Alpine permafrost could account for a quarter of thawed carbon based on Plio-Pleistocene paleoclimate analogue

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Estimates of the permafrost-climate feedback vary in magnitude and sign, partly because permafrost carbon stability in warmer-than-present conditions is not well constrained. Here we use a Plio-Pleistocene lacustrine reconstruction of mean annual air temperature (MAAT) from the Tibetan Plateau, the largest alpine permafrost region on the Earth, to constrain past and future changes in permafrost carbon storage. Clumped isotope-temperatures ($\Delta_{47}$-T) indicate warmer MAAT (~1.2 °C) prior to 2.7 Ma, and support a permafrost-free environment on the northern Tibetan Plateau in a warmer-than-present climate. $\Delta_{47}$-T indicate ~8.1 °C cooling from 2.7 Ma, coincident with Northern Hemisphere glacial intensification. Combined with climate models and global permafrost distribution, these results indicate, under conditions similar to mid-Pliocene Warm period (3.3–3.0 Ma), ~60% of alpine permafrost containing ~85 petagrams of carbon may be vulnerable to thawing compared to ~20% of circumboreal permafrost. This estimate highlights ~25% of permafrost carbon and the permafrost-climate feedback could originate in alpine areas.
Permafrost, or permanently frozen ground, underlies less than 20% of the Earth’s land area. However, soils in the permafrost zone store ~1500 Pg (petagrams, 10^{15} grams) of organic carbon, representing ~50% of global soil organic carbon (SOC) and nearly twice as much carbon as contained by the Earth’s atmosphere. While most modern permafrost is located in circumarctic regions, including subsea deposits on the continental shelves, permafrost extends through alpine areas at lower latitudes (Fig. 1a). Most of this alpine permafrost occurs on the Tibetan Plateau, which is sometimes referred to as Earth’s “Third Pole”, containing a globally significant stock of ~160 Pg of SOC. Monitoring and modelling studies reveal that across the globe, permafrost is thawing rapidly, giving rise to a potential permafrost-climate feedback. Once organic matter in permafrost is thawed, it can be decomposed by microorganisms, producing the greenhouse gases carbon dioxide (CO₂), methane (CH₄), and nitrous oxide (N₂O). However, fundamental uncertainties persist about the response of permafrost carbon to climate change, and modelling studies currently disagreeing on the magnitude and sign, of the permafrost-climate feedback. In particular, determining regions (circumarctic versus alpine regions) where permafrost carbon is more sensitive to global warming is urgently needed.

The distribution of permafrost during past climatic warm periods could shed light on the near-future response of this system to anthropogenic forcing. Geological and modelling studies indicate that the Earth was warmer than present during the mid-Pliocene Warm Period (mPWP; 3.3–3.0Ma), with atmospheric CO₂ concentrations near 400 ppmv (parts per million by volume). The mPWP is often used as a geological analogue for the near future, to study climate change impacts on sea level and extent of glaciers and ice sheets. Considering that global climate will soon reach mPWP-like conditions if current anthropogenic warming continues, paleoecological reconstruction for the mPWP and Plio-Pleistocene in regions that today contain permafrost can provide insights into the stability of circumarctic and alpine permafrost. Although over the last few decades, our knowledge of the mPWP and Plio-Pleistocene paleoclimacross the globe has improved significantly, little is known about the climate history of regions that today contain permafrost, including both circumarctic and alpine permafrost zones. The lack of comprehensive paleoenvironmental studies in modern permafrost-containing regions also represents an important knowledge gap about the response of high elevation and high latitude regions to climate change.

Multiple geochemical proxies now allow detailed reconstruction of paleoclimate from marine and lacustrine sediments. Carbonate stable isotope (δ^{18}Ow) and δ^{13}C organic carbon (TOC), total organic carbon (TN), carbon to nitrogen ratios (C/N), and grain size have been widely used for paleoenvironmental reconstruction as these proxies provide qualitative constraints on temperature, evaporation-precipitation balance, primary productivity, erosion and other climate conditions that might indirectly relate to temperature. One particularly promising temperature proxy is the carbonate-clumped isotope (Δ_{47}), thermometer, which can provide robust paleo-temperature constraints independent of the oxygen isotopic composition of water (δ^{18}Ow). Given that the persistence of permafrost depends on temperature-ecosystem interactions, reliable reconstructions of mean annual air temperatures (MAAT) from Plio-Pleistocene sediments would not only enable us to assess the presence of permafrost during the mPWP warmth, but also projects the modern permafrost regions that might be particularly vulnerable to future global warming.

Here we estimate the potential area of global permafrost thaw, and associated organic carbon in a warmer-than-present climate, by integrating paleoclimate reconstructions from geological records with climate models to explore the implications of past changes in temperature on permafrost distributions. We used the Pliocene Model Intercomparison Project 2 (PlioMIP 2) model

Fig. 1 Global map showing modern mean annual air temperature, the distribution of permafrost, and inset with the location of the study area. a Modern mean annual air temperature map showing circumarctic and alpine permafrost. b Topographic map of the Tibetan Plateau and surrounding regions showing the Kunlun Pass section on the northern Tibetan Plateau. QB: Qaidam basin, EKLM: Eastern Kunlun Mountains, WB: Weihe Basin. Note that regions north of 50°N are defined as middle to high latitude circumarctic regions. Region marked with 1–6 refers to the Tibetan Plateau-Pamir, Altai Mountains-Mongolia-Yablonoi-Sayan, Tian Shan, Rocky Mountains, Alps, and Caucasus, respectively. The image of landform made with GeoMapApp 3.6.10 (www.geomapapp.org)/CC BY.
ensemble, which stands out for its ability to provide first-order global climatic conditions for the mPWP, to further estimate the stability of modern permafrost in a global context during the warmer climate of the mPWP. Specifically, we first reconstruct paleotemperature and other paleoenvironmental conditions in an alpine permafrost region on the northern Tibetan Plateau for the Pliocene to Pleistocene (from 4.3 to 0.8 Ma) from lacustrine records near Kunlun Pass (KP) (Fig. 1b). For this work, a total of 344 samples were collected and analysed for CaCO₃, grain size, TN, TOC, C/N ratio, and δ¹³Corg. A subset of 275 samples was also analysed for carbonate δ¹⁸O and δ¹³C analyses. 57 samples were analysed for Δx, composition, which was used to estimate local MAAT, in conjunction with the other proxies. Our results reveal that a stepped cooling event occurred during the Pliocene-Pleistocene transition with warmer temperatures (MAAT > 0 °C) during the Pliocene in the northern Tibetan Plateau, indicating a potentially permafrost-free environment on the margins of the Tibetan Plateau. Using this inference as a guide, we then derived a mPWP global MAAT and mapped out the 0 °C isotherm using results from the Pliocene Model Intercomparison Project 2 (PlioMIP 2)²⁴, which we also compared with the observed paleotemperature records from the KP site, to assess the stability of permafrost across the globe in a warmer-than-present world.

