Integrated stratigraphy and palaeoclimate history of the Carnian Pluvial Event in the Boreal realm; new data from the Upper Triassic Kapp Toscana Group in central Spitsbergen (Norway)

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Abstract: The Late Triassic climate is characterized by arid conditions interrupted by a humid phase known as the Carnian Pluvial Event (CPE). This wet phase is well documented in the Tethyan realm, but evidence from the Boreal realm is scarce. Here we present evidence from quantitative palynology for the CPE from the Kapp Toscana Group on central Spitsbergen integrated with organic carbon isotope data linked to the geomagnetic polarity time scale. Our data reveal an early to mid-Julian-1 age for the Tschermakfjellet Formation. The transition to the Julian-2 is located in the De Geerdalen Formation and the Isfjorden Member is confirmed as mostly Tuvalian-3 in age. The Aulisporites astigmosus pollen assemblage zone marks the base of the CPE in the Tethys realm at the base of the Julian-2 is assigned to the Julian-1 in the Boreal region. Palaeoclimate proxy data inferred from principal component analysis indicate wetter conditions from the Julian-2 onwards, which is in agreement with the establishment of local swamp vegetation on top of a delta plain. The palaeotemperature curve indicates a period of cooler climate during the early Julian-1 followed by warming during the late Julian-1.

Supplementary materials: A list of all identified morphotaxa of pollen, spores and aquatic palynomorphs is available at http://www.geolsoc.org.uk/SUP18879.

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The Triassic was a warm period with ice-free polar regions (e.g. Hallam 1985; Dickins 1993; Frakes et al. 2005), during which all continents were assembled into the supercontinent Pangaea, which stretched from 90°S to 85°N (e.g. Ziegler 1982; Scotese 2000, 2004; Golonka 2007; Ogg 2012). The climate was dry along the western margin of the Palaeeothys, whereas the central part of Pangaea and the mid-latitudes had a strong monsoonal influence. The eastern coast of Laurasia and Gondwana and the western coast of Pangaea experienced wet and dry seasons (Parrish 1993; Mutti & Weissert 1995; Preto et al. 2010).

Simms & Ruffell (1989) described the Carnian Pluvial Event (CPE), which was a brief, but major European-wide climatic change to humid conditions during the early Carnian. The CPE, however, was rejected by some (e.g. Visscher et al. 1994) as simply an extensive fluvial depositional system in north-central Europe. More widespread evidence of a brief climatic change to humid conditions is now indicated in many marine sediments from northern, western and southern parts of the Palaeeothys (e.g. Hochuli & Frank 2000; Roghi 2004; Hornung et al. 2007a,b; Rigo et al. 2007; Breda et al. 2009; Kolar-Jurkovsek & Jurkovsek 2010; Kozur & Bachmann 2010; Archer & López-Gómez 2014; Dal Corso et al. 2015). The CPE is characterized by carbonate factory shut-down and a decrease in seawater temperature in the Palaeeothys, more widespread increases in clastic fluxes and significant faunal turnovers (Korte et al. 2005; Hornung et al. 2007b; Dal Corso et al. 2012). Synchronous events are also known from the Panthalassa Ocean (Nakada et al. 2014), probably reflecting a global climatic event in the mid-parts of the early Carnian. The CPE also occurs prior to the end of a longer-term trend, since the early Ladinian, towards increased rates of organic carbon burial (Korte et al. 2005).

The CPE has not been identified at high latitudes, but Hochuli & Vigran (2010) described a climate shift from relatively dry conditions in the early Carnian to more humid conditions in the later Carnian from the Barents Sea. This change is loosely linked with more widespread coal deposits in the Boreal Carnian. Roghi et al. (2010) suggested that the interval characterized by an increase in Aulisporites and fern spores in the Boreal region (e.g. Hochuli et al. 1989; Hounslow et al. 2007a) is contemporaneous with the CPE in the Tethys realm, but confirmation of this long-distance biostratigraphic correlation by independent stratigraphic data is lacking. The Carnian was also an interval of significantly increased sedimentation rates throughout the Barents Sea region (Riis et al. 2008), a feature loosely linked with progradation of deltaic systems across the Barents Sea shelf, which form part of a longer-term process of accommodation space filling in the Barents Sea since the early Triassic (Riis et al. 2008).

Here we aim to (1) establish a quantitative palynological study of the Carnian in central Spitsbergen, (2) interpret vegetation changes and palaeoclimate trends using principal component analysis and (3) integrate stratigraphic analysis using organic carbon isotopes and magnetostratigraphy to provide an independent detailed chronology for the palynostratigraphic correlation and palaeoclimatic interpretations.

The Boleral Late Triassic

The Middle Triassic on Spitsbergen is characterized by deposition of black organic-rich shales, the Botneheia Formation of the Sassendalen Group. Near the transition to the Upper Triassic (the base of the Kapp Toscana Group), the Botneheia Formation is followed by shales and sandstones of the Tschermakfjellet and the De Geerdalen Formations,
which were deposited in an offshore shelf, transitional to coastal and
deltaic environments (Buchan et al. 1965; Mørk et al. 1982, 1992,
1999; Mørk & Bjoroy 1984; Mørk & Worsley 2006; Riis et al. 2008;
Nagy et al. 2011). Early Norian parts of the Spitsbergen Triassic
record a return to marine depositional conditions.

Palynology is a key tool for regional correlation of Boreal
Triassic sediments as marine biostratigraphic events are rare in the
absence of conodonts and ammonites in shallow marine deposition
environments (Bjercke & Dypvik 1977; Bjercke & Manum 1977;
van Veen 1985; Hochuli et al. 1989). However, independent
detailed chrono-dating of these palynostratigraphic ranges is largely
lacking owing to a scarcity of ash-beds. Hounslow et al. (2007a,b,
2008) published magnetostratigraphic data from various
locations in Spitsbergen integrated with palynostratigraphic data.

Location and geological setting

This study focuses on the Kapp Toscana Group, but uses data from the
late Ladinian part of the Botneheia Formation (Sassendalen Group)
to the lowermost Knorringsjellet Formation. The interval
covers a regressive cycle from offshore marine shelf to prodelta
conditions, into overlying delta plain and coastal plain conditions,
and back to shallow marine conditions. Sections at Juvdalskampen,
Botneheia and Vendomdalen in central Spitsbergen have been
studied (Fig. 1).

Geological setting

Deposition in the Svalbard archipelago (and the Barents Sea) during
the Late Triassic took place on a stable platform that existed between
the mid-Carboniferous and Mesozoic (Mørk et al. 1982; Harland 1997). During the Late Triassic delta systems discharged locally from the west, but mainly from the SE into the area of central Spitsbergen (Mørk et al. 1982; Dallmann 1999; Riis et al. 2008). The gradual progradation of sediment from the SE resulted in diachronous lithostratigraphic boundaries across the Barents Sea (Fig. 2; Riis et al. 2008).

The uppermost Botneheia Formation in the Isfjorden area of
central Spitsbergen contains late (but not latest) Ladinian ammo-
noids some 4 m below the base of the Kapp Toscana Group
(Korčinskaja 1980, 1982; Dagys et al. 1993). It was deposited in a
prodelta setting to an open shelf setting (Mørk et al. 1982, 1999;
Krajewski 2008; Nagy et al. 2011). Latest Ladinian ammonoid
faunas (Nathorstites lindstroemi biozone) are known only from
Bjørnøya, and some parts of eastern Svalbard (Weitschat & Dagys
1989; Dagys et al. 1993).

