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Perspective

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Geological Society of London Scientific Statement: what the geological record tells us about our present and future climate

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Executive Statement

Geology is the science of how the Earth functions and has evolved and, as such, it can contribute to our understanding of the climate system and how it responds to the addition of carbon dioxide (CO_2) to the atmosphere and oceans. Observations from the geological record show that atmospheric CO_2 concentrations are now at their highest levels in at least the past 3 million years. Furthermore, the current speed of human-induced CO_2 change and warming is nearly without precedent in the entire geological record, with the only known exception being the instantaneous, meteorite-induced event that caused the extinction of non-bird-like dinosaurs 66 million years ago. In short, whilst atmospheric CO_2 concentrations have varied dramatically during the geological past due to natural processes, and have often been higher than today, the current rate of CO_2 (and therefore temperature) change is unprecedented in almost the entire geological past.

The geological record shows that changes in temperature and greenhouse gas concentrations have direct impacts on sea-level, the hydrological cycle, marine and terrestrial ecosystems, and the acidification and oxygen depletion of the oceans. Important climate phenomena, such as the El Niño–Southern Oscillation (ENSO) and the monsoons, which today affect the socio-economic stability and food and water security of billions of people, have varied markedly with past changes in climate.

Climate reconstructions from around the globe show that climate change is not globally uniform, but tends to exhibit a consistent pattern, with changes at the poles larger than elsewhere. This polar amplification is seen in ancient warmer-than-modern time intervals like the Eocene epoch, about 50 million years ago and, more recently, in the Pliocene, about 3 million years ago. The warmest intervals of the Pliocene saw the disappearance of summer sea ice from the Arctic. The loss of ice cover during the Pliocene was one of the many rapid climate changes observed in the record, which are often called climate tipping points. The geological record can be used to calculate a quantity called Equilibrium Climate Sensitivity, which is the amount of warming caused by a doubling of atmospheric CO_2, after various processes in the climate system have reached equilibrium. Recent estimates suggest that global mean climate warms between 2.6 and 3.9°C per doubling of CO_2 once all slow Earth system processes have reached equilibrium.

The geological record provides powerful evidence that atmospheric CO_2 concentrations drive climate change, and supports multiple lines of evidence that greenhouse gases emitted by human activities are altering the Earth's climate. Moreover, the amount of anthropogenic greenhouse gases already in the atmosphere means that Earth is committed to a certain degree of warming. As the Earth's climate changes due to the burning of fossil fuels and changes in land-use, the planet we live on will experience further changes that will have increasingly drastic effects on human societies. An assessment of past climate changes helps to inform policy decisions regarding future climate change. Earth scientists will also have an important role to play in the delivery of any policies aimed at limiting future climate change.

Introduction

Geological processes influence the climate system over a wide range of time-scales, from years to billions of years, and across all aspects of the Earth system, including the atmosphere (from the troposphere to the exosphere), cryosphere (encompassing snow, ice and permafrost), hydrosphere (encompassing oceans, streams, rivers, lakes and groundwater), biosphere (including soils), lithosphere (e.g. the rock cycle) and global carbon cycle. For example, on long time-scales, plate-tectonic processes control rates of volcanism (a key factor affecting the rate of supply of CO_2 to the atmosphere), and also the size and position of continents, mountains and ocean gateways, which influence oceanic and atmospheric circulation patterns, silicate

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weathering (which extracts CO_2 from the atmosphere by reacting it with rocks), biological productivity (plants extract CO_2 from the atmosphere for growth during photosynthesis) and weathering of organic rich material which returns CO_2 to the ocean–atmosphere system. Geological processes also impact climate on much shorter time-scales, including the cooling effects of major volcanic eruptions such as the 1991 eruption of Mount Pinatubo. Earth's climate is also influenced by extraterrestrial processes. On the longest time-scale, this includes the gradual progression of the Sun through its life cycle, as it slowly brightens over hundreds of millions of years. On medium time-scales, these include changes in the Earth's orbit (eccentricity) and changes in the tilt of the Earth's axis (obliquity). These operate on cycles of multiple periods, including 100 000, 40 000 and 20 000 years and work, together with the long-term carbon cycle and uneven distribution of Earth's land masses between the Northern and Southern Hemisphere, to control large-scale climate cycles such as glaciations.

Shorter time-scale changes in the Sun's output include the 11-year sunspot cycle and variations in that cycle with frequencies of 88, 208 and 2300 years among others. Crucially, for the majority of Earth's history, natural geological and extraterrestrial processes have directly affected climate, not least by producing major variations in atmospheric greenhouse gas concentrations. The long-term geological perspective on climate change makes it apparent that, through changes in weathering and volcanism, and Earth system feedbacks resulting from changes in the Earth–Sun relationship, the greenhouse gas concentration of Earth's atmosphere is the planet's long-term climate regulation system.

Variations in the Earth's climate state have, in turn, left their imprint throughout the geological record, from impressive features resulting from glacial erosion and deposition to subtle changes in the isotopic chemistry of marine microfossil shells (see Box 1). Palaeoclimate scientists can use these various records of Earth's climate to understand how the climate system operates over a range of time-scales, including providing vital information about the consequences of the current sharp rise in greenhouse gas concentrations and the resultant enhancement of the natural greenhouse effect. This contribution of palaeoclimatology to climate science is becoming more important as global temperatures continue to increase, along with ice melting and sea-level rising, in response to a climb in atmospheric CO_2 concentration to the highest levels in at least the past 3 million years. Indeed, because the geological past provides convincing examples of how the Earth functioned in different climate states (both colder and warmer than present), palaeoclimate research is making increasingly valuable contributions to successive IPCC Assessment Reports and is influencing policy decisions. This statement is organized around nine questions, all focused on the interactions between geology and climate change.

1. What does the geological record of climate change look like?

Sedimentary evidence indicates that there has been running water on Earth's surface since 3.8–3.4 billion years ago, which tells us that Earth's climate has remained within habitable bounds over this time, despite large changes in the Sun's energy output. For most of its history, Earth has been in a *greenhouse* state, significantly warmer than the present, with no polar ice caps and high sea-levels. This greenhouse state is sometimes referred to as a *hothouse* state instead, to distinguish it from the greenhouse effect that is always present on Earth because of the greenhouse gases present in the atmosphere (e.g. H_2O, CO_2, CH_4, N_2O), which maintain Earth's temperature at a level capable of supporting life. For example, the Cretaceous chalk forming the cliffs at Dover was deposited when the oceans flooded continental shelves during a greenhouse interval. Earth's greenhouse states have been punctuated by colder *icehouse* states, such as our current state and/or colder than today, with ice caps on both poles. (The icehouse state refers to a state of reduced, but not zero, greenhouse effect.) Other notable icehouse states include the first Snowball Earth event around 2.4 billion years ago, when ice expanded from the poles down to low latitudes (Evans *et al.* 1997). A further pair of Snowball Earth events occurred during the Neoproterozoic between 717 and 635 million years ago (Hoffman *et al.* 2017), and another less-extreme icehouse state characterized by polar ice sheets occurred during the Carboniferous about 300 million years ago, before the present icehouse state (Fig. 1).

Although these states typically last for tens of millions of years, the transitions between them can be relatively rapid, revealing the presence of climate tipping points. For instance, the most recent greenhouse-to-icehouse transition occurred 34 million years ago at the Eocene–Oligocene boundary, when the Antarctic ice sheet grew in two sharp steps each around 40 000 years in duration (Coxall *et al.* 2005).

