and its cooling potential is only about 1.2 °C before it reaches the freezing point of seawater (−2 °C), which is not enough to fulfil the constraints on MOT during the LGM from different lines of evidence (the noble gas record provided here, and refs 4 and 9). Just from this simple consideration it is obvious that the LGM ocean temperature pattern must have been different from today's. To account for this aspect we split the ocean box into three boxes representing AABW, NADW and all other waters (RES). We set the temperature, volume and salinity of AABW and NADW according to ref. 20 (AABW: −0.88 °C, 35% of total ocean volume, 34.641 PSS; NADW: 2.3 °C, 20% of total volume, 34.886 PSS) and set the RES ocean such that the averaged ocean corresponds to today's average conditions (3.53 °C, 1.34×10^{18} m^3, 34.72 PSS)[70]. In a first experiment we change the temperatures of the different volumes equally as long as AABW does not reach −2 °C. If this happens AABW temperature is set to −2 °C (non-freezing) and the remainder of the cooling is compensated by the other water masses to equal shares. This requires a lower LGM MOT of −0.2 °C owing to the nonlinearity of the solubility functions and gives a sense of how strong the effect of a changing temperature distribution can be on our MOT reconstruction.

The non-freezing AABW experiment described above follows a somewhat artificial path of the ocean temperature/volume distribution. A more realistic scenario is that AABW volume was larger in glacial times, similar to what ref. 22 describes. We use a scenario in which AABW during LGM was 40% bigger than it is today and shrank linearly over the course of the LGM–Holocene transition (17,900–11,550 yr BP) to the current situation found in ref. 20. We choose 40% because it roughly compensates the reduced AABW cooling/warming potential with its change in volume at the expense of the other (warmer) water masses. This more realistic (but still arbitrary) scenario halves the effect of a change in the temperature distribution on the LGM–Holocene MOT difference to −0.1 °C.

We use this three-ocean box model version including all elements and the AABW volume change scenario described so far for our MOT reconstructions shown in the main text. The analytical uncertainties and uncertainties of the sea-level change record are propagated to our MOT estimate, creating 3,000 Monte-Carlo MOT realizations for each data point. The same procedure is done using the two firn thermal gradient scenarios and undersaturation scenarios described earlier. This results in 12,000 MOT record realizations for each ratio and 36,000 MOT record realizations in total. Our best-estimate record is derived based on all these realizations, which provides an objective representation of all uncertainty elements discussed here. For our LGM–Holocene MOT change estimate (see averaging periods in Fig. 3) we also make use of all these realizations while we interpret the propagated measurement and the sea-level change uncertainties as of stochastic nature and treat them as normally distributed uncertainties. However, the uncertainty introduced by the Xe (and Kr) undersaturation effect we treat as non-stochastic because it represents equally likely scenarios. This source of uncertainty represents the largest contribution to the overall uncertainty and with this approach we find a LGM–Holocene MOT difference of 2.57 ± 0.24 °C.

In Extended Data Table 1 we list three more elements that are not included in our MOT records, but are discussed here for completeness. As described in ref. 22, the glacial ocean circulation might have been characterized by an approximately 1PSS saltier AABW cell owing to missing fresh water input from melting sea ice in the Southern Ocean. As the salt content can be assumed to be conserved in the ocean on these timescales, the additional salt in AABW has to be provided by NADW and RES. Owing to the salinity dependency of the solubility functions, such a salinity redistribution leads to different weights of the differently warm water masses in the MOT reconstruction. We tested this effect by a salinity anomaly of 1PSS applied to our AABW cell (compensated by NADW and RES by equal shares) and find a small effect of only −0.02 °C on the LGM MOT estimate.

Another aspect we test is the potential bias caused by a large floating ice shelf. Noble gases are basically only dissolving in the liquid phase of the ocean but the sea-level change record does not capture the corresponding liquid ocean volume change as opposed to ice that is stored on land. We assume an ice shelf with the extent of the modern winter sea ice around Antarctica and a thickness of 200 m. This seems gigantic, as we do not have any evidence that such a large ice shelf could have existed. The effect of such an ice shelf on the LGM MOT estimate would only be −0.1 °C and shows that this potential bias is also of minor relevance.

The last row in Extended Data Table 1 shows the effect of the applied 2% correction of the Xe solubility function compared to the case in which we do not apply this correction. Mass conservation of the noble gases in the model means that this temperature-independent change in the solubility function of Xe leads to a slight change in the MOT sensitivity of the ratios, including Xe (Xe/N_2 and Xe/Kr). The effect on the LGM MOT estimate, however, would only be 0.04 °C and 0.07 °C, respectively, showing that the results presented here are not much affected by this existing uncertainty in the Xe solubility. Kr is about a factor of two less soluble in sea water than Xe and the solubility function of Kr is better constrained[15] than is Xe. For these reasons, the effect on the LGM MOT estimate of the uncertainty in the Kr solubility function is much smaller than what is shown for Xe in Extended Data Table 1 and can therefore be neglected.

