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Phanerozoic paleoenvironmental and paleoclimatic evolution in Svalbard

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Abstract. Sedimentary rocks can provide information about the Earth paleoenvironment and are studied extensively to understand the causes and consequences of global climate changes in deep time. They facilitate long-time perspectives that

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constrain climate models and provide analogues for how Earth systems may respond to, and recover from, intervals of profound environmental change, including projected anthropogenic change. The Norwegian Svalbard archipelago offers an extensive Phanerozoic stratigraphic record that reflects the geological evolution of the northern flanks of continental assemblages that include Laurentia, Eurasia, and Pangea. Svalbard's Phanerozoic sedimentary and paleoclimatic archive is controlled largely by Svalbard's overall northward plate-tectonic motion from equatorial to high latitudes, but also by regional to local formation of topography and basins in response to long-term plate reorganization, as well as the near- and far-field influence of large igneous province activity on the tectono-stratigraphic and paleoclimatic development. Various sedimentary and geochemical proxies, such as bentonite beds and carbon isotope excursions associated with the far-reaching environmental effects of the Siberian Traps, the High Arctic Large Igneous Province, and the North Atlantic Igneous Province are present in Svalbard's near complete geological record. As such, Svalbard is unique in that these and numerous other global environmental perturbations are recorded within a relatively restricted study area, with most of the key events preserved and recorded in easily accessible drill cores and well-exposed outcrop sections. Here we review deep-time paleoenvironmental and paleoclimate research in Svalbard by summarizing 148 peer-reviewed scientific articles. The review builds on the well-established tectono-stratigraphic and lithostratigraphic framework, as well as state-of-the art environmental reconstructions to provide insights into the Earth system during the Phanerozoic northward drift of Svalbard and the many major biotic crises in the geological past. We focus on globally significant events including i) the expansion of Devonian vegetation, ii) the Carboniferous-Permian response to icehouse conditions during the Late Paleozoic Ice Age (LPIA), iii) the End-Permian Mass Extinction (EPME) and the subsequent Triassic recovery, the iv) Carnian Pluvial Episode, v) Jurassic-Early Cretaceous climate perturbations including the Volgian Isotopic Carbon Excursion (VOICE) and the Aptian Ocean Anoxic Event 1a (OAE1a), and vi) the Paleocene-Eocene Thermal Maximum (PETM). We present and synthesize existing core and outcrop data that preserve biological and geochemical proxies and climate sensitive sedimentary facies that reflect environmental change in terrestrial and marine settings. Finally, we discuss the Phanerozoic climate recorded in Svalbard and its role in providing high latitude calibration points for several global paleoclimate events to provide a higher latitude perspective to complement the dominance of mid- and low-latitude locations and datasets in the literature.

1 Introduction

The recent Intergovernmental Panel on Climate Change report (IPCC; Pörtner et al., 2022) and the Intergovernmental Science-Policy Platform on Biodiversity and Ecosystem Services (Watson et al., 2019) highlight the challenges humanity is facing due to ongoing and projected climate change. Human-environment interactions have accelerated dramatically through the industrial revolution, and the human species is now considered to be a dominant geological force on the planet (Stewart, 2016). The rate of change the planet is experiencing is unprecedented for at least the last 66 Ma (Zeebe et al., 2016), which could lead to a biodiversity crisis of similar amplitude to those experienced during previous hyperthermals throughout the Phanerozoic. However, it is still uncertain when ecological tipping points occur, at which an ecosystem can no longer cope with

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environmental change, will be reached. It is thus critical to understand and define the trajectories and pace of ecological change that is the result of a major climate perturbation.

Studying episodes of past climate change, as recorded in the geological record, can provide insights into the response of Earth system processes to climate perturbations. Deep-time paleoclimatology, here considered to be pre-Quaternary (i.e., older than 2.58 Ma; all absolute ages refer to the International Stratigraphic Chart 2023/09), in its broadest sense refers to deciphering how and why the climate changed in the past, and the consequences of those changes to life on Earth. Throughout the pre-Quaternary Phanerozoic (ca. 538.8 Ma to 2.58 Ma), mass extinctions or smaller scale biodiversity crises occurred repeatedly, often in response to rapid climate change (Fig.1.; e.g., Bond and Grasby 2017; Kemp et al., 2015). Understanding past climate trends and episodes of major environmental perturbations will better constrain our understanding of the causes and consequences of future change (e.g., Soreghan 2004; Jansen et al., 2007). Many proxies exist to constrain past paleoenvironmental and climatic settings, grouped into biological, chemical and geophysical including climate sensitive sedimentary facies. Different proxies are suitable for quantifying various paleoclimatic signals, including volcanic activity, atmospheric gas concentration, land/sea temperature, seasonality, precipitation, and ocean oxygenation.

Significant perturbations to global climate are often related to relatively short-lived catastrophic events (from minutes to 1-2 Myr), including meteorite impacts, volcanic and kimberlite eruptions, the emplacement of large igneous provinces (LIPs; Bryan and Ernst, 2008; Green et al., 2022) and the biogeochemical cascades that these events cause. These events disturb Earth's biosphere by causing extreme changes in temperature, precipitation, wildfire frequency, sea level, oxygen level, and saturation states of biologically important elements (e.g., aragonite saturation state of the oceans; Hönisch et al., 2012), release of deleterious substances such as mercury (e.g., Grasby et al., 2011; Sanei et al., 2012; Galloway and Lindström, 2023), and trophic knock-on effects, affecting both the atmosphere and marine realms (Jenkyns, 2010). LIP volcanism in particular directly perturbs the climate system via release of gasses directly to the atmosphere, including SO_2, CH_4 and CO_2, as well as HCl, halocarbons, and Hg. LIP emplacements are in turn often associated with global environmental changes, including mass extinctions and smaller scale biotic crises such as oceanic anoxic events (OAEs; Grasby et al. 2011; Bond et al, 2014; Ernst and Youbi, 2017; Jones et al., 2016; Svensen et al., 2019). LIPs, by definition (Bryan and Ernst, 2008), involve significant igneous volumes if igneous material (>0.1 million km^3) emplaced or erupted over large areas (>0.1 million km^2) in an intraplate setting and within a short duration (1-5 Myr pulse for >75% of the volume, 50 Myr maximum lifespan), although some igneous activity broadly accepted as LIPs have protracted histories (e.g., the High Arctic Large Igneous Province; HALIP; Dockman et al., 2018; Heyn et al., 2024).

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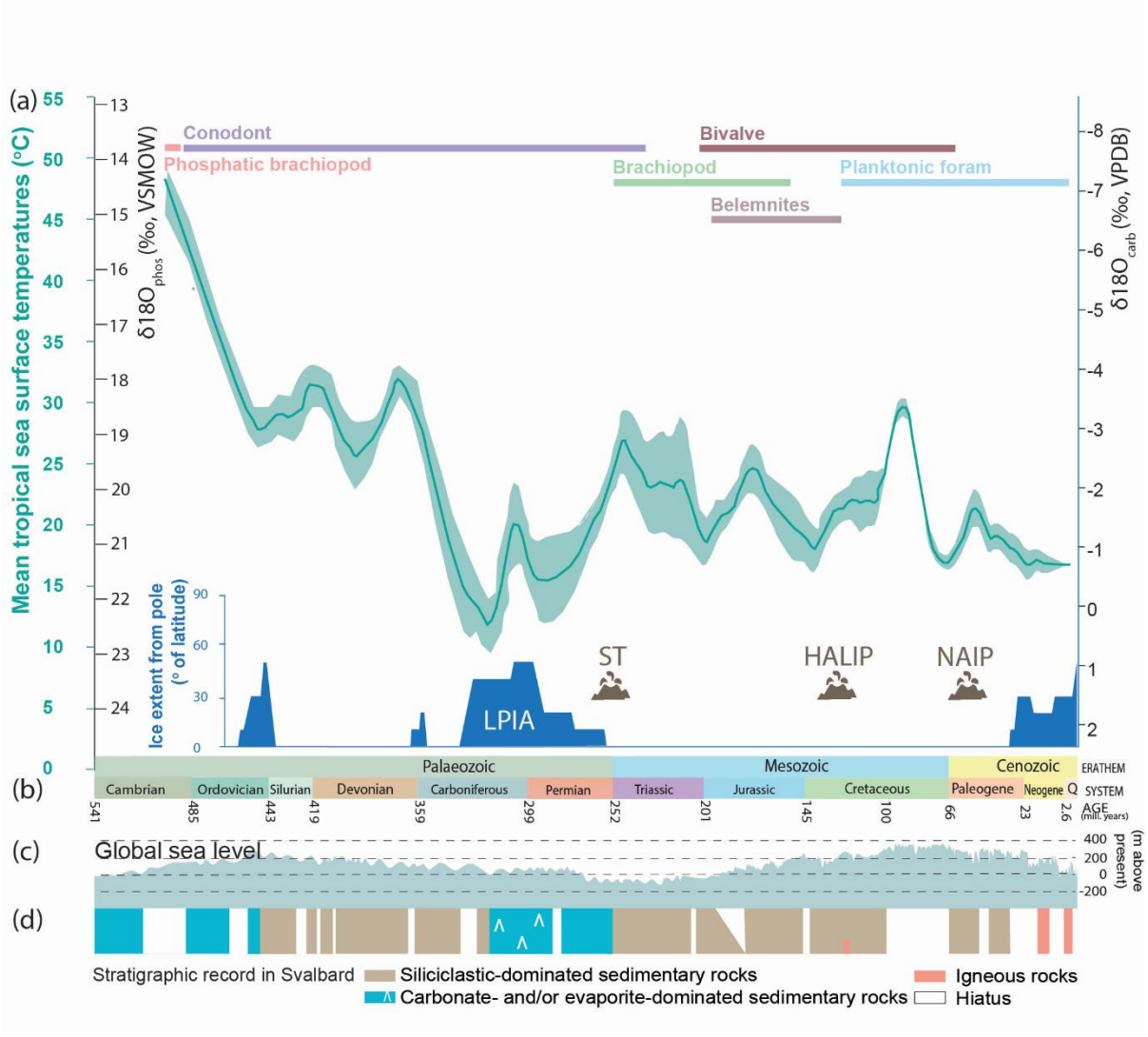

Figure 1: The stratigraphic record of Svalbard in a global deep-time climate context. (a) global data coverage including mean tropical sea surface temperatures per Myr marked as a green curve, with shaded area 95% confidence intervals, based on oxygen isotopes from phosphatic and carbonate fossils from Scotese et al., (2021), after Song et al., (2019); The scale of $\delta^{18}O_{Phos}$, black axis on the left represents phosphatic fossils (phosphatic brachiopod, conodont, and fish). The scale of $\delta^{18}O_{Carb}$ black axis on the right, represents carbonate fossils, (belemnite, bivalve, brachiopod, planktonic foraminifera). The Ice extent from pole after Macdonald (2020); major LIPs recorded in Svalbard: ST-Siberian Traps, HALIP- High Arctic Large Igneous Province, NAIP- North Atlantic Igneous Province, (b) Phanerozoic time scale, (c) Sea level curve (after Dallmann (ed) (2015), based on work of the International Commission on Stratigraphy,) (d) overview of the geological record of Svalbard (modified from Dallmann (ed) (2015)).

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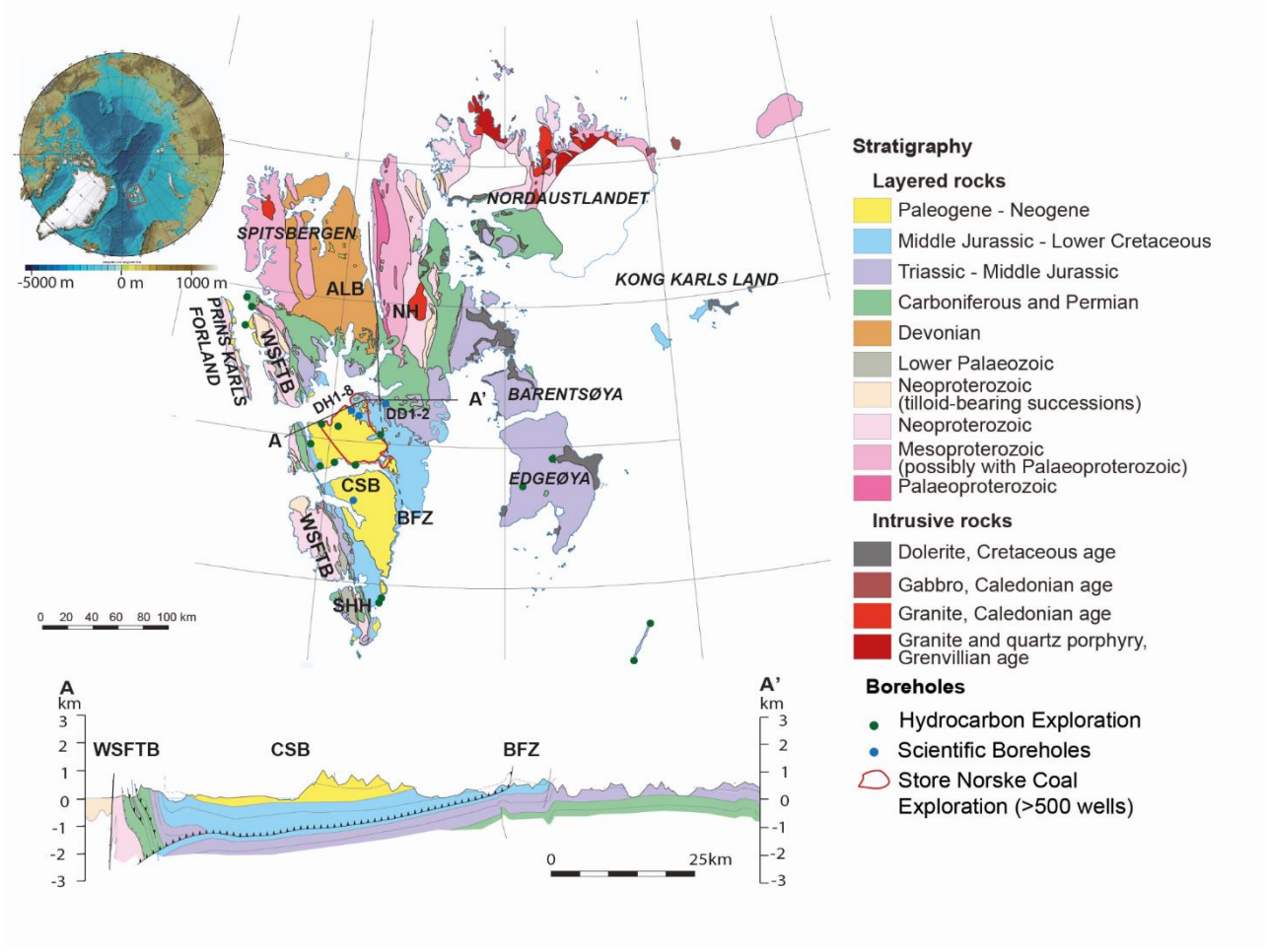

Figure 2: Geological map of Svalbard and a regional cross-section illustrating the major structural elements of central Spitsbergen and spatial coverage of industrial and research boreholes of relevance to deep time paleoclimatic studies. Upper left: International Bathymetric Chart of the Arctic Ocean (IBCAO; Jakobsson et al. (2012)). Geological Map of Svalbard and cross-section (A-A'; bottom) from Dallmann (ed) (2015), Norwegian Polar Institute. Location of boreholes from Senger et al., (2019). NH- Nordfjorden High, SHH- Sørkap Hornsund High, ALB- Andrée Land Basin, BFZ- Billefjorden Fault Zone, WSFTB- West Spitsbergen Fold and Thrust Belt.

Svalbard is a Norwegian archipelago comprising all islands between 74°–81°N and 15°–35°E, including the largest island of Spitsbergen (Fig. 2). Extensive amounts of geological data have been acquired from outcrops and drill cores across Svalbard. This data collection was largely triggered by the importance of Svalbard as an equivalent to the sedimentary successions in the offshore areas of the Norwegian Barents Sea (Steel and Worsley 1984; Olaussen et al., 2025 and references therein). The stratigraphic succession in Svalbard is almost complete (Fig. 1) with distinctive shifts in deposition over time, indicating a genetic link between deposition and paleolatitudinal position (Fig. 3), as initially identified by Steel and Worsley (1984). Svalbard's Phanerozoic paleoenvironmental evolution is largely controlled by two main factors: 1) the northward tectonic

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motion of Svalbard from equatorial to polar latitudes (Fig. 3), and 2) the influence of proximal and distant LIPs (Fig. 1). Svalbard's lower Paleozoic geological record expresses its affinity to Laurentia. Broadly speaking, Svalbard was part of the Laurasian or Eurasian plates during its post-Devonian history. The overall northward motion from an equatorial position in the Devonian and early Carboniferous to its present polar location during the Late Cretaceous (Fig. 3) was a response to absolute plate movement and relates to the break-up of the supercontinent Pangea near the end of the Jurassic. The LIPs influencing the depositional record in Svalbard (Fig.3) include the Siberian Traps, implicated as the causal factor of the end-Permian mass extinction (EPME; ca. 252 Ma; Reichow et al., 2009; Burgess et al., 2017), the High Arctic Large Igneous Province (HALIP) that influenced paleoenvironments of Svalbard and much of the circum-Arctic area during the Early Cretaceous (Midtkandal et al., 2016; Vickers et al., 2019; Galloway et al., 2022; 2023; Galloway and Lindström, 2023), and the North Atlantic Igneous Province (NAIP), associated with the opening of the North Atlantic and the warming of the Paleocene-Eocene Thermal Maximum (PETM; started ca. 56 Ma; Charles et al., 2011). Many of these regional to global-scale events can be directly studied on the exceptional sedimentary rock exposures of Svalbard, notably the vertically tilted Festningen section in western Spitsbergen (e.g., Grasby et al., 2015; Vickers et al., 2019; 2023; Senger et al., 2022). These and other events have also been recorded in drill cores collected for coal exploration, research purposes, and CO_2 storage in central Spitsbergen (e.g., Dypvik et al., 2011; Midtkandal et al., 2016; Grundvåg et al., 2017; Olaussen et al., 2019; Senger et al., 2019; Zuchuat et al., 2020; Jelby et al., 2025).

The Arctic has warmed twice to nearly four times faster than the rest of the globe in recent decades (Rantanen et al., 2022). This phenomenon, known as polar amplification, is largely due to oceanographic and atmospheric feedback processes (e.g., Screen and Simmonds, 2010). Polar amplification is apparent from the geological past as extreme climates that are inexplicable in current model gradients (Evans et al., 2018; Price et al., 2020). Due to the current position of Svalbard and its paleogeographic history, the nearly continuous Phanerozoic record on Svalbard provides an ideal site to study polar amplification; and prior to the Cretaceous, to elucidate the effects of the break-up of supercontinents and subsequent northward plate tectonic movement (Fig.3).

Currently, there are no compilations of paleoclimate and paleoenvironmental research from Svalbard addressing the pre-Quaternary Phanerozoic depositional record that is available. As a first, this study synthesizes the stratigraphic record covered by drill core material and high-quality outcrops with reliable geochronological constraints. This contribution also provides an overview of the range of proxies used to reconstruct Svalbard's paleoenvironmental evolution. Owing to its own unique tectonostratigraphic evolution, we do not include the successions exposed on Bjørnøya (the southernmost island of the Svalbard archipelago) in this review (see papers by Worsley et al., 2001; Grundvåg et al., 2023; Janocha et al., 2024, for details on this succession). To provide a framework for the review of Svalbard's deep-time climate history, the five-step paleoclimate classification of Zhang et al., (2016) was used which recognizes five major climates. These are: A- Tropical, B-Dry, C-Temperate, D-Continental and E-Polar. This division enables the deep-time climate classification estimated primarily from climatically sensitive deposits and paleontological evidence supplemented by geochemical proxies including isotope data. We systematically compile published literature (n=148) with relevance for deep-time paleoclimate and paleoenvironments in

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Svalbard. The synthesized data proxies (e.g., TOC, $\delta^{18}O$, and $\delta^{13}C$) reflect environmental changes in terrestrial and marine ecosystems that are presented in the broader context of pan-hemispherical and global climate events. The overall evolution of the paleoclimate preserved in the geological record of Svalbard is discussed and compared with the paleo-position of Svalbard and global average temperature trends.

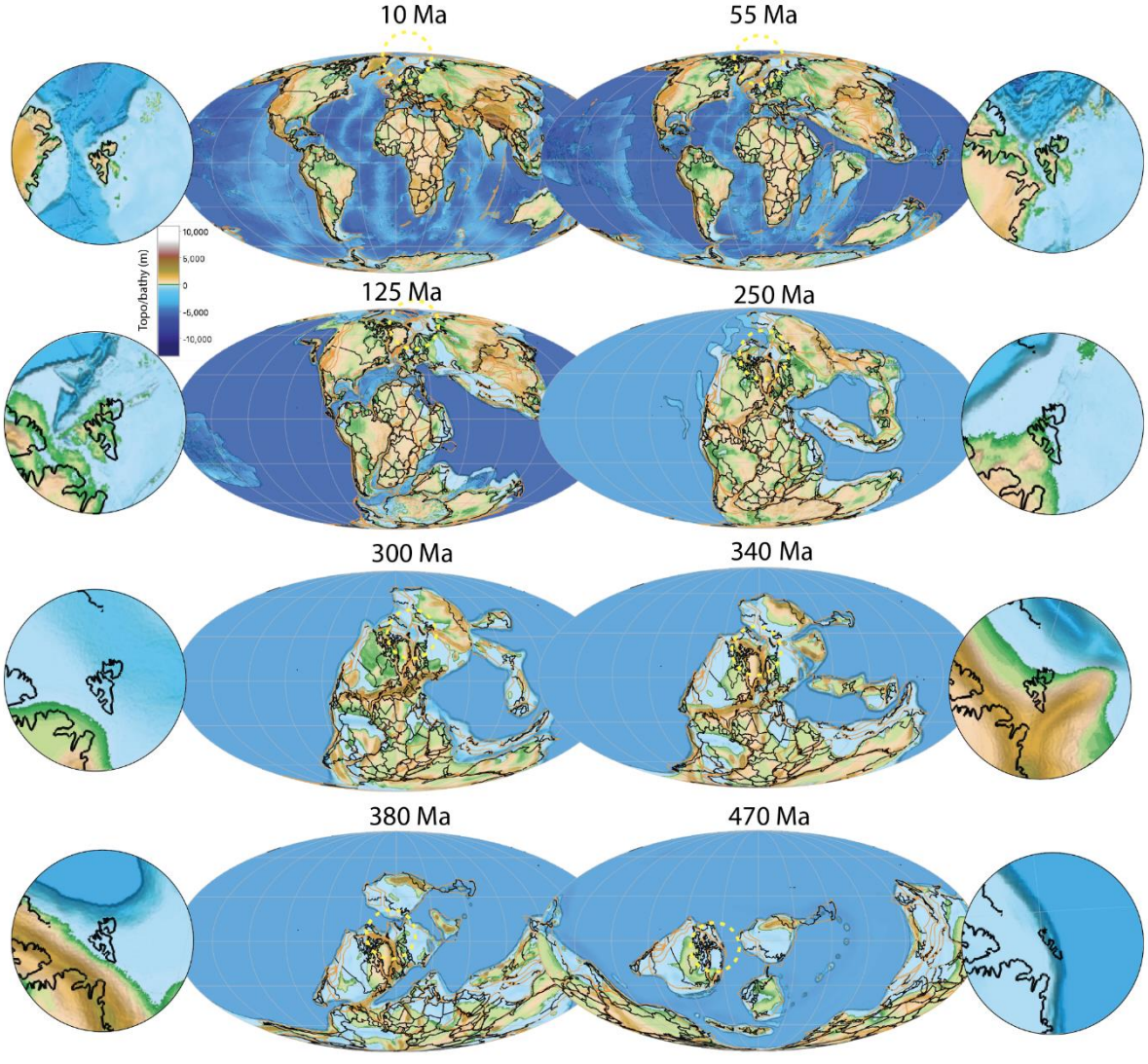

Figure 3: Global paleogeography (PaleoDEMs) from Scotese and Wright (2018), redrawn via export from GPlates (v2.5 Müller et al. 2018) for selected timesteps. Terrane boundaries in orange and political boundaries and present-day coastlines in black. Yellow dashed ring in the global Mollweide projections identifies the approximate location of Svalbard at the selected time periods with zoom in shown to the side of global maps. As this is a global model there may be discrepancies from regional Svalbard paleogeography whereby the reader is directed to Dallmann (ed) (2015) for regional resolution.

