pollen-spore abundances during the Carnian in the Wiscombe Park-1 core and in the Strangman’s Cove outcrop section, Devon. Lithological log modified from Gallois (2007) and Gallois & Porter (2006). The stratigraphical subdivision is inferred from palynology and chemostratigraphy. The boundary between substages is indicated by a dashed line due to the uncertainty of the boundary position. Only the most abundant taxa are shown. The grey area of the curves is an exaggeration (3X) of the abundances plotted in black. The unidentified palynomorphs and encountered Lycopodium grains are shown as counts. A detailed lithological log with the exact position of the palynological samples is available in the Supplementary Data (Fig. S1).

Fig. 3. Lithostratigraphy, sample positions and the relative pollen-spore abundances during the Carnian in the Lipe Hill and Sutton Mallett sections, Somerset. Lithological log is modified after Ruffell & Warrington (1988). See Fig. 2 for details.

Fig. 4. Selected palynomorphs identified from the Mercia Mudstone Group. Taxon name is followed by lithostratigraphic unit and locality. Sample and slide numbers are in parentheses. Scale bar 10 µm: 1-13, 16-37. Scale bar 20 µm: 13-16. 1 Brodispora striata, Branscombe
Mudstone Formation, Strangman`s Cove (WE203/A). 2 Porcellispora longdonensis,

Branscombe Mudstone Formation, Strangman`s Cove (WE203/A). 3 Thomsonisporis toralis,

Sidmouth Mudstone Formation, WP-1 (109.11 m/1). 4 Calamospora tener, Sidmouth

Mudstone Formation, WP-1 (111.98 m/1). 5 Kyrtomisporis sp., Sidmouth Mudstone

Formation, WP-1 (111.98 m/1), 6 Aratrisporites sp., Sidmouth Mudstone Formation, WP-1

(111.98 m/1). 7 Aratrisporites fimbriatus, Sidmouth Mudstone Formation, WP-1 (111.98

m/1). 8 Cycadopites sp., Sidmouth Mudstone Formation, WP-1 (111.98 m/1). 9

Lunatisporites acutus, Dunscombe Mudstone Formation, Lipe Hill (LH 2/2). 10 Aulisporites

astigmosus, Sidmouth Mudstone Formation, WP-1 (111.98 m/1). 11 cf. Aulisporites

astigmosus, Sidmouth Mudstone Formation, WP-1 (109.11 m/1). 12 Brachysaccus

neomundanus, Dunscombe Mudstone Formation, Strangman`s Cove (WE017/A). 13

Ovalipollis lunzenis, Branscombe Mudstone Formation, Strangman`s Cove (WE203/A). 14

Alisporites grauvogeli, Dunscombe Mudstone Formation, WP-1 (57.86 m/1). 15 Ovalipollis

ovalis, Dunscombe Mudstone Formation, Strangman`s Cove (WE305/B). 16 Triadispora

aurea, Sidmouth Mudstone Formation, WP-1 (111.98 m/1). 17 Triadispora plicata,

Dunscombe Mudstone Formation, Lipe Hill (LH 2/2). 18 Triadispora verrucata, Sidmouth

Mudstone Formation, WP-1 (111.98 m/1). 19 Triadispora obscura, Sidmouth Mudstone

Formation, WP-1 (111.98 m/1). 20 Duplicisporites granulatus, Dunscombe Mudstone

Formation, Strangman`s Cove (WE003/A). 21 Patinasporites densus, Branscombe Mudstone

Formation, Strangman`s Cove (WE203/A). 22 Triadispora crassa, Sidmouth Mudstone

Formation, WP-1 (111.98 m/2). 23 Triadispora modesta, Sidmouth Mudstone Formation,

WP-1 (111.98 m/1). 24 Partitisporites novimundanus, Dunscombe Mudstone Formation,

Strangman`s Cove (WE003/A). 25 Praecirculina granifer, Dunscombe Mudstone Formation,