Superimposed on these grand greenhouse–icehouse cycles, and often pacing transitions between states, are more rapid and frequent oscillations caused by variations in the Earth's orbit around the Sun and the tilt of the Earth's axis, known as Milankovitch cycles. Due to the gravitational interaction between the Earth and the other planets in the solar system, the obliquity (tilt) and precession (wobble) of the Earth's rotation varies on a time-scale of 41 000 years and 19 000–23 000 years, respectively, as does the circularity/eccentricity of its orbit (operating on 100 000- and 400 000-year cycles; Laskar *et al.* 2004). The impact of these cycles, which become amplified by a series of internal climate feedbacks, are perhaps most clearly seen in the geological record during the icehouse climate state of the late Cenozoic (over the last 34 million years; Zachos *et al.* 2001) where oscillations between more and less extreme glaciation (known as glacial–interglacial cycles) are clearly paced by Milankovitch cycles (Lisiecki and Raymo 2005; Huybers 2006; Fig. 1).

On even shorter millennial and sub-millennial time-scales, the most rapid geological changes in climate occur. These rapid changes include (i) rapid warming/cooling events during the most recent transitions from glacial-to-interglacial climates associated with dramatic changes in ocean circulation (e.g. Shakun *et al.* 2012); (ii) rapid warming events known as hyperthermals driven by vast outpourings of CO_2 and other greenhouse gases from Large Igneous Provinces, the most recent example of which, the

Box 1: How is past climate change written in the rocks?

Evidence for climate change is preserved in a wide range of geological settings, including marine and lake sediments, ice sheets, fossil corals, stalagmites and fossil tree rings. Advances in field observation, laboratory techniques and numerical modelling allow geoscientists to show, with increasing confidence, how and why climate has changed in the past. For example, cores drilled through the ice sheets yield a record of polar temperatures and atmospheric composition ranging back to 120 000 years in Greenland and 800 000 years in Antarctica. Oceanic sediments preserve a record reaching back tens of millions of years, and older sedimentary rocks extend the record to hundreds of millions of years. Some lines of evidence for climate change are direct (e.g. glacial landforms indicating the past presence of ice), whereas others are indirect (e.g. geochemical proxies for past changes in temperature, pH and ocean circulation). One example is the oxygen isotopic composition of marine microfossils, which provides information on past ocean temperature and salinity. This proxy is sensitive enough to record changes in global ocean salinity caused by growth and decay of continental ice sheets. While ice cores provide a direct means of measuring atmospheric CO_2 concentrations for the past 800 000 years, there are several indirect means of reconstructing atmospheric CO_2 concentrations further back in time (e.g. foraminiferal boron isotope ratios and alkenone carbon isotope ratios). Commonly, geologists use multiple independent proxies to increase their confidence in past climate reconstructions.

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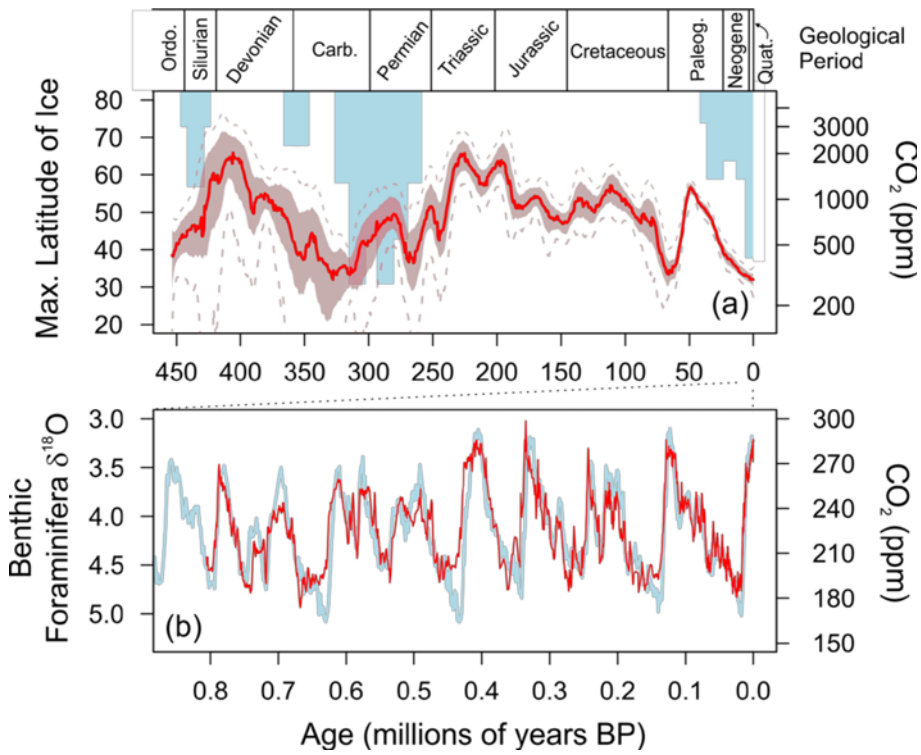

Fig. 1. Carbon dioxide and climate through the Phanerozoic. (**a**) Latitudinal extent of continental ice deposits (Crowley 1998) shown as blue bars (left axis) and atmospheric CO_2 content (red, right axis) from the compilation of Foster *et al.* (2017) augmented with the data from Witkowski *et al.* (2018). The dotted line shows the 95% confidence interval and the shaded band the 68% confidence interval. Note the close correspondence between extensive continental glaciation (i.e. cold icehouse climates) and intervals of low CO_2 content (e.g. the Carboniferous), and times of no or limited continental ice (i.e. warm greenhouse climates) with elevated CO_2 (e.g. the Cretaceous). (**b**) A close-up of the last 0.8 million years comparing CO_2 from ice cores (red, right axis; Bereiter *et al.* 2015) with benthic foraminiferal oxygen isotope ratios ($\delta^{18}O$) (blue, left axis; from Lisiecki and Raymo 2005). Benthic foraminiferal $\delta^{18}O$ reflects a combination of bottom water temperature and ice volume, cold glacial climates are towards the bottom of the diagram and warm interglacial intervals are at the top of the diagram. Again a close correspondence exists between cold intervals and low CO_2 (and vice versa). In (b) the relatively rapid cycling between low CO_2 glacials and high CO_2 interglacials is paced by orbital forcing (Milankovitch cycles, see text and Box 2). The geological time intervals are shown on the top of panel (a). Abbreviations are as follows: Carb., Carboniferous; Ordo, Ordovician; Paleog, Paleogene; Quat., Quaternary.

Paleocene–Eocene Thermal Maximum, warmed the Earth by 5°C within 2000–20 000 years (Frieling *et al.* 2017; Turner 2018); (iii) catastrophic changes caused by the mass extinction-inducing impacts with meteorites and comets, for example at the Cretaceous–Paleogene boundary (e.g. Henehan *et al.* 2019); and (iv) Dansgaard–Oeschger events in and around Greenland during Pleistocene glacials, caused by instabilities in the ocean–ice–atmosphere system at times of intermediate ice volume (e.g. Dokken *et al.* 2013).

The geological record reveals that the Earth's climate varied substantially over the last 4 billion years and confirms the importance of greenhouse gases in determining the climate state and habitability of our planet. The geological record of climate change has revealed evidence for feedbacks and tipping points

Box 2: Which came first – the CO_2 or the temperature?

On long time-scales there is a clear relationship between past CO_2 and global temperature (Fig. 1). However, at times when CO_2 and temperature are changing rapidly, for example during the glacial–interglacial transitions of the last 1 million years, this direct relationship can appear temporarily disrupted due to the varying time-scales of the feedbacks that couple CO_2 and temperature together.