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2. Tectonic and stratigraphic development

The sedimentary successions preserved in Svalbard record a changing climate controlled to a large degree by the paleo-latitude of Svalbard along with global climatic transitions (e.g., Steel and Worsley, 1984). Since the start of the Paleozoic, Svalbard gradually drifted from near-equatorial latitudes to its present position at 74–81°N (Fig. 3; Scotese et al., 1979; Torsvik et al., 2002; Torsvik and Cocks, 2019).

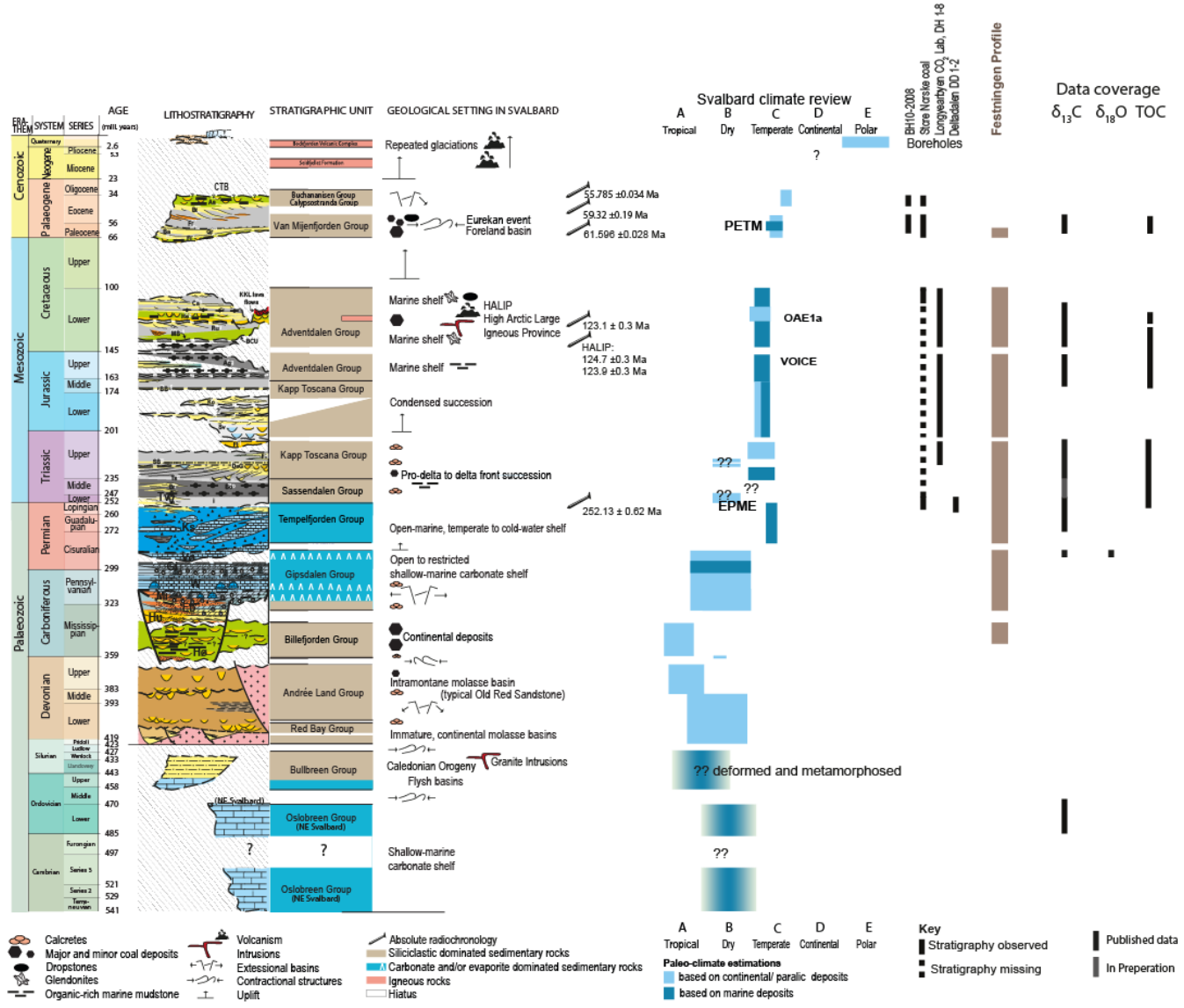

Figure 4: Stratigraphic column and climate summary of Svalbard highlighting the nearly continuous sedimentary record from the Devonian to the Neogene. The stratigraphic coverage of research boreholes is indicated. Post-Devonian lithostratigraphic column after Olaussen et al. (2025). The Climate zones from Zhang et al., (2016). Illustration of data coverage is based on Gruszczyński et al., (1989); Mii et al., (1997); Wignall et al., (1998; 2016); Galfetti et al., (2007); Cui et al., (2011); Buggisch et al., (2012); Mueller

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et al, (2014); Bond et al., (2015); Grasby et al. (2015a); Koevoets et al., (2016); Midtkandal et al., (2016); Vickers et al., (2016; 2019a; 2023), Hammer et al., (2019); Doerner et al., (2020); Jelby et al., (2020); Zuchuat et al. (2020); Wesenlund et al., (2021); Blattmann et al., (2024); Leu et al., (2024).

2.1. Paleozoic (Cambrian to Permian ca. 538-252 Ma)

The early Paleozoic succession of the Oslobreen Group (Fig. 4) preserved in northeastern Svalbard (Fig. 2) was deposited on the northern margin of Laurentia in a post-rift to a passive margin setting (Fig. 3; Smelror et al., 2024). Later, the Caledonian Orogeny started with the closure of the Iapetus Ocean and subsequent collision of Laurentia and Baltica in the Early Ordovician-Early Devonian (ca. 485–410 Ma; Barentsian Caledonides in Gee et al., 2006; 2008; Gee and Tebenkov, 2004; Harland et al., 1974). Caledonian deformation, metamorphism, and late crustal magmatism impacted mainly western and central-northern Svalbard, leaving the north-eastern parts of the archipelago practically undeformed (Johansson et al.2004; 2005; Smelror et al., 2024).

The syn- to post-Caledonian, Upper Silurian (Pridoli?) to Upper Devonian (Fransian) Old Red Sandstone (ORS) succession (ca. 423–372 Ma) represented by the Siktefjellet, Red Bay and Andrée Land groups (Fig. 4; Friend et al., 1997; Blomeier et al., 2003a; 2003b) is preserved in post-orogen collapse basins located in the central part of northern Spitsbergen (Fig. 2; Piepjohn et al., 2000; Blomeier et al., 2003; Braathen et al., 2018; Smelror et al., 2024). Significant deformation of the ORS succession subsequently took place during the Late Devonian compressional Svalbardian Event that is correlated with Ellesmerian Orogeny in Arctic Canada (McCann, 2000; Piepjohn, 2000; Bergh et al., 2011; Piepjohn and Dallmann 2014; Piepjohn and von Gosen, 2018; Beranek et al., 2020).

The Tournaisian (?)- Viséan (ca. 359-330 Ma) continental coal-bearing deposits of the Billefjorden Group are widespread across Spitsbergen (Fig. 2) and unconformably overlie the deformed Devonian and pre-Caledonian succession (Piepjohn et al., 2000). The thickness of the Billefjorden Group reaches up to 250 m in central Spitsbergen (Cutbill and Challinor, 1965; Gjelberg and Steel, 1981) and 40-100 m in northeastern Spitsbergen (Lauritzen and Worsley, 1975; Scheibner et al., 2012). The thickness along the west coast of Spitsbergen is uncertain due to younger deformation and repetition of the succession. The Billefjorden Group is unconformably overlain by the Serpukhovian (upper Mississipian) to Artinskian (Cisuralian) mixed siliciclastic-carbonate-evaporite deposits of the Gipsdalen Group (ca. 331–284 Ma). The lower Gipsdalen Group consists of syn-tectonic units filling up rift basins, the Billefjorden, Lomfjorden, St Jonsfjorden, and Inner Hornsund troughs, developed along north-south striking long-lived lineaments formed in response to regional-scale extension (Fig. 4; Holliday and Cutbill, 1972; Gjelberg and Steel, 1981; Johannessen and Steel, 1992; Faleide et al., 2008; Braathen et al., 2012). The thickest (> 1.5 km) and best-preserved basin fill occurs in the Billefjorden Trough while corresponding succession is missing on the structural highs (Fig. 2; Cutbill and Challinor 1965; Johannessen and Steel, 1992; Braathen et al., 2012; Smyrak-Sikora et al., 2018; 2021). The syn-rift units of the lower Gipsdalen Group were subsequently overlain by an up to 500 m thick warm-water carbonate-platform deposits of the upper Gipsdalen Group (Hüneke et al., 2001; Blomeier et al., 2011; Ahlborn and Stemmerik

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2015; Sorento et al., 2020). The Gipsdalen Group is overlain by upper Artinskian (Cisuralian) to Changhsingian (Lopingian) cool-water carbonate and spiculitic platform sediments of the Tempelfjorden Group (ca. 284 (?)–252 Ma). The thickness of Tempelfjorden Group varies across Spitsbergen from 6 m to 460 m (Blomeier et al., 2013; Uchman et al., 2016; Matysik et al., 2018), including complete absence of Permian deposits on the Sørkapp–Hornsund High, where Carboniferous fluvial deposits are unconformably overlain by Lower Triassic continental conglomerates (Zuchuat, 2014). These thickness variations indicate ongoing uplift of the Nordfjorden High and the Sørkapp–Hornsund High that can be linked with the late Permian rift event along the western Barents shelf margin (Faleide et al., 2008; Olaussen et al., 2025).

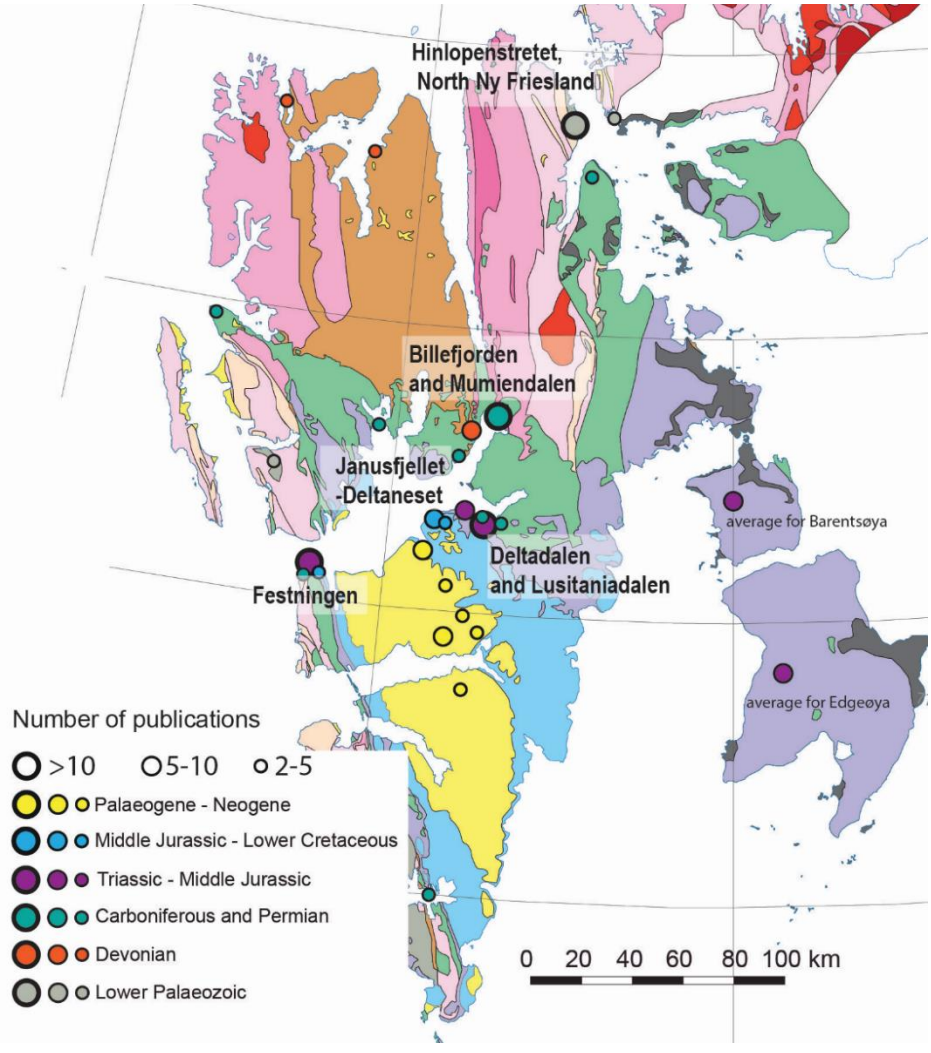

Figure 5: Location of data presented in the reviewed articles summed up in Table A1 (Supplementary table) highlighting the most studied sections in Svalbard. The size of a circle corresponds to the number of publications addressing deep-time paleoclimate proxies. For legend to the geological map see Fig. 2.

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2.2. Mesozoic

In the Early to Middle Triassic (ca. 252–237 Ma), Svalbard was part of a shallow shelf that experienced significant subsidence and which was filled with up to 700 m of sediments sourced from west and east (the Sassendalen Group; Fig. 4; Mørk et al., 1982; 1999; Wesenlund et al., 2022; Bjerger et al., 2023). By the end of the Middle Triassic (ca. 237 Ma), deltaic systems sourced in the Uralides and the Fennoscandian Shield reached and probably traversed Svalbard (Riis et al., 2008; Glørstad-Clark et al., 2010; Høy and Lundschien, 2011; Anell et al., 2013; Klausen et al., 2017; 2019; Gilmulina et al., 2022). Towards the latest Triassic and Early Jurassic, subsidence rates gradually decreased and sometimes even became negative, as expressed by condensed units with a 20 m-thick, shallow-marine and continental sandstone-shale unit of Rhaethian (latest Triassic; ca. 208–201 Ma) to Bathonian (?) age (Middle Jurassic; ca. 168–165 Ma), truncated by several subaerial unconformities (Drachev, 2016; Faleide et al., 2018: Olaussen et al., 2018; Rismyhr et al., 2018; Müller et al., 2019). This Upper Triassic deltaic and Upper Triassic to Middle Jurassic condensed section belongs to the Kapp Toscana Group. Subsidence rates increased again during the Late Jurassic (ca. 161–145 Ma), which led to the deposition of organic-rich marine strata of the lower Adventdalen Group (e.g., Koevoets et al., 2016; 2019).

The upper Middle and Upper Jurassic/lowermost Creatceous succession preserved in Svalbard consist of the Agardhfjellet Formation, and the Lower Cretaceous succession is represented by the Rurikfjellet, Helvetiafjellet, and Carolinefjellet formations, all assigned to the Adventdalen Group (Fig. 4). These marine to continental units reflect increased subsidence with uplift in the north and northwest that formed a continental, siliciclastic platform. This uplift is related to emplacement of the HALIP across the Arctic, including Svalbard, Franz Josef Land, the New Siberian Islands, the Barents Shelf, Sverdrup Basin, northern Greenland, and the Alpha-Mendeleev Ridge, via both subaerial eruptive and intrusive magmatism (Maher, 2001; Estrada and Henjes-Kunst, 2013; Senger et al., 2014; Evenchick et al, 2015; Polteau et al., 2016; Davis et al., 2017; Dockman et al., 2018; Naber et al., 2021; Bédard et al., 2021; Galloway et al., 2022). HALIP magmatism is thought to be derived from the arrival of a thermally elevated mantle plume that caused large volumes of mafic rocks including sills, dykes, lavas, and pyroclastic material (Maher, 2001; Senger et al., 2014; Buchan and Ernst, 2018; Bédard et al., 2021; Naber et al., 2021; Heyn et al., 2024). Robust U-Pb dating points to a short magmatic pulse affecting Svalbard at ca. 124.5 Ma (Corfu et al., 2013). However, in the adjacent Sverdrup Basin (Arctic Canada), multiple magma emplacement episodes have been identified, with pulses peaking at 122 ± 2 Ma, at 95 ± 4 Ma and at 81 ± 4 Ma (Jens et al., 2015; Kingsbury et al., 2017; Davis et al., 2017; Dockman et al., 2018; Bédard et al., 2021; Dummann et al., 2024). In Spitsbergen, the HALIP-activity triggered southward tilting of the platform, which resulted in progradation of a sand-rich fluviodeltaic system towards the south (Steel and Worsley, 1984; Gjelberg and Steel, 1995; Worsley, 2008; Midtkandal and Nystuen, 2009; Grundvåg and Olaussen, 2017; Grundvåg et al., 2017). The extent of the uplifted and eroded area increased during the Late Cretaceous (ca. 100–66 Ma) to the whole of Svalbard, which resulted in the upper middle Albian deposits (uppermost Lower Cretaceous; ca. 113–100 Ma; Hurum et al., 2016) being unconformably overlain by Paleocene strata (Jochmann et al., 2019; Helland-Hansen and Grundvåg, 2021).

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2.3. Cenozoic

The mid-Paleocene saw the recommencement of sediment deposition in Svalbard after a ~60 Myr hiatus in response to large-scale regional changes in plate tectonic configurations (Fig. 3). The base of the Paleocene strata in Svalbard has been dated to 61.8 Ma (Jones et al., 2017), close to the Danian–Selandian boundary (ca. 61.6 Ma). This age is contemporaneous with several changes around the Greenland microplate, including increased rifting between Greenland and Eurasia in the proto-Northeast Atlantic region (Abdelmalak et al., 2023), the first pulse of North Atlantic Igneous Province (NAIP) volcanism (Storey et al., 2007a), a change from carbonate- to siliciclastic-dominated sediments in the North Sea (Clemmensen and Thomas, 2005), widespread shear deformation along eastern Greenland (Guarnieri, 2015), and an increase in the rate of seafloor spreading in the Labrador Sea (Roest and Srivastava, 1989; Oakey and Chalmers, 2012). The combination of seafloor spreading in the Labrador Sea and rifting along the mid-Norwegian margin instigated compression between Greenland and Svalbard, which evolved into a dextral transpressional regime as rifting transitioned to seafloor spreading in the NE Atlantic by 55 Ma (Storey et al., 2007b). This period was also coincident with the rifting and breakup of the Eurasia Basin to the north of Svalbard. The dextral transpression along the Greenland-Svalbard margin caused localized crustal shortening and the formation of the West Spitsbergen Fold and Thrust Belt, WSFTB (Fig. 3; Harland, 1965), linked to the Eurekan deformation (ca. 63–35 Ma) and plate reorganization in the North Atlantic (Dallmann et al., 1993; Braathen et al., 1995; 1999; Maher and Braathen, 1995; Bergh et al., 1997; Gee and Tebenkov 2004; Faleide et al., 2008; Leever et al., 2011; Blinova et al., 2013; Piepjohn et al., 2015; 2016; Gion et al., 2017). A north-south trending foreland basin, known as the Central Spitsbergen Basin (CSB) or the Central Tertiary Basin (CTB) in older literature, formed east of the WSFTB and was filled with over 1.9 km thick Paleocene to Eocene (Oligocene?) deposits of the Van Mijenfjorden Group (Figs 3; Steel et al., 1981; 1985; Helland-Hansen, 1990; Müller and Spielhagen, 1990; Bruhn and Steel, 2003; Jochmann et al., 2020; Helland-Hansen and Grundvåg, 2021). During the Paleocene-Eocene transition (ca. 56 Ma), a passive margin started to develop to the north of Svalbard as a result of the opening of the Eurasia Basin. Finally, an Oligocene (ca. 34–23 Ma) transtensional rift phase eventually gave way to the formation of a passive margin west of Spitsbergen (Faleide et al., 2008; Lasabuda et al., 2018; Haaland et al., 2024).

The transpressional deformation related to the WSFTB was followed by NW–SE transtensional rifting that formed a series of grabens along the western Svalbard margin (Steel et al., 1985; Blinova et al., 2009; Kleinspehn and Teyssier, 2016; Kristoffersen et al., 2020; Haaland et al. 2024). The Forlandsundet Graben, one of the grabens cropping out between the islands of Prins Karls Forland and Spitsbergen (Fig. 2) contains between 1000-3000 m of Eocene to potentially Oligocene strata (Gabrielsen et al., 1992; Schaaf et al., 2021). The final separation between Greenland and the Barents Shelf margin eventually led to the opening of the Fram Strait. A shallow and narrow gateway was initially established around 20 Ma (Jokat et al., 2016; Fyhn and Hopper, 2025), and the transition from a restricted to fully ventilated Arctic Ocean took place around 17.5 Ma (Jakobsson et al., 2007). The establishment of a deep-water connection through the Fram Strait is currently debated, with suggested ages of 13.7 Ma (Jakobsson et al., 2007), 10 Ma (e.g., Kristoffersen and Husebye, 1985; Kristoffersen, 1990), and

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5 Ma (Lawver et al., 1990). During these times, Svalbard and the rest of the Barents Shelf margin experienced several changes in motion relative to the adjacent Greenland plate.

The present archipelago configuration of Svalbard with respect to the otherwise submerged setting of the Barents Shelf is thought to be a consequence of a combination of uplift during the Late Cretaceous, Paleocene-Eocene Eurekan deformation, and ongoing Holocene (last 11.7 kyr) isostatic rebound (Dimakis et al., 1998; Worsley, 2008; Henriksen et al., 2011; Dörr et al., 2013; Lasabuda et al., 2021). This uplift and exposed nature of Svalbard has resulted in the present-day exhumation of the metamorphic succession along the northern and western coasts of Svalbard and the younger sedimentary cover as described above (Fig. 3).

