**LATE PERMIAN–EARLIEST TRIASSIC HIGH-RESOLUTION
ORGANIC CARBON ISOTOPE AND PALYNOFACIES
RECORDS FROM KAP STOSCH (EAST GREENLAND)**

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Abstract

During and after the end Permian mass extinction terrestrial and marine biota underwent major changes and reorganizations. The latest Permian and earliest Triassic is also characterised by major negative carbon isotope shifts reflecting fundamental changes in the carbon cycle. The present study documents a high-resolution bulk organic carbon isotope record and palynofacies analysis spanning the latest Permian–earliest Triassic of East Greenland. An almost 700 meters thick composite section from Kap Stosch allowed discriminating 6 chemostratigraphic intervals that provide the basis for the correlation with other coeval records across the world, and for the recognition of basin wide transgressive-regressive events documenting tectonic activity during the opening of the Greenland–Norway Basin. The identification of the main factors that influenced the organic carbon isotope signal during the earliest Triassic (Griesbachian to Dienerian) was possible due to the combination of bulk organic carbon isotope, palynofacies and Rock-Eval data. Two negative carbon isotopic shifts in the Kap Stosch record can be correlated with negative shifts recorded in coeval sections across the globe. A first negative shift precedes the base of the Triassic as defined by the first occurrence of the conodont *Hindeodus parvus* in the Meishan reference section, and the second one coincides with the suggested Griesbachian–Dienerian boundary. This new organic carbon isotope record from the extended Kap Stosch section from the Boreal Realm documents regional and global carbon cycle signals of the interval between the latest Palaeozoic and the onset of the Mesozoic.

Keywords: East Greenland, latest Permian, earliest Triassic, $\delta^{13}\text{C}_{\text{org}}$ record, particulate organic matter, carbon cycle

1. Introduction

The Permian–Triassic boundary interval is characterized by a synchronous negative excursion in organic and inorganic carbon records on a world-wide scale (*e.g.*, Magaritz et al., 1988; Holser et al., 1989; Oberhänsli et al., 1989; Payne et al., 2004; Hermann et al., 2011). High-magnitude disturbances in the carbon cycle are known to continue throughout the entire Early Triassic. Based on organic and inorganic carbon isotopes, as well as on palynological and faunal evidence, several authors demonstrate the occurrence of severe environmental perturbations during the Early Triassic (Krull et al., 2000; de Wit et al., 2002; Payne et al., 2004; Brayard et al., 2006, 2009; Galfetti et al., 2007a,b; Hochuli et al., 2010a,b; Retallack et al., 2011; Hermann et al., 2011, 2012; Romano et al., 2012). However, expanded and well-preserved Early Triassic records are rare and are essential for the reconstruction of a comprehensive picture of environmental changes, both at global and regional scales. The most recent radio-isotopic ages show that the Early Triassic represents a time interval of about 5 million years (Ovtcharova et al., 2006; Galfetti et al., 2007a; Burgess et al., 2014; Baresel et al., 2015). Therefore the drastic environmental disturbances reflected in the carbon isotope and paleontological records took place during a relatively short time. Even more dramatic are changes within the lower part of this interval represented by the first two substages of the Early Triassic (Griesbachian and Dienerian) representing less than 1 million years (Ovtcharova et al., 2006; Galfetti et al., 2007a; Burgess et al., 2014; Baresel et al., 2015). High-resolution records of the carbon cycle are essential to fully

63 assess the timing and potential causes and effects of the environmental conditions in the
64 aftermath of the end-Permian extinction event.

65 The biogeochemical cycle of carbon is a complex interaction between the organic and
66 inorganic carbon pools. Some geochemical signatures stored in the sea water and/or atmosphere,
67 such as the CO₂ isotopic composition, can be recorded as biogeochemical proxies in the
68 sediments and can be used as accurate stratigraphic markers. Carbon isotope stratigraphy is
69 based on changes in the isotopic composition of carbonate and organic matter. After sorting out
70 regional effects and artefacts from the record, the shifts reflecting the global carbon cycle
71 represent a powerful stratigraphic tool (Weissert et al., 2008) for precise correlations of records
72 from different areas and different depositional settings.

73 Understanding of the Early Triassic environmental instability has been traditionally based
74 on moderate resolution records mainly from low latitude carbonate platforms in the Tethyan
75 Realm (*e.g.*, Clemmensen et al., 1984; Holser et al., 1989; Oberhänsli et al., 1989; Baud et al.,
76 1996; Morante, 1996). Nevertheless in the last decade an increasing number of high-resolution
77 studies were published, leading to a more wide-spread representation of Early Triassic records,
78 attempting to shed light onto this critical time period (*e.g.*, Payne et al., 2004; Galfetti et al.,
79 2007a,b; Grasby and Beauchamp, 2008; Hermann et al., 2010, 2011, Algeo et al., 2012; Grasby
80 et al., 2012, Clarkson et al., 2013).

81 The present study of material from Kap Stosch (Hold-with-Hope, East Greenland)
82 provides a detailed bulk organic carbon isotope ($\delta^{13}\text{C}_{\text{org}}$) record combined with palynofacies and
83 Rock-Eval analyses. The two latter are used to establish the respective contributions of terrestrial
84 and marine origin to the bulk organic matter, allowing to differentiate between regional and

global carbon signals expressed in the bulk isotopic data. The essentially siliciclastic succession of Griesbachian and Dienerian age from Kap Stosch provides so far the most expanded $\delta^{13}\text{C}_{\text{org}}$ and palynofacies records documented world-wide for this time interval with two horizons with marine fauna providing biostratigraphic tie points. This high-resolution record of an almost 700 meters thick section has been correlated with other records from East Greenland (Stemmerik et al., 2001; Twitchett et al., 2001), as well as from the Norwegian Trøndelag and Finnmark platforms (Hermann et al., 2010), from Arctic Canada (Grasby and Beauchamp, 2008; Grasby et al., 2012), South China (Cao et al., 2002), and the United Arab Emirates (Clarkson et al., 2013).

2. Geological setting

2.1. Overview of the Permian–Triassic paleogeography and lithostratigraphy of East Greenland

During Permian and Triassic times East Greenland was located in Northwest Pangea, on the western margin of a rift basin separating Greenland from Norway (Figure 1). During the Late Permian and Early Triassic the N-S oriented Greenland–Norway Basin was about 400 km wide (Doré, 1991). The reactivation of Late Carboniferous–Early Permian basin margin faults during the latest Permian–earliest Triassic, and the associated crustal extension created several sub-basins on both margins, which led to the deposition of thick siliciclastic sedimentary sequences (Surlyk, 1990; Peacock et al., 2000; Bugge et al., 2002; Seidler et al., 2004; Hamann et al., 2005; Oftedal et al., 2005).

Figure 1

Sedimentary rocks exposed along the East Greenland coast represent an ideal area for studying Upper Permian–Lower Triassic sequences (Figure 2A). Wordie (1927) was the first to document rocks of Permian–Triassic age in the Kap Stosch area. Subsequently, several studies dealt with general Permian–Triassic stratigraphy and paleontology (Koch, 1931; Nielsen, 1935), with the biostratigraphic subdivision of the Permian–Triassic beds based on ammonoids (Spath, 1935; Trümpy, 1961; Bjerager et al., 2006), and with palynology (Balme, 1979; Piasecki, 1984). More detailed lithostratigraphic descriptions of the formations were published by Perch-Nielsen et al. (1974), Surlyk et al. (1986), and Surlyk (1990). Numerous studies focused on the Permian–Triassic boundary (PTB) of East Greenland (Teichert and Kummel, 1972; Looy et al., 2001; Stemmerik et al., 2001; Twitchett et al., 2001; Wignall and Twitchett, 2002), and on the Greenland–Norway Basin evolution (Surlyk et al., 1984; Surlyk, 1990; Stemmerik, 1998; Seidler et al., 2004; Hamann et al., 2005; Müller et al., 2005; Oftedal et al., 2005).

Figure 2

The Late Permian lithostratigraphic succession of East Greenland comprises the Foldvik Creek Group with the Huledal, Karstryggen, Ravnefjeld, Wegener Halvø, and Schuchert Dal formations. The latest Permian–earliest Triassic succession consists of the Wordie Creek and the Pingo Dal formations (Figure 3) (Perch-Nielsen et al., 1974; Surlyk et al., 1986; Surlyk, 1990). The Permian Huledal Formation resting unconformably on Carboniferous deposits is composed of alluvial conglomerates including the onset of marine influence in its uppermost part. It is overlain by the marine hypersaline carbonates and evaporites of the Karstryggen Formation. Both the Huledal and the Karstryggen formations are of Capitanian age (Surlyk et al., 1986). In areas representing basin margins and paleo-highs the Karstryggen Formation is overlain by carbonate build-ups of the Wegener Halvø Formation. Black bituminous mudstones deposited in basinal areas have been defined as Ravnefjeld Formation, which contains bivalves (*Posidonia permica*), fishes (Nielsen, 1935) and few cephalopods (Perch-Nielsen et al., 1974). Abundant fauna such as brachiopods, crinoids and corals are known from the Wegener Halvø Formation.

Changhsingian to Dienerian time interval (Surlyk, 1990; Oftedal et al., 2005). Basin geometry, facies, and thickness of the Wordie Creek Formation vary from one sub-basin to another, reflecting the strong tectonic control on deposition (Seidler et al., 2004). In northern sub-basins (*e.g.*, Hold-with-Hope; Figure 2A) the high sedimentation rates, and the marked lateral thickness variations of the main sandstone bodies of the deltaic deposits indicate steep basin margins during the latest Permian–earliest Triassic (Oftedal et al., 2005). The deposition of the Wordie Creek Formation has been interpreted to represent the rift climax stage of the Greenland–Norway Basin (Clemmensen, 1980; Seidler et al., 2004; Oftedal et al., 2005). In some areas the base of the Wordie Creek Formation is marked by transgressive plurimetric conglomerate beds (Surlyk et al., 1984; Wignall and Twitchett, 2002; Seidler et al., 2004; Oftedal et al., 2005), and continues with the deposition of marine shales and mudstones with some intercalated sandstones and thin conglomerate beds. In the northern sub-basins this formation consists essentially of marine mudstones and shoreface sandstones, changing in its uppermost part to tidally influenced shallow marine siltstones and sandstones. In the southern sub-basins (*e.g.*, Jameson Land; Figure 2A) it is dominated by marine silts with increasing turbiditic input up section (Surlyk et al., 1984). The Pingo Dal Formation represents the top of the earliest Triassic basin infill. In the northern sub-basins its lower and its upper part are dominated by dark red, usually cross-bedded sandstones, while its middle part consists of red, pink and yellowish to white or variegated arkoses and arkosic conglomerates and breccias. In many localities where both, the Wordie Creek and the Pingo Dal formations are exposed, there is a gradual transition between the two formations (Perch-Nielsen et al., 1974).