The overlying Kapp Toscana Group is subdivided into three
formations (Mørk et al. 1982; Dallmann 1999). The lowermost
Tschermakfjellet Formation in central Spitsbergen was deposited in an
offshore marine to prodeltaic or delta front environment (Mørk et al. 1982). In the Isfjorden area it contains early Carnian
(but not earliest Carnian) ammonoids (i.e. Stolleyites tenuis
subzone of the S. tenuis zone), some 12 m above the base (Dagys
et al. 1993). In the Vendomdalen area and eastern Spitsbergen the
S. planus subzone of the Tenuis Zone is found some 0.5–3 m above
the base of the Tschermakfjellet Formation (Dagys et al. 1993;
Hounslow et al. 2007a). The overlying unit, the De Geerdalen
Formation, contains no age-diagnostic macrofauna giving suffi-
cient resolution for a Carnian subdivision (Hounslow et al. 2007a;
Nagy et al. 2011). The lower part of the De Geerdalen Formation in
central Spitsbergen was deposited in an interdistributary bay to
delta front environment, and the upper part in a delta plain to coastal
plain environment (Nagy et al. 2011). Mørk et al. (1982) described the
De Geerdalen Formation as a dominantly fluvial environment
deposited in a nearshore, paralic (lagoonal or delta plain) environ-
ment that was part of a NW-prograding delta system during the
Late Triassic. Constant input of freshwater resulted in changing
conditions from terrestrial or nearshore deltaic-dominated to shal-
low aquatic conditions with coastal reworking. In Svalbard, the ear-
liest Carnian ammonoid zone with Daxatina canadensis is known
only from Bjørnøya, about 300 km to the south (Tozer & Parker
1968). There it first occurs 40 m below the top of the 140 m thick
Skuld Formation (Mørk et al. 1992; Dagys et al. 1993), in a unit
consisting of grey shales, and siderite nodules (and some sandstone
beds), similar to the lithology of the Spitsbergen Tschermakfjellet
Formation. In central Spitsbergen, the Knorringsjellet Formation
lies above the De Geerdalen Formation and forms the base of the
Wilhelmyøya Subgroup, which was deposited in a shallow marine
environment (Dallmann 1999; Nagy & Berge 2008; Nagy et al.
2011). On Hopen Island the lower part of the Wilhelmyøya Subgroup
contains an early (not earliest) Norian ammonoid fauna (Brig
et al. 2012; Lord et al. 2014); hence part of the Wilhelmyøya
Subgroup on Spitsbergen may be Lower Norian.

Sample locations

Three locations are studied in central Spitsbergen (Fig. 1), east of
Longyearbyen: Juvdalskampen 35 km east (78°12’N, 16°40’E),
Botneheia 25 km (78°14’N, 16°47’E) and Vendomdalen 42 km
(78°12’N, 17°19’E). J. Nagy (University of Oslo) collected 60
samples from the sections at Juvdalskampen and Botneheia from
mainly mudstone intervals. Seventy-four samples from a magno-
stratigraphic study by Hounslow et al. (2007a) at Vendomdalen
were also used for evaluating a composite organic carbon isotope
stratigraphy, using these three locations.

Palynostratigraphy

The report by Van Veen (1985) for the Norwegian Petroleum
Directorate (NPD) aimed to provide a stable stratigraphic frame-
work for the Triassic of the Barents Sea, focusing on the Troms
area, and subdivided the Triassic into 16 palynozones.

Hochuli et al. (1989) subdivided the Upper Triassic into
six palynological assemblage zones based on a mix of outcrop
samples, core data and cutting samples. Vigran et al. (2014) expanded and simplified this previous work and published palyno-
morph assemblage zones for Svalbard and the Barents Sea based
on outcrop data, cores and wells. Paterson & Mangerud (2015)
conducted a detailed palynostratigraphic study of the Carnian to
Rhaetian succession on Hopen.

Whereas independent age control from ammonoids is fairly
good for the Early and Middle Triassic of Svalbard, the miospore-
based age assignment of the Late Triassic is mostly based on a
comparison with palynomorph ranges from the Germanic and
Alpine realms, owing to a rarity of other precise age diagnostic
forms. However, the age ranges of some miospore taxa are differ-
tent in the Boreal realm, and hence dating using long-distance rela-
tionships is of somewhat limited value.

Material and methods

Palynology

A total of 60 samples (31 from Juvdalskampen and 29 from
Botneheia) were processed for palynological analysis. To remove
the carbonates and silica, 10 g of each sample was treated with
HCl (27%) and HF (40%) according to standard processing proce-
dures described by Kuerschner et al. (2007). The residue was
sieved with a 250 and a 15 µm mesh. Heavy liquid separation with
ZnCl₂ was carried out to remove heavy minerals and a few sam-
ples were subsequently treated with ‘Schulze Reagent’ (KClO₃
and HNO₃) or nitric acid to eliminate some organic material such
as amorphous organic matter (AOM) and increase the palyno-
morph density. The slides are stored in the Department of
Geosciences, University of Oslo, Norway. Palynomorph preser-
vation is better for the Botneheia section than the Juvdalskampen
section. The palynomorphs have colours in the range of 3–5 on
the thermal alteration scale (TAS) of Batten (2002). Palynomorph
identification was mainly based on the works of Schulz (1967),
Morbey (1975), Bujak & Fisher (1976) and Bjærke & Manum
(1977). Photographs of selected palynomorphs are shown in
Figure 3. About 300 terrestrial palynomorphs were counted (quan-
titative analysis) per sample using a Leitz Diaplan microscope.

Photographs were taken with an AxioCam ERC 5s camera con-
ncected to a computer using Zen 2011 software. Sample processing
for the Juvdalskampen section was carried out by APT (Applied
Petroleum Technology AS, Kjeller, Norway). Relative abun-
dances were calculated and plotted using the Tilia/TiliaGraph and
TGView software (Grimm 1991–2001). Using CONISS (Grimm
1987) palynomorph assemblage zones were established by
constrained cluster analysis within Tilia (Figs 4 and 5). Two com-
plete slides per sample were completely scanned for further taxa
(qualitative analysis) to check for the presence of rare, biostrati-
graphically important taxa.

Principal component analysis (PCA) was calculated with the
program PAST (Hammer et al. 2001) based on the quantitative ter-
restrial palynomorph abundance data (Fig. 6).

Sporomorph Ecogroups and floral affinities

Although preservation, transportation and sedimentation effects
can occur, palynomorph assemblages largely reflect the composi-
tion of terrestrial plant communities (Muller 1959; Chaloner &
Muir 1968; Tyson 1995; Abbink et al. 2001; Traverse 2007). The
Sporomorph Ecogroup (SEG) model (Abbink 1998; Abbink
et al. 2001) can be used to interpret sea-level fluctuations. According
to this model, changes in sea-level directly affect vegetation. Abbink
(1998) and Abbink et al. (2004) defined six SEGs (Upland,
Lowland, River, Pioneer, Coastal, Tidal), each typical for a certain
depositional environment with different levels of stress and distur-
bance (Fig. 7). A hygrophytic/xerophytic (H/X) ratio is calculated
based on their climate affinities, following the concept of Visscher & Van der Zwan (1981), which is a first-order approximation of a
humidity signal. It was later applied by Roghi (2004), Galfetti
et al. (2007), Heimhofer et al. (2012), Hochuli & Vigran (2010)
and Roghi et al. (2010). Following this classification all spores are
classified as hygrophytes together with the
Alisporites, Aulisporites
and
Cycadopites
groups. All remaining pollen are classified as
xerophytic. Changes of these ratios are used to interpret climatic
change in the depositional settings. Table 1 provides a list of the
SEGs and encountered palynomorphs with their floral affinities.
Organic carbon isotope analysis
Carbon isotope values ($\delta^{13}C_{org}$) from bulk sedimentary organic matter ($\delta^{13}C_{org}$) for the Juvdalskampen and Botneheia sections were published by Mueller et al. (2014). In addition, 74 samples from the Vendomdalen section were analysed for $\delta^{13}C_{org}$. One gram of sediment was crushed and treated with 1M hydrochloric acid and left for 24h to remove all inorganic carbon. The samples were then neutralized with water and oven dried at 60°C. The homogenized samples were analysed with an elemental analyser–isotope ratio mass spectrometer (EA-IRMS; Europa Scientific 20-20 IRMS). Isotope ratios are reported in standard delta notation relative to Vienna PDB. The analytical precision based on routine analysis of internal laboratory reference materials indicates a standard deviation of <0.08‰. IA-R001, wheat flour was used as reference material ($\delta^{13}CV_{\text{PDB}} = -26.43\%$). The standard deviation of the standard was 0.05. The measurements were carried out by Iso Analytical Ltd (Crewe, UK).