CO_2 can drive climate change both as a primary agent and as a feedback, thanks to complex interactions between the various components of Earth's climate system. A key example of CO_2 as a (positive) feedback in response to an initial trigger elsewhere in the climate system comes from interglacial–glacial variability within the Late Pleistocene. For example, ice core and marine records covering the last 800 000 years or so show that during the transitions into glacial periods, CO_2 occasionally decreased after Antarctic cooling had commenced. Notably, however, there has been no such lag observed between rising CO_2 and rising temperature. For example, during the last deglacial transition (between *c.* 20 000 and 10 000 years ago) the rise in atmospheric CO_2 occurred in tandem with increasing temperature over Antarctica (Parrenin *et al.* 2013) and well before warming across the Northern Hemisphere (Shakun *et al.* 2012).

Understanding the connections between changes in temperature and CO_2 requires knowledge of the complex reactions within the carbon cycle, which involve not only straightforward thermodynamic relations (e.g. between ocean temperature and its capacity to sequester CO_2) but also the impact of changing biology, ocean circulation, air–sea gas exchange and the interactions between seawater and ocean floor sediments.

In addition, we have to bear in mind that the main drivers of the greenhouse effect on global temperatures are water vapour (about 50%), clouds (about 25%) and CO_2 (about 20%) (Schmidt *et al.* 2010). During glacial development the main source for water vapour (i.e. ocean evaporation) would have been restricted when (a) orbital change led to cooling, (b) ocean area decreased (thanks to the build-up of land-based ice sheets and associated sea-level fall) and (c) the decrease in land-plant coverage limited evapotranspiration. Apparently, water vapour and other feedbacks were more important during this phase of glacial build-up, meaning that decreasing CO_2 acted primarily as a feedback amplifying the later stages of cooling. On the other hand, the rapid rise in CO_2 during early deglaciation played a leading role in rising global temperatures over that period.

These observations show why it is unrealistic to expect a simple 1:1 relationship between CO_2 and temperature. What we have seen since 1950 is a reversal of the usual Pleistocene processes (in which CO_2 played a supporting role), with increasing CO_2 now becoming sufficiently abundant to dominate over (i) orbital insolation (which is stable to declining) and (ii) solar activity (with sunspots and solar irradiance in decline since 1990). As a result, CO_2 is now in the driving seat.

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in the climate system when gradual forcing can cause abrupt and sometimes irreversible changes. Geology also records the trends in evolution and anthropology that ran in parallel with climate change and may have been influenced by them. Examples include (i) the extinction of 96% of species at the Permian–Triassic boundary 252 million years ago (Benton 2018), (ii) the expansion of early *Homo* lineages out of Africa, as the climate became drier and less hospitable around 100 000–50 000 years ago (Tierney *et al.* 2017) and (iii) the emergence of human civilization during the Holocene, a relatively stable and warm interglacial phase within an overall icehouse state.

2. Why has climate changed in the past?

Earth's climate system can be represented in its simplest form as a balance at the top of the atmosphere between the amount and distribution of incoming solar radiation and the outgoing longwave radiation emitted by the Earth–atmosphere system. The expression of this balance can be written in mathematical form as

$$\sigma T_E^4 = \frac{1}{4} F_s (1 - A), \quad (1)$$

where σ is the Stefan–Boltzmann constant (5.67×10^{-8} W m^{-2} K^{-4}), T_E is Earth's 'effective temperature' (the temperature at which radiative equilibrium is achieved assuming the Earth acts like a blackbody), F_s is the total solar irradiance (currently 1361 W m^{-2}), and A is Earth's average planetary albedo (the fraction of incoming radiation scattered or reflected back into space, currently 0.29) (Trenberth and Fasullo 2012).

If equation (1) is rearranged to solve for the Earth's effective temperature, and the quantitative values above are inserted into the equation, then

$$T_E = \left[\frac{F_s(1 - A)}{4\sigma}\right]^{1/4} = 255\text{K}\,(-18\,^\circ\text{C}). \quad (2)$$

Thus, the simplest Earth's climate model predicts an effective temperature of –18°C (i.e. Earth's surface would be entirely frozen). Given that the Earth's average surface temperature was about +14°C prior to industrialization, something is needed to explain this discrepancy. The difference is due to the greenhouse gases, primarily water vapour and CO_2. Because of their multi-atomic nature, greenhouse gases are able to absorb and re-emit longwave radiation within the atmosphere, leading to atmospheric heating (Lacis *et al.* 2010). Water vapour cycles through the hydrological cycle so quickly that its concentration in the atmosphere is primarily controlled by temperature, which drives evaporation. We focus on CO_2 because water vapour therefore acts as a feedback to changing CO_2 and is not able to drive climate change itself (Lacis *et al.* 2010).

The Earth's energy budget has changed over geological time and at multiple time-scales. Over billions of years, the total solar irradiance F_s has changed as the conversion of hydrogen to helium in the Sun's core has produced a hotter, brighter Sun (Gough 1981). However, even with less solar irradiance 4 billion years ago during the 'Faint Young Sun', liquid (not frozen) water was still present on Earth and climate has generally been warmer than at present (Section 1) (Sagan and Mullen 1972). On the early Earth, an enhanced greenhouse effect is proposed to have compensated for this reduced solar forcing, driven by higher CO_2 and/or methane before 2.5 billion years ago (Ramstein 2011). Since then, increasing solar irradiance has been accompanied by an overall long-term decrease in CO_2 and methane, with variability on shorter time-scales. This reduction in greenhouse gas concentrations was caused by the silicate-weathering feedback (e.g. carbon dioxide is removed from the atmosphere by absorption in falling rain, forming weak carbonic acid and accelerating the weathering of silicate rocks) and biological evolution (e.g. the growth of plant biomass on Earth has removed carbon dioxide from the atmosphere and increased atmospheric oxygen concentrations (e.g. Berner and Kothavala 2001; Kasting 2004; Ramstein 2011; Foster *et al.* 2017).

Over millions of years, climate has changed in response to plate tectonics, biological evolution and the impact of carbon cycling on CO_2 (Fig. 1). Intervals of high CO_2 likely reflect times of higher volcanic inputs of this gas in response to tectonic changes (e.g. Mills *et al.* 2017). Silicate weathering is another important component of the geochemical carbon cycle, which removes CO_2 from the atmosphere (Raymo and Ruddiman 1992; Joshi *et al.* 2019). Changes in ocean carbonate chemistry, evident from the depth of calcite preservation in the deep sea, also influence CO_2 concentrations. The deposition and weathering of evaporite minerals is another driver of atmospheric CO_2 concentrations, through its influence on carbonate sedimentation (Shields and Mills 2020). The evolution of the terrestrial biosphere has also been particularly important in the carbon cycle (e.g. Berner and Kothavala 2001; Foster *et al.* 2017), as shown by the following two examples. First, the increased use of CO_2 by evolving land plants during the Ordovician has been linked to reduced atmospheric CO_2, cooling and glaciation (Lenton *et al.* 2012). Second, the evolution of trees and leaves during the Devonian, with progressive forest development, could have reduced atmospheric CO_2, leading to the Carboniferous glaciation (Beerling 2007). The entrapment of the remains of tropical forests in swamps as Gondwana and Laurasia came together, a process that led to the formation of the world's major coal deposits, further contributed to the decline in atmospheric CO_2 during the aptly named Carboniferous Period.