By combining these estimates with the modern and pre-industrial global MAAT record, the modern global distribution of permafrost, and reported SOC density, we finally estimate the potential area of global permafrost thaw, and associated organic carbon in a warmer-than-present climate. Our assessment of global permafrost stability highlights that alpine permafrost and the associated carbon pool could be disproportionately important in determining the global permafrost-climate feedback.

Results and discussion
Stratigraphy and age model of the Kunlun Pass Section. The KP site is located on the northern Tibetan Plateau (35°39’N, 94°03’E, elevation ~4.7 km) south of the Eastern Kunlun Mountains that separate the modern cold steppe environments (Supplementary Fig. 1a) from the hyper-arid desert climate of the Qaidam Basin to the north²⁵ (Fig. 1b). Modern glaciers are well-developed in the surrounding mountain ranges²⁵. The sequence of the KP section is dominated by laminated dark grey calcareous mudstone interbedded with sparse greyish siltstone and sandstone (Supplementary Fig. 1b–d), that have been interpreted as deep-lake deposits²⁶,²⁷.

This deep lake shrank and finally vanished in response to regional climate change during Late Cenozoic aridification of Asia²⁶, with subsequent exhumation of the Eastern Kunlun Shan tilting the Plio-Pleistocene sequence of the KP section and facilitating field sampling²⁷.

As shown on the thin sections (Supplementary Fig. 2), both endogenic carbonate (micrite) and alloogenic (detrital) carbonates occur in the KP deposits. Detrital carbonate inputs are generally >200 µm while micrites are smaller (<200 µm) in grain size. Detrital carbonate inputs in the KP section are mainly derived from the Permain-Triassic carbonate-rich basement rocks in the surrounding mountain belts (Fig. 1), with low δ¹⁸O, values –11 to –14‰, VPDB²⁹. Detrital carbonate fragments come from basement rocks that have undergone thermal events at higher than surface temperature, thus, they should show higher clumped isotope (Δx) temperatures and more negative δ¹⁸O values if they recrystallised in the presence of water of similar δ¹⁸Ow composition to surface waters³⁰. Based on the correlation between the Δ¹⁸O and Δx-derived temperature, we developed an approach to minimise the potential effect of detrital carbonates contamination on carbonate isotopic records (carbonate δ¹⁸O, δ¹³C, Δx) (see “Methods”; Supplementary Fig. 3).

Published magnetostratigraphy²⁷ and biostratigraphy²⁶ enabled the development of an age model using piecewise linear interpolation (Supplementary Fig. 4). The depositional age of the strata in the section span from 4.3 to 0.8 Ma. Spectral analysis of carbon and oxygen isotopes, as well as carbonate content records using the untuned age model resolved eccentricity, obliquity, and precession-scale variations (Supplementary Fig. 5), giving us confidence in the age estimates.

Multi-proxy framework utilised. Carbonate δ¹⁸O has been widely used as an indicator of the variability in the precipitation/evaporation balance of lakes¹⁸, while carbonate δ¹³C provides important information on the composition of dissolved inorganic carbon (DIC) and reflects biologic productivity as well as mixing of DIC from different sources (i.e., lake waters, groundwater, and the atmosphere)³¹. For a closed lake basin system like the KP site, carbonate δ¹⁸O and δ¹³C usually show strong covariance (r > 0.7)¹⁹, while the sediment CaCO₃ content should reflect changes in lake water CO₂ due to biological activity (photolith-ynthesis, respiration, and production of organic matter) or temperature³². The concentration of TOC in sediments should reflect both autochthonous and allochthonous sources³³. C/N ratios of organic matter in lake sediments are widely used to distinguish between land-plant and aquatic origins of sedimentary organic matter: algae and cyanobacteria typically have C/N ratios between 4 and 10 whereas vascular land plants have C/N ratios of over 20³⁴, though this ratio can be modified somewhat by decomposition³⁵. When aquatic productivity is the dominant source of organic matter in a closed lake basin, increases in sediment TOC and TN, as well as a decrease in C/N, imply an increase in primary productivity probably associated with climate warming³⁶. The δ¹³Corg from organic matter in lake sediments can also indicate the source of organic matter and climatic conditions³⁴. Recent studies of Tibetan alpine lakes show that measured δ¹³Corg values from lake sediments can be related to the distribution of aquatic macrophytes, which are sensitive to water depth³⁷. Submerged macrophytes with high δ¹³Corg values are dominant in shallow lakes where light penetrates to the lakebed, whereas algae with relatively low δ¹³Corg values dominate in deep lake environments, driving a negative correlation between δ¹³Corg and lake level for lakes with primarily autochthonous (internally produced) organic matter³⁷. At the same time, during cooler periods, enhanced glacial growth accelerates erosion, resulting in an increase in the grain size of sediments in areas of sediment deposition²⁰. In conjunction with the aforementioned proxies that are directly or indirectly associated with temperature, the Δx composition of endogenic carbonates from lake sediments provides quantitative constraints on lake water temperatures, which can, in turn, be used to derive MAAT³⁸.

Qualitative proxies for paleoenvironmental changes. There is a positive correlation (r > 0.7) between oxygen and carbon isotopic compositions of bulk carbonate throughout the whole section (Fig. 2) that supports the sedimentary interpretation of a closed lacustrine depositional environment for these strata¹⁹, and in closed basins, changes in evaporation and precipitation balance related to lake level variation likely drive variation in δ¹⁸O and δ¹³C₁⁸. Statistical analyses (see “Methods”) demonstrate that there is evidence for a step change at 2.7 Ma in all qualitative and quantitative proxy records for the KP site succession. For example, δ¹⁸O, δ¹³C, and CaCO₃ content all shift from consistent high values between 4.3 and 2.7 Ma to low values after 2.7 Ma (Fig. 3a–c).

