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Anthropogenic global warming from the perspective of the Phanerozoic evolution of Earth's climate

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A b s t r a c t. The paper presents an overview of published data on the Phanerozoic evolution of Earth's climate, and on examples of abrupt climate warming linked to greenhouse gas emissions. A major role of carbon dioxide in long-term climate forcing is documented by the co-occurrence of geological proxies of an elevated atmospheric CO_2 level and high surface temperatures during the Phanerozoic history of the Earth, which is related to global geochemical processes. The rapid climatic warming during the Early Toarcian (Jurassic) and the Paleocene-Eocene transition (Paleogene) was caused by emission of greenhouse gases (CO_2 and CH_4) from magmatic sources, thermal decomposition of fossil organic matter and their expulsion in secondary trigger processes from terrestrial and marine pools. The same phenomena are among the presumed effects of burning of fossil fuels by recent human activity. The expected anthropogenic global warming due to the combustion of the fossil fuel reserve is lesser than the major warming of the past; however, it can cause a significant increase in atmospheric CO_2 level and a rise in surface temperatures marking the end of the present icehouse conditions of the Earth's climate.

Keywords: *greenhouse and icehouse climate, greenhouse gases, global geochemical processes, rapid warming episodes*

INTRODUCTION

Discussions among geologists commonly concern the role of natural climatic factors on the Earth's climate forcing. In addition, major climate changes of the past, which are recorded in geological archives, are often evaluated in terms of their comparison to recent anthropogenic warming in order to predict possible changes in the future Earth's environment. This can be done assuming constancy of long-lasting geological, geochemical or astronomical processes.

The palaeoclimatic proxies are indeed an important and sometimes underestimated source of information on the temporal and spatial variations of the Earth's climate, its long-term evolution and the influence of certain physical or chemical processes on average surface temperatures. They can also help to distinguish between present-day natural and anthropogenic factors affecting global climate. This is particularly important for predicting the longer-term effects of anthropogenic global warming and the probable length of the future recovery of the global climate to pre-industrial conditions, both of which are difficult to model. An effective approach to past climate change requires, however, that all the forcing processes and the very long Phanerozoic evolution of Earth's climate be taken into account.

MODES OF PHANEROZOIC CLIMATE AND THEIR CO_2 FORCING

Two different modes of the Earth's climate in the Phanerozoic are related to the presence or absence of continental glaciation in polar regions. During greenhouse periods, high surface temperatures did not favour formation of continental ice sheets at high latitudes. This was associated with a smaller latitudinal temperature gradient, a more balanced configuration of climatic zones and the presence of warm deep ocean water masses (cf. Crowley, Zachos, 2000; Gröcke, 2009; Zhang *et al.*, 2019 and references therein). A side-effect of the lack of continental glaciation, amplified by the presumed high activity of the oceanic crust during at last some greenhouse periods, was a high sea-level. This led to the formation of extensive, warm epeiric seas, in which shallow-water organisms flourished.

Low temperatures in the high latitudes during icehouse periods favoured formation of continental ice caps, which retained huge amounts of fresh water. This led to sea-level fall, the circulation of cold deep ocean water masses that formed in polar regions, and reduction in marine biodiversity, especially at high latitudes and in the pelagic realm (cf. Thomas, 2008; Link, 2009; Zhang *et al.*, 2019 and references therein). From a palaeoclimatic point of view the most important features of the icehouse climate are probably the enhanced latitudinal temperature gradient and the susceptibility of the high latitude climate to orbital forcing, which resulted in a cyclical pattern of transitions from glacial to interglacial in subpolar regions and, to some degree, in northern parts of middle latitudes (cf. Pailard, 2006).

