Kick-starting ancient warming

Rapid global warming marked the boundary between the Palaeocene and Eocene periods 55.6 million years ago, but how the temperature rise was initiated remains elusive. A catastrophic release of greenhouse gases from the Kilda basin could have served as a trigger.

The warm period known as the Palaeocene–Eocene Thermal Maximum (PETM) was a period of climatic turmoil that lasted more than 100,000 years[1]. Ocean temperatures increased by 3–10 °C, and atmospheric concentrations of methane and carbon dioxide rose sharply[1]. Meanwhile, the oceans became more acidic and the carbon isotopic values of marine carbonates declined significantly, a sign of a disrupted carbon cycle. These latter observations suggest that several thousand gigatonnes of isotopically light carbon must have been injected into the atmosphere–ocean system over the course of the event[1,2]. Sources including the release of methane during the intrusion of magma into organic-rich sedimentary rocks, carbon dioxide outgassing from lava flows, and dissociation of methane gas hydrates in marine sediments[2–4] have all been previously invoked in the perturbation — but we still do not know how much light carbon came from each source.

We are also unsure just what triggered the initial warming[1–4]. More data on the environmental conditions of the PETM are becoming available, and it is now clear that the warming began several thousand years before the main disruption to the carbon cycle that is documented in the worldwide decrease in carbon isotope values[5] often used to define the event. Therefore, the initial warming was probably a necessary precursor of the upheaval of the carbon cycle. Because, over a 20-year period, methane is 72 times more potent than carbon dioxide as a global warmer, a rapid injection of a relatively small amount of methane into the atmosphere could have caused the initial warming without leaving an identifiable isotopic signature. Mechanisms that have been suggested to provide such a methane burst include seabed hydrothermal activity and large submarine landslides[3,6].

Here we propose an alternative: we suggest that the early warming of the PETM was initiated by catastrophic expulsions of methane and also carbon dioxide that had accumulated over time in the stratified waters of the Kilda basin between Greenland and Norway, in a mechanism similar to that seen today in African rift lakes.

Triggering early warming

Long-term global warming as seen during the PETM requires not only large but also sustained radiative forcing. If we assume that the lifetime of methane in the atmosphere at the time was not too dissimilar from the roughly 10 years of today, the climatic effect of isolated natural methane outbursts would have declined drastically after a decade or so, and the effects on long-term global climate would have been limited. The same limitation applies to carbon dioxide, although its lifetime is longer at timescales of centuries to millennia. A recurrent release of greenhouse gases is therefore required to explain the much longer-term warming in the PETM.

The period between gas release events (repeat time) needs to be comparable to, or shorter than, the atmospheric residence time of the warming gas, otherwise the warming effect of one release event will fade before the next event occurs. Repeat times for previously suggested natural methane releases are controlled by either oceanic circulation or by plate tectonic and mantle convection processes, but all these repeat times are in the region of centuries to millennia (see Box 1). Therefore a methane burst capable of triggering the PETM is needed to cause warming over a period of well over a century. During this initial period immediately after the trigger, the atmospheric methane residence time must have increased manyfold. Subsequently, the atmosphere could sustain a much larger load of greenhouse gases, possibly sourced from more than one major reservoir, which maintained the warming and generated the carbon isotope excursion throughout the PETM period.

Box 1 | Possible triggers for the Palaeocene–Eocene Thermal Maximum

Kilda gas expulsion. Overturning and release of the dissolved gases in the lower water masses of the Kilda basin could release between 69 and 113 Gt of carbon as methane and 285–477 Gt of carbon as carbon dioxide, assuming a basin volume of 300,000–500,000 km^3. Expulsion events could recur within centuries or millennia.

Sediment slumping. Sediment movement could disturb and release methane clathrates at the sea floor. A slump area of 100,000 km^2 has the potential to release up to 10 Gt of carbon as methane, but only negligible amounts of carbon dioxide. On the basis of the length of time it takes for warm surface waters to reach and destabilize the sea bed, we estimate that warming could induce new slumps within millennia.

Hydrothermal venting. Hydrothermal systems that formed above the magmatic intrusions associated with the North Atlantic Igneous Province could release volcanic and heat-generated gases into the water column. On the basis of the number of vent complexes documented in the Norwegian section of the province, we estimate that each intrusion event could have released 0.3–3 Gt and up to 0.1 Gt of carbon as methane and carbon dioxide, respectively. The estimated duration of igneous activity suggests a repeat time of 10–400 years.

Submarine lava flow. Lava flows onto the sea floor could have released only small amounts of methane, owing to rapid cooling of the lava when in contact with sea water, and from 0.009 to 0.2 Gt of carbon as carbon dioxide, assuming similar gas concentrations to those in modern lava flows. Flow events could recur within years or centuries.