Results
Palynology
Terrestrial versus aquatic palynomorphs
The Juvdalskampen and Botneheia sections are generally dominated by terrestrial palynomorphs with an increase in aquatic palynomorphs at the base and tops of the sections (Fig. 7). The aquatic fraction of the upper part of the Botneheia Formation and lower interval part of the Tschermakfjellet Formation at the Botneheia section is dominated by up to 95% freshwater algae Botryococcus. Above the freshwater algae dominated interval acritarchs of Micrhystridium spp. dominate the lower Tschermakfjellet Formation, with occurrences of up to 15% at Juvdalskampen (between 0 and 8 m) and at Botneheia (below 265 m). Above this, both sections are dominated (about 95%) by terrestrial palynomorphs. In both sections the top of the Tschermakfjellet Formation shows a second increase in aquatic palynomorphs, especially acritarchs of Micrhystridium spp. (40–60 m at Juvdalskampen and 285–230 m at Botneheia). The De Geerdalen Formation is dominated by terrestrial palynomorphs. Above 180 m in the De Geerdalen Formation at the Juvdalskampen section (to the top of the formation), freshwater palynomorphs increase in abundance to 50%. The interval above this maximum consists mainly of Botryococcus sp. This maximum coincides with a generally lean palynomorph interval. In the uppermost samples from the Knorringfjellet Formation at Juvdalskampen above 330 m, aquatic palynomorphs make up about 15% and contain mostly the acritarch Veryhachium and dinoflagellate cyst Heibergella asymmetrica. Overall, the amount of aquatic palynomorphs is slightly
Fig. 4. Relative sporomorph abundances at the Juvdalskampen section. Only the most abundant and biostratigraphically important taxa are shown. The grey-shaded portion of the curves is a 5x exaggeration of the abundances plotted in black.
Fig. 5. Relative sporomorph abundances at the Botneheia section. Only the most abundant and biostratigraphically important taxa are shown. The grey-shaded portion of the curves is a 5× exaggeration of the abundances plotted in black.
Aulisporites astigmosus is common in the lower part and its abundance decreases to the top of this assemblage. Spore abundance varies between 20 and 50%. Abundant forms are Calamospora and Deltoidispora. At 55 m there is an increase in Convolutispora and Concavisporites. The abundance of Leptolepidites decreases to the top of this assemblage.

The Concavisporites–Semiretisporites assemblage (J4) is characterized by a steady decrease in pollen, except for an increase in abundance at 85 m height in the section. Abundant palynomorphs are D. problematicus and Deltoidispora. Abundance of Calamospora and Carnisporites decreases throughout the assemblage zone. At 125 m there is an increase of K. cooksonae to 20% as well as an increase in Convolutispora, Atrarvisporites and Leptolepidites. Pollen are less abundant, with main taxa Brachysaccus and Chasmatosporites. Triadispora has its last occurrence within this assemblage zone.

The Leschikisporites–Kyrtonisporites assemblage (J5) is dominated by spore taxa. The base at around 145 m shows an acme of Leschikisporites aduncus and for the upper part of the interval above 320 m, taxa Concavisporites, Deltoidispora and Kyrtonisporites are abundant. The abundance of L. aduncus is decreasing throughout the assemblage zone. Chasmatosporites is present in moderate amounts, as well as Conbaculatisporites. The interval between 220 and 240 m is almost barren in palynomorphs.

**Terrestrial palynology of the Botneheia section**

At the Botneheia location the top of the Botneheia Formation was sampled with four samples, 20 samples were processed from the Knorringerfjellet Formation and five from the lower third of the De Geerdalen Formation. The pollen to spore ratios from the Juvdalskampen section are higher in the Botneheia section (up to 10%) than at the Juvdalskampen section.

The pollen to spore ratio for the whole section is low with less than 30% pollen for most of the section. At the base and at the top of the Tschermakfjellet Formation, two pollen maxima are recorded with up to 80% (maxima 1: 0–10 m above base and below 260 m at Botneheia; maxima 2: between 40 and 60 m at Juvdalskampen and between 290 and 300 m at Botneheia). At a stratigraphically higher level in the De Geerdalen Formation at 200 m there is another maximum of 40% pollen. Pollen abundance remains around 30% for the remainder of the Juvdalskampen section. The top of the De Geerdalen Formation is almost barren of palynomorphs. Based on cluster analysis, five assemblages were distinguished (Fig. 4), as follows.

The Striatoabietites–Protodiploxipinus assemblage (J1) is characterized by a low diversity of pollen and spores and is dominated by striate bisaccate (S. balmei) and non-striate bisaccate pollen (Aulisporites, Brachysaccus, Protodiploxipinus). Other abundant pollen taxa are Lunatisporites and Parvisaccites. Aulisporites astigmosus is also common. Spore abundance increases from base to the top from 10 to 50%. Abundant spore taxa are Concavisporites and Deltoidispora. Towards the top of this assemblage Punctatisporites becomes common. Above 5 m, the abundance of C. nathorstii increases to 4% and the abundance of Lunatisporites increases towards the top of the assemblage.

The Aulisporites–Camarozonosporites assemblage (J2) is characterized by a pollen to spore ratio of about 25%. The main characteristic for this zone is an increase in pollen A. astigmosus. Common pollen taxa are Concavisporites, Deltoidispora and K. cooksonae and main pollen taxa are Parvisaccites and Protodiploxipinus. Taxa Calamospora, Duplexisporites aduncus, Semiretisporites and Bactulisporites are also abundant. Pollen Brachysaccus is abundant in the lower part of the assemblage zone.

The Striatoabietites–Triadispora assemblage (J3) is dominated by taeniately bisaccate pollen of species S. balmei. Other abundant taxa are Brachysaccus, Aulisporites, Parvisaccites and Cycadopites. Aulisporites astigmosus is common in the lower part and its abundance decreases to the top of this assemblage. Spore abundance varies between 20 and 50%. Abundant forms are Calamospora and Deltoidispora. At 55 m there is an increase in Convolutispora and Concavisporites. The abundance of Leptolepidites decreases to the top of this assemblage.

The Concavisporites–Semiretisporites assemblage (J4) is characterized by a steady decrease in pollen, except for an increase in abundance at 85 m height in the section. Abundant palynomorphs are D. problematicus and Deltoidispora. Abundance of Calamospora and Carnisporites decreases throughout the assemblage zone. At 125 m there is an increase of K. cooksonae to 20% as well as an increase in Convolutispora, Atrarvisporites and Leptolepidites. Pollen are less abundant, with main taxa Brachysaccus and Chasmatosporites. Triadispora has its last occurrence within this assemblage zone.