3. Data

Paleoclimate research in Svalbard has traditionally relied on the exceptionally exposed and vegetation-free outcrops, typical of the present Arctic landscape. For the last two decades, research drilling across Svalbard has increasingly been utilized and provides unique drill core material suitable for high-resolution paleoenvironmental and paleoclimate research.

3.1 Key stratigraphic sections

Figure 5 shows the geographic distribution of the primary study sites referenced in the 148 key publications summed up in Table 1 and in Table 2 (listed with more details in Supplementary table). Many of these articles are centered on important sites with good chronological and lithological constraints. Most of the outcrops are also covered by high-resolution digital outcrop models freely available through the Svalbox database (Betlem et al, 2023), facilitating data integration. Four of the localities are represented by 10 or more publications and are described below.

3.1.1 Festningen

The Festningen section in western Spitsbergen offers a nearly complete stratigraphic section spanning from the Mississippian (ca. 359 Ma) to the Paleocene (Fig. 6). The ~7 km long section is easily accessible along the shoreline, with nearly-vertical sedimentary layers due to Eurekan deformation. Festningen is an important regional stratigraphic profile and routinely targeted by geologists (Hoel and Orvin 1937; Steel et al., 1978; Mørk et al.,1982; Nagy and Berge 2008; Midtkandal and Nystuen, 2009; Grundvåg et al., 2019), including those interested in deep-time paleoclimate (e.g., Grasby et al., 2015; Vickers et al., 2023). Mørk and Grundvåg (2020) offers a geological guidebook for the section, whereas Senger et al., (2022) provided an open-access digital outcrop model (DOM) of the 5 km long part of the protected section. The high-resolution (7 mm pixel resolution) DOM is suitable for planning additional sampling, quantitative structural and sedimentological analyses, and integrating existing paleoclimatic research (Fig. 6).

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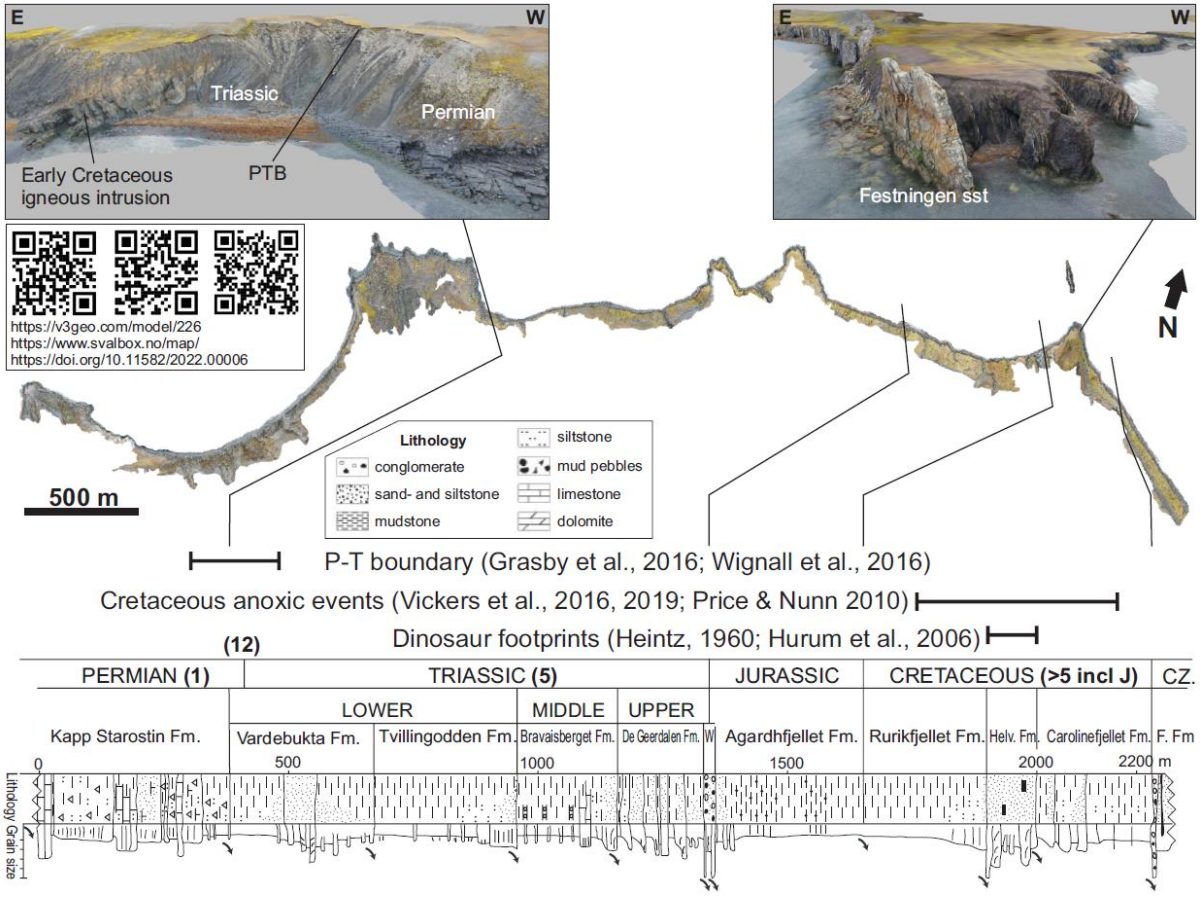

Figure 6: Stratigraphic column of the best-preserved part of the Festningen section, from Mørk and Grundvåg (2020), tied to the digital model of the entire Festningen section as presented in Senger et al., (2022). The inset images of the Permian-Triassic boundary and the Festningen sandstone illustrate screenshots of the digital outcrop model that is accessible online and freely available for download by following the QR codes and URLs. Paleoclimate-related research conducted on the section is highlighted for key events. These include amongst others the end-Permian mass extinction and the subsequent recovery phase (e.g. Wignall et al., 1998; Grasby et al., 2016), as well as several Cretaceous cooling events, anoxic events, and their associated deposits (Price & Nunn, 2010; Vickers et al., 2016, 2019; Grundvåg et al., 2019). Abbreviations: CZ = Cenozoic, W = Wilhelmøya Subgroup, Helv. Fm = Helvetiafjellet Fm, F. Fm = Firkanten Fm. The QR codes provide direct access to the digital model. Figure modified after Senger et al. (2022).

3.1.2 Janusfjellet-Deltaneset

The Janusfjellet section in central Spitsbergen exposes an Upper Triassic to Paleocene siliciclastic-dominated succession that has been extensively studied as part of the Longyearbyen CO_2 lab project (Olaussen et al., 2019). The succession includes both the Upper Triassic-Middle Jurassic sandstone reservoirs of the Kapp Toscana Group as well as the overlying Upper Jurassic-Lower Cretaceous Adventdalen Group. The Agardhfjellet Formation, the lowermost part of the Adventdalen Group, has also been extensively studied as one of the richest marine reptile sites in the world, yielding sixty specimens so far (Hurum et al., 2012; Delsett et al., 2016; 2019), along with an abundant seep fauna (Hryniewicz et al., 2015). The outcropping section dips

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gently at about 3° to the south-west and exposes the same stratigraphy as in the fully cored boreholes in Adventdalen (Olaussen et al., 2019). As an excellent analog to the Longyearbyen CO_2 lab reservoir-caprock system, the outcrops have been systematically studied with focus on sedimentology (Rismyhr et al., 2018; Jelby et al., 2020a), fault and fracture characterization (Ogata et al., 2014a; 2014b; Mulrooney et al., 2018; Betlem et al., 2024; Rizzo et al., 2024,), sandstone injectites (Ogata et al., 2023) and paleoclimatic signals (Koevoets et al., 2018; Jelby et al., 2020b).

3.1.3 Sassendalen

Sassendalen is a key region for understanding the Permian-Triassic transition and evolution of Svalbard, and within Sassendalen there are many key sections for defining different aspects of the lithostratigraphic framework of central Spitsbergen (e.g., Mørk et al., 1999). It is also a unique area to study global ecosystem recovery after the end-Permian mass extinction (see Hurum et al. 2018; Kear et al. 2023). Deltadalen and Lusitaniadalen are two valleys on the western side of Sassendalen that excellently expose the Permian Kapp Starostin to Botneheia formations. In addition is the more eastern Fulmardalen (Hammer et al., 2019; Hansen et al., 2024). The Deltadalen outcrop is directly next to the Deltadalen research boreholes, where deposits of the EPME and Permian-Triassic boundary were cored (Zuchuat et al., 2020). As such, it provides excellent borehole-outcrop correlation with the benefit of facilitating detailed sedimentological studies and high-resolution sampling away from the boreholes. In addition, a well-exposed section of Permian-Triassic transition crops out along Lusitaniadalen located around 5 km northwest from Deltadalen and has been the focus of multiple studies directly focussed on the Permian-Triassic boundary (e.g., Mørk et al., 1999a; 1999b; Foster et al., 2017; Rauzi et al., 2024).

3.1.4 Hinlopenstretet, North Ny Friesland

The Ny Friesland section in northeastern Spitsbergen is exposed along the south coast of the Hinlopen Strait (Fig. 2) and consists of a ~1 km thick Terreneuvian-Middle Ordovician carbonaceous succession of the Oslobreen Group (ca. 539–458 Ma; Hansen and Holmer, 2010; Stouge et al., 2012; Lehnert et al., 2013; Abay et al., 2022; Smelror et al., 2024). This succession, consisting of siliciclastic shoreline facies at the base, passing up to a shallow marine warm water carbonate platform deposits, belongs to the North Atlantic/Arctic warm water carbonate platform formed on eastern Laurentia (McKerrow et al., 1991; Stouge et al., 2012).

Table 1: Overview of selected drill core material and key sections (including digital sections)

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Key, selected boreholes with significant core or cuttings material

Borehole(s)	Penetrated strata (age and formation)		Depth (meters)	Drill core availability	Complementary data sets	Key reference
	Youngest	Oldest				
Gipsdalen (DD1-DD8)	Pennsylvanian Ebbadalen Fm	Precambrian Basement	91,90 - 210.50	Berlin-Spandau and Uni Oslo	Technical reports	Senger et al. (2019).
Deltadalen (DD1 and DD2)	Lower Triassic Vikinghøgda Fm	Lopingian Kapp Starostin Fm	92,9 -99,3	Oslo	Reports	Zuchuat et al. (2020); Schobben et al., (2020); Rodríguez-Tovar et al. (2021)
Tromsøbreen II	Lower Cretaceous Carolinefjellet Fm	Pennsylvanian (?) Minkinfjellet Fm (?)	2337	5 cores, in Lund	Reports	Senger et al. (2019).
Reindalspasset	Lower Cretaceous Carolinefjellet Fm	Mississippian, Billefjorden Gr	2315	Cuttings, in Endalen Svalbard	Reports	Senger et al. (2019).
UNIS CO_2 lab (DH1-DH8)	Lower Cretaceous Carolinefjellet Fm	Upper triassic De Geerdalen Fm	down to 969,7	Longyearbyen	Reports, wireline logs, publications	Corfu et al., (2013): Midtkandal et al., (2016); Polteau et al., (2016); Olaussen et al. (2019): Senger et al. (2024).
SNSK boreholes, including core BH9/05	Eocene Frysjaodden Fm	Metamorphic basement	numerous	Endalen	Reports	Cui et al. (2010); Dypvik et al. (2011); Charles et al. (2011); Nagy et al. (2013); Jones et al. (2019).
Sysselmannbreen (BH2008-10)	Eocene (?) Oligocene (?) Aspelintoppen Fm	Late Paleocene Basilika Fm	1085	Endalen, Equinor lab (Bergen)	Reports	Johannessen et al. (2011); Doerner et al. (2020).

Key outcrop sections (> 10 publications)

Outcrop sections	Penetrated strata (age and formation) Youngest	Penetrated strata (age and formation) Oldest	Extent	Key reference(s)
Hinlopenstretet, North Ny Friesland	Ordovician	Cambrian	around 10 km	Hansen and Holmer (2010); Stouge et al. (2012); Lehnert et al. (2013); Abay et al. (2022).
Billefjorden	Lower Permian	Upper Devonian	over 20 km of composite sections	Cutbill and Challinor (1965); Gjelberg and Steel (1981) Ahlborn and Stemmerik (2015) Berry & Marshall (2015); Davies et al. (2021); Smyrak-Sikora et al., (2021).

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Festningen	Paleogene Firkanten Fm	Carboniferous Billefjorden Gr	5-6 km	Hoel & Orvin (1937); Steel et al. (1978); Mørk et al. (1982); Nagy & Berge (2008); Midtkandal & Nystuen (2009); Grasby et al. (2015); Grundvåg et al. (2019); Mørk & Grundvåg (2020); Senger et al. (2022); Vickers et al. (2023).
Deltadalen and Lusitaniadalen	Lower Triassic Vikinghøgda Fm	Lopingian Kapp Starostin Fm	around 7 km	Dagis & Korcinskaja (1987); Mørk et al. (1999); Foster et al., (2017); Zuchuat et al., (2020);
Deltaneset-Janusfjellet	Lower Cretaceous Carolinefjellet Formation		around 7 km	Hurum et al., (2018); Rismyhr et al., (2018); Olaussen et al. (2019); Koevoets et al. (2018); Jelby et al. (2020a; 2020b).

3.1.5 Billefjorden and Mumiendalen

The inner part of Billefjorden exposes the Billefjorden Fault Zone, where the several km-thick Devonian Old Red Sandstone (ORS) deposits on the west are faulted against the several metamorphic and sedimentary successions in the east. These include pre-Devonian metamorphic basement, up to 250 m of Mississippian units and over 2 km of Pennsylvanian to Permian deposits (Braathen et al., 2012; Smyrak-Sikora et al., 2018; 2021). In Mumiendalen, the lowermost Upper Devonian deposits of the Andrée Land Group expose plant fossils that represent terrestrialization and evolution of one of the oldest forests in the world, that likely thrived in a monsoonal climate (Berry and Marshall, 2015; Davies et al., 2021). Above the fossil Devonian forests units, Mississippian siliciclastics with coal seams belonging to the Billefjorden Group were deposited in a humid and tropical climate (Cutbill and Challinor 1965; Gjelberg and Steel, 1981). The transition to the overlying mixed siliciclastic-carbonate-evaporite units of the Gipsdalen Group illustrates that the climate changed from subtropical semi-arid to an arid climate prevailing in through the Pennsylvanian to Cisuralian (Holliday and Cutbill, 1972; Gjelberg and Steel, 1981; Johannessen and Steel, 1992; Blomeier et al., 2011; Ahlborn and Stemmerik, 2015; Sorento et al., 2020; Smyrak-Sikora et al., 2021). Up section, the Bashkirian to Sakmarian part of the Gipsdalen Group consisting of interbedded evaporites and carbonates formed in response to glacioeustatic sea level fluctuations (Stemmerik, 2000; 2008; Ahlborn and Stemmerik., 2015; Sorento et al., 2020; Smyrak-Sikora et al., 2021) related to the late Paleozoic Ice Age (LPIA; Gastaldo et al., 1996; Montañez et al., 2007; Isbell et al., 2008).

3.2 Drill cores

Table 1 summarizes the drill cores presently available from Svalbard. They are mostly from coal exploration by Store Norske Spitsbergen Kulkompani (SNSK) and CO_2 storage research drilling in Adventdalen by The University Centre in Svalbard (UNIS); Olaussen et al., 2019; Senger et al., 2024). In addition, stratigraphic research boreholes, drilled purely out of scientific interests also exist, including the Sysselmannbreen (Johannessen et al., 2011) and Deltadalen (Zuchuat et al., 2020) boreholes. Limited core material from past petroleum exploration efforts are known to exist (eighteen boreholes were drilled from 1960

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to 1994; see review by Senger et al., 2019), but these have, hitherto, not been used in any paleoclimate investigations. Nonetheless, the associated wireline data from these wells are important calibration points for regional correlation of outcrops and onshore seismic reflection data.

3.2.1 Coal drilling

Exploration coal drilling focused on Paleocene stratigraphy of the Van Mijenfjorden Group in the Central Spitsbergen Basin (CSB) and Mississippian coal-rich strata of the Billefjorden Group near Pyramiden. Lower Cretaceous and Eocene coal bearing strata were targeted in minor campaigns. Cores from the Russian/Soviet coal mining company Trust Arktikugol including sites near Barentsburg, Colesdalen and Pyramiden, are not available and likely lost as evidenced by defunct core sheds scattered around Svalbard. However, reports from these drill cores (Verba, 2013) have been used locally to constrain surface geological mapping, such as in Mimerdalen (Piepjohn and Dallmann, 2014) and in the Billefjorden Trough (Smyrak-Sikora et al., 2021). The Norwegian coal mining company SNSK stores most of its cores in Endalen near Longyearbyen with drill dates ranging from the late 1960s to 2014, the last year of Norwegian coal exploration. These cores were investigated in several paleoclimate-related studies, particularly across the PETM and North Atlantic Igneous Province (NAIP)-related ash deposits (Table 1; Dypvik et al 2011; Jones et al, 2019).

3.2.2 Longyearbyen CO2 lab, DH1 to DH8

Eight boreholes were fully cored near Longyearbyen from 2007 to 2013 to characterize a potential CO_2 storage site (Braathen et al., 2012; Olaussen et al., 2019; Senger et al., 2024). Four of the boreholes reach the planned Upper Triassic units of the Kapp Toscana Group target storage unit at 670-700 m, while the other boreholes focus on the cap rock and overburden succession of the Adventdalen Group. The full coring across the shale-dominated cap rocks provides important constraints on the stratigraphy of the Jurassic-Cretaceous strata (Koevoets et al., 2016; Midtkandal et al., 2016; Jelby et al., 2020a; 2020b; Śliwińska et al., 2020) and also contributed to refining the global geological time scale (Zhang et al., 2021). Senger et al. (2024) provide a full overview of the data sets generated by the project including a live database of the resulting publications, including those focusing on deep-time paleoclimate.

3.3.3 Sysselmannbreen, BH10-2008

The BH10-2008 (also known as Sysselmannbreen) research borehole was drilled and fully cored in 2008 to recover a full section of the world-famous Eocene-Oligocene (?) clinoform succession of the Van Mijenfjorden Group in the CSB (Johannessen et al., 2011). The 1085 m-long core was split with one half stored in a container in Endalen, outside of Longyearbyen, and the other half stored in Equinor's laboratory in Bergen (Doerner et al., 2020).

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3.3.5 Deltadalen, DD-1 and DD-2

The most recent research drilling in Svalbard was conducted in 2014 at Deltadalen specifically to target the uppermost Permian part of the Tempelfjorden Group and Lower Triassic succession of the Sassendalen Group with a specific interest in the EPME and its aftermath (Zuchuat et al., 2020). The two ca. 100 m deep boreholes were drilled and fully cored over a 1-week period, with all equipment transported onsite by helicopter. The drill cores are stored at the University of Oslo.

Table 2. Summary of proxies based on Table A1 (Supplementary material), which reviews 148 selected papers. The data set is plotted in Figure 13.

N of articles	**N of papers**	10	11	27	22	30	30	18
	-boreholes	0	0	0	3	2	10	9
	-outcrops	10	11	27	21	30	27	11
Proxies and data	**Biological indicators**	8	8	15	14	22	24	10
	Climate sensitive facies	3	7	18	2	2	3	9
	Carbon isotopes $\delta^{13}C$	2	2	8	8	6	13	6
	Oxygen isotopes $\delta^{18}O$	1	1	4	0	0	4	1
	TOC/Rock-Eval	2	2	1	3	6	7	6
	Mercury, Hg	0	0	0	2	1	1	1
Stratigraphic intervals		**Cambro-Silurian**	**Devonian**	**Carboniferous-Permian**	**Permian-Triassic boundary**	**Triassic**	**Jurassic-Cretaceous**	**Paleogene**

Deep-time paleoclimate in Svalbard

4.1 Early Paleozoic (Cambrian, Ordovician, Silurian, 538.8-419.2 Ma)

4.1.1. Cambrian to Middle Ordovician- the Great Ordovician Biodiversification Event

The early Paleozoic registered two of the greatest evolutionary events in the history of life: the Cambrian Explosion (ca. 540–510 Ma), and the Great Ordovician Biodiversification Event (GOBE; ca. 497–445 Ma.; Webby et al., 2004; Servais and Harper,

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2018). This upper Furongian to Upper Ordovician diversification event is linked to cooling of previously very warm tropical oceans (Webby et al., 2004; Servais and Harper, 2018).

During the Cambrian and Ordovician, Svalbard was in a near-equatorial position (Torsvik et al., 2012; Fig. 2). The Terreneuvian to Middle Ordovician Oslobreen Group (Fig. 4) strata preserved in Svalbard formed on a large carbonate platform along the northern margin of Laurentia, also exposed in the Northeast Greenland Basin and eastern North Greenland Basin (Stouge et al., 2012; Fig. 2). The mildly deformed, 1–1.2 km thick sandstones, fossiliferous limestone and dolomite are preserved in northeastern Spitsbergen (Fig. 7; Harland and Wilson, 1956; Oslobreen Series in Gobbett and Wilson, 1960; Fortey and Bruton, 1973; Stouge et al., 2011; 2012; Dallmann et al., 2015). The succession is potentially interrupted by a ~15 Myr-hiatus spanning over the Series 2, Miaolingian, Furongian, and possibly the earliest Ordovician (Fortey and Bruton, 1973; Smelror et al., 2024), although the lack of dateable fossils might affect this interpretation (Smelror et al., 2024). The Oslobreen Group shows surprisingly low maximum burial temperatures (Bergström, 1980; Abay et al., 2022) and eastward increasing tectonothermal influence linked to the Caledonian Orogeny (Johansson et al., 2004). The trilobites and fauna generally show a Pacific and Laurentian affinity (Fortey and Bruton, 1973; Hansen and Holmer, 2010; Stouge et al., 2012).

In Ny Friesland, the Terreneuvian microbial laminated limestone/dolomite contain cm-scale erratic chert nodules. The Lower to Middle Ordovician (ca. 485–458 Ma) carbonates were deposited in a paleotropical marine shelf setting experiencing episodes of water column redox-stratification (Lee et al., 2019). Stouge (2012) also interprets the Tremadocian (Lower Ordovician; ca. 485–478 Ma) environment in Svalbard and Greenland as a typical tropical shelf. Occurrence of oolite beds interbedded with domed stromatolites throughout the Tremadocian on Ny Friesland and adjacent islands (Kröger et al., 2017) is consistent with a peritidal tropical carbonate factory. Uchman and Hanken, (2024) recognise that carbonates of the uppermost part of Terreneuvian and the Cambrian Series 2 contain pseudomorphs after evaporites. Hansen and Holmer (2010) recognize strong ties of Lower and Middle Ordovician brachiopods to faunas in North America and Greenland at the generic level. Hansen and Holmer (2010) also discuss the transition from low diversity brachiopod fauna in the Tremadocian and early Floian (Early Ordovician; ca. 478–470 Ma), followed by an abrupt diversification event in the late Floian and into the Middle Ordovician. The hypersaline conditions, however, mask the expected record of the Great Ordovician Biodiversification Event (Uchman and Hanken, 2024). The Middle Ordovician part of the succession represents an overall deepening, transgressive sequence (Kröger et al., 2017; Lee et al., 2019). Kröger et al., (2017) suggest that the gradual transition to deeper deposits with more shale and local siltstone and glauconitic horizons accompanied with increased burrowing and fossiliferous, cherty mud-wackestone and skeletal grainstone are evidence of general climate cooling in the transition to the Middle Ordovician (i.e., the incipient phase of Ordovician cooling). Although changing to colder sea floor conditions, tropical carbonate production continued in an inner carbonate ramp while a cold-water carbonate factory prevailed in the outer ramp (Smelror et al., 2024).