For all the samples from 4.3 to 0.8 Ma, the mean C/N ratio is 5.6 ± 2.3 (n = 334), while >95% of C/N ratios range between 2.6
and 10, indicating a predominately algal origin for organic matter\textsuperscript{34} (Supplementary Fig. 6a). A few samples (less than 5% of the total) with elevated TOC content also have a high C/N ratio (>10), suggesting only a minor contribution of terrestrial vegetation to organic matter in lake sediments. The higher TOC and TN values before 2.7 Ma are consistent with higher algal productivity associated with a warmer climate at the KP site during the Pliocene (Supplementary Fig. 6b, c). Both C/N ratios and δ\textsuperscript{13}C\textsubscript{org} fluctuate independent of stratigraphic boundaries, indicating that the degradation of organic matter during early diagenesis had only a minor influence on δ\textsuperscript{13}C\textsubscript{org} and C/N ratios (Supplementary Fig. 6a, 6d). The mean grain size record also oscillates and exhibits lower values between 4.3 and 2.7 Ma and high values after 2.7 Ma (Supplementary Fig. 6e).

Given that the high covariance of δ\textsuperscript{18}O\textsubscript{w} and δ\textsuperscript{13}C throughout the section, the increased mean grain size of the sediments, and published data indicating regional aridification during the Plio-Pleistocene climate transition\textsuperscript{38}, we infer that cooling at the KP site at 2.7 Ma, rather than opening of the hydrological system (e.g., lake level rise overtopping the topographic basin), provides a better explanation of the observations. Cooling at 2.7 Ma would have reduced lake evaporation resulting in a lowering of δ\textsuperscript{18}O\textsubscript{w}, δ\textsuperscript{13}C\textsubscript{org} and CaCO\textsubscript{3} content values (Fig. 3a–c). Lower lake temperature would have reduced primary productivity and driven further shifts of δ\textsuperscript{13}C\textsubscript{w}, CaCO\textsubscript{3} content, TN, and TOC values to lower values (Fig. 3b, c; Supplementary Fig. 6b, c). This regional cooling coincident with the intensification of Northern Hemisphere glaciation would also have driven the development of the glaciers in the northern Tibetan Plateau\textsuperscript{39} and enhanced glacial erosion in the surrounding mountain ranges. The increased input of coarser-grained sediments (Supplementary Fig. 6e) to the KP sites would have led to the lake likely shrinking, consistent with the observed increase in δ\textsuperscript{13}C\textsubscript{org} (Supplementary Fig. 6d) that has recently been used as a lake level indicator for alpine lakes in the northern Tibetan Plateau\textsuperscript{37}. Further, the divergence between δ\textsuperscript{13}C\textsubscript{w} and δ\textsuperscript{13}C\textsubscript{org} at 2.7 Ma suggests that δ\textsuperscript{13}C\textsubscript{org} record may not always be positively correlated with productivity in closed-lake systems that are dominated by autochthonous organic matter sources. Changes in the proportion of aquatic and submerged plants with different δ\textsuperscript{13}C\textsubscript{org} values could play an important role in controlling variations in the bulk δ\textsuperscript{13}C\textsubscript{org} of lake sediments. In concert, these qualitative proxies support regional cooling at 2.7 Ma coincident with the intensification of Northern Hemisphere glaciation that in turn, played a fundamental role in changing paleoenvironments at the KP site, including by influencing evaporation rates, impacting lake area and productivity, and sedimentation.

Quantitative temperature constraints indicate stepped cooling since the Pliocene. Recent clumped isotope studies indicate that lacustrine carbonates reflect warm-season near-surface water temperature\textsuperscript{35,40,41}, and therefore we interpreted Δ\textsubscript{θ} t-derived temperatures as a proxy for summer lake surface temperatures (SLST). We would expect millennial and orbital-scale climate variability was a feature of both Pleistocene and Pliocene terrestrial paleoclimate records, as with marine records, but with variations being much larger in amplitude and proxy records noisier in terrestrial environments, particularly high-elevation localities such as the KP site. We also would predict that all SLST should be above 0 °C, because otherwise, the data would imply permanently frozen (or possibly hypersaline) lakes, where carbonate precipitation should be inhibited. We note that given the error bounds on the data, all SLST values are consistent with temperature above freezing (Fig. 3d).

The SLST reconstruction exhibits substantial, stepwise changes over the last 4.3 Ma (Fig. 3d), with temperatures fluctuating between 4.8 ± 4.5 (1σ) and 27.5° ± 4.9 °C from 4.3 to 2.7 Ma and varying between −1.4 ± 1.6 and 20.8 ± 1.3 °C between 2.7 and 0.8 Ma. The uncertainty includes both analytical error and the propagated error when converting the Δθ t values to the temperatures using the temperature-Δθ t calibration of Bernasconi, Müller\textsuperscript{42} (see “Methods” and Supplementary Data 1). Mean SLST values are 15.0 ± 0.9 °C (1σ) between 4.3 and 2.7 Ma, decreasing rapidly to 8.1 ± 0.9 °C at 2.7–2.6 Ma with a mean value of 9.0 ± 0.8 °C between 2.7 and 0.8 Ma (Fig. 2d). The mean Pliocene SLST of ~15 °C in the KP section based on Δθ t is consistent with the ca. 10–17 °C temperature estimates from aquatic plants, ostracods, and mollusk shells\textsuperscript{43,44}.

Using a published relationship between SLST and MAAT\textsuperscript{45}, we estimate that local MAAT from 4.3 to 0.8 Ma is −13–17 °C cooler than SLST (Fig. 3e). The MAAT reconstruction consequently shows the same stepwise changes over the last 4.3 Ma (Fig. 3e), fluctuating between −12.2 ± 3.7 (1σ) and 17.5° ± 3.3 °C from 4.3 to 2.7 Ma and varying between −20.1 ± 1.4 (1σ) and 9.5 ± 0.9 °C between 2.7 and 0.8 Ma (uncertainty includes analytical error, propagated error when transferring the SLST to MAAT; see “Methods” and Supplementary Data 1). Mean MAAT values average 1.7 °C from 4.3 to 2.7 Ma, and decreased at 2.7–2.6 Ma to mean values of −6.4 °C, and further decreased to −7.7 °C at 2.7–0.8 Ma (Fig. 3e), similar to the local modern MAAT (−6 to −7 °C)\textsuperscript{25,44}. These results imply >0 °C MAAT on the northern Tibetan Plateau during the Pliocene, which is consistent with PlioMIP2 climate simulations (Fig. 4). The Δθ t-based estimate of local MAAT during the Pliocene of >0 °C is also broadly compatible with previous estimates of 10 ± 8 °C derived using the δ\textsuperscript{18}O\textsubscript{w} calculated from carbonate in bone and paleo-water\textsuperscript{46}. The 2.7 Ma climate cooling is well recorded in records from across the Northern Hemisphere\textsuperscript{13,17,28,46,47}.