2.2. Lithostratigraphy of the Hold-with-Hope sub-basin

At Kap Stosch, on the Hold-with-Hope peninsula, Upper Permian–lowermost Triassic sequences are exposed between 73°60'N–21°12'W and 74°04'N–21°43'W. Several creeks numbered by Koch (1931) cut through the succession (Figure 2B). Subsequently, Koch's numbering was used by Nielsen (1935) and Teichert and Kummel (1972).

This paper presents a composite section comprising about 700 meters of sediments of Late Permian to earliest Triassic age from Kap Stosch (Figure 3). The lowermost portion of the section comprises the basinal Ravnefjeld Formation, which is unconformably overlain by the deltaic sequence of the Wordie Creek Formation. The latter displays a general shallowing upward trend with coastal deposits near the top of the sequence. Two major sandstone bodies (SB II and SB III) defined and mapped by Nielsen (1935) can be easily followed in the field and allow for regional correlations. Sandstone interfingering with conglomerate foreset beds of a delta constitutes the lower part of SB II and massive sandstone beds characterize its upper part. On the other hand, SB III and other minor sandstone beds show cm- to plurimetric-scale channel-like structures that thin laterally towards the Southeast. General trends in grain size, sedimentary structures and thickness of SB II and SB III indicate a WNW-ESE direction of transport of the clastic material. In order to optimally cover the complete Late Permian to earliest Triassic interval, four sections (Black Ridge, Blåelv, Rævekløft, and Nasutdal) have been sampled at high-resolution (Figure 3 and Supplementary data 1, 2, and 3).

The **Black Ridge section** (River 14) comprises 110 meters of Upper Permian and lowermost Triassic sediments. The first seven meters of the section correspond to the Ravnefjeld Formation of Wuchiapingian (Late Permian) age. Dark and thin bedded siltstones and shales intercalated with some limestone beds are the dominant lithologies. The unconformably

overlying Wordie Creek Formation consists of dark grey and greenish siltstones intercalated with cm- to dm-scale sandstone beds. At some places small lenses of ~20-cm thick conglomerate underlie the base of the Wordie Creek Formation. In the uppermost part of the Black Ridge section, two sandstone bodies of about 20 and 10 meters respectively, which are separated by six meters of siltstones with intercalated cm-scale sandstone beds, correspond to SB II. Above SB II this section ends with seven meters of siltstones, and is covered above. So far no age-diagnostic macrofossils have been observed within this section (Supplementary data 1).

The **Blåelv section** (River 16) comprises about 400 meters of sediments of the Wordie Creek Formation. The accessible part of the section starts about five meters above SB II. Pods of gypsum have been locally observed immediately above SB II. The section is mainly composed of dark grey and greenish siltstones with intercalations of cm- to plurimetric-scale sandstone beds. About 60 meters above the base, there is a 20-meter thick interval of reddish siltstone beds ending with a sharp change to grey coloured beds. Towards the top of the section another increase in reddish over grey siltstone beds has been observed. Near the base of the section ammonoids belonging to the *Ophiceras commune* zone (Bjerager et al., 2006) are abundant, providing a first robust ammonoid age control (Supplementary data 1). Two horizons with extraformational, rounded pebbles are found at around 180 and 430 meters. They are interpreted as spill-overs of coarse material transported through distributary channels that were embedded in the fine grained and laminated sediments of the bottom lobes of the delta. Thin sandstone beds around 230 meters above the base of the Blåelv section correspond to SB III. Near the top of the section, at around 370 meters, there are two beds containing the bivalve *Claraia* sp. (Supplementary data 1).

The **Rævekløft section** is located between the rivers 9 and 10 starting immediately above SB III. It comprises about 50 meters of dark grey-green siltstones and mudstones intercalated with cm-thick sandstone beds. It represents the most completely exposed interval above SB III, thus complementing the Nasutdal (see below) and Blåelv sections. It includes beds with the ammonoid *Bukkenites rosenkrantzi*, the common bivalve *Claraia* sp., and the gastropod *Naticopsis* sp. (Brayard et al., 2015) (Supplementary data 2). The ammonoids are characteristic of the *Bukkenites rosenkrantzi* zone and provide a second robust ammonoid age control (Bjerager et al., 2006).

The **Nasutdal section** (River 10 and adjacent western ridge) comprises an approximately 400-meter thick sequence representing the upper part of the composite section. The base of this section has been logged in River 10, and it consists of 38 meters of siltstones with intercalated sandstone beds. At this location the overlying SB III is about 60 meters thick. The section above SB III was sampled on the western ridge adjacent to river 10. This section consists of dark grey-green siltstones intercalated with cm- to plurimetric-scale sandstone beds. Ammonoids (*Bukkenites rosenkrantzi*), the bivalve *Claraia* sp. and the gastropod *Naticopsis* sp. (Brayard et al., 2015) were found about 30 meters above SB III. From the middle part towards the top of the section there is a gradual increase of reddish siltstones and sandstones. In the middle part of the section, a ca. 50-cm thick bed contains abundant, poorly-preserved plant remains. Two beds in the upper part of the section contain bivalves. The upper one contains fragments of *Promyalina* sp. and other unidentified bivalves, while in the lower one none of the bivalves could be identified. Between these two beds a few isolated metric-scale build-ups of microbial limestone have been observed. The plant remains, the bivalve beds, and the microbial limestone build-ups occur within the two *Anodontophora* zones (Bjerager et al., 2006). The topmost part of the

Nasutdal section with predominance of red sandstones and siltstones probably reflects the transition between the marine Wordie Creek and the terrestrial Pingo Dal formations. However, the presence of acritarchs in the uppermost samples suggests a marine depositional environment up to the top of the section (Supplementary data 3).

2.3. The Permian–Triassic boundary in East Greenland

In East Greenland the PTB has traditionally been placed at the lithostratigraphical boundary between the Schuchert Dal and the Wordie Creek formations based on ammonoid zones (Tozer, 1994; Trümpy, 1961). In areas where the Schuchert Dal Formation is missing, the PTB has been interpreted to fall within a significant subaerial erosional hiatus (Birkenmajer, 1977; Seidler, 2000; Stemmerik et al., 2001). Thus, at Kap Stosch the PTB was described as an unconformable boundary between the Ravnefjeld and the Wordie Creek formations (Nielsen, 1935; Teichert and Kummel, 1972).

Paleontological data from conformable but rather poorly exposed PTB sections from Jameson Land describe the end-Permian extinction horizon in the top most part of the Schuchert Dal Formation (Oberhänsli et al., 1989; Stemmerik et al., 2001; Twitchett et al., 2001), and locate the PTB, based on the first occurrence (FO) of the conodont *Hindeodus parvus*, about 20 meters above the base of the Wordie Creek Formation (Twitchett et al., 2001). They correlated their geochemical $\delta^{13}\text{C}_{\text{org}}$ and $\delta^{13}\text{C}_{\text{carb}}$ data from this area with the PTB Global Stratotype Section and Point (GSSP) at Meishan in China (Jin et al., 2000; Yin et al., 2001). In Jameson Land the $\delta^{13}\text{C}_{\text{org}}$ and $\delta^{13}\text{C}_{\text{carb}}$ negative excursion is located above the extinction event, but below the FO of *H. parvus*. Therefore, according to the $\delta^{13}\text{C}$ records and paleontological data from Jameson Land,

the onset of the Wordie Creek Formation deposition begins during the late Changhsingian (Stemmerik et al., 2001; Twitchett et al., 2001).

3. Materials and methods

From the Kap Stosch sections a total of 605 samples were collected. In order to ensure sufficient amounts of organic matter for the analyses of bulk organic carbon isotope, palynofacies and Rock-Eval, sampling focused on grey-green siltstones to fine sandstone beds. A comparatively small number of samples from reddish siltstones were collected for comparison.

To analyze the bulk organic carbon isotope composition, all samples were finely ground and treated with 3 N HCl for at least 24 hours to remove all carbonates. The residue was homogenized and analysed with a ThermoFisher Flash-EA 1112 elemental analyser coupled with a ConFlo IV interface to a ThermoFisher Delta V isotope ratio mass spectrometer (IRMS). Isotope ratios are reported in the conventional δ -notation with respect to V-PDB (Vienna Pee Dee Belemnite) standard. The system was calibrated with NBS22 ($\delta^{13}\text{C} = -30.03$) and IAEA CH-6 ($\delta^{13}\text{C} = -10.46$). Reproducibility of the measurements is better than $\pm 0.15\%$. The organic carbon isotope data of the four sampled sections have been used to recognize tie points and overlaps, to ensure an accurate correlation and produce a robust composite section.