Scaling MOT to surface temperatures based on global climate models. MOTs are set by surface ocean temperatures, which in turn are related to global surface temperatures. The connection between surface and ocean interior temperature changes is, however, also dependent on the climatology (polar amplification, ocean circulations, location of deep water formation areas, and so on), which is different for glacial and interglacial periods. The constraints on the glacial climatology are fairly weak and the realization of such climatology within a climate model can be very different from model to model. Therefore, we use several independent climate models that provide climatology for glacial and interglacial conditions and calculate the scaling factors from MOT to ASST and GAST changes, respectively (see ΔASST/ΔMOT and ΔGAST/ΔMOT in Extended Data Table 2).

Such glacial–interglacial climate model experiments are part of the Paleoclimate Modelling Intercomparison Project (PMIP), which can be accessed openly via one of the Coupled Model Intercomparison Project (CMIP) data nodes. All results found in Extended Data Table 2 are based on model output from the PMIP3 project (ensemble: r1i1p1; see ref. 71 for more details about the CMIP5/PMIP3 experiments), with the exception of the Bern3D model results which were provided for this study. From the PMIP3 project results, we used the following variables from the LGM and the Pre-industrial Control experiments: (1) global averaged sea water potential temperature (thetaoga), (2) seawater potential temperature (thetao), and (3) near surface air temperature (tas). Where available, we averaged the thetaoga data to derive MOT. If only thetao was available (three-dimensional field) we averaged over the time dimension covered by the corresponding dataset (12 months) and then over the space dimension while weighting the cell values by the corresponding cell volumes. ASST was calculated by first filtering all surface cells in thetao that are covered by more than 50% with sea ice, followed by the same temporal and spatial averaging as done for MOT. Therefore, our ASST values represent the open ocean surface temperatures excluding the areas covered by sea ice, where the heat exchange with the atmosphere is negligible and the surface ocean temperature is set to freezing temperature of the corresponding water (dependent on salinity). GAST was calculated by averaging the tas fields (two-dimensional fields).

The results in Extended Data Table 2 show that the LGM–Holocene MOT difference varies strongly from model to model mainly owing to discrepancies in the LGM values. This shows that the models provide quite different climatologies in particular for the LGM conditions. Therefore the range of these model results can be interpreted according to how much different climatologies can affect the scaling factor between the globally averaged parameters calculated here. The ΔASST/ΔMOT scaling factor varies from 0.67 to 0.89 with an ensemble average of 0.80. The ΔGAST/ΔMOT scaling factor varies from 1.96 to 2.92 with an ensemble average of 2.50.

In general, the models underestimate the MOT difference between the LGM and the Holocene with an ensemble average of 1.60 °C and a range from 0.92 °C to 1.95 °C, which raises the question of whether the large spread of the scaling factors is correlated to the absolute LGM–Holocene MOT difference and, hence, may contain a bias. However, there is no correlation between the absolute LGM–Holocene MOT difference and the scaling factors, for which reason any possible bias in these scaling factors is believed to lie within the model spread.

Hypothesis behind the Younger Dryas MOT anomaly. As discussed in the main text, our MOT record shows a phase of outstanding strong and fast warming during the first half of the Younger Dryas (referred to as YD1). Here we discuss two possible underlying mechanisms.

One condition that might underlie the strong MOT warming/heat uptake during YD1 could be the strong insolation in high latitudes associated with the phase of high obliquity around YD1 (Fig. 3). In the latitudes where deep waters are formed, the local annual averaged heat flux was about 1.5 W m^{-2} higher than during the LGM. The additional heat flux could have led to an increased warming of surface waters near the deep-water formation areas during the summer seasons, which would have then been transported into the deep ocean during the winter seasons, when deep-water formation mainly occurs. The pattern of the YD1 warming, however, is not consistent with the gradual insolation change, requiring additional processes at work. For the period before the YD1 warming and its abrupt start, the change in AMOC state can provide such an explanation: before the YD1 the strong AMOC state pulls the warm waters towards the north, preventing warming of the deep (southerly ventilated) ocean. The collapse/weakening of the AMOC at the beginning of the YD1 stopped this northward heat pull and, thus, triggers the rapid YD1 warming. But for the end of the YD1 warming, which occurs considerably before the end of the Younger Dryas when the AMOC accelerates again, the AMOC can no longer explain the observation. Note that these orbital-driven heat flux changes are fairly small with regard to the baseline flux of