The Leschikisporites–Kyrtonisporites assemblage (J5) is dominated by spore taxa. The base at around 145 m shows an acme of Leschikisporites aduncus and for the upper part of the interval above 320 m, taxa Concavisporites, Deltoidispora and Kyrtonisporites are abundant. The abundance of L. aduncus is decreasing throughout the assemblage zone. Chasmatosporites is present in moderate amounts, as well as Conbaculatisporites. The interval between 220 and 240 m is almost barren in palynomorphs.
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Additionally, the interval between 305 and 310 m contains abundant *K. cooksonae* and at 305 m pollen *Triadispora* sp. Although the interval above 375 m yields more pollen (e.g. *Lunatisporites* sp., *Protodiploxipinus* sp.), it is still dominated by spores (e.g. *Concavisporites* sp., *Deltoidispora* sp., *L. aduncus*). This section corresponds to the *Concavisporites–Semiretisporites* assemblage of the Juvdalskampen section because of the occurrence of several spores and pollen in both sections.

**Aquatic palynology**

The sampled interval of the Botneheia Formation contains dominant freshwater algae *Botryococcus* and occasional *Tasmanites* sp. The base and top of the Tschermakfjellet Formation are characterized by an increase in aquatic palynomorphs (Fig. 7). It also contains moderate amounts of acritarchs *Micrhystridium* and *Veryhachium* as well as occasional *Pediastrum* and *Pterospermella* and rare foraminifera test linings. In the De Geerdalen Formation, acritarchs *Micrhystridium* and *Veryhachium* occur in the section at Botneheia and the freshwater algae *Botryococcus* is abundant in the Juvdalskampen section with up to 40% occurrence (interval between 200 and 240 m). Towards the Knorringfjellet Formation a shift to more marine palynomorphs such as dinoflagellate cyst *H. asymmetrica* and moderate amounts of acritarch *Veryhachium* takes place. Overall the Botneheia section contains a higher total count of aquatic palynomorphs compared with the Juvdalskampen section.

**Principal component analysis**

The terrestrial palynomorph counts are shown as species scores on the first and second axes of a PCA ordination diagram (Fig. 6). The two axes are the dimensions through the dataset that explain the largest variance in species composition, interpreted to result from climatic or environmental factors that control vegetation type.

*Fig. 7. Stratigraphy, lithology, sample positions, δ¹³Corg, PCA axes, distribution of the Sporomorph Ecogroups (SEGs), ratio of hygrophytic to xerophytic floral elements, terrestrial to marine palynomorph ratio, pollen to spore ratio, aquatic palynomorph and total counts of aquatic palynomorphs for the (a) Juvdalskampen and (b) Botneheia section.*
Organic carbon isotope stratigraphy

At Vendomdalen, samples from two sections located c. 5 km apart were sampled. The bulk organic carbon isotope values ($\delta^{13}C_{org}$) for the older, east Milne Edwardsfjellet (MEE) section shows a trend from ~30‰ at 0 m to ~28‰ at 20 m (Figs 8 and 9). For the younger Dalsnuten (DA) section, the $\delta^{13}C_{org}$ has a larger variability and ranges between ~24 and ~30‰. The $\delta^{13}C_{org}$ values have two negative excursions in the Tschermakfjellet Formation whereas the De Geerdalen Formation contains four possible negative excursions to ~30‰. The spore colour is in the range 3–5 at Juvdalskampen and Botneheia, indicating that modification of $\delta^{13}C_{org}$ owing to heating from Cretaceous intrusions (e.g. on Edgeøya; see Brekke et al. 2014) is unlikely. No dolerite intrusions occur in the Triassic in Vendomdalen.

Juvdalskampen–Botneheia section correlation

Correlation between the Juvdalskampen and Botneheia sections is based on their palynomorph assemblages, lithostratigraphy and $\delta^{13}C_{org}$ stratigraphy (Fig. 7). Several bio-events can also be matched in the two sections. The correlation between the Botneheia and Juvdalskampen sections was quantified using sequence slotting (Clark 1985; Thompson & Clark 1989; Thompson et al. 2012), as implemented in the CPLSLOT Windows program (http://www.geography.lancs.ac.uk/cemp/resources/software/cplsot.htm); an objective method of quantitative correlation that can deal with multivariate datasets, including palynomorph counts. We used the palynology data to produce two likely correlation models, and independently the bulk $\delta^{13}C_{org}$ data to produce a third correlation model. The final height composite of $\delta^{13}C_{org}$ used the section heights at Juvdalskampen and correlated the heights at the Botneheia section to these, using an average of these three correlation models (Fig. 10). The negative carbon excursion from height 60 m in the Juvdalskampen section was assumed to be the same feature as seen at 296 m at Botneheia (Fig. 8). This provided the single correlation constraint for the sequence slotting. Rock-Eval pyrolysis (Mueller et al. 2014) does not reveal significant changes in the organic matter composition across these $\delta^{13}C_{org}$ fluctuations in the upper part of the Tschermakfjellet Formation, which we interpret to indicate that there is little organic matter compositional control on the $\delta^{13}C_{org}$ changes.

The success of the sequence slotting was evaluated by the $\Delta$ parameter, which takes values from greater than zero to about unity ($\Delta$ < 0.5 indicates very good similarity; Thompson & Clark 1989), and the random variable (RV) and modified RV (here called RV2) coefficients, which are somewhat similar to conventional regression coefficients (Smilde et al. 2009). The high degree of similarity in the correlation models (Fig. 10) using $\delta^{13}C_{org}$ or the palynological data independently indicates the robust nature of these correlations through the Tschermakfjellet Formation. For the palynological data the Euclidean distance metric (an equal weight metric; Gavin et al. 2003) produces the smaller $\Delta$, whereas the squared chord distance metric (a ‘signal-to-noise’ metric of Overpeck et al. 1985; Gavin et al. 2003) produces larger (i.e. better) association (RV, RV2) coefficients (Figs 9 and 10). Hence, these indicate better short-range similarity for the Euclidean model, and slightly better longer-range similarity for the squared chord model. The correlation relationships are much poorer (larger sensitivity statistics; Fig. 10) through the lower parts of the De Geerdalen Formation, which may be a reflection of the sparse data in both sections through this interval.

Paloynological correlation between sections

It is feasible to compare results from the Juvdalskampen–Botneheia sections with the semi-quantitative palynomorph counts at Vendomdalen (MEE and DA sections) of Hounslow et al. (2007a). The palynomorph assemblages from the MEE section and from the Striatoothiettes–Alisporites assemblage from Botneheia are characterized by an acme of various bisaccate pollen; mainly $S$. balmei and aquatic palynomorphs such as Tasmanites sp. and Michychrysidium. The base of the Tschermakfjellet Formation at Juvdalskampen is also characterized by an acme of various bisaccate pollen in the