Palynofacies analysis has been performed on 162 samples to cover the entire record at a medium resolution. The samples have been crushed to mm-size pieces and treated with 3 N HCl and 70% HF to remove all carbonate and silicate particles (Traverse, 2007). The residues were sieved with an 11 μm mesh screen and mounted on slides for the analysis of the particulate

organic matter (POM). For each sample a minimum of 250 particles were counted using a transmitted light microscope. The organic particles have been classified into 10 categories: amorphous organic matter (AOM), marine palynomorphs (Prasinophyceae and acritarchs), membranes, degraded terrestrial organic matter (DTOM), terrestrial palynomorphs (pollen grains and spores), cuticles, translucent woody particles, fungal remains, opaque woody particles, and inertinite. In order to calculate the marine/terrestrial ratio, two categories account for the marine fraction (AOM and marine palynomorphs), whereas seven categories account for the terrestrial fraction (DTOM, terrestrial palynomorphs, cuticles, translucent and opaque woody particles, fungal remains, and inertinite). The POM classified as membranes includes flat, structured or unstructured, thin-walled particles of unclear origin, *i.e.*, algal and land plant tissues. In the present study membranes are excluded from the marine/terrestrial ratio. In contrast to membranes, cuticles show distinct structures of land plant leaves. POM classified as DTOM includes degraded organic matter with residual structures revealing its terrestrial origin. These structures also differentiate DTOM from AOM, which is structureless under the transmitted light microscope (Tyson, 1995).

Rock-Eval analysis has been performed on 120 samples. About 100 mg of ground and homogenized sample was subject to pyrolysis followed by the complete oxidation of the residual sample. A flame ionization detector measured the hydrocarbon released during pyrolysis, while CO₂ and CO were detected by infrared absorbance during both steps. In the applied standard cycle for the whole rock analysis (IFP, 2001), pyrolysis started isothermally at 300°C for 3 minutes, after that the sample was heated to 650°C. The oxidation step started isothermally at 400°C (3 minutes) and then heated up to 850°C. Organic carbon decomposition resulted in four main peaks: peak S1 (hydrocarbons released during the isothermal phase), peak S2

(hydrocarbons produced between 300 and 650°C), peak S3 (CO₂ from pyrolysis of organic matter up to 400°C), peak S4 (CO₂ released from residual organic matter below ca. 550°C during the oxidation step), and peak S5 (CO₂ released above ca. 550°C during the oxidation step). These peaks were used to calculate the amount of total organic carbon (TOC, $0.082(S1+S2)+S4/10$), the hydrogen index (HI, $S2/TOC$), and the oxygen index (OI, $S3/TOC$). The HI and OI data are proportional to the H/C and O/C ratios of the organic matter, respectively, and can be used for the classification of the organic matter in Van-Krevelen diagrams (Espitalié et al. 1985, 1986). The relative errors associated with HI and OI of source rock samples do not exceed 1.5 and 10.5 of the measured values, respectively (Behar et al., 2001). Moreover, minerals such as calcium carbonate and clay minerals may artificially decrease HI values of Type II kerogen (Espitalié et al., 1984; Spiro, 1991). Despite these limitations, the use of HI and OI values are valuable since it allows the characterization of the same bulk organic material as that analysed for $\delta^{13}C_{org}$, by contrast to palynomorph assemblages (Sluijs and Dickens, 2012), which generally represent only a small fraction of the bulk organic carbon. The organic carbon isotope, HI, and marine/terrestrial ratio data have been correlated using the ordinary least squares method with the Past3 program.

4. Results

4.1. Organic carbon isotope ($\delta^{13}C_{org}$) and palynofacies data

The composite section has been subdivided into 6 chemostratigraphic intervals based on marked shifts and trends observed in the $\delta^{13}C_{org}$ record. The relative abundances of the marine

and terrestrial POM reflected in the palynofacies analysis are described for each interval (Figure 4). Together the Black Ridge and Blåelv sections comprise the intervals 1–5 and the base of interval 6 (Supplementary data 1). The Rævekløft section comprises interval 5 (Supplementary data 2). The Nasutdal section includes part of interval 4, and intervals 5 and 6 (Supplementary data 3).

Interval 1. Interval 1 comprises the seven topmost meters of the Ravnefjeld Formation. It is characterized by $\delta^{13}\text{C}_{\text{org}}$ values around $-23.7\text{‰} \pm 0.7\text{‰}$. The boundary between interval 1 and 2 is marked by a negative shift of about 4‰ that correlates with the unconformity between the Ravnefjeld and the Wordie Creek formations. This unconformity corresponds to a change in lithology from dark mudstones with intercalated limestones to grey siltstones. A distinct purplish colour indicating subaerial weathering of the Ravnefjeld Formation beds marks the top of the unit. The marine POM is dominated by AOM around 50% and punctually up to 70%. The terrestrial fraction is represented by abundant opaque woody particles up to 70% and translucent woody particles up to 20%. The percentage of DTOM increases towards the top of the interval reaching values of around 30%. Within this interval terrestrial palynomorphs and acritarchs are regularly observed (Figure 4 and Supplementary data 1).

Figure 4

Interval 2. The basal part of the Wordie Creek Formation is characterised by a negative trend observed over ca. 80 meters of section with $\delta^{13}\text{C}_{\text{org}}$ values changing from -27‰ to -33‰ with an amplitude ranging from 2 to 4‰. This negative trend shows two $\delta^{13}\text{C}_{\text{org}}$ peaks, one at around 40 meters (-32‰), and another at around 90 meters (-33‰) above the base. The latter marks the top of the interval (Figure 4 and Supplementary data 1). In comparison with the underlying

Ravnefjeld Formation, POM assemblages of interval 2 show a distinct change. DTOM ranges from 20% to 60% within the interval 2, except at the very top where DTOM abundances are <10%. Terrestrial palynomorph abundances reach values up to 20%, and membranes reach up to 10%. Marine palynomorphs are regularly observed. Opaque woody particles abundances are reduced to ca. 10%, and translucent woody particles are observed in relative quantities of less than 10%. AOM abundances vary from 10 to 60%. Changes in AOM and DTOM abundances are reflected in variable marine/terrestrial ratio values, with an overall increasing trend and values around 70% at the top of interval 2 (Figure 4 and Supplementary data 1). Interval 2 ends within the siltstone beds intercalated in the SB II.

Interval 3. The base of interval 3 observed within the upper beds of SB II is represented by a rapid, positive shift from $\delta^{13}\text{C}_{\text{org}}$ values around -33‰ to -26‰ (Figure 4 and Supplementary data 1). Subsequently, $\delta^{13}\text{C}_{\text{org}}$ values continue in a positive trend from -27‰ to -24‰ before they abruptly drop again to values around -32‰ within the last seven meters of the interval. At the base of the interval, beds containing ammonoids belonging to the *Ophiceras commune* zone have been found (Bjerager et al., 2006) (Figure 4 and Supplementary data 1). POM assemblages are characterized by a dominance of terrestrial OM with high abundances of DTOM and common occurrence of opaque woody particles and terrestrial palynomorphs. At the base of the interval DTOM shows a distinct increase from ca. 20% to 40–50%, and an overall decrease of AOM from ca. 50% to 20%. At the top of the interval AOM abundances rise up to 90%, whereas DTOM abundances drop to <10%. Terrestrial palynomorph abundances show values between 10–20% and translucent woody particles ca. 10%. Opaque woody particles abundances increase from values around 10% to 40%. Terrestrial palynomorphs, translucent and opaque woody particles abundances drop close to 0% at the top of the interval. At the base of interval 3 the

marine/terrestrial ratio decreases from 60% to 10%. The values remain around 10–20% in the middle part of the interval. Coinciding with the sharp $\delta^{13}\text{C}_{\text{org}}$ negative shift near the top of the interval there is a significant increase in the marine/terrestrial ratio reaching up to 80–90%.

Interval 4. Interval 4 corresponds to increasing $\delta^{13}\text{C}_{\text{org}}$ values from around -32‰ to -25‰ over almost 200 meters of section, with high amplitude fluctuations up to 4‰. Interval 4 ends on top of SB III (Figure 4 and Supplementary data 1 and 3). POM assemblages are dominated by AOM near the base, followed by AOM fluctuating values in the middle and terrestrial dominance towards the top. POM assemblages of the first 50 meters show clear dominance of AOM up to 90–95% and DTOM values are mostly <10%, only occasionally reaching 20%. This results in a clear increase in the marine/terrestrial ratio up to values around 80–90%. Percentages of membranes, terrestrial palynomorphs, and opaque woody particles are close to 5%. In the middle part of the interval (from 200 to 280 meters in Figure 4) AOM decreases to values around 10–30% and then displays fluctuating values between 5–60%, while DTOM percentages increase up to ~40%, and shows also fluctuating values between 10% and 50%. Membranes and terrestrial palynomorphs percentages are around 10–15%, whereas translucent woody particles increase up to 10% and opaque woody particles up to 50%. In this part of the section the marine/terrestrial ratio is alternating between low and high values reflecting the changing abundances of AOM and DTOM (Figure 4). In the upper part of interval 4 (from 280 to 350 meters in Figure 4) POM assemblages show AOM abundances around 10%, increasing up to 50% in the uppermost portion of the interval. DTOM shows steady decreasing abundances from ca. 50% to <10%. Membranes and terrestrial palynomorphs abundances are <10%. Opaque woody particles show a significant increase from ~10% to 80%, followed by a decrease to ~20% in the topmost part of the interval. The marine/terrestrial ratios in this upper part of interval 4 are characterized by low

values (between 10–20%), increasing up to 50–60% in the uppermost part of this interval (Figure 4 and Supplementary data 1).

Interval 5. This interval is most completely exposed in the Rævekløft section. There is an overlap with the Blåelv and Nasutdal sections. $\delta^{13}\text{C}_{\text{org}}$ has been measured in all three sections (Figure 3 and 4, and Supplementary data 1, 2 and 3). In the Rævekløft section interval 5 starts directly on top of SB III. $\delta^{13}\text{C}_{\text{org}}$ values show a negative trend changing from ca. -25‰ to -34‰, with increasing amplitude from 2‰ to 5‰. The most negative values coincide with the fossiliferous beds containing the ammonoid *Bukkenites rosenkrantzi*, the bivalve *Claraia* sp. and the gastropod *Naticopsis* sp. (Brayard et al., 2015). These beds consist of finely bedded dark green-grey siltstones alternating with mudstones. POM assemblages of this interval show an increase in AOM from 10% to 90%, resulting in a significant increase in the marine/terrestrial ratio. DTOM values fluctuate around 10%, and decrease to values around 1% at the top of the interval. Terrestrial palynomorphs and membranes percentages decrease from ca. 15–20% to 5%, and opaque woody particles from 40% to 10% (Figure 4).