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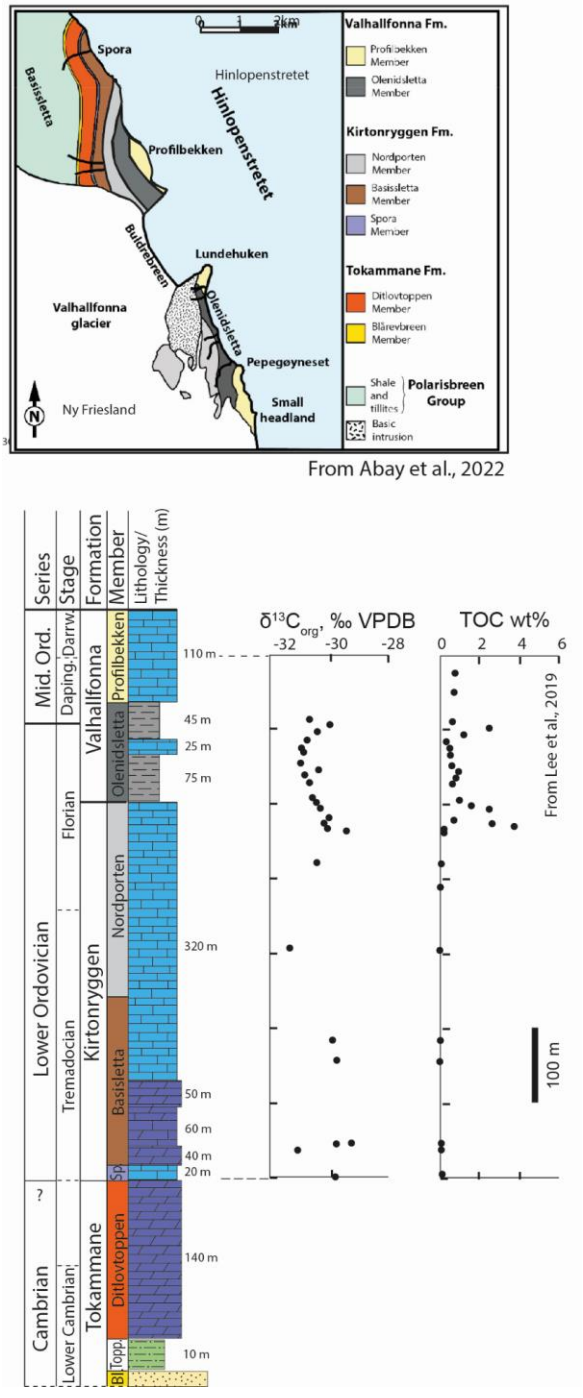

Figure 7 Stratigraphic section of the Cambrian and Ordovician Oslobreen Group succession in Ny Friesland, northern Spitsbergen (see Fig. 5 for location) After Fortey and Bruton (1973); Harland (1997); Stouge et al. (2012); Lehnert et al. (2013); Lee et al., (2019). Bulk organic carbon isotopes ($\delta^{13}C_{org}$, in ‰ VPDB); Total Organic Carbon (TOC, in weight percent); Bl.- Blårevbreen Member; Topp.- Topiggane Member; Sp.- Spora Member.

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4.2 Late Paleozoic (Devonian, Carboniferous, Permian, 419.2-251.9 Ma)

The Devonian, Carboniferous, and Permian periods record the only complete greenhouse to icehouse to greenhouse cycle (LPIA) on a vegetated Earth (cf. Isbell et al., 2008). In Svalbard, the relatively complete Devonian to Permian sedimentary succession, which encompasses the Old Red Sandstone, Billefjorden, Gipsdalen, and Tempelfjorden groups (Fig. 4), provides a unique opportunity to study responses of the tropical and near-tropical depositional systems with the terrestrial and shallow-marine settings to the LPIA glaciations. Svalbard occupied a near equatorial position for most of the Devonian and Carboniferous and from Permian started northwards drift (Fig. 2; Torsvik and Cocks, 2019).

4.2.1. Devonian: Old Red Sandstone, terrestrialization, and first forest

The advent of terrestrial vascular plants in the latest Silurian-earliest Devonian influenced weathering processes, soil formation, and strongly impacted the CO_2 cycle and global climate (Gensel et al., 2008 and references therein; Kenrick et al., 2012). The impact of vascular plants can be observed in Svalbard in the Uppermost Silurian to Upper Devonian Old Red Sandstone succession (Friend et al., 1997; Blomeier et al., 2003a; 2003b). This >8 km thick succession is restricted to extensional collapse basins formed in pure extensional (Piepjohn and Dallmann, 2014), or more likely, transtensional settings (Braathen et al., 2018). The Andrée Land Basin exposed in central-north Spitsbergen was filled mainly by a terrestrial succession, with marginal-marine conditions recorded in the northernmost part (Blomeier et al., 2003a). It notably includes red and gray-green fluvial, alluvial, lacustrine, and coastal sedimentary strata arranged into fining-upward units, with abundant plant material (Friend, 1965; Moody-Stuart, 1966; Blomeier et al., 2003a; 2003b; Piepjohn and Dallmann 2014). The succession recorded indications of long-term climatic variability, such as shifts in paleosols from calcretes and vertisols to coal and preservation of in situ tropical forests (Berry and Marshall, 2015). The biological evidence of environmental conditions recorded in the Old Red Sandstone come from plant fossils (Berry 2005; Berry and Marshall 2015; Davies et al., 2021), palynomorphs (Vigran, 1964; Allen, 1965; 1967; Friend et al., 1997), vegetation-induced sedimentary structures (Davies et al., 2021), and scarce marine-influenced fauna including ostracods and bivalves (e.g., Friend, 1961; Worsley, 1972). Bulk geochemistry along with extraction and biomarker analysis of Middle Devonian coal indicate a terrestrial plant origin with high liptinite content (Vogt, 1941; Blumenberg et al., 2018). The flora evolved over time from diminutive plants in the Middle Devonian to the first in situ forests of lycopsids and archaeopterids in the early Frasnian (Late Devonian; ca. 383–372 Ma; Berry and Marshall 2015; Davies et al., 2021). Based on sedimentological and biological evidence for highly variable seasonal discharge and scarcity of thick calcretes, Davies et al., (2021) suggested that the precipitation regime in the Devonian was tropical and monsoonal, and that the stratigraphic partitioning into red bed and gray-green strata attest to long-term fluctuations in drainage and oxidizing conditions.

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4.2.2. Carboniferous to Cisuralian: Late Paleozoic Ice Age (LPIA)

The Late Paleozoic Ice Age (LPIA) is one of the most important climatic events of the Phanerozoic that significantly influenced climate and depositional systems on Earth (Gastaldo et al., 1996; Montañez et al., 2007; Isbell et al., 2008). The LPIA is the closest analog to present climate conditions, characterized by discrete periods of glaciations separated by warm interglacials (Montañez and Poulsen, 2013). The LPIA glaciations started ca. 347 Ma at the onset of the Viséan (Mississippian), and lasted at least until ca. 285 Ma during the middle of the Artinskian (Cisuralian), potentially extending until ca. 260 Ma around the Guadalupian-Lopingian transition in the more Alpine settings in eastern Gondwana (Montañez and Poulsen, 2013). During this time, Svalbard drifted from a tropical position into the northern subtropical warm arid zone (Fig.2; e.g. Torsvik and Cocks, 2019).

The early stage of the LPIA coincided with deposition of the Mississippian Billefjorden Group. This succession unconformably overlies the folded Paleo-Neoproterozoic and Devonian successions. The terrestrial deposition occurred on broad floodplains and included abundant coal seams deposited under the humid tropical climate (Fig. 3; Gjelberg and Steel, 1981; Fairchild et al., 1982; Steel and Worsley, 1984; Lopes et al., 2019). The coal-bearing succession reaches up to 55 m thickness in eastern Spitsbergen (Scheibner et al., 2012), and 350 m in central Spitsbergen (Gjelberg and Steel, 1981; see borehole SLE 116 in Smyrak-Sikora et al., 2021). Over 1200 m of cumulative thickness is reported along the west and south parts of Spitsbergen (Gjelberg and Steel, 1981), where this succession is repeated several times due to the Paleocene and Eocene WSFTB (Maher et al., 1995; Braathen and Bergh 1995; Fig.10 in Horota et al., 2022). The fluvio-lacustrine coal deposits were commercially mined in Pyramiden from 1910 to 1998. The coal is characterized by relatively heavy $\delta^{13}C$ values, a low gammacerane index and high Pr/Ph ratios, distinctive from the Pennsylvanian coals associated with evaporites (Nicholaisen et al., 2019). Based on spores and plant fossils, Scheibner et al., (2012) suggested that the Billefjorden Group strata were deposited in a humid climate, in accordance with a paleogeographic position 10-15 °N (Fig. 2).

The shift from humid tropical to warm, arid to semi-arid depositional environments occurred during the late Serpukhovian (Mississippian) at the boundary between the Billefjorden and Gipsdalen groups (Fig. 3; Holliday and Cutbill, 1972; Gjelberg and Steel, 1981; Johannessen and Steel, 1992) and coincides with the initiation of regional-scale rifting in Svalbard and the Barents Shelf (Nøttvedt et al., 1993; Faleide et al., 2008; Braathen et al., 2012; Smyrak-Sikora et al., 2018; 2021). The following mixed siliciclastic-evaporite-carbonates succession of the lower Gipsdalen Group was deposited during the Pennsylvanian in an array of north-south striking rift basins (Gjelberg and Steel 1981; Smyrak-Sikora et al., 2021). The shift from the Billefjorden Group to the Gipsdalen Group is abrupt across most of Svalbard, and the boundary likely represents a period of non-deposition or erosion, especially on the structural highs. Contrastingly, in the inner part of one of the Billefjorden Trough, the transition is more gradual and occurs in a fluvial succession where the only changes recorded in the meandering river system is the shift from humid to arid-climate soil profiles (Olaussen et al., 2023). This change in climate setting does not correspond to recognized northward drift of Svalbard (Torsvik and Cocks, 2019) and the reasons for this change are poorly

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understood. The Billefjorden Trough began as a continental rift basin followed by the opening of a connection to the ocean in the Bashkirian, which made the preposition sensitive to glacio-eustatic sea level variations (Smyrak-Sikora et al., 2019; 2021).

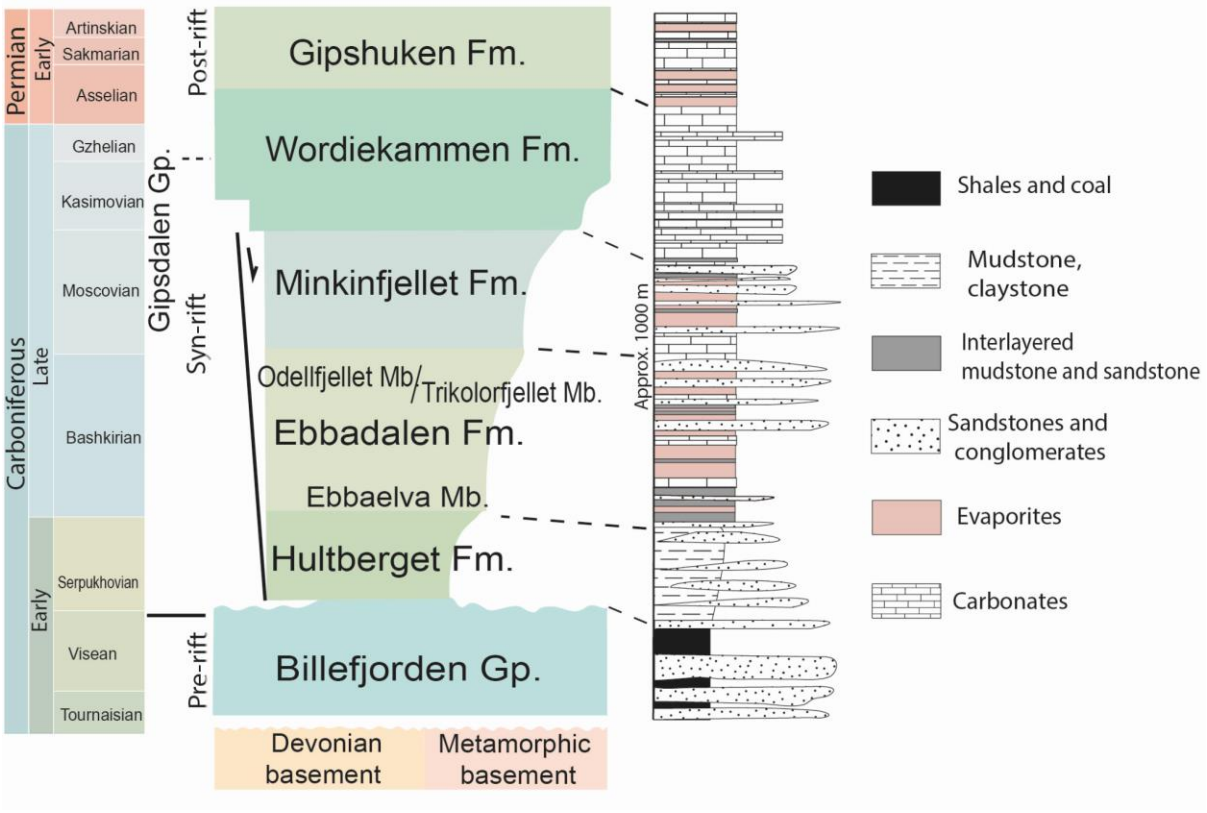

Figure 8. Carboniferous stratigraphy and simplified lithological profile demonstrating a shift from humid-tropical climate during deposition of the coal-bearing Billefjorden Group to the semi-arid to arid climate of the Gipsdalen Group seen as a change of climate sensitive facies from coal-bearing units to red siliciclastics, evaporites and warm water carbonates. Modified from Braathen et al. (2011) and Smyrak-Sikora et al. (2021).

The impact of the LPIA glaciations and deglaciations is most readily recognized in the Bashkirian (Mississippian; onset ca. 323 Ma) to Sakmarian (Cisuralian; ca. 293–290 Ma) part of Gipsdalen Group, namely the paralic to marine syn- to post-rift succession comprising the upper Ebbadalen to Gipshuken formations (Fig. 8). Glacio-eustatic sea-level variations related to LPIA significantly impacted sedimentation in shallow shelf and coastal environments. Episodes of sea-level lowstands are represented by terrestrial siliciclastics, gypsum strata that precipitated in salinas and sabkhas, karst and exposure surfaces which are interbedded with restricted to open-marine carbonate deposits formed during the sea level highstands (Stemmerik, 2000; 2008; Ahlborn and Stemmerik., 2015; Sorento et al., 2020; Smyrak-Sikora et al., 2021). The number of cycles in the Bashkirian to Sakmarian part of the Gipsdalen Group exceeds 130 cycles (Ahlborn and Stemmerik, 2015; Sorento et al., 2020;

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Smyrak-Sikora et al., 2021), however, the lack of good stratigraphic control limits cyclostratigraphic constraints. The LPIA in Svalbard is manifested also by Asselian (peak icehouse) atmospheric dust load estimated to be higher than the Moscovian (moderate icehouse; Oordt et al., 2020). This is consistent with the record from the Russian Platform that shows a $\delta^{18}O$ maximum during the glacial maximum in the Asselian (Grossman et al., 2008). Asselian to Artinskian ca. 2.5‰ decrease in $\delta^{18}O$ in the Southern Urals is attributed to ca. 4-7 °C increase in temperature and used as evidence for glacial retreat (Korte et al., 2005). This glacial retreat is questioned by Grossman et al., (2008) who show increasing $\delta^{18}O$ values related to aridification. A major deglaciation event is, however, seen in the Sakmarian as regional deepening of the carbonate platforms and temporary stop of glacio-eustatic cyclic deposition (Ahlborn and Stemmerik, 2015; Sorento et al., 2020).

4.2.3. Late Permian: cold-water carbonate platform

A transition from warm-water carbonate-evaporite deposition to the temperate to cool-water mixed siliceous–carbonate ramp occurred in the upper Artinskian (Cisuralian; ca. 285 Ma), and corresponds to the transition from the Gipsdalen Group to the Tempelfjorden Group (Ezaki et al., 1994; Stemmerik, 2000; 2008; Hüneke et al., 2001; Stemmerik and Worsley, 2005; Blomeier et al., 2009; 2013; Buggisch et al., 2012; Dustira et al., 2013; Sorento et al., 2020; Olaussen et al., 2025). The transition is attributed to the continued northern drift of Svalbard, closure of the Uralian seaway to the warmer Tethys to the southeast (Stemmerik 2008) and was likely also the result of a deepening of the entire shelf (Blomeier et al., 2013). For the remainder of the Permian, cool- to cold-water conditions prevailed along the northwestern margin of Pangea, leading to the deposition of a ca. 460 m thick succession dominated by spiculitic chert and cool-water carbonates (Cutbill and Challinor 1965; Blomeier et al., 2013; Uchman et al., 2016; Matysik et al., 2018).

Extensive oxygen isotopic data has been derived from brachiopods from the Artinskian-Changhingian (middle Cisuralian to Lopingian; ca. 285-252 Ma) Kapp Starostin Formation (e.g., Gruszczynski et al., 1989; Mii et al., 1997; Korte et al., 2005; Nielsen et al., 2013). However, the high variability in the isotopic data from these Permian brachiopods (e.g., Gruszczynski et al., 1989; Korte et al., 2005) is inconsistent with the marine habitat of these taxa (Grossman et al., 2008). Diagenetic alteration accounts for the low $\delta^{18}O$ values of most samples (Mii et al., 1997). To exclude diagenetically altered brachiopods, many researchers have used geochemistry and petrography (e.g., Mii et al., 1997) and targeted the best-preserved parts of shells. Excluding these potentially altered values, brachiopod $\delta^{18}O$ values are generally -2 to -7 ‰ for the Kungurian-Wuchiapingian (upper Cisuralian-Lopingian; ca. 283–254 Ma) interval of the Kapp Starostin Formation (Mii et al., 1997). The Guadalupian-Lopingian $\delta^{13}C$ maximum of 7.5‰ represents the highest spiriferid brachiopod $\delta^{13}C$ values in the Phanerozoic (Gruszczynski et al., 1989; Mii et al., 1997) and may reflect changes in global storage of organic carbon (Mii et al., 1997). Matysik et al. (2017) investigated the multistage diagenesis of the Kapp Starostin Formation, at medium burial depths with deep-burial overprinting. The cement in Kapp Starostin Formation shows $\delta^{18}O$ values between +2‰ and 30‰ and demonstrates a peak at temperatures of 23 to 43°C corresponding to burial depths of 0.6 to 1.3 km (Matysik et al., 2018).

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4.2.4. Capitanian Crisis

Within the Kapp Starostin Formation, it has also been proposed that a less severe middle Permian (Capitanian) mass extinction is recorded (Bond et al., 2015). This Capitanian crisis is thought to be indicated by a negative $\delta^{13}C_{org}$ isotope excursion, a lithofacies change (i.e., the loss of carbonate beds), and a drop in species richness (Bond et al., 2015). In addition, these changes have been correlated to similar changes in the Sverdrup Basin, Arctic Canada (Bond et al., 2020). These changes are also associated with redox proxies (pyrite framboids, Th/U and V/Al) suggesting the development of anoxic conditions, and the loss of carbonates from the Kapp Starostin Formation was interpreted to be the consequence of a sustained interval of ocean acidification (Beauchamp and Grasby, 2009; Grasby et al., 2015), making this transition consistent with other hyperthermal events. The timing of these changes is also consistent with the changes observed at tropical paleolatitudes (Sun et al., 2010; Wignall et al., 2012), suggesting that the Capitanian crisis was a global event. However, the interpretation that the Capitanian crisis is recorded in the Kapp Starostin Formation is disputed, owing to the lack of biostratigraphical data confirming the rock are of Capitanian age (Nakazawa 1999; Shen et al., 2005; Lee et al., 2022). A reanalysis of the same sections using brachiopod data suggested that this event is not the Capitanian crisis, but instead a faunal turnover that occurred during the Kungurian (Lee et al., 2022). Moreover, the development of ocean acidification is interpreted to have persisted for millions of years (Grasby and Bauchamp, 2009), although the large carbonate buffering capacity of ocean water suggests that ocean acidification is unlikely to persist for such long intervals of time (Hönisch et al., 2012). It, therefore, remains equivocal whether the Kapp Starostin Formation records the Capitanian Crisis.