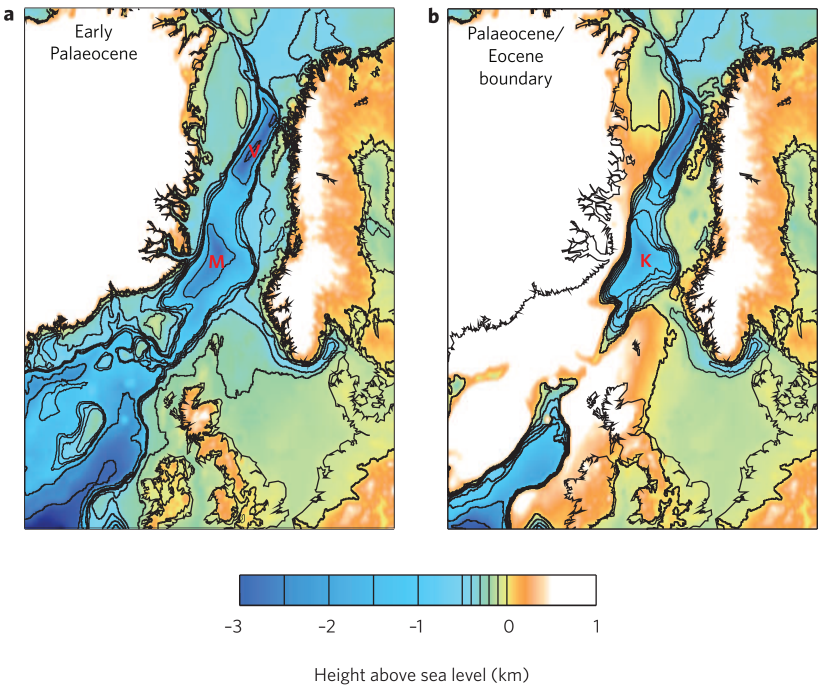

Figure 1 | The evolution of the Kilda basin. **a**, Before uplift, deep and intermediate waters flowed freely between the Vøring (V) and Møre (M) regions and the open ocean. **b**, As volcanic activity severed the connection between the Kilda basin (K) and the ocean, the hydrothermal input of brine increased the density of the bottom waters and prevented mixing with the shallower surface waters entering the basin. Methane and some carbon dioxide produced by volcanic activity and the breakdown of organic matter began to build in the deep waters. We propose that the overturning of these waters could have released these greenhouse gases to the atmosphere, thereby triggering the Palaeocene–Eocene Thermal Maximum.

Methane hydrates in marine sediments have been implicated in maintaining the PETM warming[2], but it is not clear whether they could also have provided the trigger. Methane in this reservoir is released by catastrophic sediment slides on the continental slope, or from progressive degassing[6]. Yet it is difficult to sustain a succession of such events through a positive feedback effect: it takes a millennium or more for the atmospheric warming from each methane pulse to return to deep ocean waters where it can mobilize the hydrates buried beneath the seabed[6–8], far longer than the atmospheric residence time of the original methane. For seabed hydrates to be able to trigger the PETM, the initial catastrophic sediment slides must have been forced to recur much more rapidly, at intervals of about a decade, and this period of forced repetition must have been sustained over perhaps a millennium, until additional methane from mobilized hydrate could maintain the new warm climate[9] (see Supplementary Information).

Methane release from hydrothermal vents has also been suggested as the warming trigger[3]. During the late Palaeocene and early Eocene epochs, magma from the North Atlantic Igneous Province intruded through sedimentary rocks rich in organic matter. As the rocks were baked by the hot magma, methane was produced and then released through hydrothermal vents above the active intrusions. This methane probably contributed to the prolonged warming[3]. But the repeat time of gas pulses from these vents depended on the rate of magma generation and was most probably greater than a century (see Supplementary Information).

Large bursts of methane entering the atmosphere could have increased the atmospheric lifetime by decreasing the concentrations of the OH radicals that oxidize methane in today's atmosphere. But releases of only a few billion tonnes of methane — the largest pulses expected from either methane hydrates or hydrothermal vents — could only extend methane residence times to about 20 years[7] (see Box 1), which is much shorter than the repeat times for such events. In view of this, we conclude that hydrothermal activity or submarine landslides are unlikely to have triggered the PETM.

The difficulty of decadal repeat times, and thus a necessarily high frequency of events, can be overcome by drastically increasing the mass of the initial methane burst. A release significantly exceeding ten billion tonnes, delivered in less than a decade, could increase methane's normal decadal lifetime by over half a century. One potential source of such a catastrophic release is a large, anoxic water body.