Δδ\textsuperscript{18}O\textsubscript{w} values were calculated for each sample using Δθ t-derived SLST estimates, measured δ\textsuperscript{18}O\textsubscript{w}, and a published calibration\textsuperscript{48} (Fig. 3f). Reconstructed lake water δ\textsuperscript{18}O\textsubscript{w} has a mean value of −4.3 ± 1.8‰ from 4.3 to 2.7 Ma, and decreased to −7.4 ± 2.6‰ at 2.7–2.6 Ma that partly overlaps with the value of local modern meteoric waters (−11.9 to −7.7‰)\textsuperscript{23} (Fig. 3f). Evaporative enrichment of 18O in a closed lake system results in elevated δ\textsuperscript{18}O\textsubscript{w} values\textsuperscript{15}. The decrease in lake water δ\textsuperscript{18}O\textsubscript{w} value at the KP site at 2.7–2.6 Ma may be associated with reduced evaporation in this region.
Collectively, multiple proxies indicate paleoclimate change at 2.7–2.6 Ma at the KP site in the northern Tibetan Plateau, likely related to regional cooling that was coincident with the intensification of the Northern Hemisphere Glaciation (NHG) (Fig. 3h–j)\(^1\). An ~8.1 °C decrease in local MAAT at 2.7–2.6 Ma is derived from the \(\Delta T\) paleothermometer and transfer function that relates late surface temperature to MAAT.

**Global permafrost stability and affected carbon.** Global permafrost is mainly distributed in circumarctic and alpine regions, with the Tibetan Plateau being the largest alpine permafrost region on the Earth (Fig. 1a). The presence and stability of permafrost are fundamentally sensitive to changes in energy balance\(^2\). Assuming that our Earth would evolve to a mPWP-like climate in the near future, permafrost regions that exceed the temperature threshold for permafrost persistence will eventually thaw, exposing soil carbon to decomposition and lateral export. Thus, understanding the sensitivity of the carbon stocks in both circumarctic and alpine permafrost regions in a warmer-than-present climate is of critical importance regionally and globally.

To explore potential implications of paleo and near-future climate change of permafrost climate feedbacks, we combined our paleoclimate reconstruction with mPWP climate model results to...
Fig. 3 Paleoclimate records of the Kunlun Pass (KP) section, northern Tibetan Plateau, and global proxies. a Lacustrine δ18Oc record from the KP section. b Lacustrine δ13Cc record from the KP site. c Carbonate content from the KP section. d Surface lake summer temperature (SLST) from the KP section based on clumped isotope thermometry. e Mean annual air temperature (MAAT) from the KP section estimated by applying a transfer function relating SLST and MAAT\(^{45}\). f δ18Ow (SMOW) record from the KP section. The dashed line in a–f is the bootstrap plot (1 Myr loess regression) derived from each detailed record. The error bars in d–f represent the 1σ error of SLST, MAAT, and δ18Ow of each sample. g Normalised plots showing trends in paleoclimate proxy records from the KP section. These trends are bootstrap plots (1 Myr loess regression) derived from detailed records shown in (a–f). h Magnetic susceptibility (MS) record from the Chinese Loess Plateau (CLP)\(^{28}\), showing aridification at 2.7–2.6 Ma due to intensified winter monsoon associated with the global cooling. i MAAT record from the Weihe basin\(^{46}\). j Biogenic opal mass accumulation rates (MAR) at ODP Site 882\(^{16}\). k Sea Surface Temperature reconstruction for ODP Site 982\(^{47}\). l Global benthic δ18O stack\(^{17}\). Note that the cooling event at 2.7–2.6 Ma on the northern Tibetan Plateau is simultaneous with the intensification of Northern Hemisphere Glaciation (blue shaded area). The orange shaded area demarcates the mPWP (3.3–3.0 Ma). Modern MAAT, SLST, and meteoric water δ18Ow (SMOW) values at the KP section from previous work are shown\(^{25}\). Calculation of mean values and error bars (1σ) are in Supplementary Data 1. The numbers marked in a–g represent the mean value of each proxy during the interval 4.3–2.7 Ma and 2.7–0 Ma, respectively.

Fig. 4 Simulated mean annual air temperature (MAAT) during the mid-Pliocene Warm Period (mPWP). a Pliocene Model Intercomparison Project Phase 2 (PlioMIP2)-based climate model simulation\(^{68}\), showing MAAT and the 0°C isotherm during the mPWP and pre-industrial period, respectively. Region marked with 1–6 refers to the Tibetan Plateau-Pamir, Altai Mountains-Mongolia-Yablonoi-Sayan, Tian Shan, Rocky Mountains, Alps, and Caucasus, respectively.
assess global permafrost stability and affected carbon in a warmer-than-present climatic condition. Specifically, for both circumarctic and alpine permafrost regions, we assessed the distribution of permafrost, permafrost carbon storage, and potential permafrost-climate feedbacks under a mPWP-like climate. Because surface vegetation, moisture, and soil parameters can substantially affect the propagation of surface energy into the subsurface\textsuperscript{22,23}, the precise MAAT of permafrost formation and stability can vary from several degrees below zero to a few degrees above depending on local conditions\textsuperscript{22,23}. To determine which modern MAAT best represent the modern distribution of the permafrost region, we compare multiple modern MAAT isotherms (i.e., $-2$, $-1.5$, $-1$, $-0.5$, and $0$ °C) with the modern distribution of the global permafrost region (Supplementary Fig. 7). At a global scale, the modern $0$ °C MAAT isotherm best matches the distribution of permafrost (see “Methods”), and we thus use $0$ °C as the conservative temperature threshold for global permafrost formation and persistence. While not perfect, this value is also widely used to model the modern permafrost area\textsuperscript{49}.