Over the last 40 million years, four phases of increasing glaciation have been linked to decreasing atmospheric CO_2: glaciation in Antarctica in the earliest Oligocene and the middle Miocene (Pearson *et al.* 2009; Pagani *et al.* 2011; Foster *et al.* 2012) and the appearance of Northern Hemispheric ice sheets in the late Pliocene (Martinez-Boti *et al.* 2015) and mid Pleistocene (Chalk *et al.* 2017). Superimposed on these long-term drivers of climate are variations that reflect changes in the distribution of incoming solar radiation, driven by Milankovitch cycles, and amplified by climate feedbacks including the greenhouse effect (Box 2). For example, a series of hyperthermals during the Paleocene and Eocene were paced by the approximate 100 000-year changes in the eccentricity of Earth's orbit and the release of CO_2 to the atmosphere from a changing ocean circulation (Sexton *et al.* 2011). Antarctic ice-sheet advances and retreats have also been linked to Earth's orbital cycles (e.g. Naish *et al.* 2001; Galeotti *et al.* 2016; Liebrand *et al.* 2017). In the Pleistocene, feedbacks from ice albedo and carbon storage in the deep ocean amplified and modulated the initial pacing of regional and global climate changes by orbital forcing of incoming solar radiation (e.g. Yin 2013; Martinez-Boti *et al.* 2015; Lear *et al.* 2016).

Volcanic eruptions contribute greenhouse gases and also affect the amount of sunlight reaching Earth through the emission of particles into the atmosphere. Plateau basalt eruptions in Large Igneous Provinces have been responsible for temporary cooling caused by opaque ash clouds, and longer-term warming due to the emission of CO_2 (Kidder and Worsley 2010). For example, the volcanic eruptions of the North Atlantic Igneous Province were coincident with the Paleocene–Eocene Thermal Maximum 56 million years ago. At this time, volcanic emissions dramatically increased atmospheric CO_2 concentrations over less than 20 000 years. The resulting warming of about 5°C was sustained for about 150 000 years until CO_2 concentrations gradually declined (Section 3; Zachos *et al.* 2001; Sluijs *et al.* 2007). Large eruptions by individual volcanoes can also inject particles into the stratosphere, causing temporary cooling for up to 5–10 years (Sigl *et al.* 2015). Although climatically important in the past and on geological time-scales, volcanic activity on land and in the ocean provides only a fraction of CO_2 globally – 135 times less than all human emissions

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(in 2010) and about as much annually as all human activities in Florida (Gerlach 2011).

Climate change on decadal to centennial time-scales is well recorded in tree rings, corals, bivalves, marine and lake sediments, cave deposits such as stalactites, ice cores, borehole temperatures, glacier fluctuations and early documentary evidence from cultural archives. These records reveal the overwhelming importance of greenhouse forcing and stochastic climate variability in controlling the short-term climate anomalies of the recent past (the last 2000 years, referred to as the Common Era). The Common Era geological record reveals that climate anomalies of significance have occurred on multi-decadal time-scales in the recent past, chiefly the Little Ice Age (LIA; Matthews and Briffa 2005), Medieval Climate Anomaly (MCA; Bradley *et al.* 2003), the Dark Ages Cold Period (DACP; Helama *et al.* 2017) and the Roman Warm Period (RWP; Ljungqvist 2010). These globally asynchronous anomalies were forced by a combination of solar and volcanic changes, stochastic variability in the Earth's climate system and associated feedbacks (Mann *et al.* 2009). Common Era climate anomalies, occurring before the period of human enhancement of the greenhouse effect, are characterized by a lack of global coherence (Neukom *et al.* 2019). Within the Common Era, volcanic and solar climate forcing has, at no point, been strong enough to produce globally synchronous extremes of temperature at multi-year time-scales. Human-induced changes to the Earth's atmosphere have, in the twentieth and twenty-first centuries caused spatially consistent warming not seen at any other point in the last 2000 years (Neukom *et al* 2019).

Within the geological record, several key drivers of climate change can be identified that operate over a range of time-scales. These rarely occur in isolation because the Earth system contains many feedback processes that dampen or amplify climate change. Throughout geological history, atmospheric CO_2 has acted as both a driver of – *and* feedback to – global climate change. For example, the glacial–interglacial cycles of the Pleistocene are paced by subtle changes in planetary orbits, but the magnitude of these climate transitions were sensitive to changes in atmospheric CO_2 concentrations. The concentration of atmospheric CO_2 is now at its highest level in at least the past 3 million years, indicating that our present situation has geological analogues. Thus, geology offers a powerful opportunity to understand complex feedback processes and predict how the climate system may respond in the future.

3. Is our current warming unusual?

Given the record of past climate change (Section 1), the magnitude of recent observed climate change is not unusual. But how does the *rate of forcing* provided by human-influenced greenhouses gases such as CO_2 compare with that observed in the geological record? Here, we describe five examples of rapid geological climate events.

(i) *At the Cretaceous–Paleogene boundary:* 66 million years ago, a meteorite about 11 km in diameter hit the Earth on the northern boundary of what is now Mexico's Yucatan Peninsula (Alvarez *et al.* 1980; Hildebrand *et al.* 1991). A world-wide clay layer rich in iridium (which is otherwise rare on Earth) attests to the immediate global impact of the collision (Alvarez *et al.* 1980; Schulte *et al.* 2010). Instantaneous conversion of sediments and meteorite fragments to a stratospheric dust veil cooled the Earth for at least a decade, whereas vaporization of marine carbonates injected CO_2 into the atmosphere and warmed the world for some 100 000 years after the dust had gone (Schulte *et al.* 2010; Renne *et al.* 2013; MacLeod *et al.* 2018; Henehan *et al.* 2019). The impact caused the world-wide extinction of non-bird-like dinosaurs on land and ammonites in the ocean, among others. Extinction of most calcareous plankton reduced photosynthesis, helping to keep CO_2 in the atmosphere (Hay 2013). The plateau basalt eruptions of India's Deccan Traps straddle the Cretaceous–Paleogene boundary, but their major CO_2 outgassing ended prior to the mass extinction (Hull *et al.* 2020).

(ii) *At the Paleocene–Eocene boundary:* 56 million years ago, several billion metric tonnes of carbon were injected into the atmosphere in less than 20 000 years (Gutjahr *et al.* 2017). The event appears to have been driven, at least in part, by eruptions of the North Atlantic Igneous Province where CO_2 was supplied by volcanic eruptions and metamorphism of organic-rich sediments. The increased greenhouse effect caused a geologically rapid warming event (the Paleocene–Eocene Thermal Maximum) in which temperatures rose by about 5–6°C globally and by as much as 8°C at the poles (Zachos *et al.* 2001, 2008; Sluijs *et al.* 2007; Jaramillo *et al.* 2010). The warming was accompanied by ocean acidification, ocean deoxygenation, about 12–15 m of sea-level rise, major changes in terrestrial biota and the hydrological cycle, and one of the largest extinctions of deep-sea seafloor-dwelling organisms of the past 90 million years. At its peak rate, carbon was added to the atmosphere at around 0.6 billion tons of carbon per year (Gingerich 2019). It is important to note this is an order of magnitude less than the current rate of carbon emissions, of about 10 billion tons of carbon per year (Turner 2018; Gingerich 2019). The Earth system took between 100 000 and 200 000 years to recover, as organic carbon feedbacks and chemical weathering of silicate minerals slowly removed CO_2 from the atmosphere (Foster *et al.* 2018).