Interval 6. Interval 6 shows a positive trend of $\delta^{13}\text{C}_{\text{org}}$ values over the last 270 meters of section, increasing from around -31‰ to -24‰. The base of this interval corresponds to a 40-meter thick sandy interval with relatively positive $\delta^{13}\text{C}_{\text{org}}$ values around -25‰. Due to the rare occurrence of siltstone layers, palynofacies data from only one sample is available (at around 430 meters in Figure 4). This POM assemblage is dominated by opaque woody particles (ca. 80%). Above this level POM assemblages show variable abundances of AOM between 60% to <10%, with an overall decreasing trend in the AOM abundance. At some levels in the upper part of the interval marine palynomorphs show values up to 10%. DTOM abundances vary between 30% to <10%

and terrestrial palynomorphs between 25% to <10%. Translucent woody particles are usually represented in numbers below 10%, while opaque woody particles display an increasing trend from ca. 20% to almost 90% of the assemblage. The overall marine/terrestrial ratio fluctuates between 60% and <10%, with a decreasing trend to <10% at the top of the section. In the middle part of the interval, with $\delta^{13}\text{C}_{\text{org}}$ values fluctuating between -30‰ and -25‰, a ca. 50-cm thick sandy bed containing plant remains can be found (Figure 4 and Supplementary data 3). At about 560 and 620 meters there are two beds containing bivalves. The lower one contains the bivalve *Promyalina* sp. and other unidentified bivalves, and the upper one only unidentified bivalves (Figure 4 and Supplementary data 3). The lower bivalve-bearing bed shows $\delta^{13}\text{C}_{\text{org}}$ values around -24‰, whereas the upper ones show values around -27‰. Between the two bivalve-bearing beds a few isolated patches of microbial limestone have been observed.

4.2. Rock-Eval pyrolysis data

The combination of organic carbon isotope and palynofacies data provides a good control of the origin of the POM. However, the finer fraction (<11 μm) of the total organic matter present in a sample is lost during the POM preparation (see Materials and Methods). Figure 4 and 5 includes TOC, HI, and OI data from the Kap Stosch composite section. TOC values are generally very low, below 0.1%. Exceptionally, a few samples reach TOC values between 1% and 2%, at the boundary between interval 2 and 3 (ca. 90 meters) and at the base of interval 4 (ca. 180 meters). The HI follows a slight positive trend throughout the whole section from values ~50 to ~240 and punctually increasing up to ca. 450 at about 180 meters above the base of the section (base of interval 4). The HI data show higher variability in the upper part of the section. The OI values scatter around 200 in the lower 400 meters of the section, showing a small peak of ~500

at around 120 meters above the base (interval 3). In the upper part of the section, from 400 to 680 meters (interval 6), OI values show higher variation, between ca. 200 and 1000 (Figure 4). The cross-plot between HI and OI displayed in Figure 5a shows that the majority of the samples fall between OI values up to 600 and HI values up to 200. Just few samples show OI values above 600 and HI values above 200. The OI-TOC cross-plot (Figure 5e) shows a negative logarithmic relationship. The highest TOC values yield the lowest OI values. On the other hand, the HI-TOC cross-plot shows a linear positive trend (Figure 5f) with the highest TOC values corresponding to the highest HI data. The relation between $\delta^{13}\text{C}_{\text{org}}$ values and TOC and HI (Figure 5c) shows a correspondence between the most negative $\delta^{13}\text{C}_{\text{org}}$ values and the highest HI and TOC values. Figure 5d displays a linear relationship between the S2 and TOC, with S2 values ranging from 0 to 8. Samples with the highest TOC values (ca. 2%) show S2 values around 8. Finally, figure 5g represents the relation between the marine/terrestrial ratio and the HI. High HI values correspond to high marine/terrestrial ratios.

5. Interpretation and discussion

5.1. Controlling factors of the $\delta^{13}\text{C}_{\text{org}}$ record and significance of depositional environments

For the interpretation of a bulk organic carbon isotope record it is essential to understand which factors may influence the $\delta^{13}\text{C}_{\text{org}}$ values. In our record we identified the following:

Sampling and lithology. The sampling focused on grey-green siltstone and fine sandstone beds in order to ensure a sufficient amount of organic matter per sample. Few reddish siltstone beds

have been sampled to avoid major gaps in the section and to assess the degree of preservation bias of the organic matter from deposits exposed to oxidation. In the studied section palynofacies of sandstone dominated intervals usually show a strong terrestrial signal reflected by high abundance of opaque woody particles, indicating increased oxidation and/or reworking of the organic matter (see Figure 4). The most labile marine organic matter components (*e.g.*, AOM) are prone to a faster degradation when exposed to oxic conditions; therefore refractory terrestrial organic matter particles are enriched in reddish beds. Conversely, palynofacies from siltstone dominated intervals commonly show assemblages of well-preserved mixed marine and terrestrial POM. The increase in red over grey siltstone and sandstone beds in the upper part of the section indicating well oxygenated marine conditions is also reflected in higher OI values.

Organic carbon pools. Marine and terrestrial organic matter has different ranges of carbon isotopic values, therefore the origin of the organic matter must be considered when interpreting bulk $\delta^{13}\text{C}_{\text{org}}$ records. Furthermore, the isotopic composition of the atmosphere and the isotopic composition of the dissolved carbon in the ocean have a direct influence on the isotopic composition of the organic matter accumulated in the sediments. Changes in the atmospheric and ocean carbon pools cause synchronous changes of different magnitude in the $\delta^{13}\text{C}_{\text{org}}$ values of the marine and terrestrial end-members (Weissert et al., 2008). In the studied material the marine/terrestrial ratio tends to follow the main trends and shifts of the $\delta^{13}\text{C}_{\text{org}}$ record (Figure 4). In the observed pattern the most negative $\delta^{13}\text{C}_{\text{org}}$ values coincide with the highest marine/terrestrial ratios, and vice versa. For instance, in a cross-plot between the marine/terrestrial ratio and the $\delta^{13}\text{C}_{\text{org}}$ values with a r^2 value of ca. 1 would imply a total influence of the POM on the $\delta^{13}\text{C}_{\text{org}}$ values. Figure 5b shows the cross-plot between the $\delta^{13}\text{C}_{\text{org}}$ values and the marine/terrestrial ratio of the Kap Stosch record. The samples scatter between

$\delta^{13}\text{C}_{\text{org}}$ values of -21‰ and -33‰ and marine/terrestrial ratios between 0 and 95%. The regression line has an r^2 value of about 0.3, indicating that the origin of the POM influences the $\delta^{13}\text{C}_{\text{org}}$ signal partly. Additional vital effects in fractionation may explain the scatter of the data. However, considering the complexity of the studied system, an r^2 value of about 0.3 can be considered significant. Factors of global extent, such as the isotopic composition of the CO_2 and/or changes of the CO_2 concentrations in the atmosphere, are also affecting the carbon isotopic composition of marine and terrestrial organic matter. Another relevant correlation concerning the influence of the POM on the bulk organic carbon isotope values is shown in the cross-plot between the HI and the marine/terrestrial ratio (Figure 5g). The regression line has an r^2 value of about 0.4, indicating some degree of correlation between high HI values and high marine/terrestrial ratio. Therefore, some bias in the organic carbon isotope values might be due to the origin of the POM. If changes in $\delta^{13}\text{C}_{\text{org}}$ values are measured without knowledge of concomitant POM data it will remain unclear whether variations are related to biotic or environmental changes. Although some authors question the validity of $\delta^{13}\text{C}_{\text{org}}$ curve for global correlations, several studies (*e.g.*, Menegatti et al., 1998; Gröcke et al., 1999; Heimhofer et al., 2003; Weissert et al., 2008; Schneebeli-Hermann et al., 2013) show that $\delta^{13}\text{C}_{\text{org}}$ curves can be used for reliable chemostratigraphic correlations.

Most of the samples with TOC values below 0.15% show a strong terrestrial origin of the OM (kerogen type III) as seen in the HI/OI (Figure 5a) and in the S2/TOC plots (Figure 5d). The few samples showing TOC values between 0.15% and 2% fall within the kerogen type II field (Figure 5d), indicating mixed marine-terrestrial derived origin of the OM (Tyson, 1995). The highest TOC values coincide with the highest marine/terrestrial ratios and the most negative $\delta^{13}\text{C}_{\text{org}}$ values (Figure 5c). Samples with the most negative $\delta^{13}\text{C}_{\text{org}}$ and highest TOC and HI

values (Figure 5c) belong to kerogen type I and indicate marine origin of the OM (Figure 5b). Therefore the strongest marine influence is documented at the base of interval 4 and on top of interval 5. However OI data from samples with low TOC values ($<2\%$) are not reliable enough according to Cooper (1990) because low TOC values can yield a wide range of OI values (Figure 5e). In the Kap Stosch record only samples with TOC values above 0.15% show low variability in the OI values. Due to the general low TOC values in the Kap Stosch record, only TOC data showing $>0.15\%$ values is considered to yield reliable information.