Figure 4 shows a map of simulated mPWP global MAAT and the location of simulated mPWP and pre-industrial $0$ °C MAAT isotherms. In the alpine Tibetan region, the mPWP $0$ °C isotherm is in the centre of the plateau, while the pre-industrial $0$ °C isotherm is situated at the margin of the plateau (Fig. 4). Outside the Tibetan Plateau, the mPWP $0$ °C MAAT isotherm is generally located above 60°N with a few exceptions in North America and Eurasia, while the pre-industrial $0$ °C MAAT isotherm is generally located above 50°N. From the mPWP to the pre-industrial period, the $0$ °C isotherm on the Tibetan Plateau expands outwards, nearly doubling its area, while the $0$ °C MAAT isotherm in the middle- to high-altitude migrates southwards by $\sim10^\circ$ on average (Fig. 3a).

Using this $0$ °C MAAT isotherm as a conservative limit of permafrost persistence, the northern and southern edges of the Tibetan Plateau would become unstable under a mPWP-like climate. The estimated low to middle alpine permafrost regions in this study mainly consist of Tibetan Plateau-Pamir and Altai Mountains-Mongolia-Yablonoi-Sayan in addition to Tian Shan, Rocky Mountains, Alps, and Caucasus (Fig. 4). Using the present-day-permafrost distribution\textsuperscript{1} and the simulated mPWP $0$ °C MAAT isotherm (Fig. 4), we find that about $\sim20\%$ of the circumarctic permafrost would be destabilised ($\sim3.9 \times 10^{12}$ m$^2$), but $\sim60\%$ of alpine permafrost ($\sim1.9 \times 10^{12}$ m$^2$) would be destabilised (Fig. 4).

Using carbon estimates from modern permafrost regions\textsuperscript{2,3}, we calculate the total carbon potentially affected under a future climate scenario similar to that of the mPWP (see Calculation of permafrost thawing area and carbon release in the “Methods” section). In our scenario of a mPWP-like climate, $\sim254 \pm 23$ Pg of organic carbon is potentially affected in the circumarctic permafrost zone and $\sim85 \pm 60$ Pg of organic carbon is affected in the alpine permafrost zone (Fig. 5).
elevations during global temperature change), although the difference may not be statistically significant.

To assess the possibility that surface uplift in the northern Tibetan Plateau occurred at ~2.7 Ma, we provide a first-order estimate that teasing out the global temperature change effect on the local MAAT in the northern Tibetan Plateau, based on our knowledge of the correlation between MAAT variation in the northern Tibetan Plateau and global MAAT change in recent years due to anthropogenic greenhouse gas emissions. The last 50 years of temperature monitoring records a ~2.0 °C MAAT increase at high elevation on the northern Tibetan Plateau and a ~1.3 °C global temperature increase52–54, with a linear correlation between warming of high-altitude temperatures on the northern Tibetan Plateau and global temperatures (Supplementary Fig. 8). According to the transfer function established from this correction (see “Methods”), a global decrease in temperature of ~2–4 °C since 2.7 Ma53,14,24,53 should locally be expressed as a 7.8 ± 1.5 °C MAAT decrease on the northern Tibetan Plateau, which is within the error of 8.1 ± 7.5 °C observed cooling at 2.7–2.6 Ma at the KP site. Given the 4–5 °C/km modern lapse rate on the northern Tibetan Plateau55, this ~0.3 °C difference reflects approximately 0.1 km of local surface uplift (See calculation is in Supplementary Data 1). We thus infer that although minor surface uplift of the northern Tibetan Plateau cannot be ruled out for some of the observed cooling at the KP site at 2.7–2.6 Ma, most of the record reflects the regional expression of global climate change with high-elevation temperature amplification.

We also compare results with other published clumped isotope data and with PlioMIP mPWP model simulations. Our MAAT reconstruction from the KP section indicates an average MAAT of 1.7 °C in the northern Tibetan Plateau from 4.3 to 2.7 Ma (Fig. 3f), while in the southern Tibetan Plateau, clumped isotope analysis based on aquatic gastropods in the Zhada basin suggest a MAAT > 0 °C during the Pliocene.50 Our MAAT reconstruction from the KP section indicates an average MAAT of ~0 °C at the KP site during the mPWP (Fig. 3f; Supplementary Data 1). This is consistent with the PlioMIP mPWP MAAT simulation result which shows that the 0 °C MAAT isotherm line during the mPWP goes by the KP site (Fig. 4). These consistencies between geological records from local studied sites and global mPWP MAAT simulation results not only give us confidence in both temperature proxy results and the simulations, but also provide insights on using the mPWP 0 °C MAAT isotherm simulation to explore the stability of modern permafrost in the near future (see “Methods”).

Surprising vulnerability and importance of alpine permafrost carbon. Though alpine permafrost regions only contain 14% of organic matter stocks in the global permafrost zone, our calculations, albeit for a simplified analog-based warming scenario, suggest they could play an outsized role in determining the permafrost climate feedback. Enhanced thermal vulnerability of alpine permafrost regions has been supported by the monitoring of global MAAT, which shows greater warming in low-latitude, high-elevation regions compared with high-latitude circumboreal regions52. Given the current 400+ ppmv CO2 level in the atmosphere (equivalent to ~800 Pg of carbon)9, the mPWP-based results indicate a large quantity of carbon could be thawed in circumboreal and alpine permafrost regions. Because of its climatic vulnerability, alpine permafrost region could disproportionately account for ~25% (i.e., ~85 ± 60 Pg) of the carbon affected by a transition from modern to mPWP-like climate conditions (Fig. 3c).