The early-middle Paleozoic (except for the Late Ordovician–Early Silurian) and the Mesozoic are considered as typical greenhouse periods. The Mesozoic greenhouse climate, despite the possible occurrence of short-term cold snaps, is well-documented by the widespread distribution of thermophilic floras and faunas, specific facies patterns, including the paucity of ice-related deposits, and palaeotemperature estimates based on isotope proxies (e.g., Hallam, 1994; Sellwood, Price, 1994; Price, 1999; Rees *et al.*, 2000; Leinfelder *et al.*, 2002; Sellwood, Valdes, 2006; Fletcher *et al.*, 2008; Scotese *et al.*, 2021). During the Cenozoic a gradual switch to an icehouse climate occurred. This is well-documented by the oxygen isotope record of benthic

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foraminifera showing strong shifts to more positive $\delta^{18}O$ values in the lower-middle Oligocene and, after a short-term recovery, starting from the middle Miocene onwards (Zachos *et al.*, 2001; Cramer *et al.*, 2011; Mudelsee *et al.*, 2014; Zhang *et al.*, 2019). The $\delta^{18}O$ shifts are attributed to a combined effect of a decrease in deep-sea temperatures and an increase in continental ice-volume, first in Antarctic and then in the Northern Hemisphere. However, the Cenozoic Antarctic glaciation was additionally stimulated by the circumpolar position of this continent and its isolation, starting from the Miocene, as a result of the full opening of the Drake Passage and the formation of Antarctic Circumpolar Current (cf. Zhang *et al.*, 2010).

Geological proxies recorded in geochemical and palaeobotanical archives show a close correspondence of high surface temperatures, typical of greenhouse periods with elevated atmospheric CO_2 levels (>1000 ppm) during the Phanerozoic (Pearson, Palmer, 2000; Retallack, 2002; Royer *et al.*, 2004; Royer, 2006; Glikson, 2008; Mills *et al.*, 2019; Judd *et al.*, 2024; Fig. 1). In addition, results of geochemical cycle modelling show distinct correlation of the well-documented greenhouse periods of the Phanerozoic era with high concentrations of CO_2, which is regarded as a powerful and long-lived greenhouse gas (Berner, 1991, 1994, 2006; Berner, Kothovala, 2001). Primary drivers of the Phanerozoic CO_2 level and the global climate are identified as: (1) an increased rate of chemical weathering of land areas and high organic matter burial due to a rise of vascular plants starting from the Devonian, which culminated in the formation of large Carboniferous coal deposits and (2) an increased rate of silicate weathering in the Cenozoic, after the Mesozoic standstill. Both processes resulted in a distinct drop in the late Carboniferous-Permian and late Paleogene-Neogene atmospheric CO_2 levels leading to icehouse conditions and continental glaciations (Royer, 2006). Comparable results, indicating slightly lower Mesozoic CO_2 levels, being 3 to 7 times higher than the present one, have also been obtained using the carbon-oxygen-phosphorus-sulphur cycling model of Bergman *et al.* (2004), which neglects direct modelling of the erosion rate.

In this context, the cause of the Late Ordovician–Early Silurian glaciation occurring under inferred high CO_2 level remains unclear due to the lack of precise palaeoclimatic data. It may have been facilitated by the circumpolar position of Gondwana, which made the Earth system sensitive to brief drops in atmospheric CO_2 unrecorded in the sedimentary archive (cf. Gibbs *et al.*, 2000; Lv *et al.*, 2022). On the other hand, there is no proof for continental glaciation during other brief Mesozoic cold snaps as those of the early Middle Jurassic or Jurassic/Cretaceous transition (cf. Price, 1999; Scotese *et al.*, 2021). Therefore, they seem to be of regional importance only.

The long-term process of organic matter burial and of the formation of organic matter-rich deposits is generally well-known. It might have been partially reversed by natural weathering and magmatic processes, or by present-day human activity involving combustion of fossil fuels. Silicate weathering, which is the most efficient geological process of CO_2 sequestration, has relied on the liberation of calcium and magnesium from igneous rocks over the whole Phanerozoic history of the Earth and the formation of thick and relatively weathering-resistant carbonate deposits. This process may be represented by the following chemical formula involving CO_2 binding:

$$CO_2 + Ca(Mg)SiO_3 \rightarrow Ca(Mg)CO_3 + SiO_2 \qquad [1]$$

The increased rate of Cenozoic silicate weathering is usually regarded as resulting from the uplift of Himalayas or the arrival of the weathering-prone basaltic Deccan Traps in the equatorial humid belt, both of which occurred 45–50 million years ago (Retallack, 2002; Kent, Muttoni, 2013). Further acceleration of the silicate weathering during the Cenozoic may be linked to the emplacement of the Ethiopian Traps near the equator, magmatic extrusions in SE Asia and the glacial weathering of Antarctic rocks (Kent, Muttoni, 2013; Stoll *et al.*, 2024). A passage from the low rate of continental erosion during the Mesozoic to the gradually increasing rate of the continental erosion during the Cenozoic is reflected in the seawater $^{87}Sr/^{86}Sr$ ratio. This ratio is affected by inputs of