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4.3. Mesozoic (Triassic, Jurassic, Cretaceous, 252.2–66 Ma)

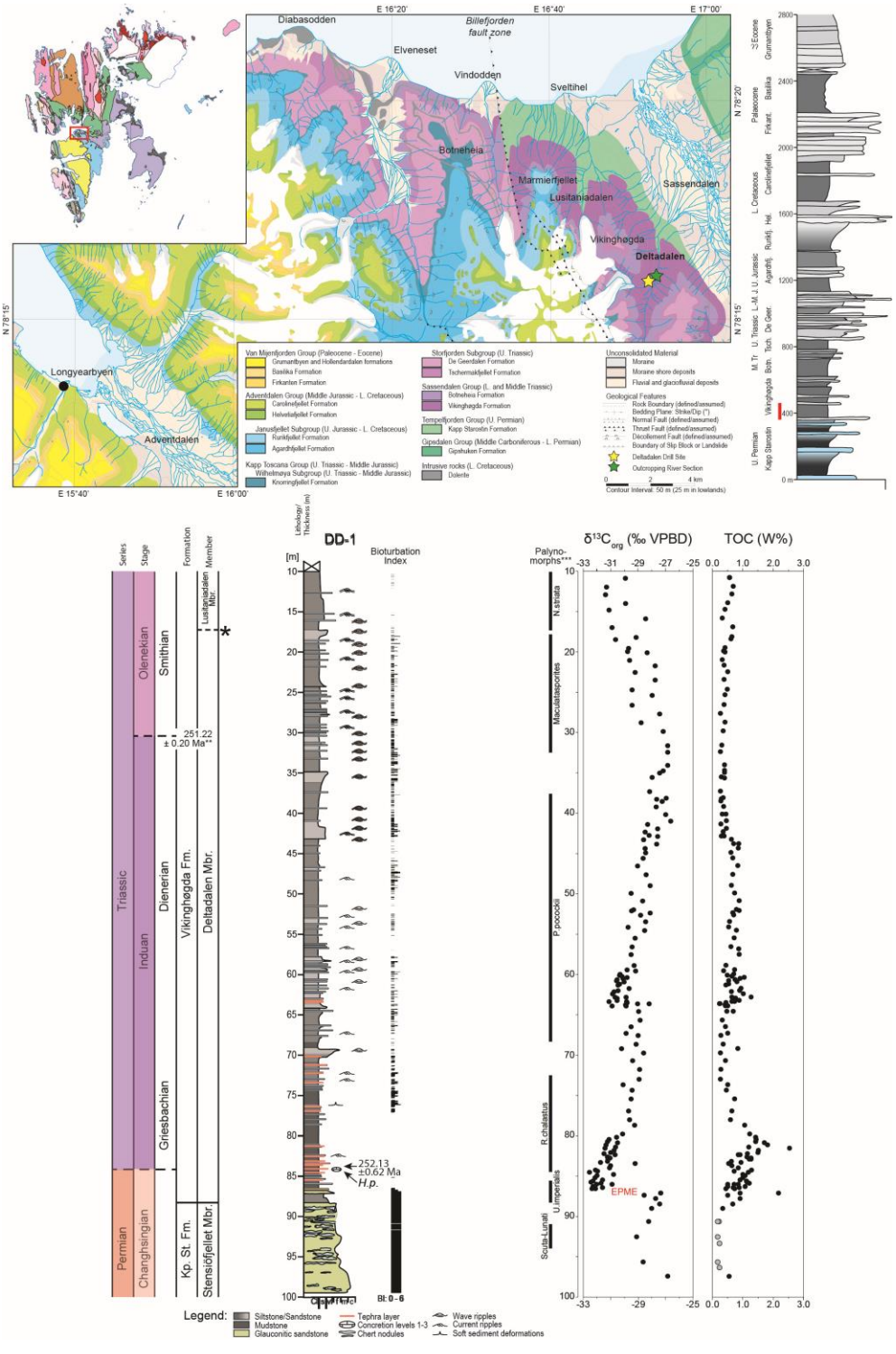

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Figure 9. Top: Geological map of central Spitsbergen highlighting the drill site (yellow star) and the adjacent Deltadalen river section (green star; geological map adapted from Major et al., 1992). Overview of the sedimentary section illustrating the lithostratigraphic formations in central Spitsbergen (after Dallmann et al. (1999), Mørk et al. (1999b), Midtkandal et al. (2008), Nagy and Berge (2008), Dypvik et al. (2011), Blomeier et al. (2013), Lord et al. (2014), Koevoets et al. (2016), and Smelror and Larssen (2016). Bottom: The Permian/Triassic boundary as it appears in the Deltadalen DD-1 drill core. Modified after Zuchuat et al. (2020). EMPE – End-Permian Mass Extinction Event; Kp. St. Fm. - Kapp Starostin Formation; Mbr. – member; Fm – formation; H.p - Hindeodus parvus; * - member boundary after Mørk et al. (1999b); **Induan-Olenekian boundary age after Burgess et al., (2014); ***palynological data from Vigran et al. (2014).

4.3.1 The Permian-Triassic transition

The End-Permian Mass Extinction Event (EPME) at ca. 252 Ma (Burgess et al., 2014) was the most catastrophic extinction event of the Phanerozoic, which decimated 75% of terrestrial species (Hochuli et al., 2010) and 81% of marine species (Stanley, 2016). This extinction is associated with a marked and continuous global negative carbon isotope excursion (CIE; Korte and Kozur, 2010). The cause(s) of the EPME are debated; both a bolide impact, and the emplacement and eruptions of the Siberian Traps LIP are implicated as causal mechanisms (Svensen et al., 2009; 2018; Grasby et al. 2011; Sanei et al.2012; Ogden and Sleep, 2012; Ivanov et al., 2013; Burgess and Bowring, 2014; Wu et al., 2021). Even though the intrusive and extrusive character of the Siberian Traps LIP is generally accepted as the extinction trigger that led to the cascading environmental changes, it is not fully understood which environmental changes led to the collapse of terrestrial and marine ecosystems (e.g., Hochuli et al., 2010; Korte and Kozur, 2010; Black et al., 2014; Grasby et al., 2015; Joachimski et al., 2020; Scotese et al., 2021; Wu et al., 2021; Galloway and Lindström, 2023a).

In Svalbard, the mass extinction event is expressed differently compared to equatorial Tethyan carbonate successions. In west and central Spitsbergen, the pre-extinction interval belongs to the uppermost part of the Kapp Starostin Formation, which is usually devoid of any skeletal fossil material (Fig. 9; Bond et al., 2015; Grasby et al., 2015; Lee et al., 2022). The only exception is the poorly preserved lingulid brachiopod species documented (e.g., Gobbet 1964) and rare impressions of large brachiopods and bivalves (WJ. Foster, pers. obvs.) so far, making it virtually impossible to robustly reconstruct diversity dynamics during this important interval (Uchman et al., 2016; Foster et al., 2022). The Permian Kapp Starostin Formation is, therefore, poorly age-constrained, and no index fossils of Changhsingian (pre-extinction) age have yet been identified. Based on sedimentological evidence as well as on the nature of the sharp negative $\delta^{13}C_{org}$ excursion, the upper part of the Kapp Starostin and the overlying Triassic Sassendalen Group (separated in the Vardebukta Formation in the west and Vikinghødga Formation in the east) have been interpreted to represent continuous deposition across the Permian-Triassic boundary (Wignall et al., 1998; Schobben et al., 2020; Zuchuat et al., 2020). This transition is also associated with the abrupt disappearance of cemented, highly bioturbated, spiculite- and chert-bearing mudstones and sandstones, conformably overlain by easily weathered, usually laminated, scarcely bioturbated, silica-poor mudstones (e.g., Mørk et al., 1993; 1999a; 1999b; Uchman et al., 2016; Rodriguez-Tovar et al., 2021). The EPME has, therefore, been interpreted as a single horizon at the base of the Vardebukta Formation in

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west Spitsbergen (Wignall et al., 1998) and ~1.6 m into the Vikinghøgda Formation in central Spitsbergen (Mørk et al., 1999; Nabbefeld et al., 2010).

The post-extinction sediments, which in Svalbard are assigned to the Lower Triassic Vardebukta and Vikinghøgda formations, yield a scarce and low-diversity ichno-assemblage (Wignall et al., 1998; Uchman et al., 2016; Rodriguez-Tovar et al., 2021), as well as abundant macrofossils, including key Triassic index fossils (ammonoids and conodonts), which have been useful in inferring the timing of the extinction and the recovery. From the base of both formations, a diverse assemblage of ammonoids have been recorded including *Otoceras boreale*, *Glyptophiceras nielseni*, *Ophiceras spathi*, *O.* cf. *compressum*, *O.* cf. *kochi*, *O.* cf. *poulseni*, *Paravishnuites paradigma*, and *P. oxynotus* (Mørk et al., 1999), which occur ca. 6 m above the base of the Vikinghøgda Formation (Nakrem et al., 2008) and 2.5 m into the Vardebukta Formation (WJ. Foster, pers. obvs.). Mørk et al. (1999), Nakrem et al. (2008), and Zuchuat et al. (2020) also described conodonts from the base of the Vikinghøgda Formation, which suggest that the interpreted extinction horizon in Svalbard is time equivalent with the onset of the mass extinction at the Global Stratotype Section and Point (GSSP) in Meishan, China, and the Permian-Triassic boundary occurs 4.1 m into the Vikinghøgda Formation. This is similar to other high-latitude clastic Permian-Triassic successions in Jameson Land, Greenland (Twitchett et al., 2001), the Sverdrup Basin in Canada (Henderson and Baud, 1997), and South Verkhoyansk region, Russia (Biakov et al., 2016). Furthermore, in South Verkhoyansk, the Changhsingian sandstones record bivalve communities with large Intomodesma species that suddenly go extinct at the base of the Otoceras concavum ammonoid zone (Biakov et al., 2016), coincident with the change in bioturbation record in Svalbard (Uchman et al., 2016; Rodriguez-Tovar et al., 2021).

Numerous geochemical and sedimentological studies have investigated the environmental changes recorded in Svalbard associated with the EPME. The negative $\delta^{13}C_{org}$ and $\delta^{13}C_{carb}$ isotope excursions, which occur just prior to the Permian-Triassic boundary, reflects a rapid influx of isotopically light carbon into the atmosphere, while the influx of heavy metals and the presence of abundant tephra layers, including one just above the first appearance datum (FAD) of *Hindeodus parvus* and dated at 252.13 ± 0.62 Ma (Zuchuat et al., 2020), which is in good agreement with the tephra beds from the Induan GSSP section in Meishan (Burgess et al., 2014) that have been inferred to link the Siberian Traps LIP and the mass extinction in Svalbard (Gruszczynski et al., 1989; Grasby et al., 2015; Zuchuat et al., 2020). The reduced Iron (Fe)/Potassium (K) elemental ratio that accompanied the extinction horizon in the Vikinghøgda Formation seems to suggest that the tropical atmospheric circulation (Hadley Cell) could have expanded towards the poles, associated with an increased aridity in the hinterland of the basin (Zuchuat et al., 2020). Redox proxies, including lipid biomarkers (Summons et al., 2022), Fe and P speciation (Schobben et al., 2020), trace-metal data (Grasby et al., 2015; Uchman et al., 2016; Wignall et al., 2016; Zuchuat et al., 2020), and pyrite framboid sizes (Dustira et al., 2013; Wignall et al., 2016) also suggest that the mass extinction is associated with the expansion of oxygen minimum zones in the ocean, bringing anoxic and euxinic conditions into shallow-marine settings, as well as subsequent pulses of redox changes throughout the Early Triassic (Rodriguez-Tovar et al., 2021). Isotopic signatures of lipid biomarkers suggest frequent phytoplankton blooms, and phosphorus speciation data indicate an increase in nutrient supply and the remobilization of biologically available P as a consequence of the mass-extinction event initiating feedback that further developed anoxic conditions (Nabbfeld et al., 2010; Schobben et al., 2020).

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Thermal stress and ocean acidification are also widely considered as key factors in the EPME, with global average temperature increases reaching 7°C (Kidder and Worsley, 2004; Svensen et al., 2009; Stordal et al., 2017; Burger et al., 2019), potentially as much as 9–12°C (Joachimski et al., 2012; 2020; Schobben et al., 2014; Chen et al., 2016). In Svalbard, there are currently no published geochemical investigations of the environmental changes associated with this hyperthermal event, but the presence of warm-water taxa such as red algae (Wignall et al., 1998), conodont genus *Clarkina* (Nakrem et al., 2008) and both ostracod and radiolarian species that were equatorial during the Changhsingian (Foster et al., 2023) suggest that higher paleolatitudinal settings were unusually warm following the mass extinction. The cessation of carbonate rocks at the top of the Kapp Starostin and across the Boreal Realm have also provided an alternative hypothesis that ocean acidification developed and persisted for most of the Late Permian (Beauchamp and Grasby, 2009; Grasby et al., 2015). This hypothesis, however, requires the persistence of undersaturated conditions for millions of years, which is inconsistent with some Earth system models that suggest that ocean acidification events cannot persist for this length of time (e.g., Hönisch et al., 2012). In addition, the lack of dissolution and repair marks on well-preserved mollusks from the extinction aftermath have also been interpreted to suggest that ocean acidification was not severe enough to have impacted skeletal calcification in the Boreal realm, at least at the onset of the Triassic (Foster et al., 2022). More research is, therefore, required to understand the role of thermal stress and ocean acidification in high-latitude marine extinctions.

The impact of EMPE on terrestrial ecosystems can also be investigated from Svalbard's marine successions. The Permian to Triassic palynological record of Svalbard and the Barents Shelf has been intensely investigated (e.g., Mangerud and Konieczny, 1993), in-part due to their utility for petroleum exploration (e.g., Vigran et al., 2014). A spore spike demise of gymnosperms, malformed spores and pollen, a drop in abundance of acritarchs, and compound specific isotopes of algal and land-plant-derived biomarkers all coincide with the mass extinction event, suggestingnear-synchroneity between effects in marine and terrestrial realms (Stemmerik et al., 2001; Nabbfeld et al., 2010; Uchman et al., 2016). The presence of Permian plant taxa, including major Paleozoic plant groups in the Lower Triassic successions of Svalbard and the Barents Shelf, however, have led to some authors interpreting that the EMPE only had a minor impact on plant communities (Hochuli et al., 2010; Vigran et al., 2014). Aberrant pollen and spores reported from the Barents Sea and elsewhere have been suggested to be a consequence of severe atmospheric pollution and increased UVA-B radiation due to emissions from emplacement of the Siberian Traps (Black et al., 2014; Hochuli et al., 2017; Galloway and Lindström, 2023 and references therein).

4.3.2. Early Triassic ecosystem recovery

The post-EPME survival and recovery of marine organisms recorded from Svalbard and the Boreal realm was unique. Within the early Griesbachian H. parvus conodont zone in central Spitsbergen (Lusitaniadalen and Deltadalen), a diverse assemblage of macro and microfossils have been recorded, including the only documented fully silicified marine assemblage of the Early Triassic (Foster et al., 2017), the oldest record of post-extinction silica organisms globally (radiolarians and siliceous sponges; Foster et al., 2023), and the presence of an ecological complex assemblage of trace fossils (Nabbefeld et al., 2010; Rodríguez-Tovar et al., 2021). In addition, across Svalbard, the Lower Triassic succession preserved many groups, including bryozoans

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(Nakrem and Mørk, 1991), algae (Wignall et al., 1998), conodonts (Nakrem et al., 2008), bivalves and gastropods (Buchan, 1965; Tozer and Parker 1968; Foster, 2015; Foster et al., 2017), ammonoids (see Nakrem et al., 2008), ostracods (Olempska and Błaszyk 1996), echinoderms (Salomon et al., 2015), and trace fossils (Wignall et al., 1998). Whilst sedimentation-rate calculations suggest marine ecosystems only required ca. 150 kyr to recover from the mass extinction (Rodríguez-Tovar et al., 2021), based on index conodonts, there is a distinctive pulse of environmental and ecological recovery in the Dienerian (Hatleberg and Clark, 1984; Wignall et al., 1998; Mørk and Worsley, 2006; Salamon et al., 2015).

The Lower Triassic succession of Svalbard is also fundamental for understanding the evolution and radiation of marine vertebrates following the Permian-Triassic transition. The Triassic succession of Svalbard has long been well-known for four described vertebrate fossil horizons (in stratigraphic order): The Fish Niveau, Grippia Niveau, Lower Saurian Niveau, and Upper Saurian Niveau (Wiman, 1910; Wiman, 1928). These bonebeds correspond to the Lusitaniadalen (the Fish Niveau) and Vendomdalen (the Grippa and Lower Saurian Niveau) members of the Vikinghøgda Formation that span the Smithian-Spathian transition (Lower Triassic; ca. 249.2 Ma), and the Ladinian-age (Middle Triassic; ca. 242–237 Ma) Blanknuten Member of the Botneheia Formation (the Upper Saurian Niveau; Maxwell and Kear, 2013; Hurum et al., 2018). Recent work on Early and Middle Triassic ecosystems in Svalbard reveals an exceptionally rapid diversification among marine vertebrates andichthyosaurs likely evolved prior to the EPME (Kear et al., 2023). The bonebeds reveal a much more complex foodweb than previously thought (Hurum et al., 2014; 2018; Bratvold, 2016; Delsett et al., 2017; Ekeheien, 2018; Engelschiøn et al., 2018; Økland et al., 2018; Roberts et al., 2022; Kear et al., 2023), suggesting that the Boreal Sea served as a climatic refuge after the EPME (Kear et al., 2023; Foster et al., 2023).

It has been hypothesized that the recovery from the EPME was delayed due to subsequent crises throughout the Early Triassic (Payne et al., 2004; Ware et al., 2011; Song et al., 2012; Foster et al., 2017b; Zuchuat et al., 2020; Wu et al., 2021), despite the degree of ichnofacies diversity and intensity reached pre-extinction level in ca. 150 Kyr (Rodríguez-Tovar et al., 2021). The first crisis occurring after the EPME is the “late Dienerian biotic crisis” (late Early Triassic; ca. 251 Ma), which is recognized by a negative CIE in the Vikinghøgda Formation in Deltadalen, central Spitsbergen, where it is sandwiched between two thin tephra layers. This benthic crisis was associated with dysoxic conditions in the water column and the seafloor (Zuchuat et al., 2020) and was first documented from the Tethyan with the Werfen Formation, northern Italy (Hofmann et al., 2015; Foster et al., 2017), as well as from subtropical latitudes along the Panthalassic margin (Ware et al., 2011; Hofmann et al., 2011; 2014). The palynological record from Greenland suggests that this CIE coincides with a more significant turnover in plant communities than at the end-Permian mass extinction (Hochuli et al., 2016). This indicates that the late Dienerian Crisis might have been of global significance. The presence of the two tephra horizons directly above and below the negative CIE of the late Dienerian biotic crisis suggest a volcanic trigger for the event, but the exact nature of this potential biotic crisis in Svalbard and globally requires additional research.

The most prevalent of the biotic crises in the Triassic is the Smithian-Spathian transition, which is associated with a negative CIE greater than at the EPME (Payne et al., 2004; Grasby et al., 2013; 2016b). In Svalbard, the Smithian-Spathian transition is recorded as a regression and subsequent transgression, with numerous fossiliferous horizons (e.g., Hoel and Orvin, 1937;

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Buchan, 1965; Tozer, 1967; Weitschat and Lehmann, 1978; Mørk et al., 1999; Foster, 2015; Hammer et al., 2019; Hansen et al., 2024; Leu et al., 2024). While the Smithian-Spathian transition has received considerably less attention than the Permian-Triassic transition, recent research has highlighted the multi-factor nature of this event on the dynamic of the C and P cycles in the Arctic (Hammer et al., 2019; Blattmann et al., 2024). Nevertheless, many outstanding questions remain unanswered both globally and on Svalbard: What was the timing of the event? What caused the CIE? What environmental changes are associated with the CIE? How were terrestrial and marine ecosystems impacted by the event?

At the Festningen section, Hg concentrations and Hg/TOC both show a noticeable spike in the Tvillingodden Formation of the Sassendalen Group, which have been associated with Hg loading associated with increased activity from the Siberian Traps (Grasby et al., 2016). This peak has also been correlated with Hg loading recorded at the Smithian-Spathian boundary at Smith Creek, Arctic Canada, even though the Smithian-Spathian boundary is associated with a carbon isotope peak at most sections globally. In the equivalent Vikinghøgda Formation at the Wallenbergfjellet section, however, the Smithian-Spathian boundary (the top of the Wasatchites tardus ammonoid zone) is associated with a positive CIE (Galfetti et al., 2007; Hammer et al., 2019) and the Hg loading occurred prior to the Smithian-Spathian boundary in the middle to late Smithian (Hammer et al., 2019). This is consistent with the Festningen section, which suggests the Hg loading occurred synchronously during the Smithian (Grasby et al., 2016a). At the Kongressfjellet and Vikinghøgda sections, just above the base of the Vendomdalen Member of the Vikinghøgda Formation, a significant shift in the palynological record is recorded (Mørk et al., 1999; Galfetti et al., 2007). This Smithian-Spathian turnover is also recorded in the shallow cores drilled at the Svalis Dome, central Barents Sea (Hochuli and Vigran, 2010). This turnover in the palynological record has been associated with the re-establishment of diverse woody gymnosperm ecosystems, marking a recovery signal from both the Permian-Triassic climate crisis and a "late Smithian Thermal Maximum". This is supported by the synchronous recovery in the latitudinal diversity gradient of ammonoids (Brayard et al., 2009) and the recovery of equatorial benthic marine communities (Twitchett and Wignall, 1996; Chen et al., 2011; Pietsch and Bottjer, 2014; Hofmann et al., 2014; 2015; Foster et al., 2015; 2017; 2018; 2023a). The Smithian-Spathian transition, therefore, appears to be marked by a late Smithian Thermal Maximum and a subsequent Smithian-Spathian boundary cooling and associated biotic recovery in Svalbard, and pan-Arctic.

4.3.3 Middle to Late Triassic: organic-rich mudstones rich in phosphate

The Middle Triassic succession is dominated by organic-rich mudstone to siltstones of the Anisian-Ladinian (Middle Triassic; ca. 247–237 Ma) Botneheia Formation of the upper Sassendalen Group (Mørk and Bjorøy, 1984; Krajewski, 2008, 2013; Wesenlund et al., 2021). The upper Blanknuten Member of the Botneheia Formation (Ladinian) reaches 12 wt.% TOC, and certain stratigraphic intervals, particularly in the underlying Muen Member (Anisian), contain abundant phosphorite nodules (e.g., Krajewski, 2013; Wesenlund et al., 2021; 2022; Engelschiøn et al., 2023a). In the Anisian, nutrient saturated runoff from continental areas, particularly associated with the approaching delta system from the southeast (see next section), coupled with upwelling of nutrient-rich waters from the Panthalassic Ocean, resulted in extensive algal blooms and the formation of oxygen

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minimum zones, which promoted dysoxia and anoxia and preservation of organic matter and precipitation of phosphate (Krajewski, 2013; Vigran et al., 2014; Wesenlund et al., 2022; Engelschiøn et al., 2023a). Repeated transgression-regression events, influenced by the emerging delta system, likely contributed to fossil-preservation potential of the Middle Triassic strata as the relatively shallow offshore environment was temporarily punctuated by anoxic events (Mørk et al., 1989; Krajewski, 2013; Engelschiøn et al., 2023a). Fossil preservation has also occurred by complete barium sulfate (barite) pseudomorphing, possibly by sulfate remobilization from the organic-rich shales (Engelschiøn et al., 2023b). Moreover, the high TOC values in the Ladinian Blanknuten Member suggest deposition under euxinic conditions, possibly governed by restricted water circulation due to shallowing of the basin, as well as water mass stratification caused by the increasing influx of riverine waters into the marine basin (Wesenlund et al., 2022).

4.3.4. Late Triassic: the Carnian Pluvial Episode and the world's largest delta plain

The Carnian Pluvial Episode (CPE, ca. 233 Ma) marks a period of climate and biotic changes spanning the Julian 2 and Tuvalian 1 substages of the Carnian (Simms and Ruffell, 1989; Breda et al., 2009; Dal Corso et al., 2018). The CPE was first recognized as concomitant to global carbonate platform environments perturbations associated with an increased terrigenous input into sedimentary basins (Dal Corso et al., 2018) and the rapid diversification of dinosaurs (Benton et al., 2018). The CPE is also associated with several isotope perturbations in terrestrial and marine C and Hg records, which potentially reflects multiple pulses of volcanic activity of the Wrangellia LIP (Dal Corso et al., 2018; Jin et al., 2023). Analyses of palynological assemblages, chemical weathering indices, $\delta^{18}O$ apatite records, and redox-sensitive elements suggest that the CPE was characterized by an extremely humid climate (e.g., Roghi, 2004; Baranyi et al., 2019), warm temperatures (Rigo and Joachimski, 2010; Rigo et al., 2012; Sun et al., 2016) and widespread marine dys- and anoxia (Soua, 2014; Sun et al., 2016; Tomimatsu et al., 2021). Shallow-water carbonate production switched as a result of climatic variations and eustatic sea-level fall (Jin et al., 2020), which impacted both the marine and continental biosphere, with high extinction rates in ammonoids and conodonts (Rigo et al., 2007; Dal Corso et al., 2022), rapid extinction of terrestrial tetrapods and a subsequent diversification of dinosaurs (Bernardi et al., 2018), mammals (Benton et al., 2018), scleractinian corals (Stanley, 2003), calcareous dinoflagellates and plants (e.g., Dal Corso et al., 2022).

