Lack of evidence for a substantial sea-level fluctuation within the Last Interglacial

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During the Last Interglacial, global mean sea level reached approximately six to nine metres above present. This period of high sea level may have been punctuated by a fall of more than four metres, but a cause for such a widespread sea-level fall has been elusive. Reconstructions of global mean sea level account for solid Earth processes and so the rapid growth and decay of ice sheets is the most obvious explanation for the sea-level fluctuation. Here we synthesise published Last Interglacial geomorphological and stratigraphic indicators, and find no evidence for ice-sheet regrowth within the warm interglacial climate. We also identify uncertainties in the interpretation of local relative sea level data that underpin the reconstructions of global mean sea level. Given this uncertainty, and taking into account our inability to identify any plausible processes that would cause global sea level to fall by four metres during warm climate conditions, we question the occurrence of a rapid sea-level fluctuation within the Last Interglacial. We therefore recommend caution in interpreting the very high rates of global mean sea level rise that are in excess of three to seven metres per 1,000 years and that have been proposed for the period following the Last Interglacial sea-level lowstand.

Last Interglacial sea-level change

There is broad consensus that the highstand in GMSL during the Last Interglacial (LIG: Marine Isotope Stage (MIS) 5e, ~129-116 ka) was likely 6-9 m higher than present[1], implying a smaller than present global ice volume. Relative sea level (RSL), as recorded by proxy records, is locally variable[2] due to spatially non-uniform variations in the height of the geoid and the solid Earth and, as a result, local records of RSL do not reflect GMSL[3]. To better understand the structure of the GMSL highstand during the LIG, Kopp et al.[4] developed a global sea-level database which they statistically analysed to produce a posterior probability distribution of GMSL over the LIG (Figure 1). That study, and further probabilistic analyses[5], highlight an interesting structure within the GMSL highstand: a 95% probability of an fluctuation (sea-level fall and rise), of greater than 4 m (67% probability). Given that the Kopp analysis is corrected for local perturbations to the height of the land and the sea surface, the primary mechanism invoked to explain the GMSL fall during the middle of the LIG is growth of one or more ice sheets[2]. However, sea-level data from the present interglacial provide no evidence for comparable (>4 m sea-level equivalent) ice-sheet regrowth[6]. Understanding the potential mechanism(s) for the LIG GMSL sea-level fall is important because the rates of GMSL rise that are inferred to have followed the lowstand are high[5] (*likely*, i.e. 67% probability, between 3 and 7 m/kyr) and suggest a period of rapid ice-sheet collapse. Given the societal importance of the risks of future sea-level rise[7], it is vital to understand whether, and how, any lowstand occurred, because this constrains the subsequent high rates of GMSL rise in the LIG[4, 8], which currently inform future climate adaptation assessments and strategies[7, 9].

To understand the nature of LIG sea-level change, and to assess the plausibility of the reconstructed GMSL lowstand, requires knowledge of four-key elements, which we now consider in turn: 1) the solid Earth response to ice-sheet loading and unloading during and following the preceding glacial; 2) ice-sheet histories during the LIG; 3) thermosteric sea-level change, and 4) the quality of local RSL data which underpins the reconstruction.

Solid Earth processes

To reconstruct GMSL it is important to identify the regional solid Earth processes that impact the elevation of former RSLs[10]. Kopp et al.[4, 5] undertook this by running 250 alternative ice-sheet histories and randomly selected Earth viscosity profiles and using them in their Bayesian inversion to generate the GMSL curve shown in Figure 1. In contrast, Düsterhus et al.[11] used a massive ensemble approach to analyse the same geological dataset, and performed 39,000 model runs, where the output of each run is a glacial-isostatic adjustment (GIA) based GMSL prediction associated with a specific ice-volume history and Earth model, which was subsequently evaluated by its fit to the RSL observations. Within the range of the most-probable runs, they found low variability within the interglacial and no notable GMSL lowstand. However the variability and character of the ice-model history can have a large impact, and some of Düsterhus et al.'s[11] lower-probability runs show more sea-level variability within the interglacial.

The ice-history model inputs are important because RSL changes within an interglacial are greatly influenced by processes associated with the gradual relaxation of the solid Earth[10] following the collapse of the major ice sheets that defined the preceding glacial[3] (MIS 6). For example, the presence of large mid-latitude Northern Hemisphere ice sheets during MIS 6 will have led to the formation of extensive uplifted regions ('peripheral bulges') surrounding the ice sheets, predominantly within ocean areas. Following the demise of the ice sheets, and if there was no further change in global ocean mass, gradual subsidence of the ocean-based peripheral bulges would result in far-field sea-level fall via a process known as ocean syphoning[12] (which partly explains a local sea-level fall recorded in Western Australia[13]). If this local sea-level fall driven by the solid-Earth response was then overprinted by melting of the remaining ice sheets late in the interglacial, sufficient enough to cause a local sea-level rise, a RSL lowstand would be recorded in some locations[13].