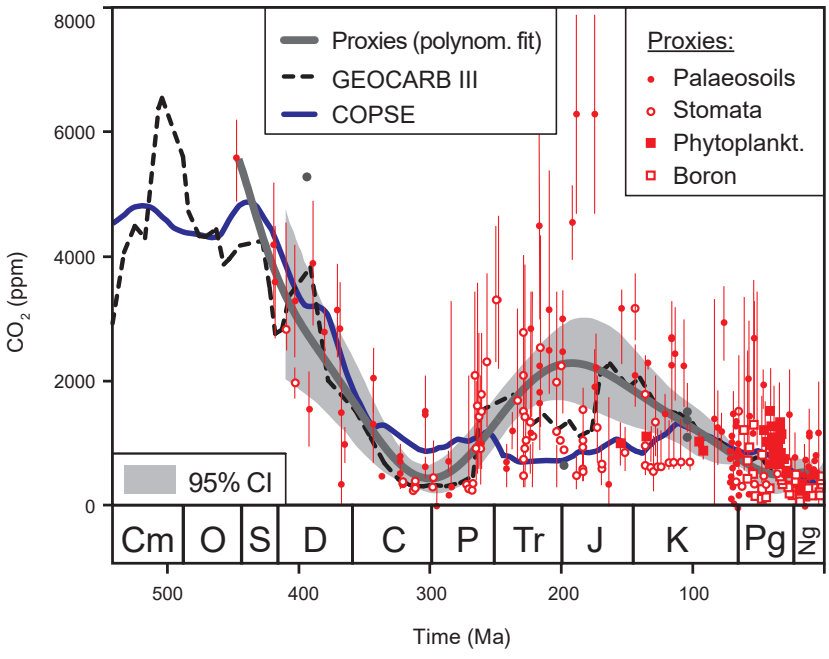

Fig. 1. Atmospheric CO_2 concentrations during the Phanerozoic estimated on the basis of geochemical modelling i.e. the GEOCARB III model of Berner and Kothavala (2001) and the COPSE model of Bergman *et al.* (2004) as well as direct determinations of proxies for ancient CO_2 level according to the database of Royer *et al.* (2004), which comprises CO_2 estimates based on interpretation of $\delta^{13}C$ values of the carbonate fraction of palaeosoils, stomatal density of plants, long-chained alkenones of haptophytic algae and $\delta^{11}B$ values of marine carbonate fossils. A tenth degree polynomial function has been fitted to proxy data (R = 0.69), with given 95% confidence limits (CI), to show a general trend of observed Phanerozoic CO_2 variations. Customary labels of the geological periods of the Phanerozoic era and the time scale are marked at the bottom of the diagram

strontium derived from global oceanic and continental pools characterized, respectively, by low and high strontium isotope ratios (Wierzbowski, 2013). Jurassic and Early Cretaceous seawater was characterized by a low $^{87}Sr/^{86}Sr$ ratio, with the Phanerozoic minimum of ~0.70683 falling close to the Middle–Late Jurassic transition (Wierzbowski *et al.*, 2017; McArthur *et al.*, 2020). A general rise of $^{87}Sr/^{86}Sr$ ratio, interrupted by a slight Paleocene-Eocene decrease, commenced in the Late Cretaceous (Coniacian; McArthur *et al.*, 2020). It culminated in the high present day seawater $^{87}Sr/^{86}Sr$ ratio of 0.70917 (cf. El Meknassi *et al.*, 2020).