Figure 5

Relative sea-level and transgressive-regressive cycles. In the present study the combination of the marine/terrestrial ratio inferred from the POM assemblages and sedimentological observations are used to identify transgressive and regressive episodes. Decreasing or low marine/terrestrial ratios define regressive episodes, and conversely, increasing or high marine/terrestrial ratios reflect transgressive episodes. Applying this relationship, three transgressive-regressive (T–R) cycles in the Wordie Creek Formation are shown in Figure 6. Interval 2 shows an overall increasing trend in the marine/terrestrial ratio and a parallel trend towards more negative $\delta^{13}\text{C}_{\text{org}}$ values. This interval is dominated by grey siltstones and includes the lower part of SB II. Interval 2 can be considered as a first transgressive episode (T1) in the studied section modulated by a short regressive phase (Figure 6). A maximum flooding surface is interpreted at around 90 meters. Almost the entire interval 3 is considered a regressive episode (R1 in Figure 6). It shows a decreasing trend in the marine/terrestrial ratio and less negative $\delta^{13}\text{C}_{\text{org}}$ values. R1 includes the top part of SB II and the short interval above comprises lithologies with increasing reddish over grey sediment colour. Beds containing ammonoids belonging to the *Ophiceras commune* zone (Bjerager et al., 2006) are located within the lower part of this regressive episode. The top of interval 3 is characterized by a sharp shift in the marine/terrestrial ratio and in the $\delta^{13}\text{C}_{\text{org}}$ values (ca. -8‰). It is interpreted as a second transgressive episode (T2 in Figure 6). A drastic change in the colour from reddish to grey siltstones occurs within T2 coinciding with a peak in TOC values up to ca. 2%. Interval 4 shows first high marine/terrestrial ratios, followed by variable ratios in the middle, and ending with low marine/terrestrial ratios coinciding with an overall decreasing trend of $\delta^{13}\text{C}_{\text{org}}$ values from ca. -32‰ to ca. -24‰. It has been interpreted as second regressive episode (R2 in Figure 6). Grey coloured siltstones dominate at the onset of this regressive episode, while in the upper part there

is an increase in reddish beds. The top of R2 is reached at the top of SB III at around 350 meters. The third transgressive episode starts above SB III and includes interval 5 (T3 in Figure 6). It is characterized by dark green-grey siltstones and mudstones intercalated with thin sandstone beds. This interval shows an increase in the marine/terrestrial ratio and a $\delta^{13}\text{C}_{\text{org}}$ negative trend decreasing from values of ca. -24‰ to ca. -34‰. This strong marine influence is also documented by the occurrence of the ammonoid *Bukkenites rosenkrantzi*, the bivalve *Claraia* sp., and the gastropod *Naticopsis* sp. (Brayard et al., 2015). Interval 6 starts with the deposition of a massive sandstone body, followed towards the top by grey and reddish siltstones with increasing predominance of reddish over grey coloured sediments. Oscillating marine/terrestrial ratios show a general decreasing trend and $\delta^{13}\text{C}_{\text{org}}$ values become more positive towards the top of the section. Thus interval 6 has been interpreted as a third regression episode (R3), which is modulated by short transgressive episodes.

The identification of three T–R cycles within the Wordie Creek Formation allows the interpretation of the Kap Stosch section in terms of depositional environments and basin evolution. The sedimentary sequence at Kap Stosch is characterized by shallow marine deltaic settings influenced by relative sea-level changes due to increased rifting activity (Teichert and Kummel, 1972; Surlyk, 1990; Oftedal et al., 2005). The deposition of the Wordie Creek Formation occurred partly during the synrift phase (Griesbachian) of the opening of the Greenland–Norway Basin (Surlyk et al., 1986; Seidler et al., 2004; Oftedal et al., 2005) and changes in the sedimentary sequence and depositional environment are reflected in the marine/terrestrial ratio (Figure 6). In the present study, distal depositional environments are interpreted as transgressive episodes, while proximal depositional environments are interpreted as regressive episodes (Figure 6).

Figure 6

The three above described T–R cycles of the Wordie Creek Formation are compared with studies of Triassic sequence stratigraphy in the Arctic (Figure 7). For the Triassic six second-order sequence boundaries have been defined in the Sverdrup Basin (Embry, 1997). Correlation with other globally distributed Triassic sections allowed Embry to consider the second-order boundaries observed in the Sverdrup Basin as global sequence boundaries. Applying this scheme the sedimentary hiatus between the Ravnefjeld and the Wordie Creek formations, traditionally known as the PTB, can be related to the second-order sequence boundary located close to the PTB in the Sverdrup Basin (Embry, 1997) as shown in Figure 7. According to this author the next sequence boundary was interpreted to be a third-order of latest Dienerian age. This excludes a correlation of the upper T–R cycles observed at Kap Stosch with any subsequent sequence defined by Embry (1997).

On the other hand, Skjold et al. (1998) defined five second-order sequence boundaries for the Triassic of the Southwest Barents Shelf. Their first second-order sequence boundary can also be correlated with the sedimentary hiatus between the Ravnefjeld and the Wordie Creek formations. Within this first second-order sequence (latest Permian to Spathian–Anisian in Figure 7), Skjold et al. (1998) described five third-order sequences, of which the first two comprise the Griesbachian and the Dienerian, respectively (Figure 7; T1-1 and T1-2 in Skjold et al., 1998). Thus, the first and the second T–R cycles from Kap Stosch fall within the range of the first third-order sequence of Griesbachian age (T1-1). The pronounced transgressive event T2 suggests a correlation with the maximum flooding surface of T1-1, and T3 might be correlated with the maximum flooding surface of the second third-order sequence of Dienerian age (T1-2 of Skjold et al., 1998) (Figure 7).

Figure 7

Mørk et al. (1989) and Mørk and Smelror (2001) applied the sequence stratigraphic scheme of Embry (1988, 1997) to the Svalbard–Barents Shelf and to the Eastern Siberian successions. These authors recognized 10 transgressive episodes along the northern margin of Pangea during the Triassic. All of them are also recognizable in the Sverdrup Basin (Embry, 1988; Mørk et al., 1989). Four of these episodes are well dated as earliest Griesbachian, earliest Smithian, earliest Anisian, and earliest Carnian, respectively and can be followed from the Arctic Canada (British Columbia and Sverdrup Basin) to Svalbard–Barents Shelf and to Eastern Siberia (Mørk, 1994). The earliest Griesbachian transgression can be correlated with the transgression in the basal part of the Wordie Creek Formation (T1, Figure 7). A Dienerian transgression has been

observed on Bjørnøya (Mørk et al. 1990) and on southernmost Spitsbergen (Worsley and Mørk, 1978; Nakrem and Mørk, 1991). This transgressive event seems to be coeval with T3 in the Kap Stosch section (Figure 7) suggesting that T3 corresponds to an at least regional and probably eustatically controlled event (Mørk, 1994).

5.2. Comparison of the $\delta^{13}\text{C}_{\text{org}}$ record from Kap Stosch and $\delta^{13}\text{C}$ records from Jameson Land (East Greenland)

Correlations of $\delta^{13}\text{C}$ isotope records from various areas enable us to differentiate between regional (basin-wide) and global signals. Figure 8 shows the correlation between two PTB sections in Jameson Land - East Greenland (Stemmerik et al., 2001; Twitchett et al., 2001) and the Kap Stosch section.

Figure 8

Two $\delta^{13}\text{C}_{\text{org}}$ records from stratigraphically complete sections from Fiskegrav (Jameson Land) cover the uppermost Permian Schuchert Dal Formation and the transition to the overlying Wordie Creek Formation (Stemmerik et al., 2001; Twitchett et al., 2001) (Figure 8). The depositional environment at Fiskegrav has been described as bioturbated, grey off-shore marine siltstones (Stemmerik et al., 2001). Biostratigraphic control of the latest Permian interval is provided by the spore peak marking the palynological PTB of Stemmerik et al. (2001), as well as faunal evidence including agglutinating foraminifera, brachiopods, and ammonoids (Stemmerik et al., 2001; Twitchett et al., 2001). Both $\delta^{13}\text{C}_{\text{org}}$ records from Jameson Land start with stable values around -24‰ in the Schuchert Dal Formation corresponding to values found in interval 1 in the Ravnefjeld Formation at Kap Stosch (Figure 8). Due to the sedimentary gap, interval 1 ends in the unconformity at Kap Stosch. However, as recorded in Fiskegrav, stable $\delta^{13}\text{C}_{\text{org}}$ values around -24‰ continue in the overlying Schuchert Dal Formation (Stemmerik et al., 2001; Twitchett et al., 2001). The record of Stemmerik et al. (2001) shows an essentially continuous decrease from $\delta^{13}\text{C}_{\text{org}}$ values around -24‰ to around -29‰ in the upper part of the Schuchert Dal Formation. The end of this first decrease in the $\delta^{13}\text{C}_{\text{org}}$ values has been correlated with the base of interval 2 at Kap Stosch. A second drop to $\delta^{13}\text{C}_{\text{org}}$ values around -32‰ above the boundary between the Schuchert Dal and the Wordie Creek formations occurs within interval 2. The $\delta^{13}\text{C}_{\text{org}}$ record of Twitchett et al. (2001) also shows values of around -24‰ in the upper part of the Schuchert Dal Formation. Near the base of the Wordie Creek Formation the values drop to about -33‰ similarly to the two-step negative shift shown by Stemmerik et al. (2001). The base of interval 2 is correlated with the part with $\delta^{13}\text{C}_{\text{org}}$ values between -25‰ and -30‰. The $\delta^{13}\text{C}_{\text{carb}}$ record of Twitchett et al. (2001) follows the general trend of their $\delta^{13}\text{C}_{\text{org}}$ record and drops from about -4‰ to about -10‰ near the formational boundary. Compared to the $\delta^{13}\text{C}_{\text{org}}$ curve, the

$\delta^{13}\text{C}_{\text{carb}}$ record includes slightly younger sediments showing a minor positive excursion to around -6‰ towards the top of the Fiskegrav section (Twitchett et al., 2001). The isolated occurrence of *Hindeodus parvus* reported by Twitchett et al. (2001) indicates a Griesbachian age for the upper part of interval 2 (Figure 8) (Zhang et al., 2014; Brosse et al., 2015; Yuan et al., 2015). Both curves from Fiskegrav have been correlated with the GSSP PTB from Meishan (China) (Yin et al., 2001) and the end of the two-step negative shift has been interpreted as a global signal of the carbon cycle.

5.3. Comparison of the $\delta^{13}\text{C}_{\text{org}}$ record from Kap Stosch with other Early Triassic $\delta^{13}\text{C}$ records from the Boreal and Tethyan realms

Figure 9 shows the correlation of five globally distributed $\delta^{13}\text{C}$ records, three from the Boreal Realm: Norway (Hermann et al., 2010; Hochuli et al. 2010b), and Arctic Canada (Grasby and Beauchamp, 2008; Grasby et al., 2012), as well as two Tethyan records, one from South China (Cao et al., 2002) and a $\delta^{13}\text{C}_{\text{carb}}$ record from United Arab Emirates (Clarkson et al., 2013).