(iii) *The Eocene–Oligocene Climate Transition:* 34 million years ago is also known as Earth's latest greenhouse–icehouse transition, marking the rapid expansion of the Antarctic ice sheet following a gradual cooling of global climate. Small changes in the balance between volcanic CO_2 emissions and chemical weathering rates had led to a slow decline in atmospheric CO_2 concentrations, from about 1000–1500 ppm at 50 million years ago to about 400–900 ppm by 34 million years ago (Pearson *et al.* 2009; Beerling and Royer 2011; Pagani *et al.* 2011; Anagnostou *et al.* 2016). This reduction in CO_2 concentrations led to a gradual global cooling of about 4–7°C in the tropics, with an amplified response at high latitude (Zachos *et al.* 2008; Kent and Muttoni 2013; Froehlich and Misra 2014; Inglis *et al.* 2015; Cramwinckel *et al.* 2018). Although it used to be thought that Antarctic cooling resulted from its physical isolation by the Antarctic Circumpolar Current (Kennett 1977), geological evidence from the Southern Ocean suggests that the current did not fully develop until much later, during the Oligocene or Miocene (Pfuhl and McCave 2005; Dalziel *et al.* 2013). It is now thought that the cooling climate through the Eocene eventually came close to Antarctica's threshold for glaciation, with ocean circulation changes perhaps having a secondary influence, and a spate of cool summers driven by orbital parameters determining the exact timing of the glaciation (Coxall *et al.* 2005, 2018). Once glaciation began, a set of positive feedbacks (increasing albedo and cooling of the elevated ice surface) caused rapid ice-sheet growth in two steps around 40 000 years each in duration and a sea-level fall of several tens of metres (DeConto and Pollard 2003; Lear *et al.* 2008; Gulick *et al.* 2017). The establishment of the Antarctic ice sheet in less than 0.5 million years following a gradual cooling over 15 million years is a classic example of a tipping point in the climate system (DeConto and Pollard 2003; Francis *et al.* 2008).

(iv) *During the last deglaciation:* (20 000 to 11 700 years ago), CO_2 concentrations measured in bubbles of ancient air from Antarctic ice were tightly coupled to Antarctic air temperatures (Parrenin *et al.* 2013; Beeman *et al.* 2019) (Box 2). The deglaciation was driven by increasing Northern Hemisphere insolation; CO_2 supplied by the warming and de-stratifying ocean provided an important positive feedback (Shakun *et al.* 2012). The concentration of atmospheric CO_2 rose from 190 to 280 ppm (an average rate of *c.* 0.01 ppm a^{-1} but including centennial changes of *c.* 0.1 ppm a^{-1} (Marcott *et al.* 2014)). Again, it should be stressed that this is far slower than the modern

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Table 1. *Rates of change of CO_2 concentration, temperature and sea-level for the Dansgaard–Oeschger Events, the last deglaciation and the present day*

	Maximum rate of CO_2 change (ppm a^{-1})	Maximum rate of temperature change (°C a^{-1})	Maximum rate of sea-level change (mm a^{-1})
Millennial change during last glacial	0.008	0.32 (regional – Greenland)	
Last deglaciation	*c.* 0.01–0.1	*c.* 0.002 (global)	*c.* 50
Present day	>2	0.018 (global)	>3

Maximum rates of change for Dansgaard–Oeschger Events and the last deglaciation are measured over decades or centuries, but are expressed here as annual rates to facilitate comparison with present-day change

anthropogenic rate of more than 2 ppm a^{-1}. At the same time, Antarctic temperature increased by 9°C with the fastest warming between 12 700 and 11 700 years ago, at a rate of about 0.004°C a^{-1}. As the global change in temperature during the deglaciation was about 4–6°C (Shakun *et al.* 2012; Tierney *et al.* 2020), the global rate of warming was roughly half that of Antarctica, substantially slower than the present rate of global temperature change of 0.018°C a^{-1} (based on 1970–2020 data) (Table 1). Between 20 000 and 7000 years ago, melting continental ice caused sea-level to rise 130 m at an average rate of about 10 mm a^{-1} (IPCC 2013; Clark *et al.* 2016), but reached about 50 mm a^{-1} during Meltwater Pulse 1A (14 650–14 300 years ago), a brief period of exceptionally rapid loss of ice (Deschamps *et al.* 2012; IPCC 2013). These rates are much higher than the current rate of rise of 3.6 mm a^{-1} for 2006–15 (IPCC 2019), but the average is close to the maximum forecast rate of 10–20 mm a^{-1} at 2100 (IPCC 2019). These rapid rates of sea-level rise show that ice melt can respond rapidly to changes in the Earth system driven by small rises of CO_2 and temperature.

(v) Aside from the Cretaceous–Paleogene boundary, the fastest climate changes in the geological record are those of the *Dansgaard–Oeschger (D–O) Events* recorded in Greenland ice cores, where abrupt rises in regional temperatures of up to 16°C took place within a few decades (a rate of up to 0.32°C a^{-1}) (Masson-Delmotte *et al.* 2012; IPCC 2013), remained stable for 500–1000 years, then cooled (Steffensen *et al.* 2008). D–O events occurred during periods of intermediate ice cover when climate stability was weakest. The D–O cooling phases occurred at the same time as weak rises in temperature of about 1.2°C over 2500 years in Antarctica (a rate of 0.0005°C a^{-1}), associated with rises in CO_2 of about 20 ppm (a rate of 0.008 ppm a^{-1}) (Ahn and Brook 2007). D–O events, and their Antarctic counterparts, followed a cycle of about 1000–5000 years, driven by natural oscillations within the Atlantic Meridional Overturning Circulation (Broecker 2001). The fast warming of D–O events represents a regional tipping-point response in the climate system.

The geological record provides us with evidence of rapid, naturally occurring climate change. However, in terms of global-scale events, in the entire geological record, only the instantaneous Cretaceous–Paleogene meteorite impact event occurred more rapidly than the current human-induced global warming. A particular strength of palaeoclimate research is the identification of feedbacks and linkages between different parts of the climate system, which are important for understanding the future response of our planet to anthropogenic CO_2 emissions.

4. What does the geological record indicate about global v. regional change?

Although CO_2 emissions occur locally, the Earth's atmosphere mixes them relatively quickly and thoroughly. Thus, local emissions of CO_2 lead to global changes in climate. Through the different feedback systems outlined above (Sections 2), global climate change is then translated into regional climate changes, which are the conditions directly experienced by different communities across the world. Furthermore, some regions play a powerful role in creating feedbacks or tipping points in the climate system, which can, in turn, have global impacts. Palaeoclimate records have been used to gain critical insights into these processes. For example, during the Pleistocene, cycles of ice-sheet advance and retreat led to larger temperature changes in the polar regions than in the tropics (Brigham-Grette *et al.* 2013), in part due to changes in the reflection of incoming solar radiation by the changing extent of ice. This sensitivity to albedo is one cause of *polar amplification*, the increased signal of global warming in the polar regions compared with the tropics (CAPE-Last Interglacial Project Members 2006). The geological record shows that polar amplification also occurs in ice-free climates, driven by other atmospheric processes (Cramwinckel *et al.* 2018). Understanding polar amplification is critical for predicting future changes to Arctic sea ice and stability of carbon stored in high-latitude permafrost.

Regional seasonal weather patterns can also be imprinted in the geological record, including the West African and Indian Monsoons, which are crucial for the socio-economic stability and food and water security of billions of people (Dilley *et al.* 2005; Turner and Annamalai 2012). The geological record shows a dynamic history of these monsoon systems, caused by changes in orbital parameters and feedbacks in the climate system that are the subject of ongoing research (Gebregiorgis *et al.* 2018; Pausata *et al.* 2020; Williams *et al.* 2020). Palaeoclimate studies also indicate that past changes in the Indian and West Africa Monsoons may have triggered abrupt events in other regions ('induced tipping') and may have had a domino effect impacting climates in areas as far away as the Arctic (Nilsson-Kerr *et al.* 2019; Pausata *et al.* 2020).

One of the most important climatic phenomena on our planet is the El Niño–Southern Oscillation (ENSO), which influences flooding, droughts, food supplies and wildfires across the world. ENSO is an oscillation in atmospheric-pressure patterns and sea-surface temperatures that brings alternating warmth and rainfall variations mainly in and around the tropical Pacific, but with connections to the mid-latitudes, and represents the largest source of year-to-year global climate variability. Palaeoclimate proxies have been used to reconstruct ENSO variability over the past 7000 years, providing empirical support for recent climate-model projections, indicating an intensification of ENSO associated with anthropogenic global warming (Grothe *et al.* 2019).