Under the low CO_2 levels (<500 ppm), typical of the late Cenozoic icehouse, high latitudes became susceptible to climatic forcing linked to Milankovitch cycles (cf. Pearson, Palmer, 2000; Royer, 2006). A predominant ~100 Ka cyclicity related to orbital eccentricity is found to affect Quaternary climate. Although, the role of eccentricity changes in the summer insolation of high latitudes is insignificant in itself, the presence of an additional mechanism of glacial dust formation, which reduces the albedo of ice, likely played a major role in the 100 Ka cyclicity of build-up and loss of Quaternary continental ice sheets (Ganopolski, Calov, 2011; Ganopolski, Brovkin, 2017). The glacial-interglacial cycles coincide with variations in surface temperatures, deep ocean temperatures, and fluctuations in $\delta^{18}O$ values of seawater related to the periodical increases and decreases in ice volume. These parameters are recorded in the oxygen isotope values of carbonate oozes from oceanic cores as well as the oxygen and deuterium isotope profiles of Antarctic and Greenland ice cores (cf. Petit *et al.*, 1999; Shackleton, 2000; Glikson, 2008; Lüthi *et al.*, 2008 and references therein; Fig. 2). Interestingly, the orbitally controlled temperature oscillations of the middle Pleistocene–Holocene interact with similar in-phase variations in the atmospheric CO_2 level deciphered from ice-cores (Fig. 2). Although the coupling of ice sheets and carbon cycle remained, for a long time, unclear and was a source of confusion regarding the role of CO_2 in the Quaternary climate forcing, recent models, which take into account physical ocean-atmosphere interactions, simulate well the magnitude and timing of glacial-interglacial atmospheric CO_2 fluctuations (Ganopolski, Brovkin, 2017; Ganopolski, 2024). Only minor long-term temporal changes