We recognise that there is not a well-established relationship between amount of permafrost thaw and amount of greenhouse gas release11, and that the distribution of modern permafrost is influenced by the legacy of the last ice ages56,37. Although these factors hamper a precise calculation of how much carbon will be released when thawing permafrost, our simple scenario allows projection of which permafrost-containing regions are likely to become unstable under climatic conditions similar to the mPWP. As regional MAAT in areas of permafrost storage exceed the temperature threshold for permafrost persistence, the total carbon that is stored in the permafrost soil column will eventually thaw and destabilise organic matter, with microbial production of greenhouse gases that could be taken up by biota and/or ultimately released into the atmosphere. Ultimately, the amount of CO2, CH4, and N2O released to the atmosphere from these regions will depend on specific ecological conditions, including slope, hydrological status, duration of seasonal thaw, disturbance, and amount of gradual versus abrupt thaw2,10. However, upland areas—including on the Tibetan Plateau—are often more vulnerable to carbon release following permafrost degradation because of aerobic respiration and thermo-erosion10, suggesting that alpine permafrost carbon might be dually responsive to climate change. This implies that detailed long-term monitoring of alpine regions is urgently needed to quantify organic matter stocks, lateral transport (e.g., fluvial and colluvial processes), and vertical greenhouse gas flux. Constraining alpine temperature amplification and carbon cycle dynamics under different global warming scenarios will be critical to determining the amount of additional warming that this carbon release from permafrost could cause.

Methods

Kunlun Pass site and age model. The KP section is the same section that was studied previously for magnetostratigraphy with an approximately two-metre resolution27 and multiple faunal layers by Li Xie26. Given problems in determining reliable magnetostratigraphy within conglomerate layers with sand-sized matrix, we focused on fine-grained lacustrine deposits in the middle to the upper parts of the section and re-correlate the observed polarities using the Geomagnetic Polarity Time Scale (GPTS)58. The coarse yellowish sandstone in the upper parts of the section and the black organic-rich mudstone layer that preserve leaves enable us to correlate our newly measured section to a previous study27. We correlate two short normal intervals of N2 and N3 to chron C1r.1n and C1r.2n, respectively. The long normal interval N4 can be correlated to the C2An.1n-C2An.3n. Four short normal intervals of N5, N6, N7, and N8 to chron C3n.1n, C3n.2n, C3n.3n, and C3r.4n, respectively. The age tie points used to establish the age model for the section are given in Supplementary Fig. 4. We thus determined the age model by piecewise linear interpolation.

To confirm the age model (4.3–0.8 Ma), a total of 344 samples were collected for carbonate (CaCO3) content, δ18Oc, and δ13Cc measurement (Fig. 3a–c). The obvious fluctuations in all three records indicate potential orbital cycles (Fig. 2a–c). If our magnetostratigraphy-based age model is correct, we should resolve orbital cycles from this record. We used these high-resolution records (Supplementary Fig. 3a–c) to carry out time-frequency analysis. Power spectral analysis and Fast Fourier transformation (FFT) were performed using Aycycle software9. Using the untuned age model, spectral analysis on CaCO3 content, carbon, and oxygen isotope records resolves orbital cycles, indicating eccentricity, obliquity, and precession (Supplementary Fig. 5b, c), giving us confidence in the age model. To better explore the paleoclimatic change during the Plio-Pleistocene climate transition, the sampling rate of carbonate, δ18O content and isotope system (δ13C) (12–15 kyr) for the time interval from 4.3 to 1.8 Ma. For the time interval between 1.8 and 0.8 Ma, the sampling rate of CaCO3 content varies between ~12 and ~20 kyr while the sampling rate of isotope samples ranges from ~40 to ~130 kyr (Supplementary Fig. 5).

Sample measurements. Approximately 50 g of freeze-dried sediment from each specimen was homogenised for carbonate δ18Oc and δ13Cc, clumped isotope (Δ13Cy), carbonate content (CaCO3), organic δ13Corg, TOC and TN measurements. Carbonate δ18Oc and δ13Cc measurements. δ18Oc and 13Cc of carbonate samples were analysed at the Stable Isotope Ratios in the Environment Analytical Laboratory, University of Rochester. Results are reported in per mil (‰), relative to the Vienna PeeDee Belemnite (VPDB) standard. Around 0.3 g of each sample that display no visible veins or recrystallisation were sieved through a 200-mesh screen, homogenised and subsequently reacted with 50 ml 3% H2O2 for 5 h to remove organic matter. Each sample was then rinsed with deionized water. After drying at room
temperature (~20 °C), the powder was weighed, sealed in a glass vial, and flushed with helium gas. The powder was then reacted with pure phosphoric acid at 70 °C to release CO2 gas. Analyses were performed using a Thermo Delta plus XP CF-IRMS with a GasBench II peripheral device. To calculate isotopic ratios, 3 in-house standards calibrated to international standards (namely NBS-18, NBS-19, and L-SVEC) were used. Analytical errors for δ13O and δ18O are within ±0.1‰ (1σ) and ±0.06‰ (1σ), respectively. A total of 275 δ13O and δ18O values were obtained.

**Clumped isotope (Δc) measurements.** Clumped isotope analyses were carried out in the Tripathi Lab at the University of California–Los Angeles, following the methods of Upadhyay, Lucarelli60. A total of 57 samples were analysed. During the pretreatment, approximately 0.5 g of each sample that displays no visible veins or recrystallisation was reacted through a 70-μm sieve, homogenised and subsequently reacted with 50 ml 3% H2O2 for 5 h to remove organic matter. Each sample was then rinsed with deionized water and dried at 40 °C for 24 h. Samples were then ground and homogenised using an agate mortar and pestle. Before analysing on a Thermo 253 Gas Source ratio mass spectrometer, three to seven aliquots of each sample were reacted at 90 °C with 103% phosphoric acid for ~25 min. The resulting CO2 was cryogenically purified.

**Organic C (δ13Corg) and oxygen isotope (δ18Ow) measurements.** Organic C (δ13Corg) and oxygen isotope (δ18Ow) measurements were determined at State Key Laboratory of Loess and Quaternary Geology, Institute of Earth Environment, Chinese Academy of Sciences calibrated to international standards (namely NBS-18, NBS-19, and L-SVEC) were used. A total of 344 data were obtained. The samples were determined using an elemental analyser (vario EL cube). The analytical precision of the CaCO3 content is ±0.5%.

**TOC and TN measurements.** TOC and TN measurements were made at Qinghai Institute of Salt Lakes, Chinese Academy of Sciences. A total of 344 data were obtained. Approximately 0.5 g of each sample was ground using an agate mortar and pestle, sieved through a 200-μm mesh screen, and homogenised. Each sample was then treated with 2 M HCl for 24 h at room temperature to remove carbonate. Samples were rinsed with deionized water and dried at 40 °C for 24 h. The concentrations of total TOC and TN in the samples were determined using an elemental analyser (vario EL cube). The analytical error for both the TOC and TN content is ±0.1%.