For the future, many of the regional patterns that we see in the geological record are predicted to continue. As human-induced warming continues, changes at the poles will be larger than elsewhere, and summer sea ice is predicted to eventually disappear from most of the Arctic, as was the case during the Pliocene *c.* 3 million years ago. The ENSO may intensify, influencing the frequency of droughts and flooding in many areas.

5. When Earth's temperature changed in the past, what were the impacts?

The geological record also contains evidence of the wider impacts of climate change, which have relevance for humans, including changes to the hydrological cycle, ecosystems, ocean oxygen levels and pH, glaciers and ice sheets, and sea-level. Here we give some examples of past changes in precipitation, sea-level and ecosystems that accompanied past changes in climate.

The warming at the Paleocene–Eocene Thermal Maximum (Section 3) was associated with dramatic re-organization of the global hydrological cycle, more episodic rainfall (especially in arid

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regions), re-organization of fluvial systems and increased export of sediments to marginal marine locations (Schmitz and Pujalte 2007). On more recent glacial–interglacial time-scales, astronomical forcing of temperature appears to have a strong impact on the strength of monsoon systems via expansion of the Intertropical Convergence Zone (ITCZ) during warm periods and its contraction during cold periods. A dramatic change in the mid-Holocene is provided by the Saharan humid period, when lakes and grasses supported a flourishing trans-Saharan ecosystem (deMenocal and Tierney 2012). About 5000 years ago, the Northern Hemisphere cooled as insolation declined, and the Saharan region became a desert as the northern boundary of the ITCZ moved towards the equator (deMenocal and Tierney 2012). Overall, palaeoclimate records provide empirical support for projections of a globally enhanced but regionally uncertain hydrological cycle, resulting in both flooding and drought across the world.

The geological record shows changes in sea-level over the past 35 million years that were largely caused by changes in global ice volume as continental ice sheets waxed and waned. The largest changes in ice volume occurred in response to changing concentrations of atmospheric CO_2. For example, sea-level fell several tens of metres as the Antarctic ice sheet formed, starting about 34 million years ago at the Eocene–Oligocene Climate Transition, once a long-term decline in atmospheric CO_2 concentration reached a glaciation tipping point (Lear and Lunt 2016) (Section 3). During the Pliocene (about 3 million years ago), CO_2 concentrations were similar to modern levels (about 400 ppm), much of East Antarctica was covered by an ice sheet, but one smaller than that of today, and sea-levels were 6–20 m higher than they are now (Miller *et al.* 2012; Dumitru *et al.* 2019). Extensive Northern Hemisphere glaciation began 2.7 million years ago, as CO_2 concentrations declined further and the global climate cooled. During the coldest intervals of the Pleistocene, enormous ice sheets covered much of North America, Britain, Scandinavia and parts of the Siberian coast, and sea-level was at times 135 m lower than today (Lambeck *et al.* 2014). Orbital cycles were responsible for the timing of the growth and decay of these major Northern Hemispheric ice sheets, but the magnitude of change was strongly amplified by the roughly 100 ppm glacial–interglacial CO_2 changes recorded in ice cores. As the major Northern Hemispheric ice sheets retreated following the Last Glacial Maximum, sea-level rose about 1 m per century for about 100 centuries. However, this sea-level rise was not uniform; pulses of ice-sheet collapse caused some intervals of extremely rapid sea-level rise (e.g. about 5 m per century, equivalent to 50 mm a^{-1} (Table 1)) (Deschamps *et al.* 2012). Such abrupt geological events highlight the non-linear response of ice sheets to warming; contributions to sea-level are highly dependent on ice–ocean interactions and individual ice-sheet behaviour. Today, sea-levels are rising nearly 4 mm a^{-1}, and this rate is accelerating due to increasing ice loss from glaciers and ice sheets (Oppenheimer *et al.* 2019). Due to the prolonged response time of ice sheets, we are already committed to future substantial sea-level rise resulting from historical CO_2 emissions and their associated warming. The geological record is also consistent with predictions that the long-term magnitude and rate of future sea-level rise will be highly sensitive to future CO_2 emission scenarios.

In addition to its impact on Earth's climate, CO_2 concentration has a direct impact on the Earth system and especially on its biota. The rapid emission of CO_2 over time-scales shorter than 10 000 years inevitably leads to an acidification of seawater. Such an impact is clear from the carbonate-poor layers of the deep ocean during the Paleocene–Eocene Thermal Maximum, which were caused by ocean acidification resulting from rapid emissions of CO_2 to the atmosphere (Zachos *et al.* 2005). The geological record is consistent with experiments that suggest that the effect of ocean acidification on organisms depends on the degree of biological control of the organism over mineralization (Gibbs *et al.* 2006). Those organisms that exert least control suffer the most harm, which includes organisms with a vast range of ecological roles, including foraminifera, gastropods, molluscs and corals. Enhanced CO_2 concentrations also impacted marine and terrestrial photosynthesizers. Over the past 10 million years, the ability of land plants to process the CO_2 needed for growth (called C3 or C4) has changed from domination by C3 plants (most of the flowering plants) to a marked increase in C4 plants (many of them grasses) (Tipple and Pagani 2007). The physiology of C4 plants and their associated biochemistry mean that they can grow more efficiently than C3 plants at relatively low levels of CO_2, consistent with the natural decline in CO_2 over the Cenozoic. This change was accompanied by a great expansion of savannahs, typically covered by grasses.

The geological record is consistent with predictions that the long-term magnitude and rate of future sea-level rise will be highly sensitive to future CO_2 emission scenarios and may include intervals of very rapid rise. It also provides empirical support for projections of a globally enhanced but regionally uncertain hydrological cycle, resulting in both flooding and drought and affecting water security across the world. Marine and terrestrial ecosystems will also be affected by climate change, with implications for global food supplies.

6. How does the geological record inform our quantification of climate sensitivity?

The Equilibrium Climate Sensitivity, or ECS, is a key metric that encapsulates the response of the entire Earth system into a single number. It is defined as the global mean surface temperature change, given a doubling of atmospheric CO_2, once the system has reached equilibrium. Its value largely dictates the amount of warming the planet will experience for a given increase in greenhouse gas concentrations. It is a key input into many applications, including climate impacts assessments, and is used to inform government policy on climate change. As such, it is of great importance that the 'best-estimate' of ECS, and its uncertainty, are robustly constrained. ECS can be partly constrained by recent observations and through evaluating the strength of multiple Earth system feedback processes. However, key additional constraints come from measurements from the geological record.

The Earth's temperature response to an increase in CO_2 is determined by a complex range of feedbacks. Feedbacks acting over 10–100-year time-scales include water vapour, cloud cover, aerosols (including dust) and sea-ice/snow cover. Uncertainty in the strength of these fast-acting feedbacks creates uncertainty in the value of ECS. The IPCC (2014) gave a 66% probability that the ECS value was between 1.5 and 4.5°C. On the longer time-scales more familiar to geologists (>1000 years), additional feedback processes, including the growth and decay of ice sheets, ocean circulation changes and vegetation dynamics, all add to the ECS. Therefore, information from geological record has led to the concept of 'Earth System Sensitivity' (ESS), which can be used to describe the response of the Earth system to a doubling of atmospheric CO_2 on time-scales of 10 000– 100 000 years (Lunt *et al.* 2010; Rohling *et al.* 2012), and which is complementary to ECS.