Table 1. Classification of the sporomorph ecogroups (SEGs) and floral affinities

| SEG          | Taxon                     | Floral affinity* |
|--------------|---------------------------|------------------|
| Upland       | Angustiscusites klausii   | Xerophytic       |
|              | Brachysaccus sp.          | Xerophytic       |
|              | Lunatisporites spp.       | Xerophytic       |
|              | Parvisaccites radiataus   | Xerophytic       |
|              | Podosporites sp.          | Xerophytic       |
|              | Protodiploxypinus ssp.    | Xerophytic       |
|              | Quadraculina annelleiformis|                 |
|              | Striatoothiettes baltica   | Xerophytic       |
|              | Tripadiospora sp.         | Xerophytic       |
| Upland or tidal| Aulisporites spp.         | Hygrophytic      |
| Upland or unknown | Ovalipollis sp.         | Xerophytic       |
| Lowland      | Aulisporites astigmosus    | Hygrophytic      |
|              | Baculatisporites sp.      | Hygrophytic      |
|              | Calamospora nathorstii    | Hygrophytic      |
|              | Conbaculatisporites sp.   | Hygrophytic      |
|              | Convisporites sp.         | Hygrophytic      |
|              | Deltiodisporespia sp.     | Hygrophytic      |
|              | Duplexisporites problematicus|                 |
|              | Kyrtomospores sp.         | Hygrophytic      |
|              | Punctatisporites sp.      | Hygrophytic      |
|              | Chasmatosporites spp.     | Hygrophytic      |
|              | Cycadopites sp.           | Hygrophytic      |
|              | Porcellispora longdonensis| Xerophytic       |
| Lowland or river | Convolusporas sp.      | Hygrophytic      |
|              | Kraeuwelisporites sp.     | Hygrophytic      |
|              | Leschkesporis aduncus     | Hygrophytic      |
|              | Punctatisporites walkomii | Hygrophytic      |
|              | Zehrasporites sp.         | Hygrophytic      |
|              | Tetrassacus sp.           | Xerophytic       |
| Lowland, river or tidal | Camarozonosporites spp. | Hygrophytic      |
|              | Carnisporites sp.         | Hygrophytic      |
|              | Semiretisporites sp.      | Hygrophytic      |
| River        | Annullispora sp.          | Hygrophytic      |
|              | Apatitisporites sp.       | Hygrophytic      |
|              | Leptolepidites major      | Hygrophytic      |
| Tidal        | Densospores sp.           | Hygrophytic      |
|              | Retritiles sp.            | Hygrophytic      |
| Unknown      | Gibesporisporites hirsutus | Hygrophytic      |
|              | Thomosporites toralis     | Hygrophytic      |
|              | Schizaeosporites worsleyi | Xerophytic       |

* Affinities from Visscher & Van der Zwan (1981), Abbink (1998), Abbink et al. (2004), Roghi (2004), Galletti et al. (2007), Barron et al. (2010), Hochuli & Vigran (2010), Roghi et al. (2010), Schrank (2010), Gedi & Zija (2012) and Petersen & Lindström (2012).


**Boreal climate and stratigraphy during CPE**

*Striatoabietites*–*Alisporites* assemblage. Terrestrial palynomorphs dominate the *Aulisporites*–*Camarozonosporites* (Juvdalskampen) and *Aulisporites*–*Concavisporites* (Botneheia) assemblages and spores make up about 85% of the terrestrial palynomorphs. *A. astigmosus*, *Concavisporites* and *Deltoidispora* show an increase in abundance. *A. astigmosus* has a maximum at 25 m at Juvdalskampen, at 270 m at Botneheia and at 280.7 m at the Dalsnuten section. The Juvdalskampen and Botneheia sections at greater heights are characterized by a low pollen to spore ratio and less negative δ13Corg (between 280 and 290 m at Botneheia and between 10 and 45 m at Juvdalskampen). *Kraeuselisporites cooksonae* has its first occurrence close to the base of this interval in the Juvdalskampen and Botneheia sections. The interval at 55 m at Juvdalskampen contains a similar palynomorph assemblage to the interval at 295 m at Botneheia (*Striatoabietites*–*Triadispora* and *Deltoidispora*–*Striatoabietites* assemblages). They are characterized by a maximum in bisaccate pollen of species *S. balmei*.

The palynomorph assemblages from the base of the De Geerdalen Formation are characterized by a low pollen to spore ratio and the assemblages consist almost entirely of terrestrial palynomorphs. The amount of *Aratripsorites* and *Concavisporites* increases at the base of both sections. At Botneheia, the last occurrence of *Triadispora* at 305 m is at a similar level within the formation to the last occurrence of *Triadispora* at 95 m at Juvdalskampen. Aquatic palynomorph abundance increases higher in the De Geerdalen Formation with mainly freshwater algae *Botryococcus* (200–203 m at Juvdalskampen and above 375 m at Botneheia). The palynological correlation between the sections for the De Geerdalen Formation is challenging owing to the low sampling resolution and differing sample spacing from the Dalsnuten and Juvdalskampen sections.

*Nagy et al.* (2011) described potentially reworked foraminifera from the base of the Knorringsfjellet Formation; this could also be the case for the palynomorphs at this level at Juvdalskampen. Comparison with *Hounslow et al.* (2007a) suggests that some of the samples from the upper Isfjorden Member may contain reworked taxa that would influence their palynological age assignment.

**Correlation to the Vendomdalen sections**

Correlation between the Juvdalskampen–Botneheia δ13Corg composite and δ13Corg at east Milne Edwardsfjellet was visually evaluated because of limited stratigraphic overlap. In the MEE section (which crosses the Botneheia–Tschermakfjellet Formation boundary), δ13Corg is lower compared with the other sections, which is interpreted to indicate a slightly older part of the Botneheia Formation than that seen in the Botneheia section. We matched the δ13Corg values at 261 m from Botneheia with the similar δ13Corg values at 16 m from the MEE section, to produce the composite through the Ladinian–Carnian transition (Fig. 9). The Botneheia Formation–Tschermakfjellet Formation boundary in central Spitsbergen is typically characterized by a hiatus (or highly condensed interval; *Hounslow et al.* 2007a), which explains the rapid reduction in δ13Corg values across this boundary.

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*Fig. 8. Stratigraphic correlation of the CPE, PCA axes and some Late Triassic spore and pollen markers of sections from Spitsbergen and time equivalents from Lunz am See in Austria (Mueller et al. 2015) based on δ13Corg bulk isotope values. Vendomdalen palynology from *Hounslow et al.* (2007a) and magnetostratigraphy from Hounslow & Muttoni (2010). (Note: sections not drawn to same scale.)*
At Dalsnuten we matched the $\delta^{13}C_{\text{org}}$ peak at $c.$ 270 m with the similar positive excursion at $c.$ 75 m in the Juvdalskampen section, along with the overlying match of negative excursions at $c.$ 290 and $c.$ 90 m, respectively. Attempting sequence slotting for matching to Dalsnuten was not feasible owing to widely varying spacing and sparse data at Juvdalskampen. The $\delta^{13}C_{\text{org}}$ height composite (Fig. 10) uses the relative heights of remaining samples at Dalsnuten, with respect to these correlation points. The correlation defined by the $\delta^{13}C_{\text{org}}$ stratigraphy indicates that the base of the De Geerdalen Formation is diachronous between these sections. This is not unexpected, as the definition of the base of the De Geerdalen Formation is defined by lithostratigraphy at the first significant sandstone bed. Diachronous Late Triassic lithostratigraphic boundaries have also been widely inferred in the Barents Sea successions (Riis et al. 2008; Hoy & Lundslien 2011; Lundslien et al. 2014; Red et al. 2014). Higher in the De Geerdalen Formation (above $c.$ 390–480 m at Dalsnuten), channel sandstones predominate in the section, and probably complicate an already data-sparse interval, owing to complex lateral relationships. The variations could also reflect carbon storage on the De Geerdalen floodplain, or a climatic imprint on the $\delta^{13}C_{\text{org}}$. Rock-Eval pyrolysis (Mueller et al. 2014) does not reveal significant changes in the organic matter composition across these $\delta^{13}C_{\text{org}}$ changes at Juvdalskampen.