The Kilda basin capacitor

During the late Palaeocene to early Eocene, a suitable enclosed marine basin for the storage of such enormous amounts of methane briefly existed between Norway and Greenland[10] (Fig. 1), which we call the Kilda basin. By analogy with the lake overturning events observed in the modern stably stratified lakes in the African great rift zone, we suggest that the PETM was triggered by pulses of methane and carbon dioxide released from this basin. Like energy stored in a capacitor, greenhouse gases would have been stored (up to the point of sudden and catastrophic release) in the bottom layer of the basin, sequestered from the atmosphere through a stable stratification of the waters in the enclosed water body.

The Kilda basin formed as the sea floor rose rapidly in isostatic response to the emplacement of hot mantle beneath the region[10] (Fig. 1). This was the same mantle convection event that generated the voluminous North Atlantic Igneous Province magmas. The related uplift of 0.5–1.5 km grew throughout the Late Palaeocene and, shortly before the Palaeocene/Eocene boundary, severed the deep-water connection with the Atlantic in the vicinity of the Faroe–Shetland basin. Subsequent decay of the mantle thermal anomaly lowered the sea floor through the early Eocene, gradually allowing Kilda's deep water to circulate (Supplementary Information). The Kilda basin was comparable in dimensions to the modern Red Sea. Also like the Red Sea, the basin probably experienced periods as both a silled basin with a connection to the open ocean and an isolated lake, depending on the relative sea levels.

In its tectonic and biogeographic setting, the Kilda basin resembles the strongly stratified great lakes of the African rift, such as Malawi and Tanganyika. Although smaller in scale, the rift lakes occupy deep basins, some more than 1 km deep, which is similar to the reconstructed depth of Kilda. They are abundantly supplied with organic debris from surrounding forests, just as Kilda was flanked by moist mountains covered in rich forests[11]. They also experience lake surface temperatures of 20–30 °C, not dissimilar to the temperatures of 18–23 °C reported for sites north of Kilda[12].

The water masses within some African great rift lakes do not mix vertically under normal circumstances. As a result, lakes Kivu and Nyos are known to store abundant methane and carbon dioxide, which can be released in catastrophic events in which the layers of the lake overturn (Supplementary Information). For the analogy to work, the Kilda water body must have been similarly stably stratified, after the documented interruption of the deep-water connection between the basin and the Atlantic. Looking to a modern setting, the virtually enclosed Black Sea is stably stratified, with about 1.5 times the estimated deep-water volume of Kilda.

In both the modern African lakes and the Black Sea, the vertical distribution of salt is crucial in controlling stratification. Supplies of fresh river run-off to the surface waters and hydrothermal brine to the bottom waters generate the large density differences between the bottom and upper water masses that make stratification permanent. The Kilda basin also had both types of water influx. Large river systems drained Greenland and Norway, bringing fresh water to the surface, and more than 700 hydrothermal vents dated to the latest Palaeocene period have been mapped in the Vøring and Møre regions of the Kilda basin alone[3] (Fig. 1), covering perhaps one-quarter of Kilda's deep-water area. This is suggestive of a greater flux of brine than is found today in the great African lakes. Sediment records recovered from the Faroe–Shetland basin, which lay within the ancient Kilda basin (Fig. 1), show that the bottom waters became isolated from the Atlantic and turned anoxic during the latest Palaeocene[13], implying stratification. Unfortunately, the regional uplift that formed Kilda also led to the erosional removal of PETM sediments in basins where cores are available; consequently we lack direct evidence for fluctuations in anoxia that could be associated with surface water overturns at the beginning of the PETM.

There were two sources of methane to the deep waters of the Kilda basin (see Supplementary Information). Organic debris delivered by rivers from its thickly forested flanks[14] would have supported vigorous methanogenesis as microbes broke down the organic matter. Meanwhile, magma of the North Atlantic Igneous Province generated methane by flowing as lava over organic-rich mud at the seabed itself, as well as by intruding into and baking organic-rich mudrocks kilometres beneath the seabed to supply the hydrothermal vents[3]. Melting of local gas hydrates would not necessarily be required, although hydrate dissociation driven by the falling pressure at the sea floor as it was raised may also have contributed[10]. Carbon dioxide would have been produced in generous quantities in the deep water as anaerobic microbes oxidized methane and from hydrothermal systems. As in modern Lake Kivu, most of the gases released from Kilda's bed would have been trapped in the lower layers of the water column, and Kilda's deep water would have become increasingly saturated with these dissolved gases.