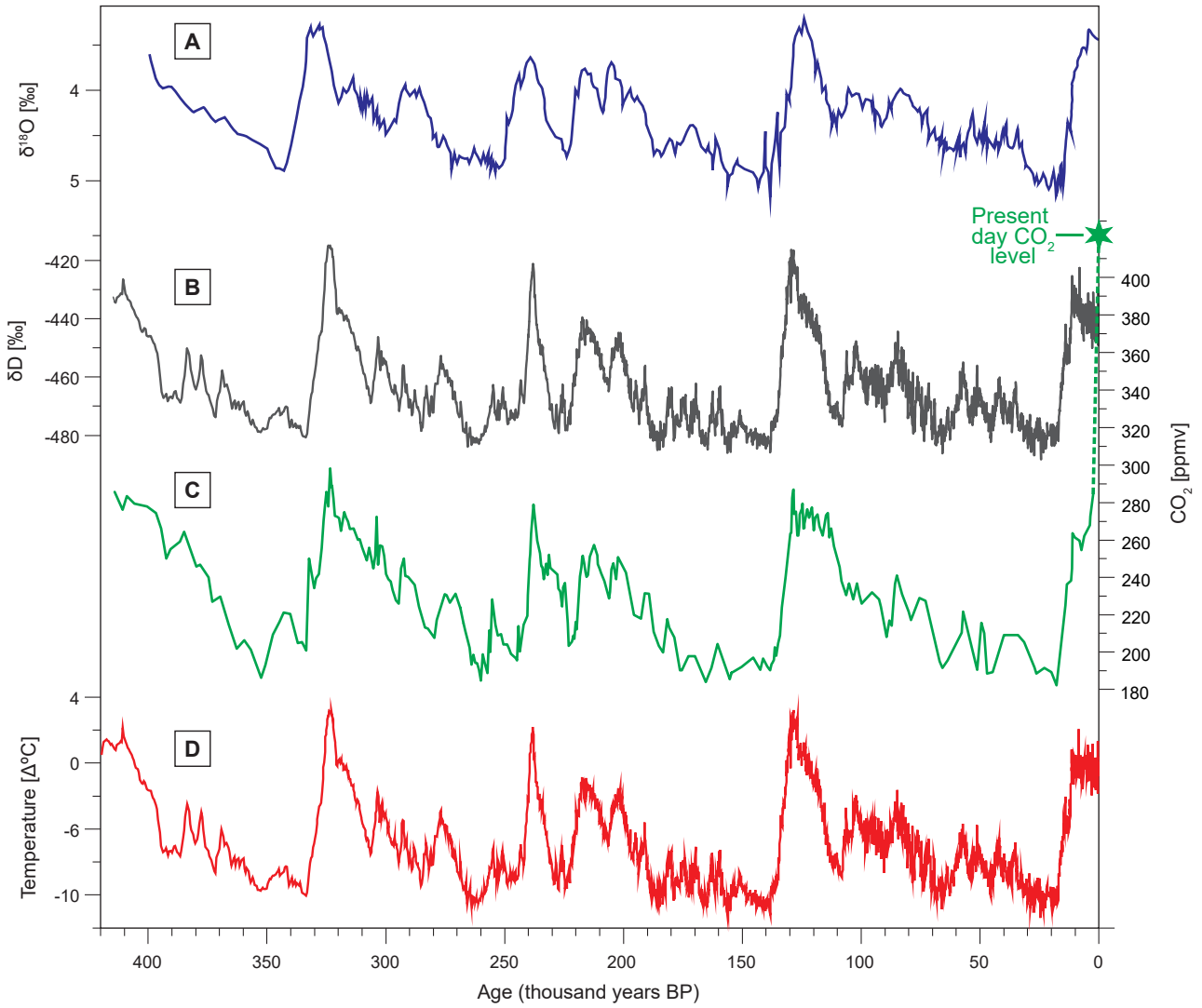

Fig. 2. Climatic proxies of the middle Pleistocene–Holocene. **A** – benthic $\delta^{18}O_{carbonate}$ curve from oceanic cores V19-30 and V19-28 according to Shackleton (2000). **B** – deuterium isotope profile of the Vostok core after Petit *et al.* (1999). **C** – variations in CO_2 concentrations in air bubbles of the Vostok core (solid line; after Petit *et al.*, 1999) supplemented with an industrial-era increase in atmospheric CO_2 content (dashed line); the present-day atmospheric CO_2 concentration is marked with asterisk. **D** – temperature variations estimated based on the δD signal of the Vostok core (after Petit *et al.*, 2001)

in atmospheric CO_2 concentrations of the last 800 Ka may represent a feedback of the Earth system to variations in the global weathering rate (Lüthi *et al.*, 2008).

Notably, human activity is resulting in a rapid increase in atmospheric CO_2 level. Its present-day value (425 ppm) significantly exceeds the pre-industrial atmospheric CO_2 concentration (280 ppm), which is close to the maxima of the past four interglacial ages recorded in ice-cores (Fig. 2). It brings Earth's climate closer and closer to the so-called "cool non-glacial conditions" predicted for the CO_2 range between 500 and 1000 ppm (Royer, 2006). The question arises as to whether similar rapid climatic phenomena occurred in the geological history of the Earth and whether the past natural climate change may be used as a key for the interpretation of the present anthropogenic one.

RAPID WARMING EVENTS IN THE PHANEROZOIC

Severe climate changes occurred during the Phanerozoic history of the Earth. These were mostly linked to perturbations in the global carbon cycle, which are recorded in the $\delta^{13}C$ values of marine carbonates as well as the $\delta^{13}C$ values of marine and terrestrial organic matter. A significant role of rapid greenhouse gas emissions, caused by volcanic processes, and its impact on the Mesozoic-Cenozoic Earth's climate is well-studied and discussed as regards the Permian–Triassic mass extinction event (PTME) end-Triassic mass extinction (ETE), early Toarcian Oceanic Anoxic Event (TOAE), early Aptian Selli Event (or OAE 1a), Cenomanian–Turonian boundary Bonarelli event (or OAE2), Cretaceous–Paleogene (K–Pg) extinction event, Paleocene–Eocene thermal maximum (PETM) and the Eocene Thermal Maximum 2 (ETM-2). The course of geological phenomena responsible for the emergence of these events has been reviewed in detail (e.g., Keller *et al*., 2011; Davies *et al*., 2017; Remírez, Algeo, 2020; Joo *et al*., 2020; Harper *et al*., 2020; Hull *et al*., 2020; Kender *et al*., 2021; Dal Corso *et al*., 2022; Jiang *et al*., 2022 and references therein). From the point of view of anthropogenic global warming, the well-studied "supergreenhouse" episodes of the Early Jurassic TOAE (also called the "Jenkyns Event") and the Paleocene-Eocene PETM are particularly interesting. Both episodes are considered as related to emissions of CO_2 and CH_4 gases that are well-recorded in sedimentary archives. They resulted in strong perturbations of the global carbon cycle and global warming.