Because the geological record contains the results of numerous climate 'experiments' associated with changes in CO_2, it provides an important means of estimating the value of ECS, contributing to our estimates of future warmth (e.g. Goodwin *et al.* 2018). To estimate ECS from the geological record, quantitative paired records of atmospheric CO_2 and global temperature from proxies are needed (Fig. 2; Box 1). Taking account of the offset between ESS (which is recorded in the geological record) and ECS (which informs climate policy), many studies of the geological past have provided support to the canonical range for ECS of 1.5–4.5°C (e.g. Skinner 2012;

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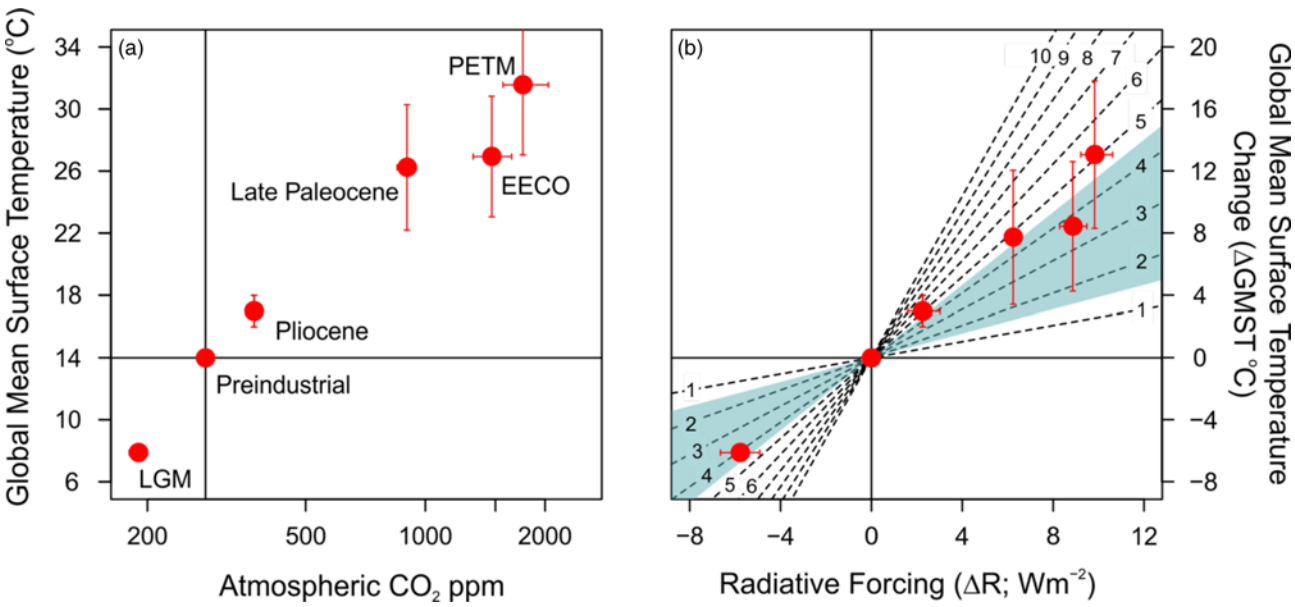

Fig. 2. (**a**) The relationship between atmospheric CO_2 concentration and global mean surface temperature (GMST) for five time intervals where both variables have been recently well constrained by geological data (Anagnostou *et al.* 2020; de la Vega *et al.* 2020; Inglis *et al.* 2020; Sherwood *et al.* 2020; Tierney *et al.* 2020). Error bars reflect 68% confidence intervals and in some cases are smaller than the symbol. LGM, Last Glacial Maximum; EECO, Early Eocene Climatic Optimum; PETM, Paleocene–Eocene Thermal Maximum. (**b**) The relationship between radiative forcing (ΔR in W m^{-2}) and global mean surface temperature (ΔGMST) relative to pre-industrial values. Contours show equilibrium climate sensitivity from 1 to 10°C per CO_2 doubling, and the blue band shows the canonical IPCC range of Equilibrium Climate Sensitivity of 1.5 to 4.5°C per CO_2 doubling. Note that data in panel (b) are corrected for slow feedbacks (see text) following the methods detailed in Tierney *et al.* (2020) and Inglis *et al.* (2020).

Martinez-Boti *et al* 2015; Anagnostou *et al* 2020; Inglis *et al* 2020; Tierney *et al* 2020; Fig. 2). Or, equivalently, our best estimates of past CO_2 change, and the canonical IPCC estimate of ECS, can explain the majority of the warming/cooling seen in the geological record. A recent study by the World Climate Research Program (Sherwood *et al.* 2020), which includes estimates from the geological record, has narrowed the uncertainty of ECS to between 2.6 and 3.9°C, meaning that 'low' ECS (<2°C per CO_2 doubling) is no longer consistent with our best modern and geological observations.

Overall, the geological record provides additional supporting evidence for the latest best-estimates of Equilibrium Climate Sensitivity and also allows us to estimate the importance of long-term feedbacks that are not straightforward to investigate using models or direct observations. With estimates of past radiative forcing and values of ECS between 2.6 and 3.9°C, we are able to explain the majority of the warming/cooling seen in the geological record. Through providing key constraints on ECS, the geological record is having an important input into policy decisions.

7. Are there past climate analogues for the future?

To understand current and future climate change, we can also look at intervals in the past when climate was similar to or warmer than today. These intervals are referred to as 'analogues'. There is no such thing as a perfect past analogue for our current climate or that of the future because the continents were not in the same location as they are now, which affects patterns of atmospheric and ocean circulation. Nevertheless, examining the forcings and responses of past warm climates provides us with important information about regions and processes that are sensitive to global warming. The geological record also reveals environmental responses operating across a variety of time-scales, so that climate-system feedbacks operating at scales longer than centuries can be understood. Our insights into the relative importance of different forcings, and their environmental responses, have come from both geological data and climate simulations of these past warm climates.

Recent interglacials of the Pleistocene (including the Holocene) provide key information concerning natural climate variability in the geologically recent past, where configurations of oceans and continents and the height of mountain belts were more or less similar to today. The relative importance of greenhouse-gas forcing v. orbitally forced insolation on temperatures varies, however, and in turn influences other climate parameters, for example sea-ice extent. For the slightly warmer interglacials of the last 1 million years, there is a strong influence of orbitally forced insolation on the global, high-latitude and seasonal expressions of warmth (Yin and Berger 2012), and warm interglacials can thus offer important regional analogues for a warmer-than-present climate. These intervals provide key information concerning natural past interglacial climate variability prior to the Industrial Revolution. The most recent interglacial witnessed warmth at both poles and provides a key window to ice-sheet and sea-level change (Dutton *et al.* 2015). For the future, when atmospheric CO_2 (and other greenhouse gas) levels are expected to continue to drive a temperature increase by perturbing Earth's radiative balance, we have to look further back in the past.

In the mid Pliocene (3.3–3.1 million years ago), atmospheric CO_2 concentrations ranged from 389 (–8 to +38) ppm to 331 (–11 to +13) ppm (de la Vega *et al.* 2020), which is higher than pre-industrial levels of about *c.* 280 ppm and slightly lower than modern levels (*c.* 407.4 ± 0.1 ppm in 2018). Earth's continental configurations, land elevations and ocean bathymetry were all similar to today (Haywood *et al.* 2016). The Pliocene was characterized by several intervals in which orbital forcing was similar to that of modern times and so it offers us a close analogue to the climate under modern CO_2 concentrations (McClymont *et al.* 2020). During this interval, global temperatures were similar to those predicted for the year 2100 (+2.6 to 4.8°C compared with pre-industrial) under a business-as-usual scenario (i.e. with no attempt to mitigate emissions). Several lines of work suggest similarities between the model-predicted

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ocean circulation of the future and that of the mid-Pliocene warm period, with a weaker thermohaline circulation, related to upper-ocean warming and stratification, but also reduced ice sheets and sea ice, a poleward shift in terrestrial biomes and weaker atmospheric circulation (Haywood and Valdes 2004; Cheng *et al.* 2013; Corvec and Fletcher 2017; Fischer *et al.* 2018). Pliocene sea-level may have reached 20 m above the present-day value and may have varied, on average, by 13 ± 5 m over Pliocene glacial–interglacial cycles, in association with fluctuations in the extent of the Antarctic ice sheet (Grant *et al.* 2019).