**Discussion**

**Carnian carbon isotope stratigraphy**

The carbon isotope stratigraphy through the Carnian is not completely known; the most complete stratigraphic coverage comes from the $\delta^{13}C_{\text{carb}}$ of Korte et al. (2005), with more limited, but detailed, stratigraphic coverage of $\delta^{13}C_{\text{carb}}$ data from Dal Corso et al. (2012, 2015). These datasets show increasingly more positive values from the late Ladinian into the Julian-1 (Figs 9b and 11), a feature that is clearly displayed in our data. A negative carbon isotopic excursion (CIE; –1) in $\delta^{13}C_{\text{org}}$ may mark the Ladinian–Carnian boundary, or perhaps more complex relationships in the $\delta^{13}C_{\text{carb}}$ data (Fig. 11). A broad positive CIE in $\delta^{13}C_{\text{carb}}$ in the mid and upper parts of Julian-1 is matched with flat $\delta^{13}C_{\text{carb}}$ values, with some evidence of smaller positive excursions in $\delta^{13}C_{\text{carb}}$ (Fig. 9b). Our composite data through the Tschermakfjellet and lowest De Geerdalen Formation are inferred to match this broad positive CIE, in the age interval mid- to late Julian-1 (Fig. 9). This is compatible with the
inferred biochronology, derived from the magnetostratigraphy of Hounslow et al. (2007a). Our data also appear to display the positive CIE (+1) seen in the δ13C_{carb} data, which is also shown in δ13C_{org} in the Balatonfüred core (Fig. 11). The single sample spike in the δ13C_{carb} data from the St Cassian section may be anomalous, but the low Mn and high Sr content of the calcite suggests that it is diagenetically unaltered (Korte et al. 2005). This isotopic excursion may be amplified by the coeval changes in palynological composition in the upper Tschermakfjellet Formation.

A strong negative CIE (–2) during the early parts of Julian-2 is seen in δ13C_{carb}, δ13C_{org} (Figs 8, 9b and 11) and marine algal biomarker material (Dal Corso et al. 2015). The CIE (–2) appears to be displayed in our data from the lower parts of the De Geerdalen Formation (Fig. 9a). The recovery of δ13C_{org} from the CIE (–2) is less clear. The Milières–Dibona section appears to show complete recovery to more positive values by the late Julian-2 (Fig. 11), but the δ13C_{carb} shows continuation of the CIE until early Tuvalian-1 (Fig. 9b). This may reflect the disconnection between marine carbonate and atmospheric carbon sources (Dal Corso et al. 2015).

Following CIE (–2) the bulk δ13C_{org} shows rather more complex changes (Fig. 11), which show only partial recovery, involving much variability, but involving initial recovery after CIE (–2) to a weak positive CIE (+2). This behaviour is tentatively related to what we see in the mid parts of the De Geerdalen Formation, but is incompletely defined by our sparse data through this interval (Figs 9a and 11). The channel sandstones at Dalsnuten in the lower part of this interval may have given rather complex interrelationships to the Juvalsdalen section, so it is not clear if the simple height matching we used is appropriate in this part of the sections. The Julian-1 to Julian-2 boundary previously inferred in the De Geerdalen Formation was based on application of conodont biostратigraphy of the magnetostratigraphically matched sections of Hounslow & Muttoni (2010). The Lunz section of Dal Corso et al. (2015) has probably the best defined (i.e. by conodonts and ammonoids) Julian-1 to Julian-2 boundary interval, and therefore this is chosen as the best compromise in our final age model.

Following Lord et al. (2014), primarily using the magnetostratigraphy, we relate the Isfjorden Member at Dalsnuten to the latest Tuvalian-2 to Tuvalian-3 interval. Using this appears to show a correspondence between the δ13C_{carb} changes at Dalsnuten and the δ13C_{carb} changes of Korte et al. (2005) through this interval (Fig. 9), relating to the CIE numbered +3 and –3 in the Tuvalian. The δ13C_{carb} data from Silická Brezová showing the CIE (–3) excursion (with δ13C_{carb} < 3‰) are mostly from brachiopod data showing some diagenetic alteration (Korte et al. 2005), so there is less confidence about the reliability of this excursion. However, as Korte et al. (2005) noted, their apparently non-diagenetically altered data brachiopod data follow similar trends to diagenetically altered data through the Triassic, so the impact of the alteration may be small. Hence, we infer from these relationships that the magnetochron UT10 represents most of the Isfjorden Member at Dalsnuten, with the interval represented by UT6r to UT9n (late Julian-2 to late Tuvalian-1) missing at the disconformable base of the Isfjorden Member. The tentative δ13C_{carb} correlations proposed here allow for a refined age model for the Isfjorden Member–De Geerdalen Formation from that inferred by Lord et al. (2014).

**Palynological-based age assignment in the Boreal Carnian**

Several palynomorph range charts for the Boreal Triassic exist (e.g. van Veen 1985; Hochuli et al. 1989; Vigran et al. 2014), varying in their details through the Carnian. The assemblages from the lower Tschermakfjellet Formation correlate with assemblage F of Hochuli et al. (1989). Key forms present are Angusticoculates klausii, Podosporites sp., Protopilylinodina sp., S. balmei and Triadispora sp. These forms are typical for assemblages younger than Anisian according to Hochuli et al. (1989). However, Hochuli et al. (1989) based their age assignment on extrapolated ranges from the Tethys although they speculated that the age ranges in the Boreal region could be different, owing to the different latitudinal affinity having an impact on the true age ranges. The Tschermakfjellet Formation in

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**Fig. 10. Sequence slotting correlations between the Juvdalskampen and Botneheia (a) Correlation models using the palynological data (euclidean and squared chord distance metric models) and the δ13C_{carb} data (Euclidean distance metric). A blocking length of 3 was used in all models, and the low in the isotope curve in the upper part of the Tschermakfjellet Formation was used as an approximate initial correlation constraint. Δ = slotting statistic of Clark (1985), RV/RV2 = association coefficients (Smilde et al. 2009). Rp, Rs are conventional Pearson and Spearman correlation coefficients respectively for the isotope data. (b) Sensitivity statistic (reduction in combined path length, CPL), showing how sensitive the correlations are, if data from each level at Juvdalskampen are successively removed, and then the analysis re-run. Slotting most robust for palynological data below 60 m (Juvdalskampen) and less robust above.**
the MEE section contains both ammonoids and conodonts indicating the early Carnian, and when linked with the magnetostratigraphy in the Vendomdal sections (Hounslow et al. 2007a) indicates a Julian-1 (I) age for the sampled base of the Tschermakfjellet Formation from the Botneheia section.

The assemblages from the Striatoabietites – Protodiploxipinus (Juvdalskampen) and Striatoabietites – Alisporites assemblages (Botneheia section) belong to zone VIII of van Veen (1985) and assemblage F of Hochuli et al. (1989) based on the previously mentioned forms and the first occurrence of Chasmatosporites (Botneheia at 264 m). Taeniata bisaccate pollen (e.g. Lunatisporites, Striatoabieites) are common in palynozone VIII and were used for correlation (Fig. 2). The younger Aulisporites – Concavisporites (Juvdalskampen) and Aulisporites – Concavisporites (Botneheia section) assemblages are characterized by an increase in A. astigmosus. Hounslow et al. (2007a) reported this form to be dominant at 280.7 m at Dalsnuten. It correlates well with assemblage VII of van Veen (1985) to which an early Carnian (Julian) age was assigned. In addition, various forms of Kyrtomisporites have their first occurrence in this assemblage.

The Striatoabietites – Triadispora assemblage from the Juvdalskampen section records an increase in Brachysaccus and has a resemblance to assemblage D of Hochuli et al. (1989). Assemblage E is distinguished on quantitative criteria which were not recognized in the Juvdalskampen section.

The Concavisporites – Semiretisporites assemblage corresponds to zone VI of van Veen (1985), characterized by the presence of Striatoabietites spp., Schizaeosporites worsleyi, Triadispora sp. and Aratrisporites spp. The lower part of the De Geerdalen Formation (lower part of the Concavisporites – Semiretisporites assemblage) with the last occurrence of K. cooksonae and an increase in Triadispora corresponds to the top of assemblage D of ‘mid-Carnian’ age.