Crucially, there is very little evidence to constrain the limits and volume of the MIS 6 ice sheets and, as such, the ice histories used by Kopp et al.[4, 5] and Düsterhus et al.[11] are reliant upon scaling oxygen-isotope curves[14]. As the authors note, other effects, such as temperature, are present within these curves[15], so they do not provide a direct analogue for total ice cover over time and, importantly, nor do they provide information on the spatial distribution of ice. Future research would therefore clearly benefit from improved constraints on the MIS 6 ice-sheet histories in order to constrain the magnitude and timing of the LIG GMSL highstand[16, 17]. Recent work has also demonstrated that global-scale dynamic topography, driven by convective mantle flow, is relevant on the timescales of the LIG, and may impact upon the reconstructed peak GMSL by several metres[18]. However, due to the slow nature of mantle flow[19], this process cannot account for multi-meter scale changes in sea level within the interglacial.

Kopp et al.[4] account for solid Earth processes when developing the GMSL reconstruction in Figure 1, but this is challenging due to currently limited ice-sheet constraints, resulting in modelling uncertainties. As a response, we instead consider whether there is any direct evidence for ice-sheet regrowth during the LIG, which may drive a >4 m GMSL fall.

Possible mechanisms of ice-sheet (re)growth

In order for there to have been an increase in global ice volume of >4 m sea-level equivalent during the LIG (see Figure 1), the Antarctic and/or Greenland ice sheets must have experienced positive mass balance and/or mid-latitude ice-sheet regrowth must have been initiated early in the LIG whilst temperatures were warmer than today[20] (Figure 2 and supplementary information). Kopp et al.[5] suggested that it is *very likely* (within 95% confidence limits) that the fastest rate of sea-level fall, prior to the final LIG sea-level decline, was 2.8 to 8.4 m/kyr (Kopp, pers. comm). We estimate that 1.15 – 3.45 million km^3 of net ice volume gain across Greenland and/or Antarctica over 1000 years (equivalent to uniform thickening at a rate of 73-220 mm/yr across both ice sheets) would be necessary to explain this *very likely* range of sea-level fall rates (see supplementary information). To place these rates in context, the volume of the modern Greenland Ice Sheet is only 2.9 million km^3. Note that these estimates assume ice discharge remained constant throughout the period of sea-level fall, and they therefore reflect minimum estimates for the rate at which ice sheets gained mass during the warmer climate and higher insolation conditions of the LIG (Figure 2b).

Mechanisms of ice-sheet inception and growth are poorly constrained and largely theoretical, but there are five inter-related hypotheses that have been invoked (Figure 3): i) enhanced precipitation[21], ii) ice saddle growth[22], iii) instantaneous glacierization[23-25], iv) marine ice transgression[26], and v) solid Earth feedbacks[27], which we now consider in turn.

Early work on the North American ice sheet complex[21] emphasised the importance of enhanced precipitation in the 'highland origin, windward growth' model (Figure 3a). This suggests that ice sheets originate as highland snow-fields that grow fastest towards the prevailing wind direction. The saddle formation hypothesis (Figure 3b) emphasises the role of topography[22] such that where two high elevation ice masses coalesce, the formation of a saddle is likely to lead to rapid growth. This involves positive feedbacks between albedo and ice-sheet elevation/precipitation, and requires the initial development of independent regional ice centres. The instantaneous glacierization mechanism (Figure 3c) emphasises the importance of upland plateaus that are close to the threshold of glaciation[23-25]. Climatic cooling could lower the regional snowline, such that plateau snowfields rapidly increase in number and size, and the expanding snowfields then thicken to become ice sheets[26]. In contrast, the marine ice transgression hypothesis[26] (Figure 3d) emphasises the importance of sea-ice formation across shallow marine basins. Here, climatic cooling leads to sea-ice formation, which in turn lowers the regional albedo. This process could theoretically allow the sea ice to thicken to form ice shelves (*sensu lato*) that become grounded in shallow water, creating marine ice domes that expand to form an ice sheet[26]. Finally, solid Earth rebound in response to ice loss (Figure 3e), particularly in regions underlain by low upper mantle viscosity (e.g. West Antarctica[28]), could lead to grounding line advance. The formation of ice rises[29] during this process could also help to stabilize the ice sheet due to their buttressing effect[30]. Such solid Earth feedbacks have recently been invoked to explain grounding line re-advance in West Antarctica during the Holocene[31], but the magnitude of the associated sea-level fall is predicted to have been an order of magnitude smaller than suggested by the Kopp et al.[4, 5] for the LIG. In considering whether one or more of these mechanisms could be important for ice-sheet regrowth during the LIG, it is important to note that these mechanisms are not mutually exclusive, and that their relative importance may vary between ice masses and/or through time. Furthermore, these are hypothesised mechanisms and the rates and magnitudes of change, under differing climate conditions, are largely unknown. Given these potential mechanisms, we evaluate the geomorphological, glaciological and modelling evidence for ice-sheet regrowth during the LIG.