**Grain size measurements.** Grain size measurements were determined at State Key Laboratory of Loess and Quaternary Geology, IEECAS. A total of 344 samples were analysed. During the pretreatment, 10% HCl and 10% H2O2 were used to remove organic matter and carbonates in the samples. All the samples were then put under deionized water and kept overnight in beakers. Before grain size analysis, water was aspirated off and each sample was then dispersed in an ultrasonic bath in a 10 ml vial in a 10% (NaPO3)2 solution for 10 minutes. Finally, grain size distributions were obtained using a Malvern 2000 laser instrument.

**Statistical analyses confirm a climate change event at 2.7 Ma at the KP site.** All δ13C, δ18O, δ13Corg, CO2 content records shifted to a lower value at 2.7 Ma. We note this shift at 2.7 Ma is noisy in KP site SLST and MAAT records. To evaluate if there is robust evidence for a shift in regional climate at 2.7 Ma, we conducted a Student’s T-test. Using the δ13C, δ18O, CO2 content, TOC, TN, δ13Corg grain size, SLST, MAAT, and δ13Corg records. This method determines the probability that two populations are the same with respect to the variable tested. The fundamental criterion of the Student’s T-test is if the p value is less than 0.05, there is a >95% level of confidence that the two age distributions are not the same. Lower p values provide a higher probability of rejecting the null hypothesis that there is no difference between the two populations.

We divided each record using two bins (i.e., 2.7, 10, 1.5, 2.0, 2.5, 3.0, 3.5, and 4.0 Ma) and conducted a Student’s T-test. This was done with δ13C, δ18O, CO2 content, TOC, TN, δ13Corg grain size, SLST, MAAT, and δ13Corg records. The Student’s T-test results, shown in Supplementary Table 1, reveal the p values derived from all records are smallest if 2.7 Ma is the time when there is a shift, indicating that the 2.7 Ma boundary is robust.

Using 1 million years as an age window, we also used the bootstrap plots (loess regression) to evaluate the trends in δ13C, δ18O, CO2 content, TOC, TN, δ13Corg grain size, SLST, MAAT, and δ13Corg records (Fig. 3; Supplementary Fig. 6). The normalised plots also show a shift at 2.7 Ma (Fig. 3g), consistent with a change in regional climate at 2.7 Ma at the KP site.

Collectively, our statistical analyses confirm that a climate change event at 2.7 Ma is recorded in multiple proxy records from the KP section, with the direction of change consistent with cooling. This inference of cooling is observed in other proxy records from across Eurasia and the Northern Hemisphere11–17,28,46,67 (Fig. 3h–l).

**Sample petrography, carbonate facies description, and diagenetic screening.** All KP section samples were examined under reflected light using a microscope. Fifteen representative samples were impregnated with epoxy (to hold them together) and prepared as thin sections. These sections were examined using cross-polarised light microscopy at the University of Rochester to evaluate the origin and potential for diagenesis of carbonate. Thirteen of the fifteen samples are predominantly fine-grained (clay to silt-sized), containing micrite, clay, and quartz (Supplementary Fig. 2a). We also examined 2 moderately sorted, very fine sandstones that show quartz, detrital carbonate, and metamorphic grains, with a clay–silt-sized matrix (Supplementary Fig. 2b, c). The minor amount of quartz and detrital carbonates grains are derived from locally exposed basement rocks including carbonate outcrops in the surrounding regions23. Carbonate in over 90% of samples is dominantly micritic (Supplementary Fig. 2a-c). Minor blocky sparite or microspar in pore spaces was observed in the samples, and likely accounts for the minimal lithification of the sediment. The lack of lithification and dominantly micritic texture of the carbonates demonstrates little to no recrystallisation associated with diagenesis (Supplementary Fig. 2a). We interpret the minor sparite association with pore spaces as diagenesis associated with carbonate precipitation in the sediment. This inference is further supported by the shallow depth of burial of sediments (<300 m).

**Methods of removing detrital carbonate component.** We developed a new approach to reduce the effect of detrital carbonates contamination on carbon isotopes. Two test samples (16KLI102 and 16KLI152) were sieved through a 230-μm mesh-screen and a 70-μm mesh-screen, respectively. Each test sample was divided into four groups (non-sieved, <212 μm, 212–63 μm, <63 μm) depending on grain size. Materials from each group were then homogenised and subsequently reacted with 50 ml 3% H2O2 for 5 h to remove organic matter. Materials from each group were rinsed with deionized water and dried at 40 °C for 24 h, then ground and homogenised in an agate mortar. Two to three replicates from each group were weighed and then analysed on a Thermo 253 Gas Source isotope ratio mass spectrometer. Measurement and normalisation of these samples followed the procedure outlined in the Clumped Isotope (Δc) Measurements section of this paper, stated above.

**Collective, our statistical analyses confirm that a climate change event at 2.7 Ma is recorded in multiple proxy records from the KP section, with the direction of change consistent with cooling. This inference of cooling is observed in other proxy records from across Eurasia and the Northern Hemisphere11–17,28,46,67 (Fig. 3h–l).**
A total of 16 replicates for sample 16KL152 and 13 replicates for sample 16KL102 were analysed and results are displayed in Supplementary Fig. 3. The non-isothermal boxmolds with the detrital grain, the smallest at the same time to represent the results of this sample. For these two samples, their 1\(^8\)Oc values are 9.4 to 7.7 °C while 1\(^8\)Oc values, respectively. Two samples (16KL160 and 16KL331) yield a better characterises the distribution of modern permafrost, and the pre-industrial 0 °C MAAT isotherm line generally defines the extent of the modern permafrost in different regions. We thus define 0 °C as the conservative temperature threshold for global permafrost and 1 °C MAAT isotherm lines are located inside the extent of the modern permafrost region (Supplementary Fig. 4). This contradiction indicates these MAAT isotherms might not represent the threshold for global permafrost formation or persistence. On the other hand, the modern 0 °C MAAT isotherm better characterises the distribution of modern permafrost, and the pre-industrial 0 °C MAAT isotherm line generally defines the extent of the modern permafrost in different regions. We thus define 0 °C as the conservative temperature threshold for global permafrost enables us to provide a first-order quantitatively estimate of the areas that permafrost could be affected in a warmer than today climate scenario in the near future. The PlioMIP2-based climate simulation shown in Fig. 4 depicts the global mPWP MAAT with the mPWP and pre-industrial 0 °C MAAT isotherms. These two isotherms models the residual permafrost region in both middle to high latitude circumarctic permafrost and low to middle latitude alpine permafrost regions during the mPWP and pre-industrial periods, respectively. We loaded these two modelled 0 °C MAAT isotherm lines into ArcGIS 10.1 and calculated the circumarctic and alpine permafrost areas within these isotherms that correspond to mPWP-like and pre-industrial-like climates, respectively. Comparing the modelled stocks in the pre-industrial-like climate with the modern permafrost area, allows us to estimate the amount of permafrost area that has thawed since the pre-industrial period (Supplementary Data 3). These two modelled permafrost thaw areas (including Pamir, Altai Mountains-Mongolia-Yablonoi-Sayan, Tian Shan, Rocky Mountains, Alp, and Caucasus), are shown in both Figs. 1 and 4, with their area calculated using ArcGIS 10.1 (Supplementary Data 3).