The bottom of the Leschikisporites – Kyrtomisporis assemblage at Juvdalskampen is characterized by an acme of L. aduncus (145 m). This acme was not recognized at the Botneheia section. Hounslow et al. (2007a) described L. aduncus to be common in the higher De Geerdalen Formation and dominant in the Isfjorden Member. The findings are in accordance with Van Veen’s zonation scheme of being typical for assemblage Va, to which he assigns a late Carnian (Tuvalian) age. A similar assemblage was described by Hochuli & Vigran (2010) in their floral phase 12, which is dominated by monolete spores based on a well from the southern Barents Sea. Higher in the section the abundance of L. aduncus decreases and the top of the De Geerdalen Formation is characterized by almost sporomorph barren samples (180–240 m Juvdalskampen). Van Veen (1985) described a similar sequence for assemblage Va. Lord et al. (2014) and Paterson & Mangerud (2015) also described a L. aduncus acme palynozone for the De Geerdalen Formation from Hopen. The last occurrence of
species *Thomsonisporites toralis* at 180 m at the Juvdalskampen section suggests an age older than that of assemblage C of Hochuli and of assemblage IV of van Veen. The occurrence of *S. balmei* at 380 m at the Botnehea section, together with *S. worsleyi* at 200 m at the Juvdalskampen section and *T. toralis* at 180 m at Juvdalskampen, suggests a Carnian age for the top of the sampled interval from the De Geerdalen Formation.

Vigran et al. (2014) introduced 15 composite assemblage zones for the Triassic of the Barents Sea and Spitsbergen. Most of the sections from this study correlate with their *Aulisporites astigmosus* Composite Assemblage Zone. This zone is, most of all, characterized by an abundance of *A. astigmosus* and *A. klausii*. Also, *L. aduncus* is common to abundant for some intervals. From the Barents shelf this assemblage was recognized in the middle part of the Snadd Formation (Hochuli & Vigran 2010). Vigran et al. (2014) correlated this zone to the assemblages G to D of Hochuli et al. (1989) and assigned an early to mid-Carnian age. Vigran et al. (2014) also recognized this assemblage from other sections of the Svalbard Archipelago (e.g. Isfjorden in Spitsbergen and Blankenhus in Edgøya).

The Knorringfjellet Formation in the upper part of the *Leschikisporites–Krytomsiporis* assemblage is characterized by an increase in *Kyrtomisporites spp.*, presence of *K. cooksanae* and the first occurrence of *Quadreaculina unellaiformis*. This suggests an affiliation to zone IVb of van Veen and *A. asymmetrica*. Also, an abundance is common to abundant for some intervals. From the Barents shelf this assemblage was recognized in the middle part of the Snadd Formation (Hochuli & Vigran 2010). Vigran et al. (2014) correlated this zone to the assemblages G to D of Hochuli et al. (1989) and assigned an early to mid-Carnian age. Vigran et al. (2014) also recognized this assemblage from other sections of the Svalbard Archipelago (e.g. Isfjorden in Spitsbergen and Blankenhus in Edgøya).

The interval from the late Ladinian to early Carnian could correspond to the *Aulisporites astigmosus* Composite Assemblage Zone of Vigran et al. (2014). The lower *Rhaetogonyaulax* spp. Composite Assemblage Zone below was not recognized, which is probably due to the hiatus between the top of the De Geerdalen Formation and base of the Knorringfjellet Formation at Juvdalskampen. The *Limbosporites landbladii* Composite Assemblage Zone is characterized by sporotypes and regular occurrences of dinoflagellates such as *H. asymmetrica*. Vigran et al. (2014) also assigned a Norian age to this assemblage.

**Development of the depositional environment**

The Kapp Toscana Group in Spitsbergen consists mainly of deltaic deposits forming an overall regressive setting (e.g. Mørk et al. 1982; Riis et al. 2008; Nagy et al. 2011). The top of the Botnehea Formation contains terrestrial palynomorphs and abundant aquatic palynomorphs such as *Botryococcus* and *Tasmanites*. This suggests deposition in a restricted shallow marine environment with a reduced salinity (Guy-Olson 1992; Krajewski 2008, 2013). A palynofacies analysis of this section (Mueller et al. 2014) shows abundant AOM typical for a marine environment with a restricted circulation (Tyson 1993, 1995). Marine palynomorphs include *Micryhystridium* and *Veryhachium*. The terrestrial palynomorph component records an increase in typical upland vegetation following the Sporomorph Ecogroup model (SEG; Abbink 1998; Abbink et al. 2004) in the *Striatoabietites–Protodiploxyphus* and *Striatoabietites–Aulisporites* assemblage zones (Fig. 7). Because upland vegetation consists mostly of bisaccate pollen grains it tends to be transported easily by wind (‘Neves effect’ described by Chaloner & Muir 1968; Abbink et al. 2001). This leads to a selective enrichment of upland vegetation forms in a marine setting and explains the apparent contradiction. The marine influx gradually decreases higher in the succession, with an exception for the top of the Tschermakfjellet Formation where marine palynomorphs such as *Micryhystridium* and *Veryhachium* increase in abundance. At this level upland vegetation becomes more abundant again (Deltoidispora–Striatoabietites and *Striatoabietites–Triadispora* assemblage zones). This interval could represent a short marine incursion within the main regressive depositional trend as already suggested by Nagy et al. (2011) for their biofacies MB2 representing a prodelta setting. However, we note that mainly only one bisaccate pollen type, *Striatoabietites*, is affected in this interval whereas other bisaccates remain rather stable. If the abundance of bisaccate pollen were influenced by sea-level changes alone, we would expect a more even increase in all bisaccate pollen types rather than an increase in only *Striatoabietites*. Therefore, we suggest that although there is some taphonomical transport effect in the pollen record, the pattern in *Striatoabietites* abundance is related to climate changes as indicated by the PCA 1 curve (see also discussion below).

Marine palynomorphs in the De Geerdalen Formation above are rare at the base and absent for the mid and upper part. The palynomorph records show a mostly lowland or river SEG. This combined with sedimentological observations from sandstone bodies suggests a delta plain setting with river deposits. This setting is also the reason why a direct correlation between the channel sandstone units within the De Geerdalen Formation is challenging. The youngest unit, the Knorringfjellet Formation, yields more marine palynomorphs of types *Veryhachium* and *Cymatosphaera* and dinoflagellate cysts of *H. asymmetrica* (around 330 m at Juvdalskampen). The terrestrial component consists mostly of a lowland SEG but does also record a slight increase in upland vegetation. This suggests that the Knorringfjellet Formation is separated from the De Geerdalen Formation by a transgression, and forms the base of a new depositional sequence in a shallow marine setting.

**Palaeoclimate and the Carnian Pluvial Event**

The principal component analysis (Fig. 7) suggests a decrease in temperature (PCA 1 curve) in the lower part of the Tschermakfjellet Formation and an increase at the top, coinciding with the negative CIE preceding the positive CIE (+1). These temperature fluctuations have not been reported elsewhere in the Boreal region. However, the Tethyan Early Carnian oxygen isotope curve shows a transient rise by about 3% within the late aon and aonides ammonite Zone indicating a cooler period during the early Julian-1 (Cordevolian) followed by warming during the late Julian-1 (Korte et al. 2005). These oxygen isotope fluctuations suggest a rapid temperature decline and increase that could well correlate with those indicated in the PCA 1 curve of the present study. The PCA 1 curve is mainly influenced by the abundance of the bisaccate *Striatoabietites*. The temperature maxima correlate with peak abundance of *Striatoabietites* of about 50–60% above the normal background abundance in the miospore assemblages extracted from the Tschermakfjellet Formation. Potentially, bisaccate pollen abundance can be also biased by sea-level changes (Neves effect as discussed above). However, the pollen record shows mainly changes in the proportion of *Striatoabietites* whereas other bisaccate pollen types show only minor variations. Therefore, we suggest that most of the PCA 1 variation is related to temperature changes rather than short-term sea-level fluctuations.