Toarcian Oceanic Anoxic Event

Formation of the large Karroo-Ferrar igneous province (~183 million years ago) located in the southern Gondwana resulted in huge magmatic CO_2 and thermogenic CH_4 and CO_2 emissions in short-term pulses that are responsible for the TOAE (Ivanov *et al*., 2017; Font *et al*., 2022). Emission of thermogenic gases has been linked to the intrusion of Karoo-Ferrar sills into organic-rich sedimentary host rocks containing coal deposits. This process may be compared to the combustion of fossil fuels. The TOAE occurred after the Late Pliensbachian cooling event (Korte, Hesselbo, 2011; Gómez *et al*., 2016). Processes involved in the amplification of greenhouse forcing, which were triggered by rising surface temperatures, likely included dissociation of shallow-marine methane clathrates and decomposition of terrestrial organic matter (Jenkyns, 1988; Hesselbo *et al*., 2000; Pieńkowski *et al*., 2016).

The emission of greenhouse gases is recorded worldwide by a negative carbon isotope excursion in the lower part of the Falciferum (Serpentinum) Zone of the Lower Toarcian, which is characterized by 6–7 and ~2‰ negative shifts in organic matter and marine carbonate $\delta^{13}C$ values, respectively (e.g., Hesselbo *et al*., 2000, 2007; Kemp *et al*., 2011; Hermoso *et al*., 2012; Suan *et al*., 2015; Arabas *et al*., 2017; Them II *et al*., 2017; Xu *et al*., 2018; Boulila *et al*., 2019; Reolid *et al*., 2020; Fig. 3). The negative excursion is also documented in the Polish Basin (Hessebo, Pieńkowski, 2011; Pieńkowski *et al*., 2016, 2020). It was accompanied by global warming of 6–8°C inferred from isotopic and chemical proxies, a crisis of carbonate productivity and a marine extinction event linked to ocean acidification (Bailey *et al*., 2003; Suan *et al*., 2010; Gómez, Goy, 2011; Caruthers *et al*., 2013; Krencker *et al*., 2020). The global warming and freshening of Arctic surface seawaters likely led to the slowdown of global oceanic circulation, ocean stratification, and bottom water anoxia, which is manifested by the widespread occurrences of black shales (Dera, Donnadieu, 2012; Remírez, Algeo, 2020). An increase in pCO_2 level, from the latest Pliensbachian ~850 to ~1750 ppm during the pre-TOAE, is suggested based on botanical and geochemical proxies (McElwain *et al*., 2005; Müller *et al*., 2020). The emission of greenhouse gases during the TOAE, mostly in the form of CH_4, has been variously estimated in different studies, from 1500–2700 Gt of pure carbon by Hesselbo *et al.* (2000), to 10,000 Gt of carbon by McElwain *et al.* (2005) and 15,000–27,000 Gt of carbon in more recent papers (cf. Suan *et al*., 2015; Remírez, Algeo, 2020). The absolute timing of TOAE is, up to now, poorly constrained. Although most astrochronological calibration suggests its 300–500 Ka duration and the high-frequency carbon cycle oscillations during its initial part (Boulila *et al*., 2014, 2019; Boulila, Hinnov, 2017) the duration of TOAE has been estimated at up to ~900 Ka by other authors (cf. Huang, Hesselbo, 2014; Pieńkowski *et al*., 2024).

Paleocene–Eocene Thermal Maximum

PETM is a well-known greenhouse episode, which occurred ~55,6 Ma ago at the Paleocene–Eocene transition. It is marked by a few permil negative $\delta^{13}C$ excursion, mostly related to emission of magmatic and thermogenic gases during the formation of the North Atlantic Igneous Province, albeit evidence for this scenario is ambiguous (Svensen *et al*., 2004; Storey *et al*., 2007; Gutjahr *et al*., 2017; Kender *et al*., 2021). Therefore, the carbon isotope excursion has also been related to the emission of CH_4 and CO_2 from secondary carbon reservoirs, i.e. dissociation of marine methane clathrates or a decay of terrestrial organic matter in response to an initial warming caused by magmatic processes (Dickens *et al*., 1995; Matsumoto, 1995; Kurtz *et al*., 2003; DeConto *et al*., 2012; Kender *et al*., 2021) or linked to a bolid impact (Kent *et al*., 2003; Cramer and Kent, 2005; Schaller *et al*., 2016).