5.3.1. Norway

A record covering the latest Permian to earliest Triassic interval has been published from shallow cores from the Finnmark Platform (Hermann et al., 2010; Hochuli et al., 2010b). The Permian part of this record has been described as carbonates and siltstones deposited in shallow to outer shelf environment, while the Triassic part is interpreted as shallow, storm-influenced shelf deposits with alternating silty sandstones and sandstones (Bugge et al. 1995). From base to top it shows a two-step decrease of $\delta^{13}\text{C}_{\text{org}}$ values from about -24‰ to about -30‰. The base of the section with values of about -24‰ can be correlated with interval 1 from Kap Stosch. The

interval of continuous decrease of the $\delta^{13}\text{C}_{\text{org}}$ values (from ca. -24‰ to ca. -29‰) observed over about 30 meters in the Finnmark record is missing at Kap Stosch due to the sedimentary gap. The decrease in the $\delta^{13}\text{C}_{\text{org}}$ values below -30‰ characterizing the base of the Wordie Creek Formation (interval 2) at Kap Stosch can be observed in the Finnmark Platform (Figure 9). This pattern can also be recognised in the Trøndelag Platform (Hermann et al., 2010, see below), in Jameson Land (Stemmerik et al., 2001; Twitchett et al., 2001), and in Spitsbergen (Wignall et al., 1998). The spore peak observed in Jameson Land (Stemmerik et al., 2001) and on the Finnmark Platform (Hochuli et al., 2010b) falls between intervals 1 and 2. The Kap Stosch interval 2 correlates with the Finnmark section where the record shows $\delta^{13}\text{C}_{\text{org}}$ values around -29‰ and the second drop to values around -30‰ near the top of the section. So far the Finnmark Platform record represents the most complete and best resolved $\delta^{13}\text{C}_{\text{org}}$ dataset covering the PTB.

Figure 9

An expanded succession of Upper Permian–Lower Triassic has been recovered from the mid-Norwegian shelf (Trøndelag Platform). The sedimentary sequence has been interpreted to be a fully marine succession of sandstones, coarse-grained turbidites, shales and reworked sabkha (Bugge et al., 2002). It represents a thick sedimentary sequence similar to the one exposed at Kap Stosch according to Bugge et al. (2002). The Lower Turbidite Unit seems to be equivalent to the Ravnefjeld and Schuchert Dal formation in East Greenland, while the Upper Turbidite Unit corresponds to the Wordie Creek Formation (Figure 9). A sedimentary gap has also been described between the two units (Bugge et al., 2002). Hermann et al. (2010) subdivided the $\delta^{13}\text{C}_{\text{org}}$ record of this section into eight intervals (“a” to “h”). Their curve is closely comparable with the one from Kap Stosch (Figure 9). In both records the Late Permian section shows $\delta^{13}\text{C}_{\text{org}}$ values around -24‰ (Figure 9; interval 1 from Kap Stosch and interval “a” in Hermann et al., 2010). These values are followed by a sharp drop in the $\delta^{13}\text{C}_{\text{org}}$ values, corresponding to the sedimentary hiatus. The Kap Stosch interval 2 corresponds to the interval between the sedimentary hiatus up to the base of the sharp positive $\delta^{13}\text{C}_{\text{org}}$ shift around 320 meters corresponding to the intervals “c”, “d”, and “e” in the Trøndelag record (Hermann et al., 2010). The interval immediately above the sedimentary hiatus is characterised by $\delta^{13}\text{C}_{\text{org}}$ values around -27‰ followed by a decrease towards a minimum of ca. -32‰ at about 40 meters above the hiatus (intervals “c” and “d” in Hermann et al., 2010). The top of interval “d” is correlated with the first $\delta^{13}\text{C}_{\text{org}}$ minimum within interval 2 at Kap Stosch, and interpreted to represent the global end-Permian negative shift. The upper part of interval 2 of the Kap Stosch record shows a decreasing trend while in the Trøndelag $\delta^{13}\text{C}_{\text{org}}$ record the values remain stable around -29‰ (interval “e” in Hermann et al. 2010). In the Kap Stosch section, the upper part of interval 2 is dominated by the sandstone beds of SB II, whereas in the Trøndelag Platform succession

siltstones are the dominant lithology. Differences in the structure of the Greenland–Norway Basin may have led to different evolution of the depositional settings at the two margins, leading to different contributions of the marine and terrestrial POM fractions, as suggested by the $\delta^{13}\text{C}_{\text{org}}$ values and palynofacies records (Hermann et al., 2010). In the Kap Stosch section, interval 3 can be correlated with the Trøndelag interval “f” (Hermann et al., 2010). This interval is marked by the onset of a sharp $\delta^{13}\text{C}_{\text{org}}$ positive shift in both records, followed by a slight positive $\delta^{13}\text{C}_{\text{org}}$ trend at Kap Stosch and stable values in the Trøndelag record ending in both sections with a sharp negative $\delta^{13}\text{C}_{\text{org}}$ shift. In contrast to the hypothesis of Hermann et al. (2010), that the sharp shift in interval “f” might be due to a sedimentary gap, there are no sedimentological evidences for a hiatus, neither in the descriptions of the Trøndelag Platform well (Bugge et al., 2002) nor in the Kap Stosch section (Ofstedal et al., 2005). The stable $\delta^{13}\text{C}_{\text{org}}$ values observed in the upper part of the Trøndelag section (intervals “g” and “h” in Hermann et al. 2010) can be correlated with interval 4 from Kap Stosch. However, interval 4 shows a positive $\delta^{13}\text{C}_{\text{org}}$ trend at Kap Stosch, whereas the Trøndelag $\delta^{13}\text{C}_{\text{org}}$ values are rather stable. Similar to interval 2, the data from interval 4 suggest some differences in the carbon burial dynamics between the two margins of the Greenland–Norway Basin.

5.3.2. Arctic Canada

Two sections from the Sverdrup Basin in the Arctic Canada spanning the latest Permian to the Early Triassic are compared with the Kap Stosch record (Grasby and Beauchamp, 2008; Grasby et al., 2012). The depositional setting in the Sverdrup Basin has been interpreted to represent a deep basin surrounded by shallow shelves dominated by carbonate and chert prior to the Late Permian events (LPE) (Grasby et al., 2012), and by clastic sediments in the lowermost

Triassic (Grasby and Beauchamp, 2008). Intervals 1 and 2 can be correlated with the older section from Arctic Canada (Buchanan Lake) and the overall positive trend described as interval 6 at Kap Stosch is similar to the younger part of the Canadian section (Smith Creek) (Figure 9). Parts of the Griesbachian, corresponding to the intervals 3–5, are missing in the Canadian record (Grasby et al., 2012). The Buchanan Lake record starts with $\delta^{13}\text{C}_{\text{org}}$ values around -26‰, which are slightly more negative than the Kap Stosch interval 1 values. Despite the generally lower values at Buchanan Lake the pattern of the $\delta^{13}\text{C}_{\text{org}}$ curves are very similar. Following interval 1 the Canadian record displays a decreasing $\delta^{13}\text{C}_{\text{org}}$ values from about -26‰ to -29‰. In the Kap Stosch record this negative trend is missing, falling within the sedimentary hiatus between the Ravnefjeld and the Wordie Creek formations. However, it can be observed in the more complete sections from Jameson Land (East Greenland) (Stemmerik et al., 2001; Twitchett et al., 2001), and from the Finnmark Platform (Hermann et al., 2010). In the Buchanan Lake record stable $\delta^{13}\text{C}_{\text{org}}$ values around -27‰ correspond to the onset of interval 2, followed by a pronounced decrease to $\delta^{13}\text{C}_{\text{org}}$ values to around -30‰. This minimum has been correlated with the global end-Permian carbon negative shift (Grasby and Beauchamp, 2008) and it corresponds to the first $\delta^{13}\text{C}_{\text{org}}$ minimum within interval 2 in the Kap Stosch record (Figure 9).

For the younger Canadian record (Smith Creek) adequate age control could be obtained thanks to the common occurrence of conodonts and ammonoids. The fossil content of this section has been described by Tozer (1967), Baud et al. (2008), and Grasby and Beauchamp (2008). The Smith Creek record ranges from the latest Griesbachian to the late Smithian. For the purpose of correlation with Kap Stosch only part of this record has been reproduced in Figure 9. At Kap Stosch the presence of the ammonoid *Bukkenites rosenkrantzi* near the top of interval 5 indicates an early Dienerian age (Figure 8). For the basal part of the succession at Smith Creek

the conodonts *Clarkina planata* and *Clarkina carinata* indicate a latest Griesbachian age, whereas *Neospathodus dieneri* and *Neospathodus svalbardensis* provide an earliest Dienerian age (Orchard, 2007). Thus based on conodont assemblages Grasby and Beauchamp (2008) placed the Griesbachian–Dienerian boundary (GDB) a few meters above the base of the Smith Creek section (Figure 9). At Kap Stosch the occurrence of the Dienerian ammonoid *Bukkenites rosenkrantzi* within interval 5 suggests placing the GDB below these beds. Therefore the GDB has been tentatively placed at the boundary between interval 4 and 5 at Kap Stosch, coinciding with a pronounced turn in the $\delta^{13}\text{C}_{\text{org}}$ trend. Based only on the $\delta^{13}\text{C}_{\text{org}}$ data we correlate the Smith Creek record with the Kap Stosch interval 6. The protracted character of the Smith Creek record, and the gap between the latter and the Buchanan Lake record do not allow for precise correlations with Kap Stosch.