Concerning changes in humidity as indicated in PCA 2 curve, the climate was relatively wetter during the deposition of the Botnehea Formation (Longobardian) and changed to drier conditions during the deposition of the Tschermakfjellet Formation (Julian-1). From the lowest part of the De Geerdalen Formation the climate became increasingly humid until the top of the formation. The interval from the late Ladinian to early Carnian could correspond
to floral phases 10 and 11 of Hochuli & Vigran (2010). This interval is followed by a change at the base of the De Geerdalen Formation to dominant hydrophytic floral elements and an increase in monolete spores of type L. aduncus (floral phase 12 of Hochuli & Vigran 2010). The superabundance of L. aduncus, typical for swamp environments, along with the occurrence of coal beds and freshwater palynomorphs, such as Botryococcus, at the top of the De Geerdalen Formation, suggest a humid delta plain environment with rivers or lakes (Fig. 7).

The CPE in the Tethys has been described from the Julian-2-II A. austriacum ammonoid zone, there coincident with the A. astigmosus acme zone (e.g. Roghi et al. 2010; Dal Corso et al. 2012). However, in the Boreal region the A. astigmosus zone is located in the lower parts of the Tschermakfjellet Formation (Vigran et al. 2014; this paper), within the Tethyan T. aonoides ammonoid zone. Therefore, it appears that the A. astigmosus zone is older in the Boreal region (Fig. 8). It is likely that the Aulisporites-producing mother-plant, which was a hygrophytic gymnosperm (Bennetitales, Kräusel & Saarschmidt 1966; Balme 1993) migrated southwards over the time interval of the Carnian Pluvial Event when palaeoenvironmental conditions became favourable for its proliferation in the Tethys realm. In Spitsbergen, the mid to upper interval of the De Geerdalen Formation is contemporaneous with the CPE in the Tethys (Julian-2; e.g. Roghi et al. 2010; Mueller et al. 2015). This interval of the De Geerdalen Formation is characterized by terrestrially dominated lowland or river vegetation with the L. aduncus acme and an increase in humid conditions.

There is evidence in the Tethys realm for a vegetation shift from xerophytic floral elements in the early Julian to hydrophytic elements in the late Julian, with a return to xerophytic elements in the early Tuvalian (e.g. Roghi 2004; Hormung et al. 2007a,b; Rigo et al. 2007; Kolar-Jurkovsek & Jurkovsek 2010; Breda et al. 2009; Kozur & Bachmann 2010; Roghi et al. 2010). This study shows that time-equivalent events are reflected in the lower De Geerdalen Formation on Spitsbergen, an interval that consists mainly of spore-bearing plants that indicate a humid climate and has a Julian-1 to Julian-2 transitional age (Fig. 8). The carbon isotope perturbations documented in our Boreal sections correlate to those seen on the Palaeothetys margins (Fig. 9).

A possible cause for the CPE climate shift could be the eruption of the Wrangellia oceanic plateau (Jones et al. 1977; Greene et al. 2010; Dal Corso et al. 2012, 2015) that started in the latest Ladinian (Xu et al. 2014). This eruption resulted in the release of large amounts of carbon dioxide into the atmosphere, which probably caused the global climatic change to more humid conditions, with later perturbations of the carbon cycle represented primarily by CIE (~2). The sedimentological expression of the CPE in the Barents Shelf sediments is not entirely clear, as the prograding deltaic systems, sourced from the Urals, seem to have been stable and progressive systems since the early Triassic (Riis et al. 2008; Glørstad-Clark et al. 2010). However, the more humid event represented by the CPE may be expressed in the overall increased sedimentation rates, increased channel-sandstone development and incision rates. To some extent this is what we see at Dalsnuten, with the development of major channel-sandstone bodies immediately after CIE (~2) event, in the earliest Julian-2. However, the synchronicity of such major sandstone bodies remains to be discovered in other areas of Svalbard and the Barents shelf. There may also be coeval changes in palaeosol style and development associated with the humid CPE. On Edgeoya, major units of listric fault development characterize the lower part of the De Geerdalen Formation (Osmundsen et al. 2014), which may also be a response to delta-front overloading by high sedimentation rates stimulated by the CPE.

Conclusions

Rarity of marine index fossils prevents a detailed biostratigraphy correlation of the Late Triassic sediments from central Spitsbergen with Tethyan biozonations (Fig. 2). However, this study successfully integrates palynology, δ13Corg stratigraphy and magnetostratigraphy (Hounslow et al. 2007a) to produce a composite stratigraphy that provides a detailed understanding of the stratigraphic changes through the Kapp Toscana Group at Juvdalskampen, Botneheia and Vendomdalen (Fig. 8). The composite organic carbon isotope stratigraphy provides sufficient detail for a correlation to Carnian mid-latitude sections, in that it displays the major isotopic excursions seen in other mid-latitude sections, but in a lower Carnian interval with high sedimentation rates (Fig. 9). Palynology reveals nine assemblage zones (Figs 4 and 5), from two sections, that correlate with the existing stratigraphic schemes from van Veen (1985), Hochuli et al. (1989) and Vigran et al. (2014). The A. astigmosus acme zone from the Tethys of early Julian-2 age, which characterizes the onset of the CPE in the Tethys, has a Julian-1-1 age in the Boreal realm (Fig. 8). The integrated stratigraphy reveals a diachronous age for the beginning of the De Geerdalen Formation, the base of which is around the upper boundary of the Tethyan ammonoid Aon biozone. The remainder of the De Geerdalen Formation (below the base of the Isfjorden Member) is located in the substage interval upper Julian-1 to mid-Julian-2. Organic carbon isotopic excursions in the Isfjorden Member confirm the latest Tuvalian-2 to Tuvalian-3 age of this member, previously inferred using magnetostratigraphy (Fig. 9). Marine palynomorph taxa in the Knorringsjellet Formation suggest a latest Carnian or Norian age for the top of the sections.

Application of the Sporomorph Ecogroup (SEG) model and the ratio of marine to aquatic palynomorphs suggest a restricted marine with freshwater influx setting for the base of the sections from Botneheia (Fig. 7). The Kapp Toscana Group is characterized by an overall regressor trend with marine conditions for the base of the Tschermakfjellet Formation and change from an upland SEG to a lowland and river SEG. An interval close to the top of the Tschermakfjellet Formation records a short return to more marine influx. The De Geerdalen Formation is terrestrialydominated with mostly lowland or river SEGs. Marine palynomorphs (dinoflagellate cysts and acritarchs) in the Knorringsjellet Formation above suggest a return to shallow marine conditions at the top of the section.

Multivariate statistical analysis reveals a relatively drier and warmer climate for the Botneheia Formation and lower Tschermakfjellet Formation of Julian-1 age (Figs 7–9). This is followed by a cooler phase and a short increase in temperature during the upper negative carbon isotope excursion in the Tschermakfjellet Formation. Humidity increased around the Julian-1–Julian-2 boundary in the De Geerdalen Formation. Botanical affinities show a general dominance of hydrophytic floral elements. In Spitsbergen a dominance of spores and occurrence of local coal beds suggest a humid swamp or marsh setting during the late Julian, which is the Boreal equivalent of the Carnian Pluvial Event known from the Tethys in the De Geerdalen Formation. A clearer sedimentological expression of the CPE in the Carnian sediments of Svalbard awaits further exploration.

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