A possible analogue for a climate with even higher atmospheric CO_2 concentrations (1400 ± 470 ppm) is provided by the early Eocene climate optimum, *c.* 50 million years ago (Anagnostou *et al.* 2016). At that time, there was somewhat reduced solar output (Foster *et al.* 2017; Fischer *et al.* 2018), but there were no large continental ice sheets (and therefore a reduced planetary albedo). The Himalayan Mountains were underdeveloped and the overall configuration of the continents was unlike today, with an open Tethys Ocean, and absence of a strong Antarctic Circumpolar Current – affecting global climate. It is important to account for these differences when considering the Eocene as an analogue for our future climate, since they are likely to have impacted the sensitivity of climate to greenhouse forcing (Farnsworth *et al.* 2019). Nevertheless, palaeoclimate studies of the Eocene continue to provide useful insights into climate system dynamics, such as mechanisms for polar amplification in ice-free climate states (Cramwinckel *et al.* 2018).

Overall, the geological record can provide us with useful windows through which we can explore possible future climates and it informs us how the Earth works in different climate states.

8. How can the geological record be used to evaluate climate models?

Climate models are the primary tools used to make predictions about our future climate, a future in which CO_2 concentrations are expected to increase rapidly. These projections feed directly into bodies such as the IPCC and, as such, inform international policy through the Conferences of the Parties to the UN Framework Convention on Climate Change (UNFCCC), which is the body that agreed at its annual meeting in Paris in December 2015 that the guardrails for future warming should be set ideally at warming of 1.5°C above pre-industrial levels and, at the outside, at 2°C (UNFCCC 2015*a*); national government targets (such as Intended Nationally Determined Contributions; UNFCCC 2015*b*) are guided by this international advice.

Climate models are computer codes that represent our best understanding of the physical, chemical and biological processes that determine the operation of Earth's climate system. They are based on fundamental principles such as the Navier–Stokes equations of fluid flow and conservation equations for mass and energy. Key applications of these models are predictions of the future evolution of our climate under various scenarios of fossil-fuel use and land-use change. Evaluation of these models is typically carried out by comparing model results to global and regional recent well-observed climate changes of the last 100 years or so (e.g. Williams *et al.* 2010; Sellar *et al.* 2019), but future changes in carbon dioxide concentrations could be an order of magnitude greater than seen over this period (Meinshausen *et al.* 2020).

Geological data provide both qualitative and quantitative evidence of past climate change under high greenhouse gas concentrations and so provide a crucial out-of-modern-sample evaluation of climate models, providing testbeds that in some instances are similar to those expected for the future (Section 7). Additionally, because climate models are based on fundamental physical principles, the geological record can also provide quantitative evidence of the response of the climate system to drivers other than greenhouse gases, such as orbital variations or plate tectonic changes, against which the models can be independently evaluated (e.g. DeConto and Pollard 2003). This key role for geological data in model evaluation is increasingly being recognized by the major international modelling centres (such as the Met Office in the UK), as evidenced by the prominent role of palaeoclimate in the most recent and forthcoming reports from the IPCC (Masson-Delmotte *et al.* 2013).

The mid-Pliocene provides evidence of sea-level change under high atmospheric CO_2 concentrations, which can be used to evaluate the behaviour of ice sheets simulated by ice-sheet models (e.g. DeConto and Pollard 2016; Gasson and Kiesling 2020). The key aspect here is that, because ice sheets respond to climate change on long time-scales, the geological record is essential for giving a long-term perspective that is completely absent from, for example, satellite records of recent ice-sheet changes. In this recent work, the geological record has been used to identify the models that perform best in the deep past and to apply only these 'geologically consistent' models to the future.

A long-standing discrepancy between models and data has been the failure of models to reproduce the amount of warming towards the poles during periods of super-high CO_2 concentrations (*c.* 1000 ppm) during the warmest periods of the last 100 million years (Barron 1987). In particular, the early Eocene has proven to be a particular challenge for models, which have previously under-estimated the amount of warming seen in the geological record (Lunt *et al.* 2012). However, recent work has shown that developments in representing the properties of clouds act to amplify the modelled warming towards the poles, bringing the models into agreement with the data (Zhu *et al.* 2019).

Nonetheless, some persistent discrepancies remain. For example, the geological record shows that during the mid-Holocene, North Africa was much wetter than today and vegetation thrived in regions of the modern Sahara desert (Section 5). Current state-of-the-art climate models generally underestimate the extent of these observed changes (e.g. Williams *et al.* 2020), which were paced by orbital parameters, but involved other feedbacks including atmosphere–ocean, atmosphere–land and dust interactions (Hopcroft and Valdes 2019; Pausata *et al.* 2020). Identifying and simulating all relevant feedbacks is an ongoing target in climate-change research and it is hoped that future improvements in models, such as their representation of atmospheric convection, will reconcile these differences. In the meantime, in the majority of instances, models have *under*-estimated the changes seen in the geological record (e.g. Valdes 2011; Gasson and Kiesling 2020; Pausata *et al.* 2020; Williams *et al.* 2020).

Overall, recent successes in model evaluation using the geological record increase our confidence of future climate predictions from models. Indeed, the geological community uses climate models in applications such as oil and gas exploration (e.g. Markwick 2019) and the long-term geological storage of high-level radioactive waste (e.g. Lindborg *et al.* 2018).

9. What is the role of geology in dealing with the climate emergency for a sustainable future?

Geology contributes to understanding how we live on Earth, where and which resources we derive from it and how we discard waste into it. Geoscientists study the soil from which we grow crops, the aquifers from which we extract water and the resources from which we obtain energy and minerals. We study the risks associated with living on a dynamic planet and, hence, geoscientists will play a key role in moving society towards a sustainable low carbon future.

One of the geoscience community's most important contributions to the decarbonization necessary to deal with the climate

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emergency will be to discover the resources that will power post-fossil-fuel energy systems. Large-scale investment in renewable energy and improved electricity and heat storage will require new resources of critical metals and raw materials from the Earth's crust (Vidal *et al.* 2013). However, many of these resources are currently being found and mined at rates far too slow to support a global energy revolution (Natural History Museum 2019). Geoscientists have the vital skills needed to assess the distribution, concentration (grade) and sustainable extractability of the critical raw materials needed for decarbonization.

Other sources of energy, from nuclear to deep geothermal, both of which could be critical for the decarbonization, require geoscientific expertise, which is also indispensable for carbon capture and sequestration (Matter *et al.* 2016). Natural resource extraction and processing itself is a major contributor to greenhouse gas emissions (IRP 2019). There is a key role for geoscientists in making the discovery and extraction of these resources as efficient as possible in order to mitigate global warming.

Links between climate change and the UN Sustainable Development Goals are shown in *Geology and the UN Sustainable Development Goals* (Geological Society 2014). Many of these links revolve around challenges related to georesources. Consequently, our roles include ensuring that discovery and extraction of resources is carried out responsibly, fairly and sustainably.

Geoscientists are making vital contributions to all of the UN's Sustainable Development Goals and that includes the human response to climate change and its impacts. Although we cannot and should not address these challenges alone, we are well placed to ensure they are framed critically and robustly, recognizing the opportunities and limitations afforded by the planet on which we live. Geoscientists will play an increasingly important role in the transition to a low carbon, green economy which is necessary to prevent a worsening of the climate emergency.

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