While the CPE is well-documented in the Tethyan Realm (Dal Corso et al., 2018), evidence from the Boreal Realm remains limited. In Svalbard, preliminary palynological evidence from the Kapp Toscana Group in central Spitsbergen integrated with organic carbon isotope and paleomagnetic constraints indicate warming during the late Julian-1 (Mueller et al., 2016; Paterson et al., 2016) and suggest wetter conditions starting from the Julian-2, which occurs in the lower part of the De Geerdalen Formation (Mueller et al., 2016). In addition, the detailed study of paleosols in the De Geerdalen Formation above CPE seems to indicate the transition from humid (coal) to warm arid (caliche) climate settings (Lord et al., 2022). Nevertheless, the precise location of the CPE in Svalbard's stratigraphy remains uncertain, and increased research can help understand the exact triggering mechanism of these climate perturbations.

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4.4.2. Jurassic-Cretaceous: A greenhouse with cold snaps

The Jurassic and Cretaceous saw some of the warmest background global temperatures of the Phanerozoic (Jenkyns et al., 2012). Permanent polar ice caps were not present, although evidence for episodic cooling and the growth and decay of small, ephemeral polar ice caps has been presented (e.g., Price, 1999; Miller et al., 2005; Grasby et al., 2017; Alley et al., 2020). Oxygen isotope records of Jurassic and Early Cretaceous from Svalbard have been derived largely from belemnites (e.g., Ditchfield, 1997; Price and Nunn, 2010; Hammer et al., 2011) and an extensive dataset has been derived from Kong Karls Land (Ditchfield, 1997). The usefulness of these data, however, is hampered by modest biostratigraphic age constraints and pervasive diagenesis. Nonetheless, some well-preserved belemnites were identified, and oxygen isotope data from Hammer et al. (2011) from the upper Agardhfjellet Formation (Volgian–Ryzanian, which approximately correlates with late Tithonian to early Berriasian, Late Jurassic-Early Cretaceous; ca. 149-140 Ma) and Price and Nunn (2010) from the Hauterivian (Early Cretaceous; ca. 133-126 Ma) part of the Rurikfjellet Formation of Festningen (see Jelby et al., 2020b for a discussion of the age constraints) show $\delta^{18}O$ values ranging from -3.0 to 0.8 ‰ VPDB. The belemnite data from Kong Karls Land (spanning the Aalenian to Valanginian; Middle Jurassic to Early Cretaceous; ca. 175–132 Ma) give $\delta^{18}O$ values of -0.7 to 1.2 ‰ VPDB. These data argue for cooling and warming episodes at these high paleolatitudes and may indicate high seasonality and/or high frequency climatic variability, although the absolute temperatures they represent are hard to interpret. Common practice would be to assume that the seawater $\delta^{18}O$ was -1 ‰ SMOW (i.e., that of an ice-free world), and use either the equation for molluscan calcite (Anderson and Arthur, 1983) or experimentally derived synthetic calcite (Kim and O'Neil, 1997) to calculate the precipitation temperature of the calcite. However, Price and Nunn (2010) used the presence of glendonites in certain horizons to independently assess paleotemperature of those intervals, and backcalculated the $\delta^{18}O_{sw}$. This suggested that the seawater $\delta^{18}O$ was much lower than the global average (i.e., heavily meteorically influenced), which is supported by the depositional setting (ranging from marine to terrestrial delta plain). Furthermore, recent work using clumped isotope thermometry on belemnites suggests that the equations of Kim and O'Neil (1997) or Anderson and Arthur (1983) are not suitable for belemnite calcite, and rather, the temperature equations of Kele et al. (2015) or Daëron et al. (2019) should be used (Price and Passey, 2013; Wierzbowski et al., 2018; Bajnai et al., 2020; Price et al., 2020; Vickers et al., 2019b; 2020; 2021). Using the latterresults in warmer estimated temperatures than previously thought, even if one assumes a meteorically influenced $\delta^{18}O_{sw}$ of -2.5 ‰ SMOW. For example, the belemnite data from Kong Karls Land suggests temperatures closer to 12-20 °C rather than the 8 to 13°C originally postulated by Ditchfield (1997). The belemnites from Festningen may indicate a temperature range of around 13–30 °C rather than the 9–25 °C originally suggested by Price and Nunn (2010) and Hammer et al. (2011).

4.4.2.1. Jurassic–Cretaceous boundary and the Volgian Isotopic Carbon Excursion (VOICE)

The global carbon-isotope ($\delta^{13}C$) signal of the Upper Jurassic and Jurassic-Cretaceous (J-K, ca. 145 Ma) boundary as recognized in Tethyan, Atlantic, and Pacific sections is generally characterized by a steady decline. This decline has been attributed to progressive deceleration of the carbon cycle due to the development of more oligotrophic oceanic conditions and

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reduced marine primary production (Weissert and Channell, 1989; Weissert et al., 1998; Weissert and Erba, 2004; Tremolada et al., 2006; Price et al., 2016). In Svalbard, however, the upper Kimmeridgian-middle Volgian succession displays a prominent negative CIE termed the "Volgian Isotopic Carbon Excursion" or "VOICE" by Hammer et al. (2012). The VOICE is of much greater magnitude than the entirety of the long-lived decline of the lower-latitude records (Fig. 11; Morgans-Bell et al., 2001; Price and Rogov, 2009; Žák et al., 2011; Hammer et al., 2012; Zakharov et al., 2014; Koevoets et al., 2016; 2018; Galloway et al., 2020; Jelby et al., 2020b). This recently recognized carbon-isotope marker of Boreal sections (observed across Svalbard, northern Siberia, Arctic Canada, the Russian Platform and possibly southern UK; Galloway et al., 2020; Jelby et al., 2020b and references therein) and newly discovered occurrences in the Neuquén Basin of Argentina (Capelli et al., 2022; Weger et al., 2022) and possibly in the eastern Tethys (Fallatah et al., 2024), is characterized by a relatively abrupt negative excursion (≤6.4‰ in Svalbard) in $\delta^{13}C_{org}$ values. This excursion is followed by a positive trend in $\delta^{13}C_{org}$ values through the upper Volgian-Ryazanian and across the J-K boundary (Price and Rogov, 2009; Hammer et al., 2012; Dzyuba et al., 2013; Zakharov et al., 2014; Koevoets et al., 2016; Galloway et al., 2020; Jelby et al., 2020b; Vickers et al., 2023).

In Svalbard, the VOICE is documented in both drill core (DH-2, central Spitsbergen) and outcrop sections (e.g., Myklegardfjellet, eastern coast of Spitsbergen; Festningen, western Spitsbergen), and falls within the paper-shale-dominated Oppdalssåta and Slottsmøya members of the Agardhfjellet Formation (Koevoets et al., 2016; 2019; Jelby et al., 2020b; Vickers et al., 2023). The VOICE and positive recovery across the J-K boundary were originally considered unique to the Boreal realm, as a result of decoupling from the global carbon reservoir during the Late Jurassic. Galloway et al. (2020) and Jelby et al. (2020b) attributed this Boreally-limited CIE to restricted oceanographic connectivity between the shallow epeiric seas of the high northern latitudes and open oceans, associated with water-mass stratification and increased continental runoff (Park et al., 2024) due to an eustatic sea-level lowstand. As a result, basinal depletion of ^{13}C resulted from oxidation of terrestrial organic matter, or input of isotopically light CO_2 by respiration of marine organisms, and/or riverine dissolved inorganic carbon (DIC) (Patterson and Walter, 1994; Holmden et al., 1998). However, the recent discovery of the VOICE in the southern hemisphere (Rodriguez Blanco et al., 2022; Capelli et al., 2022; Weger et al., 2022), albeit with diachroneity seen also in the Arctic (e.g., Rogov, 2021; Rogov et al., 2023), that may reflect poor age control or a latitudinal climate gradient, indicates that the excursion is not limited to high northern latitudes. Weger et al. (2022) suggested that the VOICE was driven by changes in the input of terrestrially-derived organic matter, controlled by relative sea-level change and climate, although this hypothesis is refuted by Fallatah et al., (2024) and Galloway et al. (2024) because the VOICE may be present in a restricted setting in the Tethys (Fallatah et al., 2024) and because there is a lack of consistent changes in organic matter type (as indicted by OI and HI), indices of weather and grain size across the VOICE in the Canadian and Festningen material (Galloway et al., 2024).

Collectively, the VOICE is ascribed variously to restricted circulation (Galloway et al., 2020; Jelby et al., 2020; Śliwińska, et al., 2020, Fallatah et al., 2024), precipitation of authigenic carbonate in reducing conditions (Rodriguez Blanco et al., 2022), or variations in the input of terrestrially derived organic matter that is in turn a manifestation of relative sea level change and climate (Capelli et al., 2021; Weger et al., 2022). However, none of these models alone can explain the contemporaneous

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enrichment of Ag in Arctic basins (Galloway et al., 2024) and alternatives must be considered. An interval of silver (Ag) enrichment occurs across the VOICE in the Sverdrup Basin (Canadian Arctic, 3-6x higher than average shale) and at Festningen (6x higher than average shale). Silver is similarly enriched in black shales of Jurassic and Cretaceous age in the Barents Sea, Norwegian Shelf, and West Siberia Basin (Lipinski et al., 2003; Zanin et al., 2016). The relationship of Ag to organic matter, S, Fe, and redox-sensitive trace elements in the strata from Canada and Festningen suggest that an extra-basinal source of Ag to seawater during the Volgian existed and that the source was enhanced hydrothermal flux in the proto-Amerasia basin during rift climax, with sufficient circulation to transport high Ag-seawater to surrounding shelves (Svalbard) and within and to extensional basins (Sverdrup Basin). Europium (Eu) values show an anomaly (Eu/Eu*>1) during the VOICE both in the Canadian strata and strata at Festningen, suggesting the presence of hydrothermal fluids. It is possible that the negative carbon isotopic signature of the VOICE is, in part, associated with the putative hydrothermal systems hypothesized to have caused Ag enrichment. Nearly all negative carbon isotope excursions in the geological record, but so far not the VOICE, are interpreted to reflect episodes of massive carbon release. Its global manifestation could therefore be the result of widespread rifting and outgassing of large quantities of CO_2 (see Brune et al., 2017) associated with the breakup of Pangea (Galloway et al., 2024).

4.4.2.2. Valanginian Weissert Event

The Weissert Event is a prominent global carbon-cycle perturbation which occurred in the Early Cretaceous (ca. 133 Ma). It is expressed in carbon isotope records (Lini et al., 1992; Price et al., 2016) and is manifested in Arctic Canada (Galloway et al., 2020) and Svalbard (Jelby et al., 2020b; Vickers et al., 2023) (Fig. 11). The isotope event consists of a globally recognized positive CIE of a significant magnitude (2–5 ‰), which is widely documented in marine carbonates, fossil shell material, terrestrial plants, and organic matter (e.g., Lini et al., 1992; Gröcke et al., 2005; Aguirre-Urreta et al., 2008; Price et al., 2016; Galloway et al., 2020; Jelby et al., 2020b; Vickers et al., 2023; Fallatah et al., 2024). Regional to global climate cooling in the Valanginian (Early Cretaceous; ca. 140–133 Ma) is well documented (e.g. Pucéat et al., 2003; Weissert and Erba, 2004; McArthur et al., 2007; Bodin et al., 2015; Meissner et al., 2015; Grasby et al., 2017b), although the timing, magnitude and extent of this cooling is debated (e.g., van de Schootbrugge et al., 2000; McArthur et al., 2007), as is the mechanism for the CIE. Paleoclimatic reconstructions using biomarkers indicate a ~3°C global surface cooling across the event (Cavalheiro et al., 2021), and glacial deposits from the Eromanga Basin (Australia) suggest the transient development of a small southern polar ice-cap (Alley et al., 2020), whereas stable oxygen isotope records show mixed signals, with rising $\delta^{18}O_{belemnite}$ values suggesting cooling in the Boreal Realm from the Late Valanginian to Early Hauterivian (Podlaha et al., 1998; McArthur et al., 2004; Price and Mutterlose, 2004; Bodin et al., 2015; Meissner et al., 2015), with little change in Tethyan records (e.g., van de Schootbrugge et al., 2000; McArthur et al., 2007).

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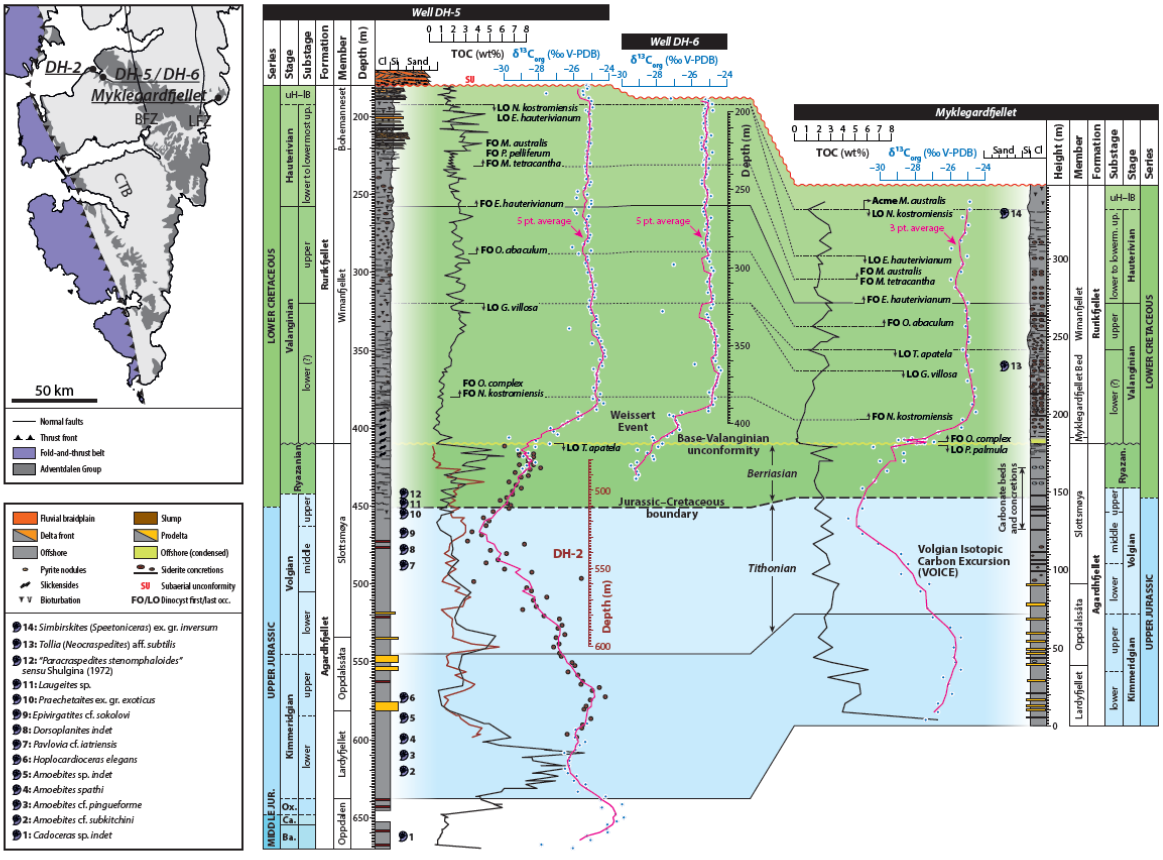

Figure 10. Correlation of organic stable carbon-isotope ($\delta^{13}C_{org}$) stratigraphy between the cored wells DH-2, DH-5 and DH-6 and the Myklegardfjellet outcrop section, calibrated by total organic carbon (TOC) trends, and dinocyst and ammonite biostratigraphy (compiled and modified from Jelby et al. 2020b; 2024, and Koevoets et al. 2016, 2018). The boundary between the Agardhfjellet and Rurikfjellet formations (conforming to a base-Valanginian unconformity and demarcated by a marked drop in TOC values) is used as a correlation datum. Note the clear expression of the Weissert Event and 'Volgian Isotopic Carbon Excursion' (VOICE) in the different sections. uH–lB, upper Hauterivian–lower Barremian. For details on the dinocyst bio-events, the reader is referred to Jelby et al. (2020b; 2024) and Śliwińska et al. (2020).

In the Arctic (including Svalbard), the widespread occurrence of glendonites around the Weissert Event interval (from the Berriasian to Hauterivian; Grasby et al., 2017b; Vickers et al., 2019a; Galloway et al., 2020; Jelby et al., 2025) has led to speculation of the cooling being decoupled from the CIE (e.g., Rogov et al., 2017), although this may partially be an artifact of age uncertainties for the Arctic successions (e.g. Vickers et al., 2019a; Jelby et al., 2020b; 2025). It may also suggest that episodic cooling punctuated the background warmth throughout the Early Cretaceous. The emplacement of the Paraná-Etendeka LIP is broadly coincident with the Weissert Event CIE (Fig.11), although uncertainties surrounding the exact relative timings have led to much debate around the causal mechanism for the event (e.g., Weissert and Erba, 2004; Duchamp-Alphonse et al., 2007; Dodd et al., 2015; Rocha et al., 2020; Gomes and Vasconcelos, 2021). Recently, it has been shown that the peak

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of Paraná-Etendeka activity coincided with the onset of the Weissert Event (Martinez et al., 2023), supporting a volcanic trigger for this CIE.

In Svalbard, the $\delta^{13}C_{org}$ record reveals an abrupt and pronounced positive excursion of up to 5.5‰ in the lower Valanginian, coincident with the base of the Rurikfjellet Formation (Fig. 11; Jelby et al., 2020b; 2025). The excursion is clearly observed in both core (DH-5 and DH-6 in Adventdalen) and outcrop (Myklegardfjellet) sections, where $\delta^{13}C_{org}$ reaches the most positive values recorded since the Callovian–Oxfordian (Koevoets et al., 2016). Glendonites are found in numerous horizons in the Festningen locality, although a patchy $\delta^{13}C_{org}$ curve and poor stratigraphic dating of this section due to local small-scale tectonism has led to uncertainty as to the relative age of the glendonites with respect to the Weissert Event (Vickers et al., 2019a; 2023; Jelby et al., 2020b; 2025). Glendonites in the Canadian successions occur in Valanginian of the Deer Bay Formation (Grasby et al., 2017b; Galloway et al., 2020).

Examination of the Hg record across this interval at both Festningen and correlative sites in the Sverdrup Basin similarly yielded uncertain results, due to the poor age constraints preventing convincing identification of the Weissert Event CIE (Vickers et al., 2023). Nonetheless, Arctic Hg/TOC ratios are observed to increase across the proposed Weissert Event intervals in both Svalbard and the Sverdrup Basin. This supports recent work indicating that Paraná-Etendeka volcanism was synchronous with the onset of the Weissert Event CIE (Gomes and Vasconcelos, 2021). In other localities on Svalbard (DH-5 borehole in Adventdalen and Myklegardfjellet in eastern Spitsbergen, Fig. 10), the Weissert Event is well dated with age-diagnostic ammonites and palynomorphs (Jelby et al., 2020b; 2025), and with the observed $\delta^{13}C_{org}$ trend being consistent with the globally recognized Weissert Event (Fig. 11; Lini et al., 1992; Weissert et al., 1998; Weissert and Erba, 2004). The onset of the Weissert Event in Svalbard appears to have occurred earlier than in other Boreal sites, but this diachroneity may reflect a depositional hiatus and/or stratal condensation in the glauconitic plastic clay of the Myklegardfjellet Bed at the base of the Rurikfjellet Formation, corresponding to the Tethyan T. pertransiens – N. neocomiensiformis ammonite zones (Jelby et al., 2020b; 2025).

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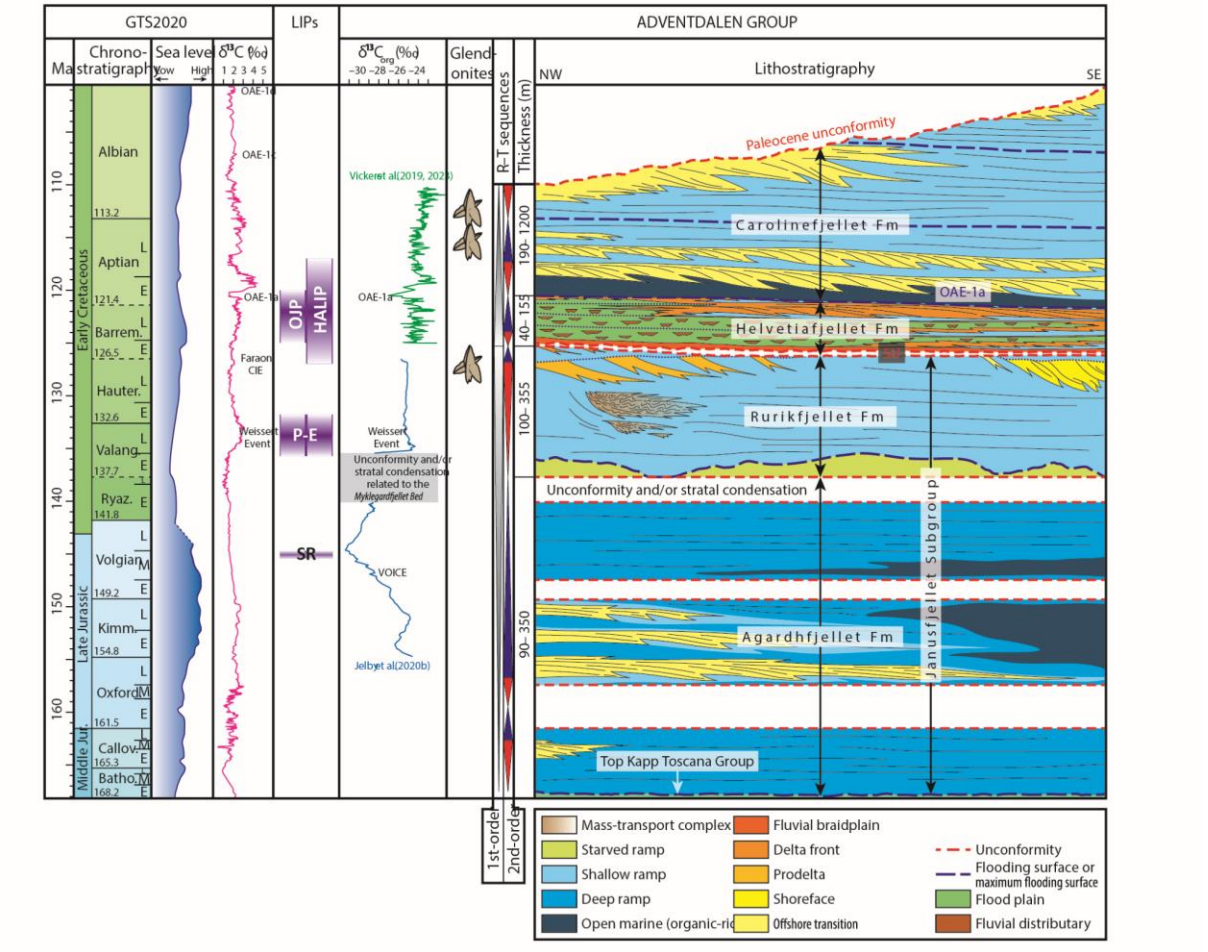

940 **Figure 11. Overview of the Upper Jurassic and Lower Cretaceous stratigraphy of Svalbard plotted against time, from Jelby et al. (2024), based on Grundvåg et al. (2017; 2019). To the left of the stratigraphic summary chart, thicknesses of the formations, first- and second-order regressive-transgressive sequences, occurrence of glendonites (Vickers et al., 2019), LIP volcanic episodes (see references in Vickers et al., 2023), and organic stable carbon isotopes (see Jelby et al., 2024 and references therein) are shown. These are plotted against global $\delta^{13}C$ trends, sea level, and ages from the Geological Time Scale 2020 (Gale et al., 2020; Hesselbo et al.,**
945 **2020).**

In summary, the Weissert Event represents an important recoupling of the carbon cycle between Boreal (including Svalbard) and lower-latitude basins (including Tethyan, Atlantic, and Pacific), probably because of re-established ocean connections in response to a global eustatic sea-level rise and changing oceanic gateway connections (Haq, 2014; Galloway et al., 2020; Jelby
950 et al., 2020b). However, Jelby et al. (2020b) recognized that the decay of the event in Svalbard spanned the late Valanginian to early Barremian (Early Cretaceous; ca. 125.77–121.4 Ma) but is negligible compared to most other Boreal and lower-latitude records, and that the signal remains relatively stable at near-peak values following the positive excursion. This indicates that

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the oceanographic reconnection between higher and lower latitudes, and thus ocean ventilation, must have been sufficiently limited to keep some Boreal basins (including Svalbard) relatively deviated from prevailing global carbon-cycle dynamics following the isotopic event due to water exchange only through narrow, shallow straits (Price and Mutterlose, 2004).