5.3.3. China

An attempt is made to correlate the record from Kap Stosch with the $\delta^{13}\text{C}_{\text{org}}$ record of the PTB GSSP at Meishan in China (Cao et al., 2002). The extremely condensed character of the Chinese section allows only the identification of the negative spike with $\delta^{13}\text{C}_{\text{org}}$ values around -31‰, and it is correlated with the first minimum $\delta^{13}\text{C}_{\text{org}}$ values in interval 2 (Figure 9). Consequently, the FO of the conodont *Hindeodus parvus* used as marker for the PTB would fall within interval 2 in Kap Stosch, after the end-Permian $\delta^{13}\text{C}$ negative shift in Meishan.

5.3.4. United Arab Emirates

A $\delta^{13}\text{C}_{\text{carb}}$ curve from the Tethyan Realm representing a shallow marine carbonate record from the Musandam Peninsula spans the Late Permian–Early Triassic interval (Clarkson et al.,

2013). Compared to the Kap Stosch record this data set comprises a comparable degree of resolution of the carbon isotope data, including also conodont age control. The Musandam Peninsula record starts with an interval of relatively positive $\delta^{13}\text{C}_{\text{carb}}$ values around 4‰. This is followed by a sharp negative shift that reaches the lowest values (ca. 1‰) in the entire record. These negative values are correlated with the first minimum in the $\delta^{13}\text{C}_{\text{org}}$ values of the Kap Stosch record (interval 2). However, in the Musandam Peninsula record *Hindeodus parvus* - the marker of the PTB - appears shortly before the minimum values, suggesting that the FO of *H. parvus* is diachronous, as suggested in other studies (Korte et al., 2010; Zhang et al., 2014; Brosse et al., 2015; Yuan et al., 2015). Relying on the chemostratigraphic evidence and using the FO of *Hindeodus parvus* as official marker of the PTB, this boundary falls probably within interval 2 or might even be represented by the sedimentary gap at Kap Stosch. After the negative shift the $\delta^{13}\text{C}_{\text{carb}}$ values of the Musandam Peninsula record recover to ca. 2‰. A second negative shift is observed near the GDB (Clarkson et al., 2013). In the Kap Stosch section this event most likely coincides with the negative values of interval 5 interpreted as the GDB (Figure 9).

5.4. Regional and global carbon cycle signals at Kap Stosch

The correlation with globally distributed carbon isotope records from the Late Permian–earliest Triassic allows differentiating regional and global carbon isotope signals in the $\delta^{13}\text{C}_{\text{org}}$ record at Kap Stosch. The most distinct features of the Kap Stosch and the Trøndelag Platform $\delta^{13}\text{C}_{\text{org}}$ records, representing the two margins of the Greenland–Norway Basin, such as the marked positive excursion and the subsequent fast decline (interval 3) have so far no equivalent in any other basin. Short-lived isotope excursions can be easily missed in coeval, low resolution records. Thus interval 3 is considered to reflect a distinct feature of the Greenland–Norway

Basin. It is interpreted as a first abrupt regression, followed by a period of relative low sea-level and finally ending with an abrupt transgression. A Griesbachian synrift phase of the opening of the basin has been postulated by Seidler et al. (2004) and Oftedal et al. (2005). The correlation between the HI and the marine/terrestrial ratio in the Kap Stosch section (Figure 5g) indicates that some degree of amplification in the most marine influenced parts of the $\delta^{13}\text{C}_{\text{org}}$ record has to be taken into account (*e.g.*, the transgression on top of interval 3 or interval 5) (Suan et al. 2015). Although interval 3 is interpreted as a regional carbon isotope signal, contribution of global carbon isotope variations cannot be completely dismissed. As summarised by Hermann et al. (2010) positive shifts close to the PTB have also been recorded in the $\delta^{13}\text{C}_{\text{carb}}$ records of Austria (Gartnerkofel, Holser et al., 1989) and China (Meishan, Xie et al., 2007). However, compared to the records in the Greenland–Norway Basin, the magnitude of the positive shifts is less pronounced in these sections, and the insufficient biostratigraphic control hampers the correlation of these isotopic events.

Two other features observed at Kap Stosch can be followed in different areas and are interpreted to represent variations of the global carbon cycle. The first is the negative shift in the carbonate and organic carbon values near the PTB. According to the chemostratigraphic correlations of the Kap Stosch section with sections with biostratigraphic control (Jameson Land, Musandam Peninsula and GSSP of Meishan), the base of the Triassic falls within the lowermost part of the Wordie Creek Formation (Figure 4). Thus, the first negative shift within interval 2 can be considered as a global carbon cycle signal. A second negative shift (interval 5) interrupting the general positive trend observed in the upper part of the section (intervals 4–6) is located just below the beds containing the ammonoid *Bukkenites rosenkrantzi* of Dienerian age. This suggests a correlation with the negative shift in the $\delta^{13}\text{C}_{\text{carb}}$ record of the Musandam Peninsula at

the base of the Dienerian (Clarkson et al., 2013). The following overall positive trend observed in the Dienerian of the latter record and in the record of the Canadian Arctic (Grasby et al., 2012) is similar to the one of interval 6 at Kap Stosch. The carbonate isotope record from South China (Payne et al., 2004) and the non-marine $\delta^{13}\text{C}_{\text{org}}$ record from Australia (Retallack et al., 2011) show a similar negative trend near the inferred base of the Dienerian. This leads to the conclusion that the shift of interval 5 represents a global negative isotope event close to the GDB.

5.5. Possible causes for the changing $\delta^{13}\text{C}_{\text{org}}$ patterns in the Kap Stosch record

The world-wide end-Permian negative carbon isotope excursion can be considered to be the first of several large perturbations of the Early Triassic carbon cycle (Payne et al., 2004; Galfetti et al., 2007a,b; Hermann et al., 2011, 2012; Retallack et al., 2005, 2011; Clarkson et al., 2013). Tethyan marine sections in South China, the North Indian Margin, and the Arabian Peninsula show major positive excursions across the Dienerian–Smithian, Smithian–Spathian, and Spathian–Anisian boundaries (Payne et al., 2004; Galfetti et al., 2007a; Hermann et al., 2012; Clarkson et al., 2013). A Gondwanan non-marine section from Australia (Retallack et al., 2011) also reported marked Early Triassic carbon cycle perturbations comparable to the marine records. Boreal sections from Spitsbergen (Galfetti et al., 2007b) and Arctic Canada (Grasby et al., 2012) show a positive excursion at the Smithian–Spathian boundary indicating that the Early Triassic carbon isotope variations are not restricted to the Tethyan realm. The magnitude of all carbonate and organic carbon isotope excursions is in the range of 4–8‰ during the latest Permian and Early Triassic.

The origin of the disturbances of the carbon cycle during the Early Triassic has been widely debated. A combination of several factors may explain the observed positive and negative excursions in the carbon records, rather than a single linear explanation (Payne and Kump, 2007). The Siberian large igneous province has been proposed to affect the stability of the carbon cycle during the Late Permian and Early Triassic (Renne et al., 1995; Berner, 2002; Reichow et al., 2009; Svensen et al., 2009). However, volcanic emissions alone would generate relatively small negative excursions in the carbon isotope record. Carbon isotope data of land plant cuticles show a negative shift illustrating a change in atmospheric carbon isotope composition across the PTB (Schneebeil-Hermann et al., 2013). Hence other sources of isotopically light carbon must have played an important role (Payne and Kump, 2007; Svensen et al., 2009). A hypothesis for the release of large amounts of ^{13}C -depleted carbon associated with the emplacement of the Siberian Trap is the metamorphism of organic-rich rock during volcanic intrusions (Svensen et al., 2009). Other concomitant factors that could cause negative carbon isotope excursions are marine productivity collapse caused by anoxia (Wignall and Twitchett, 1996), marine regression and oxidation of organic carbon on the exposed shelf (Holser et al., 1989), as well as destabilization of methane hydrates (Morante, 1996; Krull et al., 2004; Retallack and Krull, 2006). Sobolev et al. (2011) suggested that the Siberian Trap eruption alone could trigger the end-Permian mass extinction due to the massive degassing of CO_2 from recycled crust in the plume head. On the other hand, just a few mechanisms can explain rapid positive carbon isotope excursions since ^{13}C -rich sources are unknown. One mechanism could be an increase in organic carbon burial by increased primary productivity or increased preservation of organic matter under anoxic oceanic conditions (Kump and Arthur, 1999).

In the Kap Stosch record two global carbon isotope signals have been recognized. The negative shift within interval 2 has been correlated with the world-wide negative shift pre-dating the base of the Triassic (Figure 8 and 9). This negative shift is recorded in carbonate and organic carbon isotope records world-wide with nearly the same magnitude. Simultaneous rapid negative excursions in inorganic and organic carbon must have been triggered by an event affecting both the marine and terrestrial carbon pools. The most recent dating indicates that the main stage of the Siberian Trap emplacement partially pre-dates and partially is synchronous with the end-Permian events; late volcanic eruptive stages occur later in the Early Triassic. Therefore, it could have triggered the negative shift within interval 2 in the Kap Stosch record. The second global signal is the negative shift associated to the GDB (interval 5) (Figure 9). The most negative values of the Kap Stosch record (ca. -33‰) are documented at the end of interval 5. Corresponding shifts have been recorded in other sections, such as in South China (Payne et al., 2004), Australia (Retallack et al., 2011), and in the Musandam Peninsula (Clarkson et al., 2013 in Figure 9) documenting the global extent of this event. It can be explained by a renewed input of light carbon, probably from a further eruptive pulse of the Siberian Trap and/or the effect of positive feedbacks (*e.g.*, destabilization of methane hydrates) due to previous eruptive pulses. The base of interval 3 of the Kap Stosch record and the base of interval “f” of the Trøndelag Platform record (Hermann et al., 2010) represent a very fast positive excursion (from ca. -32‰ to -26‰). A possible explanation is that the carbon isotope positive shift is regionally magnified by the tectonic activity of the Greenland–Norway Basin. Likewise the negative shift occurring at the top of interval 3 may reflect a combination of regional and global causes.