The estimations of a global temperature rise during the PETM range from 3 to 8°C but most of them oscillate between 4 and 5°C (Sluijs and Brinkhuis, 2008; McInerney and Wing, 2011; Dunkley Jones *et al*., 2013; Evans *et al*., 2016). The duration of the onset of PETM, comprising a decrease to minimal $\delta^{13}C$ values, has been variously calculated from 750 years to 30 Ka (Röhl *et al*., 2007; Murphy *et al*., 2010; Wright and Schaller, 2013; Turner, Ridgwell, 2016). Recent astronomical calibration, however, suggests a 6 Ka duration of its onset (Li *et al*., 2022). The time span

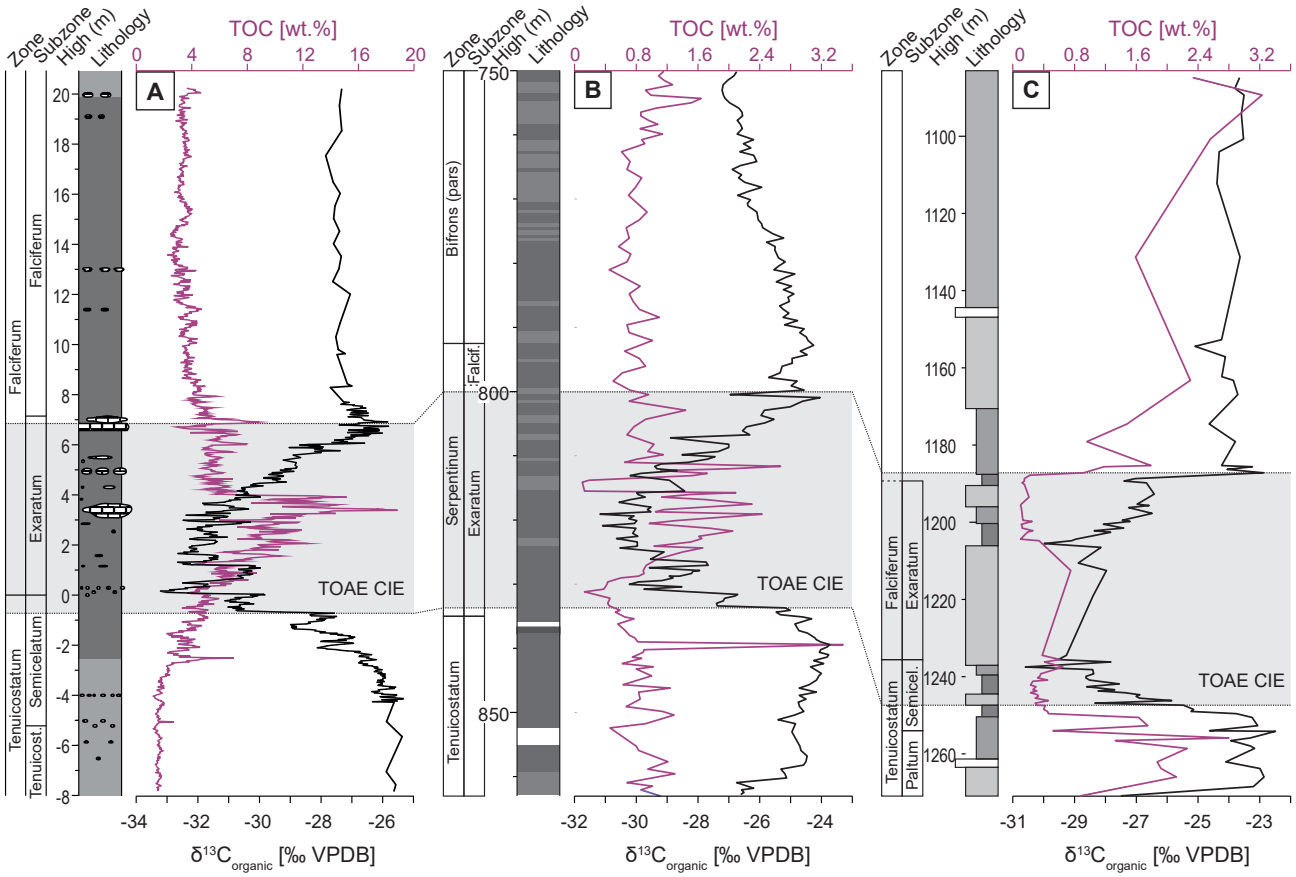

Fig. 3. Organic matter carbon isotope variations and total organic carbon (TOC) values of bulk sediments at the interval of the carbon isotope excursion (CIE) associated with the TOAE. **A** – Whitby section, Yorkshire, UK (after Kemp *et al.*, 2011). **B** – Mochras borehole, Wales, UK (after Xu *et al.*, 2018; Pieńkowski *et al.*, 2024). **C** – Kaszewy borehole, central Poland (after Pieńkowski *et al.*, 2020). The TOAE CIE excursion is marked in light grey. A lack of TOC enrichment in the CIE interval of the Kaszewy section is likely due to the shallowness of the Polish basin, which resulted in a lack of anoxic conditions and in decay of labile marine organic matter that predominated during this interval (Pieńkowski *et al.*, 2016, 2020).

of the entire PETM event from the onset until its termination and a return to cooler climatic conditions, similar to those of pre-PETM, is calculated at 170 Ka and includes a ~80 Ka period of the recovery phase (Röhl *et al.*, 2007; Murphy *et al.*, 2010; Wright, Schaller, 2013). The PETM timing constrains the assumptions of palaeoclimatic models but due to its rapid onset this event is often regarded as caused by a burst of greenhouse gas emission.

The modelling of carbon cycle perturbations during the PETM suggest a total emission of 3,000 to 17,000 Gt carbon, which may have included emissions by secondary, trigger processes (Zeebe *et al.*, 2009; Cui *et al.*, 2011; Wright, Schaller, 2013; Meissner *et al.*, 2014). The emission may have doubled the initial atmospheric CO_2 level of 800–1700 ppm (Zeebe *et al.*, 2009; Meissner *et al.*, 2014). Interestingly, the methane scenario of PETM imposes a lower peak level of pCO_2 of ~1200 ppm (Cui, Schubert, 2017). However, some calculations of carbon emission during the PETM may be underestimated due to the relatively heavy carbon isotope signatures of volcanogenic CO_2 gas, which hampers proper assessment of its amount (Gutjahr *et al.*, 2017).