4.4.2.3. The Cretaceous HALIP and OAE1a

The largest carbon cycle perturbation of the Early Cretaceous occurred in the early Aptian (a. 121–113 Ma). It is associated with Ocean Anoxic Event 1a (OAE1a, ~120 Ma), when black shale deposition occurred across multiple marine sites, indicative of widespread ocean anoxia (e.g., Jenkyns, 1980). OAE1a is characterized by a globally-recognized sharp negative CIE followed by a twin-peaked positive "recovery" CIE (Fig. 11; Jenkyns, 1995; Menegatti et al., 1998; Ando et al., 2002; Price, 2003; Weissert and Erba, 2004; Herrle et al., 2015; Dummann et al., 2021), which has also been recognized in multiple localities across Svalbard and in the Canadian Arctic (Herrle et al., 2015; Midtkandal et al., 2016; Vickers et al., 2016; 2019a; 2023; Grundvåg et al., 2019; Dummann et al., 2021). Significant perturbations in global climate are believed to have occurred along with the CIE, with global warming followed by cooling being evident from multiple proxies in sites from across the globe (e.g., Bottini et al., 2015; Bodin et al., 2015; Harper et al., 2021; Galloway et al., 2022).

In Svalbard and other Arctic localities, the occurrence of numerous glendonite horizons immediately after the CIE (inAptian - Albian strata) supports the global extent of the post-OAE1a cooling (Fig. 11; Schröder-Adams et al., 2014; Herrle et al., 2015; Grasby et al., 2017b; Rogov et al., 2017; Vickers et al., 2019a). The OAE1a is believed to be linked to LIP volcanism, as both the Greater Ontong-Java Plateau (OJP) in the Pacific Ocean and the HALIP were being emplaced approximately synchronously with the onset of the event (Fig.11; Midtkandal et al., 2016; Percival et al., 2021; Galloway et al., 2022). However, uncertainties surrounding the exact relative timings of the volcanism and the OAE1a complicate the interpretation of the cause of the event (Tarduno et al., 1991; Mahoney et al., 1993; Parkinson et al., 2002; Tejada et al., 2002; 2009; Erba et al., 2004; 2015; Chambers et al., 2004; Thordarson, 2004; Dockmann et al., 2018; Kasbohm et al., 2021; Galloway et al., 2022).

Attempts to use Hg as a proxy for volcanism to resolve the question of relative timing of HALIP and OJP with regards to the OAE1a CIE and accompanying climatic/environmental perturbations have yielded ambiguous results (Percival et al., 2021; Vickers et al., 2023), and recent work using palynology suggests HALIP-related landscape disturbances began to occur in the latest Barremian, coincident with the first pulse of the HALIP, but prior to the early Aptian onset of OAE1a (Galloway et al., 2022). The onset of the negative $\delta^{13}C$ excursion across Svalbard occurs within a sapropel-rich interval which is highly impoverished in marine palynomorphs (dinocysts), but which yield a number of reworked taxa (e.g., in the Ullaberget outcrop section, the DH2, and the DH1 cores near Longyearbyen (Midtkandal et al., 2016; Grundvåg et al., 2019; Śliwińska et al., 2020), narrowing the age to the Barremian-Aptian transitional interval. The regional flooding event that is considered to correlate globally with the OAE1a yields well preserved dinocysts and has been assigned an earliest Aptian age (Midtkandal et al.,2016; Grundvåg et al., 2017; 2019). Younger Cretaceous strata that elsewhere record other OAEs (e.g., OAE1b, OAE 2, OAE3) are absent from Svalbard (Fig. 4).

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4.5. Cenozoic

The Paleogene succession in Svalbard includes Paleocene to possibly mid-Oligocene age strata (61.8 to ~30 Ma), and thus was deposited during the global climate transition from greenhouse to coolhouse conditions (Zachos et al., 2001; Westerhold et al., 2020). Since the Late Cretaceous, Svalbard was already located north of the Arctic Circle (e.g., Harland, 1997) at paleolatitudes comparable to the present (Fig. 3). Therefore, the Cenozoic geological record in Spitsbergen provides a natural reference interval for future polar amplification to global warming in a warm Arctic. The deposits of the CSB belong to the Van Mijenfjorden Group with thicknesses up to 2200 m. The basin fill begins with the Paleocene coal-bearing successions that demonstrates a transgressive trend from delta plain to prodelta/outer shelf facies, followed by marine mudstones and intensely bioturbated sandstones that mainly were sourced from north and northeast of Svalbard (Petersen et al., 2016; Lüthje et al., 2020; Jochmann et al., 2020). The first evidence of WSFTB-derived sediments occurs in the latest Paleocene with a westerly-derived clastic wedge known as the Hollendardalen Formation. The upper part of the CSB continued to fill into the Eocene and possibly the Oligocene and consists of >800 m thick deposits of shelf and shelf-edge deltas, slope clinotherms, and basin floor fans sourced from the WSFTB to the west (Steel et al., 1981; Johannessen and Steel, 2005; Helland-Hansen and Grundvåg, 2021). Maximum subsidence occurred during the deposition of the Frysjaodden Formation, which contains a significantly expanded (>30 m) Paleocene-Eocene Thermal Maximum (PETM) sequence (Cui, 2010; Charles et al., 2011; Dypvik et al., 2011). The shallowing-upward uppermost CSB fill is seen as a transition from deepwater marine via shallow marine/delta front to coastal plain and continental strata. The Norwegian-Greenland Seaway became severely restricted at this time, isolating the Arctic from the Atlantic Ocean during the PETM and early Eocene (Blakey, 2021; Hovikoski et al., 2021; Jones et al., 2023). Overall, the Paleogene coal-bearing successions developed in several stratigraphic levels of Svalbard are an excellent archive for reconstruction of the past vegetation and climate. These intervals provide an excellent insight into the fauna/vegetation of Svalbard at these times (see section c below and 6.3).

4.5.1. Paleocene hothouse from an Arctic Circle perspective

The Paleocene was characterized by hothouse climate (Zachos et al., 2001; Westerholt et al., 2020). In Svalbard, the CSB Paleocene succession consists of the terrestrial to nearshore Firkanten Formation (yielding fossil fauna) and the offshore marine Basilika, Grumantbyen and lowermost Frysjaodden formations. The Firkanten Formation of early Paleocene age (Selandian) contains fossil plant-bearing units also known as the Barentsburg flora (see Golovneva et al. 2023). These deposits have been extensively investigated due to economic interest and exploitation of coal resources over the last century, yet there are relatively few publications describing the succession in detail (cf. Steel et al., 1981; Nøttvedt, 1985; Nagy, 2005; Jochmann et al., 2020; Lüthje et al., 2020). Coal exploration over the past few decades has provided more than 500 drill cores through the early Paleogene of Svalbard and focuses in particular on the coal-bearing Todalen Member of the Firkanten Formation. Fission-track ages from the Firkanten Formation dated the unit to 63 ± 2 Ma and 64 ± 2 Ma (Blythe and Kleinspehn, 1998). A more

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precise age is derived from an ash layer that cross-cuts the lowermost coal seam in the Firkanten Formation. It has been dated using U-Pb methods to 61.596 ± 0.028 Ma Ma (Jones et al., 2017) at the Danian–Selandian boundary, constraining the main coal deposition to the early Selandian. The Paleocene coals have recently been shown to comprise much higher resolution stratigraphic records than previously anticipated (Large and Marshall, 2015; Large et al., 2021), especially when coals are formed from peat growing in colder climates. The age model developed by Large and Marshall (2015) and an improved understanding of long-term storage of peatland carbon (Large et al., 2021) can be implemented to assess accumulation rates. For example, the 1.5 m-thick Longyear coal seam has a modeled accumulation rate of 59 kyr and 99 kyr for temperate and boreal climates, respectively, which suggests a high-resolution record that can be used to infer variation in atmospheric dust input and the effects of forest fires (Marshall, 2013). Material has been collected from the now closed coal mines in ongoing research projects to expand on these existing datasets. The Selandian coal seams provide details of the vegetation on Svalbard at that time. The paleoflora from Svalbard reveal a temperate (mean annual temperature 10.1 ±2°C), maritime, humid climate, with warm summers and cool mild winters (e.g., Ulf et al. 2007; Golovneva et al. 2023). Temperate Arctic paleotemperatures are corroborated by Pantodont tracks discovered at the upper boundary of the coal layer of the Firkanten Formation (Lüthje et al. 2010). The climatic conditions were similar in the larger Arctic region (e.g., O'Regan et al., 2011).

4.5.2. The Paleocene-Eocene Thermal Maximum (PETM)

The PETM was a transient period (~150-200 kyr) of rapid global warming that began around 56 Ma, superimposed on already greenhouse conditions of the early Paleogene (Zachos et al., 2001). The event is recognized as a global negative CIE in sections worldwide, attributed to a massive release of ^{13}C-depleted carbon to the ocean-atmosphere system (McInerney and Wing, 2011). Potential sources include surface reservoirs such as dissociation of methane hydrates from marine sediments (e.g., Dickens et al., 1995) and/or volcanic and thermogenic degassing from the emplacement of the NAIP (e.g., Svensen et al., 2004; Gutjahr et al., 2017; Berndt et al., 2023; Jones et al., 2023). Existing High Arctic records for the PETM and following hyperthermal events suggest subtropical temperatures in both marine and terrestrial realms (e.g., Suc et al., 2020; Sluijs et al., 2000). In Svalbard, the PETM and its various environmental effects has been thoroughly documented in several chemostratigraphic to biostratigraphic multi-proxy studies (Charles et al., 2011; Dypvik et al., 2011; Harding et al., 2011; Nagy et al., 2013; Wieczorek et al., 2013; Jones et al., 2019; Cui et al., 2021; Pogge von Strandmann et al., 2021). The bulk of these studies are focused on a SNSK drill core (BH09/05) from a coal-exploration well located on the eastern flank of the CSB (Fig. 2).

The timing and duration of the PETM is well constrained from Svalbard strata with a high-precision U-Pb radiometric age of an ash layer within the CIE outcropping near Longyearbyen, couple with evidence of orbital cycles within the strata of the BH09/05 core (Charles et al., 2011). Other studies explore the myriad consequences of extreme warming in Svalbard strata. The abundance of kaolinite in the PETM interval (Dypvik et al., 2011) and a large negative lithium isotope excursion coincident with the CIE (Pogge von Strandmann et al., 2021) suggests increased weathering rates in response to warmer and wetter

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conditions. Increased runoff rates resulted in a stratified water column with a freshwater surface layer and oxygen depleted bottom waters, which severely reduced the diversity of various fauna elements in the basin (Harding et al., 2011; Dypvik et al., 2011; Nagy et al., 2013). Osmium isotopes (Wieczorek et al., 2013) and mercury anomalies (Jones et al., 2019) show key variations in the activity of the NAIP at this time, supporting the hypothesis that NAIP volcanism and magmatism was at least partially responsible for the extreme carbon emissions at the PETM onset. While studies on the BH09/05 core are now plentiful, there are numerous other cores and outcrops that contain the PETM strata that have received little to no attention.

4.5.3. The decline of the Eocene greenhouse climate

The hothouse conditions of the early Eocene were followed by a gradual global cooling that culminated in the transition into the icehouse climate at around 34 Ma (e.g., Westerhold et al., 2020; Hutchinson et al., 2021). Paleobotanical records from the Arctic suggest that during the middle Eocene, the mean annual precipitation was >120 cm/yr (Greenwood et al., 2010). In Svalbard, the continental, flora-bearing units are found within the Aspelintoppen Formation (Steel et al. 1978; 1985), probably spanning from the Middle Eocene and not younger than Late Eocene or Early Oligocene (Matthiessen, 1986; Cepek and Krutzsch, 2001), and the Upper Eocene (Golovneva and Zolina, 2023) or Early Oligocene (Head, 1984) Renardodden Formation. Abundant paleoflora from the Aspelintoppen Formation and the Renardodden Formation (Manum,1962, Kvacek and Manum, 1993; Kvacek et al., 1994 and Cepek and Krutzsch, 2001; Uhl et al., 2007) indicates mean annual precipitation rates of 1423 and 1716 mm/yr, respectively (Golovneva, 2000). Angiosperm morphotypes indicate a strong seasonal precipitation pattern (from 356 to 656 mm in the three wettest months and 112 -247 mm for the three driest months; Clifton, 2012). The high rates of precipitation and increased weathering rates, in combination with active tectonism, promoted the transfer of sand into deeper settings via flood-generated hyperpycnal flows at this time (Grundvåg et al., 2023b). In the late Eocene, Svalbard was only a few degrees south of the present 78°N (Fig. 3), but fossil plant material indicates that the temperature at that time was much warmer than today, and the estimated mean annual air temperature was around +9 °C (Golovneva, 2000; Goloneva et al. 2023; Uhl et al., 2007). Several studies have evaluated the abundance and diversity of fossil plants, the occurrence of coal seams, as well as fossil insects in the terrestrial parts of the Eocene Aspelintoppen Formation (e.g. Dallmann et al., 1999; Uhl et al., 2007; Marshal et al., 2015). In other areas of the High Arctic, such as Ellesmere Island, fossil remnants of a varanid lizard, the tortoise Geochelone, and the alligator Allognathosuchus confirm warm temperatures that remained above freezing (e.g., Estes and Hutchinson, 1980; Eberle and Greenwood, 2012).

The sporadic occurrence of glendonites and outsized clasts, the latter possibly indicating rafting by temporal sea ice, in the marine parts of the succession suggests strong seasonal or temporal temperature variations in Svalbard (Kellogg, 1975; Dalland, 1977; Spielhagen and Tripati, 2009). This is in accordance with some of the paleo-floristic/insect studies that infer freezing temperatures during winter months, and an overall cooling trend for the entire interval (Golovneva, 2000; Uhl et al., 2007; Wappler and Denk, 2011). Some of the signals preserved in the sedimentary rock record may be caused by other allogenic forcing factors than climate fluctuations, such as tectonics and relative sea-level changes, but similar results from a

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range of proxies including plant morphotypes support the validity of paleoclimate reconstructions. As such, deconvolving climatic and tectonic signals in tectonically active basins is of major importance. In the restricted Arctic Ocean during the early middle Eocene (ca. 49 Ma), increased runoff caused stratification of the water column (with a fresh-water lid) and led to the well-known Azolla freshwater algal blooms that eventually contributed to the withdrawing of the atmospheric CO_2 and cooling of the global climate from the middle Eocene onwards (Brinkhuis et al., 2006; Speelman et al., 2009).

4.5.4. The Eocene-Oligocene Transition and connection with the Arctic Ocean

A major step in the long-term Cenozoic climate evolution took place at the Eocene Oligocene Transition (EOT, ~34 Ma), when decreasing of atmospheric CO_2, and changes to ocean gateways led to a development of the first permanent ice-cap in Antarctica and initiated the icehouse type of climate which exists until today (Straume et al., 2020; Westerhold et al., 2020; Hutchinson et al., 2021). However, the global scale of the transition is not fully understood, since in contrast to the well-studied deep sea sites from southern and equatorial regions, the signature of the EOT for the northern high latitudes remains poorly constrained. Climate models suggests that closing and opening the gateways to the Arctic Ocean (such as the Fram Strait) had equally large impact on the temperature development in the high northern latitudes as the CO_2 decrease (e.g., Hutchinson et al., 2019; 2021; Straume et al., 2022; Śliwińska et al., 2023). However, the number of proxy records from the northern polar regions to evaluate the history of the Fram Strait and validate the climate models is limited. The ACEX core (IODP Expedition 302) from the Arctic Ocean contains a hiatus that misses an estimated interval from 44.4 Ma to 18.2 Ma (Backman et al., 2008). The ODP site 913 from the Greenlandic Sea suffers from a hiatus at the EOT (Eldrett et al., 2004). Furthermore, the existing sea surface temperature proxy data are of extremely low resolution (Liu et al., 2009) in comparison with time-equivalent records from the Labrador Sea and the North Sea (Śliwińska et al., 2019; 2023). Molecular fossil (alkenone) records suggest at least 5°C cooling in the northern high latitudes, associated with the transition towards the coolhouse climate (Liu et al., 2009; Śliwińska et al., 2023). The existing pollen record revealed a significant cooling of ca. 5°C in cold months mean temperatures on East Greenland across the EOT (Eldrett et al., 2009).

During the EOT, Svalbard was already located at ~80°N and therefore provides unique insights into the climate evolution across the EOT in the northern high latitudes. In the CSB the youngest Paleogene unit is the Aspelintoppen Formation, which is assigned to the Late Eocene (plant fragments) or Oligocene (mollusks) (Manum and Throndsen, 1986). Unfortunately, the age model for this formation remains poorly constrained. With an improved age model, the Eocene–Oligocene succession could provide a valuable contribution to the atmospheric temperature evolution across the EOT in the northern high latitudes and be further resolved with pollen records, comparable with the Norwegian-Greenland Sea (Eldrett et al., 2009). Marine to terrestrial deposits of possible late Eocene to Oligocene age have also been reported mainly from the exposed parts of the Forlandsundet and Bellsund grabens on the West Spitsbergen margin (Gabrielsen et al., 1992; Weber, 2019; Śliwińska and Head, 2020; Schaaf et al., 2021) . However, these studies are very limited. The foraminifera assemblages collected from the Sarstangen conglomerate at the Balanuspynten profile on the eastern side of the Forlandsundet Graben reveal the presence of

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marine Oligocene strata assigned to the Buchananisen Group (Feyling-Hanssen and Ulleberg, 1984). This age has later been substantiated by palynostratigraphic analyses, which suggests an early to middle Oligocene age, at least for the sediments exposed along the eastern basin margin (Schaaf et al., 2021). The two foraminiferal zones (TA and TB) that were originally assigned to the middle to upper Oligocene, can more accurately can be assigned to the lower Oligocene (lower Rupelian; the TA zone) and the upper Oligocene (lowery Chattian; the TB zone). The early Oligocene age of the marine strata is confirmed by the presence of dinocyst *Svalbardella cooksoniae* (Manum, 1960). The appearance interval of *S. cooksoniae* in the earliest Oligocene seems to be associated with a cooling interval (Śliwińska and Heilmann-Clausen, 2011). A single sample from the Calypsostranda Group at the Renardodden section on the southern shore of Bellsund, a structural outlier interpreted to be an exposed part of the Bellsund Graben, has yielded dinocysts of late Eocene or early Oligocene age (Head, 1984; Śliwińska and Head, 2020). Based on the association of pollen in the Skilvika Formation, an upper Paleocene to Eocene age can be suggested for the lower part of the section (Weber, 2019).

4.5.5 Neogene hiatus

Svalbard experienced two uplift phases in recent times. The first and major uplift phase started in the Eocene (>36 Ma) and persisted to ca. 10 Ma. This was followed by less prominent uplift from ca. 10 Ma onwards that generated the modern topography of the archipelago (Dörr et al., 2013). These uplift events are matched by contemporaneous uplift phases in Greenland, the Barents Shelf, and Baltica (Dörr et al., 2013) and are attributed to crustal thinning and the onset of ocean spreading in the Arctic and North Atlantic driven by mantle processes related to anomalously hot mantle underlying this part of the Arctic (Green and Duddy, 2010). The presence of thick pre-glacial (Miocene and Pliocene? 23 to 2.58 Ma) and glacial (late Pliocene and Pleistocene) offshore clastic wedges along the western and northern margins of Spitsbergen (Hjelstuen et al., 1996; Lasabuda et al., 2018; Alexandropoulou et al., 2021) suggests a net denudation of ca. 3 km (Riis and Fjeldskaar, 1992; Lasabuda et al., 2021). The overall amplitude of Neogene uplift decreases eastwards with the uplift along parts of WSFTB exceeding 2.5 km and over 1.5 km in the CSB (Dörr et al., 2013). Estimates based on organic geochemical proxies suggest total uplift of 2.5 to 3.5 km (Throndsen, 1982; Marshall et al., 2015; Olaussen et al., 2019). As a result, Miocene and Pliocene sedimentary strata are not preserved anywhere on Svalbard, while only intermittent remains of Pleistocene glacial deposits are found (Ingólfsson and Landvik, 2013). Uplift of 9 mm/yr continues in central-western Spitsbergen today, with only 1 mm/yr of that attributed to isostatic rebound due to the recent Weichselian glaciation (Kierulf et al., 2022).

5. Absolute radiochronology of Svalbard stratigraphy

Despite the remarkable continuous stratigraphic successions spanning much of the Phanerozoic in Svalbard, robust absolute radiometric stratigraphic age constraints are scarce. Except for data on Devonian and older magmatic and metamorphic rocks from the pre-Caledonian “basement” (e.g., Myhre et al., 2008; Petterson et al., 2009; Majka and Kosminska, 2017; McClelland

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et al., 2019), no robust ages are published from the stratigraphy pre-dating the Permian-Triassic boundary. The first radiometric age comes from an ash layer within the Deltadalen section at the Permian-Triassic boundary (Fig. 10). A zircon U-Pb chemical abrasion isotope dilution thermal ionization mass spectrometry (CA-ID-TIMS) age of 252.13 ±0.62 Ma from a bentonite bed (volcanic ash) ca. 15 cm above the first appearance datum (FAD) of the age diagnostic *H. parvus* (onset Triassic) in the Vikinghøgda Fm. ties the biostratigraphic record to an absolute age. This age records the onset of the Triassic in Svalbard and the Panthalassic Ocean within error (Zuchuat et al., 2020). Given the sedimentation rate constraints for this section (Zuchuat et al., 2020), and the uncertainty in the age of the bentonite, there is overlap in age between the FAD of *H. parvus* in Svalbard and the FAD of *H. parvus* of 251.902 ±0.024 Ma at the Induan GSSP (Burgess et al., 2014). This indicates synchronicity of the end-Permian mass extinction in the Panthalassic and Tethyan domains at a 0.2 % (2σ) level of uncertainty.