The computation and mean value of SOC with 1 °C MAAT isotherm as the quantity of the soil carbon in the corresponding permafrost thaw region in a mPWP-like and pre-industrial-like climate are then calculated as the quantity of carbon that would be thawed in the near future and are shown in Fig. 4 and Supplementary Data 3.

Climate simulation. Required spatial information for Pliocene and pre-industrial surface temperatures have been derived from the Pliocene Model Intercomparison Project Phase 2 (PlioMIP2). PlioMIP2 co-ordinates climate modelling groups from 90°N) Arctic permafrost and low to middle latitude (between 25°N and 50°N) alpine permafrost regions (including the Tibetan Plateau- Pamir, Altai Mountains- Mongolia-Yablonoi-Sayan, Tian Shan, Rocky Mountains, Alp, and Caucasus). We also recognise that permafrost can also be destabilised due to other reasons, such as coastal erosion, infrastructure damage, landslides, ecosystem damage, etc. We believe a more complex specific modelling approach might help to gain realistic estimates; however, it is currently difficult to predict and even quantify all these parameters. Comprehensive quantitative impact of these unknown factors on the stability of global permafrost is worthy of further exploration and research. Despite the limitations, we use the conservative temperature threshold for global permafrost to provide a first-order quantitatively estimate of the areas that permafrost could be affected in a warmer than today climate scenario in the near future.

Correlation between global temperature and regional MAAT changes. We evaluate the potential correlation between changes in global temperature and MAAT in the northern Tibetan Plateau using temperature information from weather stations. Using the global temperature variations reported by Morice, Kennedy39 and the MAAT on the northern Tibetan Plateau reported by Wang, Fan32 we made a cross plot of anomalies in global temperature and MAAT in the northern Tibetan Plateau for the 1961–2010 period (both defined relative to the 1961–1990 average). This plot shows a strong linear correlation (r = 0.84) between these two anomaly data sets (Supplementary Fig. 8). A transfer function relating the global temperature variations to the MAAT variation in the northern Tibetan Plateau is established:

\[
\Delta T_{\text{NTP}} = 2.5725 \times \Delta T_{\text{GC}} + 0.0295
\]

where \(\Delta T_{\text{NTP}}\) is the variation of MAAT (°C) in the northern Tibetan Plateau in response to \(\Delta T_{\text{GC}}\) (global temperature variation, °C). This transfer function enables estimation of MAAT changes in the northern Tibetan Plateau in response to a global decrease in temperature of −2−4 °C at 2.7 Ma. We acknowledge that the relation between the \(\Delta T_{\text{NTP}}\) and \(\Delta T_{\text{GC}}\) during the Plio-Plenocene transition might not completely follow this transfer function. Despite uncertainties (e.g., contiguity effect and moisture transport), this transfer function allows us to provide a first-order estimate of potential surface uplift in the northern Tibetan Plateau.

Calculation of permafrost thawing area and corresponding carbon affected. According to definition, permafrost is ground with a temperature that remains or below 0 °C for two or more years65. Thus, the permafrost persistence is basically temperature-dependent. Although more paleotemperature data derived from different proxies are consistent because a lower boundary for the permafrost reconstruction from sediments in permafrost region (including KP site)15,29 and the simulations (Fig. 4), give us confidence in using the mPWP MAAT isotherm simulation results to explore the stability of modern permafrost across the globe in a warmer than today climate scenario in the near future.

To de
around the world in completing climate simulations for the mPWP using a unified experimental design. Full details of the experimental design and boundary conditions used for the mPWP modelling studies can be found in Haywood, Dowsett et al., and Dowsett, Dolan, respectively. In this study, we utilise the multi-model mean annual surface air temperature field from PlioMP2 for both the mPWP (Eoi400 experiment) and pre-industrial era (E280 experiment; see Haywood, Tindall et al.). These data are used to plot the spatial position of the −2,−1.5,−1.0,−0.5, 0 °C MAAT isotherms. The estimated MAAT is an average of the 1° × 1° latitude/longitude grid box from each climate model.

Data availability
All new data produced for this study are from samples from a continuous Pliocene to Pleistocene lacustrine record on the northern Tibetan Plateau. The authors declare that all data supporting the findings of this study are available within the article and its Supplementary Data and are accessible online at https://doi.org/10.6084/m9.figshare.19033130.v2.

Code availability
No code was developed for this study.

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Author contributions

All the authors have directly contributed to data acquisition or analysis of results; F.C. and C.G. acquired the carbonate stable isotope data; A.T. supported clumped isotope analyses. F.C., B.E., D.U., A.A., and A.T. acquired clumped isotope data and provided input on sample preparation. F.C., C.G., and A.T. contributed to clumped isotope interpretations. X.L. acquired the carbonate content, organic carbon isotope, grain size, total organic carbon, and total nitrogen data; F.C., X.L., and L.L. collected samples and described the section in the field. A.H. and J.T. carried out the analysis of the PlioMIP2 simulations. F.C. and G.G. developed the methods and carried out the permafrost carbon estimates, with refinements from B.A. L.W. contributed to the graphing that supported the carbon calculations. F.C. and G.G. carried out the interpretation of the data with input, advice, and refinements from U.S., A.T., W.L., F.S., and J.N. F.C. wrote the main body of the manuscript, with other co-authors contributing portions in text, reviewing, and revising the figures and text.

Competing interests

The authors declare no competing interests.

Additional information

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