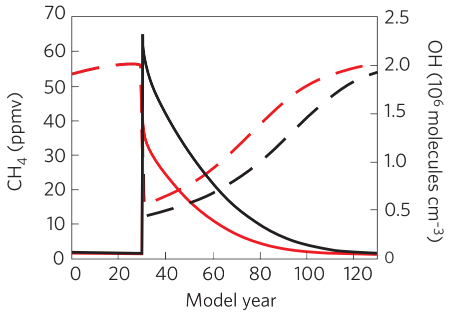

Figure 2 | The fate of Kilda's methane. Modelled CH_4 mixing ratios (solid lines) and OH concentrations (dashed lines — initial curve is model settling to equilibrium). Our model simulation shows that the initial burst of 150 Gt (black) or 90 Gt (red) of methane greatly reduces the quantity of OH radicals in the atmosphere. This increases the atmospheric lifetime of the methane, allowing it to persist at significantly higher levels for decades before decaying to background levels after about 70 years.

Considering uncertainties in the depth of the rift centre and the dynamic support model, the volume of the Kilda deep isolated water could have been between 300,000 and 500,000 km^3 (Fig. 1). The deep water of modern Lake Kivu contains 20 mmol kg^{-1} methane and 80 mmol kg^{-1} carbon dioxide (Supplementary Information). If we assume comparable concentrations for Kilda, roughly 0.3 Mt km^{-3} of methane and 3.5 Mt km^{-3} of carbon dioxide could have been stored within the deep waters of the basin. Total dissolved gas content would have thus have been on the order 90–150 Gt of methane and 1,050–1,750 Gt of carbon dioxide. Because Lake Kivu is not supersaturated in dissolved gases at the ambient pressures in its deep water, it is possible that the maximum gas content in the Palaeocene Kilda basin could have been even higher.

Kilda, like Lake Malawi[15], must have overturned periodically before starting a new charging cycle. Degassing events would have had a range of possible triggers, from submarine landslides to massive hydrothermal events or lava flows. These triggers would have spurred catastrophic gas eruptions. By analogy with Lake Nyos, the resultant eruptive fountain from Kilda may have reached a height of hundreds of metres[16].

The impact on atmospheric chemistry

We are concerned only with the trigger of the PETM, not with sustaining a warm climate over the full ~100,000–200,000 years of its duration. An estimate of the climate impact of a single large injection of methane from the Kilda basin into the atmosphere comes from the Cambridge two-dimensional global atmospheric model[17] (Supplementary Information). In our model simulations, 90 or 150 Gt of methane are released at 60° N into the Palaeocene model atmosphere, a smaller input than the 1,500 Gt inferred by Renssen *et al.*[18]. The carbon dioxide release is also assessed, but has only a minor impact for a single pulse. The methane pulse initially overwhelms OH radicals in the atmosphere, increasing the atmospheric lifetime of methane from ~10 to ~60 years initially in the 150-Gt release scenario (Fig. 2). The burst raises the concentrations of methane in the lower atmosphere by roughly a factor of 40, to ~65 ppm by volume (ppmv). This initial mixing then begins to decay back, and takes more than a century to return to background levels (Fig. 2).

The simulated climate response is determined by the radiative forcing that arises from the magnitude and duration of the methane pulse. In the scenario in which 150 Gt of methane is released into a low-CO_2 Palaeocene atmosphere, the total globally averaged radiative forcing would have risen to more than 10 W m^{-2} shortly after the Kilda pulse, dominated by the methane component. (To put the figure into context, radiative forcing from anthropogenic greenhouse gas release has at present reached 1.6 W m^{-2} over a period of 250 years[19].) Half a century after the pulse, the total radiative forcing could still have exceeded 5 W m^{-2}. In the second half of the century after the pulse, as methane was removed from the atmosphere, forcing by carbon dioxide would have become more important; after a century, the only significant climate forcing left would arise from carbon dioxide, with a value of up to 3 W m^{-2}. The very significant carbon dioxide forcing mostly originates from the vast amount of carbon dioxide injected directly by a single Kilda burst (see Box 1). The other triggers outlined in Box 1 could release at most only about one-tenth of the mass of methane and 1/1,000 of the carbon dioxide compared with that released by a Kilda burst.

Thus a single large gas expulsion from Kilda could have sustained an average 5 W m^{-2} radiative forcing for a century. Such a forcing would have resulted in average global warming of up to several degrees, with amplification at high latitudes. We considered a single giant outburst as a limiting case. Comparisons with African lakes suggest that the Kilda capacitor, with high inputs of organic debris, could have been charged and discharged multiple times in relatively short succession, perhaps on a decadal scale or less if Kilda, like modern Lake Kivu, had multiple sub-basins and widespread igneous activity.