The next absolute stratigraphic tie point occurs in the Barremian to lower Aptian Helvetiafjellet Formation, where a bentonite layer from two cores taken in Longyearbyen (DH-3 and DH-7; Fig. 3) was dated to 123.1 ± 0.3 Ma (zircon U-Pb CA-ID-TIMS; Corfu et al., 2013; Midtkandal et al., 2016). Bio- and chemo-stratigraphical evidence suggests that this bentonite layer is of mid-Barremian age (Midtkandal et al., 2016), and that it occurs ~40 m below the Barremian–Aptian boundary and the onset of the Early Aptian Oceanic Anoxic event 1a (OAE1a). Subsequent magnetostratigraphy tied this bentonite age to the magnetic polarity record and hence Zhang et al. (2021) were able to calculate an age of 121.2 ±0.4 Ma for the Barremian – Aptian boundary, accepting the M0r magnetochron as a boundary marker. Based on available ages for the HALIP in Svalbard (124.7 ±0.3 Ma to 123.9 ±0.3 Ma; Corfu et al., 2013) and Franz Jozef Land (~122-123 Ma; Corfu et al., 2013), there is no overlap in age with a HALIP magmatic pulse of this event with the OAE1a. However, mafic lavas and intrusives from the Sverdrup basin in northern Canada do show overlapping ages (Evenchick et al., 2015; Dockman et al., 2018) with the updated Barremian – Aptian boundary age not precluding a relationship between a pulse of the HALIP with the EAO1a.

To our knowledge, there are no further absolute age constraints published for the Mesozoic stratigraphy of Svalbard. However, the onset of Paleogene sedimentation in the CSB (Fig. 4) and the Paleocene to early Eocene stratigraphy is well constrained through high precision zircon CA-ID-TIMS U-Pb ages (Charles et al., 2011; Jones et al., 2017; Jochmann et al., 2020). A bentonite bed towards the base of the Firkanten Formation (see diagram in Fig. 12), dated from three different parts of the basin, yielded an age of 61.596 ±0.028 Ma overlapping with the Danian–Selandian boundary, and a bentonite from the lower part of the overlying Basilika Formation has an age of 59.32 ±0.19 Ma overlapping with the Selandian–Thanetian boundary (Jones et al., 2017). Based on these marker horizons, deposition within the CSB was estimated to start around 61.76 ±0.09 Ma. A bentonite horizon within the Frysjaodden Formation was dated to 55.785 ±0.034 Ma by Charles et al. (2011). This ash layer is a key marker bed for constraining the Paleocene–Eocene boundary as it is found within the PETM CIE in Svalbard strata. Charles et al. (2011) used this age and possible precession cycles within borehole BH09/05 to estimate an age of 55.866 ±0.098 Ma for the Paleocene–Eocene boundary (i.e., the PETM onset).

To date, there are no further robust and precise radiometric ages from Svalbard's Paleozoic strata. However, well-studied outcrop successions and abundant drill core material from much of the Paleozoic offers an excellent possibility to search for

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target volcanic rocks in the stratigraphy that may be dated by high precision zircon U-Pb methods, thus improving the chronostratigraphy not only of Svalbard but potentially also globally (e.g., Zhang et al., 2021).

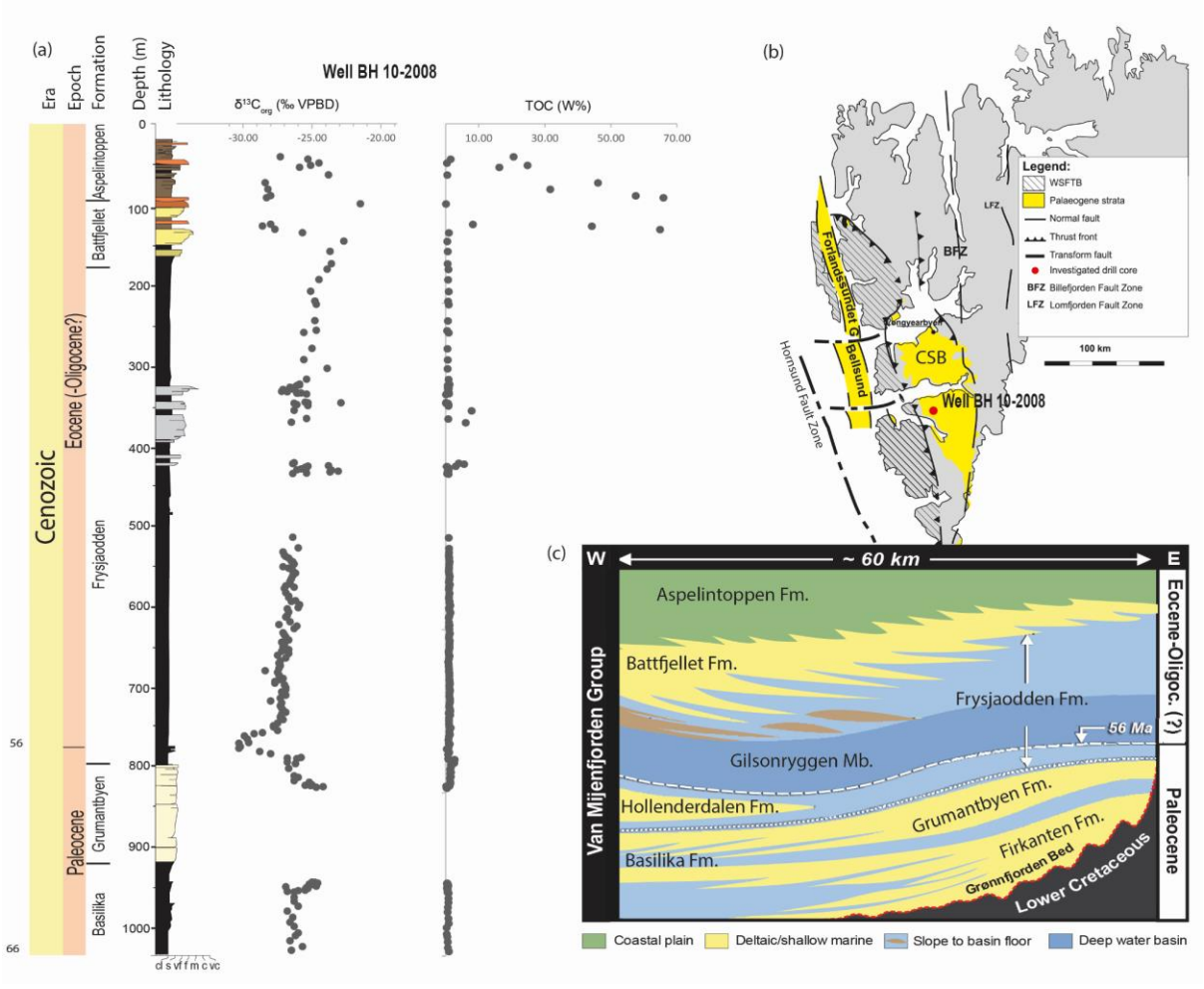

Figure 12. (a) Geochemical data from well BH 10-2008 is from Doerner et al., (2020), and borehole stratigraphy after Grundvåg et al. (2014). (b) geological map showing the outline of Paleocene deposits and position of BH 10-2008 from Helland-Hansen & Grundvåg (2021). (c) Litostratigraphic summary of the Paleogene deposits of the Central Spitsbergen Basin. Modified from Helland-Hansen & Grundvåg (2021). The age of the Paleocene-Eocene boundary from Charles et al. (2011) and Harding et al. (2011).

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6. Evolution of the Phanerozoic climate in Svalbard

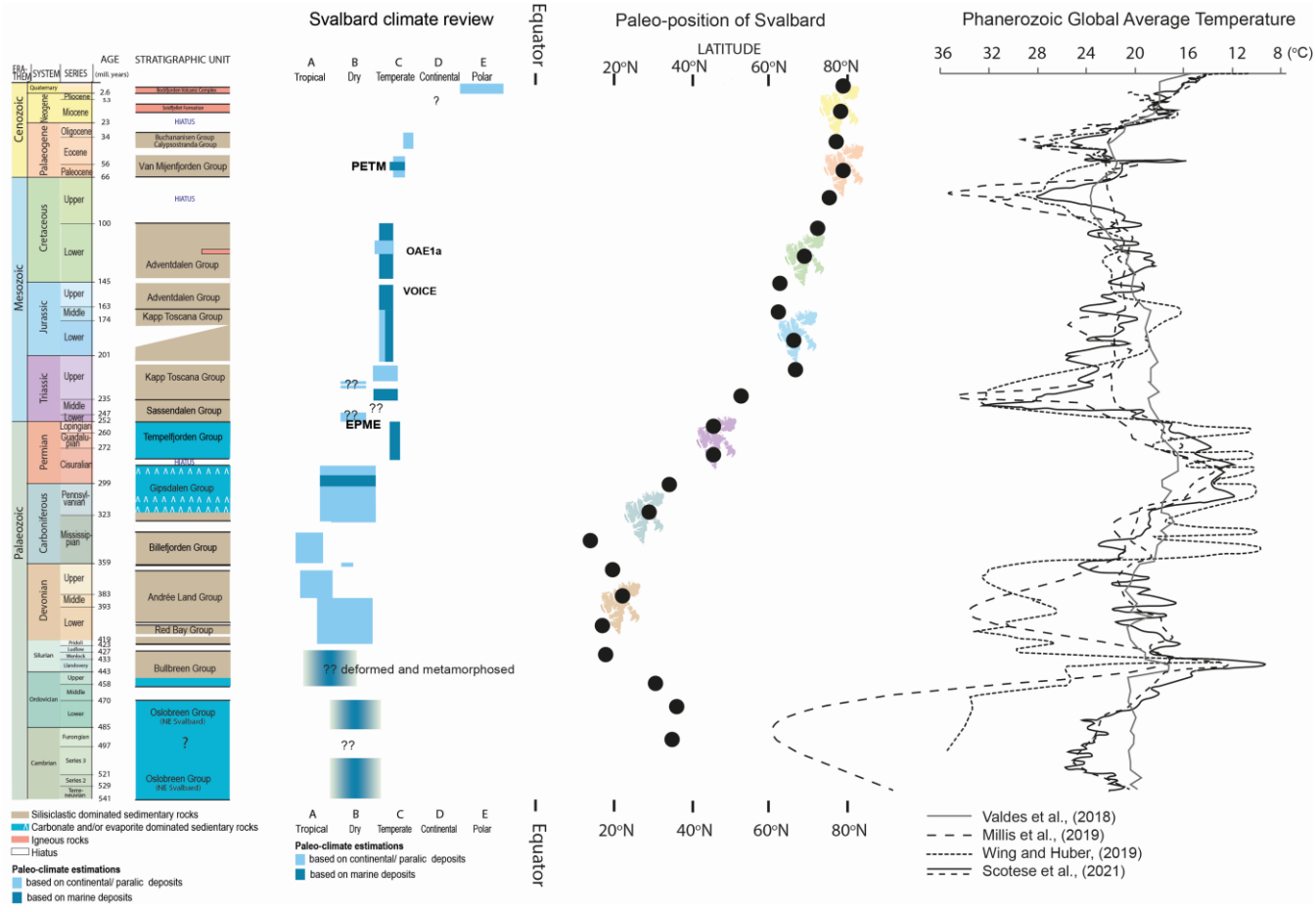

Figure 13. Correlation of the Phanerozoic climate plot based mainly on climate-sensitive facies (See Supplementary material), supplemented by biological and geochemical proxies with the paleogeographic position of Svalbard (Scotese, and Wright, 2018) and compilation of the published global average temperature curves.

The parameters controlling the Phanerozoic climate in Svalbard can be simplified to the two core elements: global distribution of paleo-climatic zones (Köppen belts; e.g., Boucot et al., 2013) and the paleo-latitude of Svalbard (e.g., Steel and Worsley, 1984; Torsvik and Cocks, 2017, Olaussen et al., 2025). The deep-time models of Phanerozoic climate either for low latitude regions between 40°N and 40°S (Fig. 13; based on oxygen isotope record (Song et al., 2019; Verard and Veizer, 2019; Veizer and Prokoph, 2015; Grossman, 2012; Royer et al., 2004) or Global Average Temperature (GAT; Wing and Huber 2019, Valdes et al. 2018, Mills et al. 2019), are characterized by a "Double Hump" pattern (Fig. 13; Fischer, 1981; 1982; 1984; Frakes et al., 1992, Scotese et al., 1999; Summerhayes, 2015). The GAT trends show high temperatures during the early Paleozoic, cooler temperatures during the late Paleozoic, followed by warmer Mesozoic and early Cenozoic temperatures, finally returning to cooler temperatures in the late Cenozoic (Fig. 13). This pattern is formed in response to breakup and accretion of

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supercontinents (Nance et al., 2014; van der Meer et al., 2014; 2017). The geochemical proxies, such as oxygen isotopes, have limited application in Svalbard due to burial, diagenetic, and hydrothermal alterations (e.g., Buggies, 2013; Matysik et al., 2017). Therefore, the overall Phanerozoic climate trends illustrated in Figure 14 are based on climate sensitive lithofacies supplemented by biological and geochemical proxies where possible. The data sets used for all deep-time paleo-climatic and paleo-environmental studies in Svalbard are coming mainly for the outcrop's investigations (See Fig. 14). Only about 18% of the studies mainly of the Mesozoic and Cenozoic strata is addressing borehole data (Fig. 14).

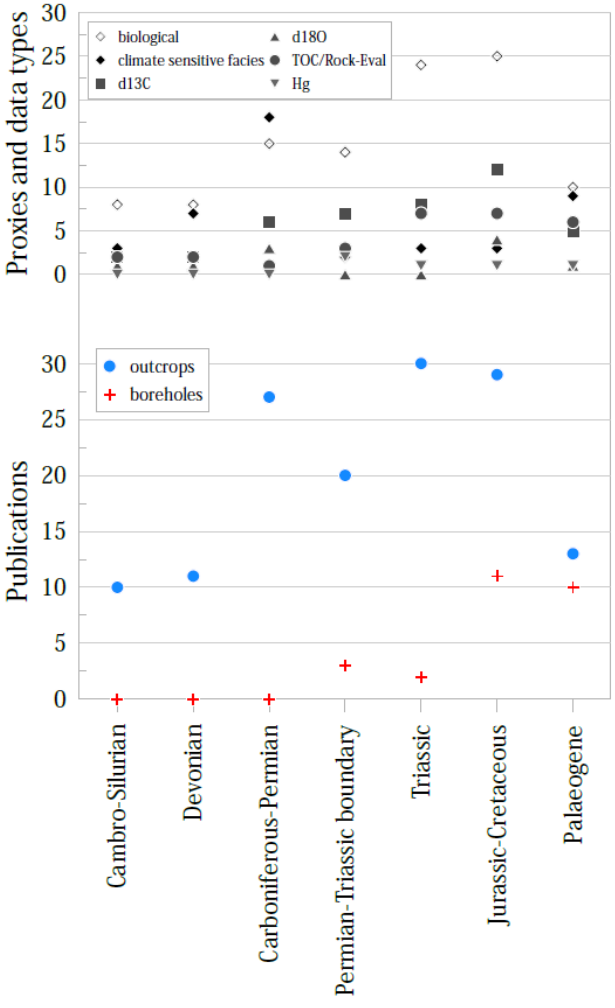

Figure 14. Compilation plot of main proxies as a function of geological time, based on Table 2.

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6. 1. Paleozoic

The warm and tropical to subtropical climate recorded in the early Paleozoic sedimentary succession in Svalbard reflects its near equatorial position (Figure 3). The Paleozoic climate slightly oscillates between the tropical and subtropical, dry climate. While the Late Ordovician cooling (ca. 460–440 Ma) is not recorded in Svalbard, likely due to a stratigraphic gap, the highest temperatures were reached in the Mississippian (Fig. 13). The northward drift of the continental plate on which Svalbard is located accelerated in the Mississippian to the late Triassic (Fig. 13; Torsvik and Cocks, 2019), while its impact on the climate can be recognized in the Carboniferous-Permian succession. The Mississippian to Pennsylvanian shift in climate seen as a transition of climate sensitive facies form the Mississippian coal-bearing deposits indicating tropical humid climate, through the Pennsylvanian semi-arid subtropical evaporites and siliciclastic red beds, and warm-water carbonate facies (chapter 4) is potentially accommodated by the changes in paleolatitude (Steel and Worsley, 1984; Torsvik and Cocks, 2019), however, the impact of global cooling related to the LPIA (e.g. Isbell et al., 2008) should be also considered. The following cooling trend expressed by the shift from cool-water to cold-water carbonate platform deposits (chapter 4) takes place despite the overall increased trend of global temperature in the Permian (Fig. 13).

6. 2. Mesozoic

Despite the general northward migration through different climate zones, Svalbard's sedimentary Mesozoic strata recorded a more complex story reflecting both global and regional climatic and environmental trends and the control from the paleolatitude position of Svalbard, which is evident in the Paleozoic, is not clearly seen. Indeed, the Mesozoic is characterized at the global scale by overall warm conditions, punctuated by several established hyperthermal events and potential ones, such as, the EPME, the late Dienerian biotic crisis, the Smithian-Spathian transition, and the Carnian Pluvial Episode. Most of these hyperthermals are associated with the emplacement of LIPs or smaller-scale volcanic activities (see chapter 4 for details). The emplacement of one LIP, however, might have triggered a global cooling rather than a global warming episode (the Weissert Event; Martinez et al., 2023). In addition to these individual events, longer-lived climate perturbations and oscillations are recorded in the Mesozoic strata of Svalbard, including periodic cooling-warming episodes during the Middle Jurassic to Early Cretaceous, as testified by the presence of cool-climate-indicators such as glendonite crystals in certain stratigraphic intervals. These climatic cycles generated (glacio-?) eustatic sea-level variations, leading to changes in global oceanic circulation as shallow seaways were periodically exposed during sea-level lowstands. These climatic and (glacio-?) eustatic sea-level variations impacted the amount and the redistribution of precipitation, runoff, temperature, salinity, water-mass stratification, nutrients and productivity between basins, as well as notably impacting the global carbon- and phosphorus cycles. One of these periodic perturbations recorded in Svalbard was the VOICE event.

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6.3. Cenozoic

The Late Cretaceous climate maximum along with the Neogene cooling (Fig. 13) are not recorded in Svalbard due to stratigraphic gaps. During the Cenozoic, Svalbard was positioned at the high northern latitudes (the Arctic Circle). Under the greenhouse conditions of the Paleocene and Eocene, paleoflora of Svalbard suggest a humid and temperate climate at that time, punctuated by hyperthermal conditions during the PETM (Chapter 4). Therefore, palynoflora from Svalbard provides a unique insight into the high latitude end member for estimating the latitudinal gradient under the greenhouse conditions of the early Cenozoic. Notably, the paleoflora suggest slightly warmer temperatures and higher mean annual precipitation during the Paleocene, than in the Eocene-earliest Oligocene. This may be an effect of the decline of the greenhouse climate. During the early Oligocene, Svalbard (located at approximately 80°N) experienced significant cooling: the presence of specific dinocysts suggests a notable cooling interval during this period. The cooling trend is a direct response to the global cooling and the transition to the modern icehouse climate state with a bipolar glaciation (Fig. 13).

Conclusions

In this contribution we synthesized the review of the Pre-Quaternary Phanerozoic (ca. 541 Ma to 2.588 Ma) deep-time paleoclimatic research conducted on Svalbard's sedimentary succession and conclude that:

- Svalbard represents an excellent location for studying multiple globally relevant paleoclimatic events within a spatially constrained area.
- Svalbard's geological record is influenced both by its northward drift and the recurring influence of large igneous provinces that affected its climate. These include, at least, the end-Permian Siberian Traps LIP, the Early Cretaceous High Arctic LIP and the Paleogene North Atlantic Igneous Province.
- Specific events that record environmental perturbations at the global scale imprinted on Svalbard's geological record include the LPIA, the End-Permian Mass Extinction, Early Cretaceous anoxic events and cold snaps, and the Paleocene-Eocene Thermal Maximum.
- These are recorded as changes in biological, lithological, and chemical proxies of past climates preserved in both outcrops and drill cores.
- Absolute radiometric ages constrain the continuous stratigraphic successions but are unevenly distributed throughout the stratigraphy.
- Of the 148 key publications, the most used proxies to quantify past environments and climate are sedimentological studies of biological indicators (104), climate sensitive facies (45), carbon isotopes $\delta^{13}C$ (42), oxygen isotopes $\delta^{18}O$ (10), and mercury (5).

This contribution serves as an important foundation for future deep-time paleoclimate studies utilizing outcrops, opportunistic drill cores or dedicated deep-time paleoclimate scientific drilling planned in the near future.

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Supplementary material

The supplementary material presents a comprehensive table summarizing 148 scientific articles. The table outlines the diverse proxies utilized in deep-time climate studies conducted in Svalbard, highlighting their respective methodologies, findings, stratigraphic succession and implications for paleoenvironmental reconstructions. It can be accessed here: https://doi.org/10.5281/zenodo.14334261.

Data availability

All data sets are coming from published scientific contributions or are part of the manuscript, including The supplement presents a comprehensive table summarizing 148 scientific articles

Authors contributions

AASS- Data Curation, Conceptualization, writing – Original Draft and Editing, visualization, Project administration
PB - Writing-Review and Editing, Visualization
VSE - Writing-Original Draft, Writing-Review and Editing
WJF - Writing-Original Draft, Writing-Review and Editing
SAG - Writing-Original Draft, Writing-Review and Editing
MEJ - Writing-Original Draft, Writing-Review and Editing, Visualization
MTJ - Writing-Original Draft, Writing-Review and Editing
GES - Writing-Original Draft, Writing-Review and Editing, Visualization
KKS - Writing-Original Draft, Writing-Review and Editing
MLV - Writing-Original Draft, Writing-Review and Editing, Visualization
VZ - Writing-Original Draft, Writing-Review and Editing, Visualization
LEA - Writing-Original Draft
JIF - Conceptualization,
JMG - Writing-Original Draft
WHH - Editing
MAJ- Writing- Review and Editing
MMJ-
EJ -Writing- Review and Editing
MK, Visualization and editing,
DK - Writing-Review and Editing
GL -Writing-Original Draft
TM -Data Curation

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AM -Writing-Original Draft

SO -Writing-Original Draft

SP - Conceptualization,

GDP - Writing-Original Draft

LS -Writing-Original Draft

KiS- Conceptualization, Writing - Original Draft, Writing-Review and Editing, visualization, Supervision, Funding acquisition

Competing interests

The authors declare that they have no conflict of interest.

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