Strangman`s Cove (WE001/A). 26 Camerosporites secatus, Dunscombe Mudstone

Formation, Sutton Mallett (SM 2/1). 27 Duplicisporites mancus, Branscombe Mudstone


Fig. 6. Correlation of the Wiscombe Park-1 core and the Strangman`s Cove section with δ¹³C_TOC bulk isotope values, calcareous marker horizons (Gallois & Porter 2006) and the range of the key biostratigraphically important pollen taxa. Horizons with increased

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Hygrophyte vegetation elements are indicated. Black diamond symbols mark the samples included in the palynological analysis. The boundary between substages is indicated by a dashed line due to the uncertainty of the boundary position. The grey area is the correlation suggested by Gallois & Porter (2006) based on the lithology. Detailed lithological log of the Strangman’s Cove section with the palynological sample locations is in the Supplementary Data (Fig. S1). SMF = Sidmouth Mudstone Formation, Jul. = Julian. Tuv. = Tuvalian. BMF = Branscombe Mudstone Formation.

Fig. 7. Conceptual model of the relationship between sampling size (lake diameter) and the relative proportion of pollen grains originating from different habitats around inferred lake deposits of the DMF as applied to Late Triassic palynomorph types. Modified after Jacobson & Bradshaw (1981).

Fig. 8. Stratigraphy, lithostratigraphy and climatic interpretation of selected palynomorph groups in the Strangman’s Cove section. The boundary between substages is indicated by a dashed line due to the uncertainty of the boundary position. For the position of the palynological samples see Fig. 3. Detailed log with the position of the palynological samples is available in the Supplementary Data (Fig. S1). The total abundance of spores, *Alisporites* spp., algae and interpreted negative isotope excursions are indicated. BMF = Branscombe Mudstone Fm.
Figure 2
Wiscombe Park-1

Stratigraphy

Spores
Bisaccate pollen grains
Triadispora group
Monosaccate pollen grains
Circumpolles group
Palynomorph assemblages

Wasilow et al. 2000

Stratigraphy

Spores
Bisaccate pollen grains
Triadispora group
Monosaccate pollen grains
Circumpolles group
Palynomorph assemblages

Strangman’s Cove

Spores
Bisaccate pollen grains
Triadispora group
Monosaccate pollen grains
Circumpolles group
Palynomorph assemblages

palynologically barren interval
<table>
<thead>
<tr>
<th>Chronostratigraphy</th>
<th>Ammonoid zones (Tethys)</th>
<th>Palynostratigraphy</th>
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<tr>
<td>Upper Triassic</td>
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<td>Norian</td>
<td>Guembelites jandianus</td>
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<td></td>
<td>Volcanic events</td>
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<td>Spore-pollen events</td>
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</tbody>
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**Figure 5**

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Figure 7

Click here to download Figure

Regional pollen rain

- conifers
- seed ferns

Total palynomorph ratio (%)

- Extra local pollen rain
  - cycads, Bennettitales, seed ferns
  - Aulisporites spores

- Local pollen rain
  - spores

Diameter of lake (m)
Figure 8

- Spore abundance
- *Alisporites* spp. abundance
- Algae abundance
- Diversity (total number of taxa)

**Palynological samples**

- We203
- We214
- We201
- We302
- We305
- We104
- We103
- We109
- We017
- We015
- We011
- We003

**Lithostratigraphy**

- BMF
- Dunscombe Mudstone Fm
- Lincombe M.
- Dunscombe M.F.

**Substages**

- Tuvalian
- Julian/Tuvalian
- late Julian

**δ¹³C_TOC**

- wetter?
- drier?
- wetter?
- drier?
- wetter?
- drier
- wetter
- drier
- wetter

**Climate**

- wetter?
- drier?
- wetter?
- drier?
- wetter?
- drier
- wetter
- drier
- wetter

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