Both the capacitor and the charge persisted into early Eocene times (Supplementary Information). Repeated methane discharges would prolong the atmospheric methane loading and consequently the radiative forcing. Kilda methane discharges could also have forced the climate through indirect effects. For example, H_2S also dissolved in the Kilda water could have reached the stratosphere if the eruption was violent. There it would have caused a thick sulphur smog that blocked ultraviolet radiation, reduced the OH concentration, and thereby further extended the atmospheric lifetime of CH_4 (see Supplementary Information). More speculatively, the persistence of Kilda capacitor conditions through the early Eocene means that it could have been involved in triggering some of the less extreme global warming events that succeeded the PETM[20].

Unlike other suggested triggers, bursts of methane and carbon dioxide from Kilda could have been large enough, and could have been repeated frequently enough, to initiate the persistent global warming throughout the PETM. Could the comparable injection of modern anthropogenic emissions induce the same response from the planet? ❐

References

1. Zachos, J. C. *et al. Science* **308,** 1611–1615 (2005).
2. Dickens, G. R. *Earth Planet. Sci. Lett.* **213,** 169–183 (2003).
3. Svensen, H. *et al. Nature* **429,** 542–545 (2004).
4. Storey, M., Duncan, R. A. & Swisher, C. C. *Science* **316,** 587–589 (2007).
5. Sluijs, A. *et al. Nature* **450,** 1218–1221 (2007).
6. Nisbet, E. G. *Phil. Trans. R. Soc. A* **360,** 581–607 (2002).
7. Thorpe, R. B., Law, K. S., Bekki, S., Pyle, J. A. & Nisbet, E. G. *J. Geophys. Res.* **101,** 28627–28636 (1996).
8. Nisbet, E. G. *Can. J. Earth Sci.* **27,** 148–157 (1990).
9. Bice, K. L. & Marotzke, J. *Paleoceanography* **17,** doi:10.1029/2001PA000678 (2002).
10. Maclennan, J. & Jones, S. M. *Earth Planet. Sci. Lett.* **245,** 65–80 (2006).
11. Collinson, M. E., Hooker, J. J. & Grocke, D. R. in *Causes and Consequences of Globally Warm Climates in the Early Paleogene* (eds Wing, S. L., Gingerich, P. D., Schmitz, B. & Thomas, E.) 333–349 (Geol. Soc. Am. Spec. Pap. 369, Boulder, Colorado, 2003).
12. Sluijs, A. *et al. Nature* **441,** 601–605 (2006).
13. Mudge, D. C. & Bujak, J. P. *Mar. Petrol. Geol.* **18,** 577–590 (2001).
14. Moran, K. *et al. Nature* **441,** 601–605 (2006).
15. Pilskaln, C. H. & Johnson, T. C. *Limnol. Oceanogr.* **36,** 544–557 (1991).
16. Zhang, Y. *Nature* **379,** 57–59 (1996).
17. Warwick, N. J., Bekki, S., Nisbet, E. G. & Pyle, J. A. *Geophys. Res. Lett.* **31,** L05107 (2004).
18. Renssen, H., Beets, C. J., Fichefet, T., Goosse, H. & Kroon, D. *Paleoceanography* **19,** PA2010 (2004).
19. IPCC. *Climate Change 2007: The Physical Science Basis* (eds Solomon, S. *et al.*) (Cambridge Univ. Press, 2007).
20. Nicolo, M. J., Dickens, G. R., Hollis, C. J. & Zachos, J. C. *Geology* **35,** 699–702 (2007).

Additional information

Supplementary Information accompanies this paper on www.nature.com/naturegeoscience.

E. G. Nisbet[1], S. M. Jones[2]*, J. Maclennan[3], G. Eagles[1], J. Moed[4], N. Warwick[4], S. Bekki[5], P. Braesicke[4], J. A. Pyle[4] and C. M. R. Fowler[1] are at [1]Department of Earth Sciences, Royal Holloway, University of London, Egham TW20 0EX, UK, [2]Department of Geology, School of Natural Sciences, Trinity College Dublin, Dublin 2, Ireland, [3]Department of Earth Sciences, University of Cambridge, Downing Street, Cambridge CB2 3EQ, UK, [4]Centre for Atmosphere Science, Department of Chemistry, University of Cambridge, Lensfield Road, Cambridge CB2 1EW, UK, [5]LATMOS-IPSL UPMC, B.102, Tour 45–46, 3ème étage 4, place Jussieu, 75252 Paris Cedex 05, France. *e-mail: e.nisbet@es.rhul.ac.uk; stephen.jones@tcd.ie*