CONCLUSIONS

This short review of palaeoclimatic data on the evolution of Earth's climate during the Phanerozoic and concerning abrupt warming episodes bring us to the conclusion that its primary drivers are greenhouse gases. Their link with long-term climatic variations and rapid warming episodes may be deciphered from geochemical archives of past sea-surface temperatures, perturbations in the global carbon cycle and ancient atmospheric pCO_2 levels. Importantly, the brief greenhouse episodes of the past, including the TOAE and PETM, show many similarities with the already initiated anthropogenic global warming. This applies to both the expected effects of rapid emissions of greenhouse gases on surface temperatures and their geochemical record in the form of negative carbon isotope excursions, which can be compared to the present-day isotope Suess effect (cf. Keeling, 1979; Eide *et al.*, 2017).

Recent estimates of carbon emissions during the major supergreenhouse episodes of the past, including the TOAE and PETM, exceed expected anthropogenic emissions (4600–4800 Gt of carbon) caused by burning of the known global fossil fuel reserves (cf. Glikson, 2008; Parker, Mainelli, 2024). However, this anthropogenic process will have significant impact on the global climate, and may result in a considerable increase in atmospheric CO_2, even >1000 ppm, to cause major global warming (cf. Parker, Mainelli, 2024). Although it is difficult to project melting of the entire Antarctica ice-cap under future global warming, due to the specific position of this continent in a circumpolar region with elevation of many parts of its interior (cf. O'Donnell, Nyblade, 2014) the process of the glacieustatic sea-level rise, even below the maximum of 60–70 m, will have significant impact on humankind.

Climatic models take into account ~300 Gt of carbon emitted since the beginning of the industrial age but there is still insufficient data to estimate the full size and speed of the expected warming. Secondary emissions from terrestrial and marine sources, which can be released by trigger processes, are additionally difficult to calculate. The rate of future global warming will, however, depend on the rate of world-wide annual anthropogenic CO_2 emissions, presently calculated at 10 Gt of carbon per year (Le Quéré *et al*., 2018; Friedlingstein *et al*., in press).

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