6. Conclusions

The high-resolution Late Permian–earliest Triassic bulk organic carbon isotope ($\delta^{13}\text{C}_{\text{org}}$) record from East Greenland combined with palynofacies and Rock-Eval data are correlated with other world-wide distributed sections. These correlations allow differentiating regional and global carbon isotope signals. The palynofacies and Rock-Eval analyses were performed for two purposes. Firstly, they were used to identify probable causes for the observed $\delta^{13}\text{C}_{\text{org}}$ variations, and secondly, they represent the basis for defining transgressive-regressive cycles. Key points in this study are:

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1. The alternation of distal/proximal settings as indicated by palynofacies assemblages allowed the identification of three transgressive-regressive cycles within the Wordie Creek Formation at Kap Stosch. The transgressive-regressive cycles have been compared with the sequence stratigraphic scheme developed for the Sverdrup Basin in Arctic Canada and for the Norwegian Barents Shelf. The latest Permian sedimentary hiatus in the Kap Stosch section, formerly recognised as PTB, comprises the second-order sequence boundary defined close to the PTB in both areas, the Sverdrup Basin and Barents Shelf. The third-order sequence boundary, placed at the Griesbachian–Dienerian boundary in the Barents Shelf, correspond to the boundary between the second and third T–R cycle defined at Kap Stosch.

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- The subdivision of the $\delta^{13}\text{C}_{\text{org}}$ record from Kap Stosch into 6 chemostratigraphic intervals allowed the correlation with world-wide distributed carbon isotope records from the Boreal and Tethyan realms and the distinction between regional and global

875 signals. Despite the fact that the marine/terrestrial ratio follows the general $\delta^{13}\text{C}_{\text{org}}$
876 trends, the origin of the organic matter is not the main driver of the $\delta^{13}\text{C}_{\text{org}}$ variations.
877 The trends observed within the studied sections are summarized as follow:

- 878 • Interval 1 shows positive and rather stable $\delta^{13}\text{C}_{\text{org}}$ values as documented from other
879 PTB sections.
- 880 • Interval 2 contains the world-wide synchronous $\delta^{13}\text{C}$ negative shift that is
881 immediately followed by the FO of *Hindeodus parvus* in the Meishan GSSP section.
882 Therefore, using the carbon isotope signal as a correlation tool between the GSSP
883 reference section and East Greenland we suggest that the PTB should be placed ca. 50
884 meters above the unconformity between the Ravnefjeld and the Wordie Creek
885 formations implying a latest Permian (Changhsingian) age for the base of the Wordie
886 Creek Formation. However, according to the FO of *H. parvus* documented by
887 Clarkson et al. (2013) in the Musandam Peninsula section, the PTB falls below the
888 world-wide $\delta^{13}\text{C}$ synchronous negative shift, suggesting diachronism of the FO of this
889 marker with respect to global isotopic changes of the carbon cycle.
- 890 • Interval 3 displays a positive and negative excursion by ca. 7‰ that can be correlated
891 with the $\delta^{13}\text{C}_{\text{org}}$ record from the Trøndelag Platform. This feature shows that the
892 opening of the Greenland–Norway Basin during the Griesbachian affected carbon
893 burial of both margins similarly, resulting in a distinct regional component of the
894 carbon isotope signal. However, a contribution of the global carbon cycle on these
895 events cannot be completely excluded. Interval 4 shows a positive trend from $\delta^{13}\text{C}_{\text{org}}$

896 values ca. -32‰ to -25‰. So far the only comparable record for these two intervals
897 comes from the Trøndelag Platform (Hermann et al., 2010).

898 • The negative $\delta^{13}\text{C}_{\text{org}}$ shift defined as interval 5 is associated with the occurrence of
899 the ammonoid *Bukkenites rosenkrantzi*. Based on the correlation of the Kap Stosch
900 record with the $\delta^{13}\text{C}_{\text{org}}$ record of Arctic Canada and the $\delta^{13}\text{C}_{\text{carb}}$ record of the
901 Musandam Peninsula, the Griesbachian–Dienerian boundary is tentatively placed at
902 the boundary between the intervals 4 and 5.

903 • Interval 6 showing an overall positive trend of $\delta^{13}\text{C}_{\text{org}}$ values from ca. -34‰ to -24‰
904 is correlative with the biostratigraphically calibrated Dienerian section from the
905 Arctic Canada (Smith Creek), confirming a Dienerian age for the upper part of the
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Figure captions

Figure 1. Global paleogeographic reconstruction for the Early Triassic after Blakey (2012) and location of the study area and other areas used for correlation: (1) East Greenland, (2) Trøndelag Platform, (3) Finnmark Platform, (4) Arctic Canada, (5) Musandam Peninsula, and (6) South China.

Figure 2. A – East Greenland coast between latitudes 70°N to 74°30'N, location of the study area and other important sites (modified from Seidler et al., 2004). B – Kap Stosch area and the sampled sections: Black Ridge (river 14), Blåelv (river 16), Nasutdal (river 10), and Rævekløft (between river 9 and 10). Numbers of rivers according to Koch (1931) (modified from Teichert and Kummel, 1972).

Figure 3. Chrono- and lithostratigraphy of the Late Permian–earliest Triassic of Kap Stosch, based on Perch-Nielsen et al. (1974) and Bjerager et al. (2006). Vertical scale is arbitrary. Sampled sections - Black Ridge (orange), Blåelv (red), Nasutdal (green), and Rævekløft (blue) - and their relative stratigraphic position in the composite section. The two regional sandstone bodies SB II and SB III defined by Nielsen (1935) were used for the correlation of the sections during field work. The Permian–Triassic boundary interval is marked by a diagonal lines area, and the Griesbachian–Dienerian boundary is marked with a dashed line. The ages are inferred from U/Pb dating of zircons, 251.8 million years for the Permian–Triassic boundary (Burgess et al., 2014) and 251.2 million years for the Early Smithian (Galfetti et al., 2007a).

Figure 4. Bulk organic carbon isotope, palynofacies, total organic carbon, hydrogen index, and oxygen index data of the Kap Stosch composite section. From left to right: chronostratigraphy, the Permian–Triassic boundary interval is marked by a diagonal lines area, and the Griesbachian–Dienerian boundary is marked with a dashed line, asterisk (*) refers to the Pingo Dal Formation, and Pale. refers to Paleogene; simplified lithology of the composite section; macrofossil content; bulk organic carbon isotope data ($\delta^{13}\text{C}_{\text{org}}$) with isotopic intervals (1–black, 2–beige, 3–violet, 4–dark green, 5–light green, and 6–orange); particulate organic matter expressed in %, each bar represents 10%, AOM – amorphous organic matter, marine palynomorphs, DTOM – degraded terrestrial organic matter, membranes, terrestrial palynomorphs, cuticles, translucent woody particles, fungal remains, opaque woody particles, inertinite (marine categories are in red, membranes in grey, and terrestrial categories in green); the marine/terrestrial ratio in dark blue; TOC – total organic carbon; HI – hydrogen index; and OI – oxygen index.

Figure 5. a – Oxygen index (OI) and hydrogen index (HI) cross-plot, and the kerogen types (I, II, and III) fields as defined in Tyson (1995); **b** – Marine/terrestrial ratio and organic carbon isotope values ($\delta^{13}\text{C}_{\text{org}}$) cross-plot inferred from the particulate organic matter counting from Kap Stosch; **c** – Total organic carbon (TOC, black dots), hydrogen index (HI, white dots), and organic carbon isotope values ($\delta^{13}\text{C}_{\text{org}}$) cross-plot; **d** – S2 and total organic carbon (TOC) cross-plot, and the kerogen types (I, II, and III) fields as defined in Tyson (1995); **e** – Oxygen index (OI) and total organic carbon (TOC) cross-plot; **f** – Hydrogen index (HI) and total organic carbon (TOC) cross-plot; **g** – Hydrogen index (HI) and marine/terrestrial record cross-plot.

Figure 6. Transgressive-regressive cycles of Kap Stosch inferred from the particulate organic matter and interpretation of the depositional environments. From left to right: chronostratigraphy, the Permian–Triassic boundary interval is marked by a diagonal lines area, and the Griesbachian–Dienerian boundary is marked with a dashed line, asterisk (*) refers to the Pingo Dal Formation, and Pale. refers to Paleogene; simplified lithology of the composite section; macrofossil content; bulk organic carbon isotope data ($\delta^{13}\text{C}_{\text{org}}$) and isotopic intervals (as in Figure 4); marine/terrestrial ratio in dark blue; relative sea-level variations (solid line) and interpreted depositional environment in grey-scale; transgressive (T) and regressive (R) episodes and sequence boundaries within the Wordie Creek Formation.

Figure 7. Sequence stratigraphy scheme. Correlation of the Kap Stosch transgressive-regressive episodes of the Wordie Creek Formation with sequence stratigraphy studies of the Sverdrup Basin (Embry, 1997), the Barents Shelf (Skjold et al., 1998), Bjørnøya (Mørk et al. 1990) and southernmost Spitsbergen (Worsley and Mørk, 1978; Nakrem and Mørk, 1991). Vertical scale is arbitrary.

Figure 8. Correlation of the $\delta^{13}\text{C}_{\text{org}}$ record from Kap Stosch with $\delta^{13}\text{C}$ records from Fiskegrav - Jameson Land (East Greenland), $\delta^{13}\text{C}_{\text{org}}$ and $\delta^{13}\text{C}_{\text{carb}}$ records on the left (Twitchett et al., 2001), and $\delta^{13}\text{C}_{\text{org}}$ record in the middle (Stemmerik et al., 2001). The end-Permian C-isotope negative shift is used as baseline.

Figure 9. Correlation of the $\delta^{13}\text{C}_{\text{org}}$ record from Kap Stosch with $\delta^{13}\text{C}$ records from other areas. Boreal Realm: Trøndelag and Finnmark platforms (Hermann et al., 2010), Arctic Canada (Grasby and Beauchamp, 2008; Grasby et al., 2012); Tethyan Realm: South China (Cao et al.,

1300 2002), and the Musandam Peninsula (Clarkson et al., 2013). LPE: Late Permian events, GDB:
1301 Griesbachian–Dienerian boundary, DSB: Dienerian–Smithian boundary. The end-Permian C-
1302 isotope negative shift is used as baseline.