Evidence for LIG ice-sheet (re)growth

Compared to its MIS 6 and 2 glacial maxima, the Greenland Ice Sheet (GrIS) was much reduced and largely terrestrial during the LIG[32], with marine-terminating glaciers likely only present in the east[33]. Model simulations of the LIG GrIS[33-36] indicate that increased air temperatures and higher insolation led to a negative surface mass balance across much of the ice sheet[37]. Radar data suggest that ice was present in central and northern Greenland, but identifying the full extent of LIG ice is methodologically challenging[38]. Offshore records suggest a restricted LIG GrIS based on pollen concentrations that are five-times higher than in the Holocene[39], while geochemical proxies indicate sustained ice surface melt[40-42]. Ice rafted debris (IRD) with a southern Greenland provenance[43] suggests centennial-scale oscillations and glacial meltwater inputs until 124 ka [44] (Figure 2f) . The low input of IRD is consistent with a smaller ice sheet compared to glacial maxima, but the presence of IRD indicates that some outlet glaciers reached the ocean. However, none of the Greenland records or models point unequivocally to periods of pronounced ice regrowth during the LIG.

Direct evidence for the extent and thickness of the Antarctic Ice Sheet during the LIG is limited[45-47]. Marine sediment records in the Ross and Weddell Seas detail evidence for collapse (and regrowth) of the West Antarctic Ice Sheet (WAIS) during the mid- and late-Pleistocene, but dating precision is insufficient to confidently attribute this to the LIG[48]. The ANDRILL record in the Ross Sea indicates that multiple cyclic fluctuations between subglacial, ice-proximal, ice-distal and open marine conditions occurred during the Pleistocene[49] but it does not appear to indicate any evidence of ice-sheet regrowth during the LIG, instead showing a transition from proximal grounding line sedimentation to distal and perhaps open marine conditions. Records from the Amundsen and Bellingshausen Seas show no evidence for interglacial WAIS collapse (nor expansion) since at least MIS 13[50-52]. Onshore evidence points to only minor changes in ice surface elevation in the central parts of the WAIS during the LIG[46, 53]. Thus, there is little direct evidence for significant changes in WAIS extent during the LIG.

Like the WAIS, direct evidence for any changes to the East Antarctic Ice Sheet (EAIS) extent during the LIG are not well constrained. In the McMurdo Dry Valleys, Taylor Glacier expanded slightly, depositing the Bonney drift during the LIG[54] when the Taylor Dome ice core also shows evidence for regional thickening of up to ~70 mm/yr[55]. In the Larsemann Hills, Princess Elizabeth Land, lake cores reveal warm biota and an active hydrological system, implying ice-free conditions for ~10,000 years prior to an abrupt transition to glacial conditions at around 120 ka[56]. Extrapolating these limited local records to understand the behaviour, and possible LIG regrowth, of the entire EAIS is not possible. Some changes in isotopic-derived temperature records within EAIS ice cores could reflect ice-elevation changes associated with ice sheet mass balance[57]. However, existing ice cores are not optimally located to unambiguously identify elevation changes associated with changes in either the WAIS, nor in the Aurora, Wilkes or Recovery basins of the EAIS, where any changes might first be expected[57-60].

Two recent Antarctic modelling studies simulated significant interglacial ice-volume change. One includes rapid ice-shelf loss and subsequent retreat via an ice-cliff fracturing mechanism coupled with atmospheric warming[60]. The outputs are generally consistent with Kopp et al.'s[4] estimate for the Antarctic contribution to GMSL during the LIG, but produces no notable mid-LIG RSL lowstand. The second proposes that a combination of marine ice-sheet instability, warm subsurface ocean temperatures, and variations in surface accumulation, could result in a double-peaked contribution of WAIS melt to LIG sea level[61]. By invoking a potential subsurface ocean-cooling episode, flanked by increases in the surface mass balance, the model outputs suggest a potential recovery in ice volume between the two stages of collapse. However, this requires a +2 to +3^{o}C Southern Ocean temperature anomaly, which is greater than the +1.2^{o}C recorded by the palaeo data[62], but perhaps not implausible given the average 2.6^{o}C 2-sigma uncertainty stated for those proxy records[62].

The LIG was warmer than present, and models that explore Antarctic ice-sheet response to future warming suggest that enhanced precipitation, particularly in the EAIS, would increase surface mass balance[63] but that this would be counteracted by enhanced surface melt near the coast[64] resulting in limited net change in ice-sheet volume. Thus, although there is some evidence for a very minor advance of marginal areas of the EAIS during the LIG, there is no unequivocal evidence of significant regrowth. Only by invoking ocean warming >2^{o}C is it possible to simulate a double-peak in sea level due to changes in Antarctic ice volume[61].

Beyond the polar regions, the general consensus is that there was no Laurentide Ice Sheet (LIS) during the LIG[65], with most evidence pointing towards inception after 120 ka[66], i.e. after the period of proposed GMSL fall (Figure 1). An abrupt drainage event recorded in a proximal marine core in the Labrador Sea is attributed to former glacial Lake Agassiz in the Hudson Bay region at c.124.5 ka[67], which may provide indirect evidence of some Laurentide ice early in the LIG, but not for regrowth within the LIG. The primary evidence for a general absence of LIG Laurentide ice comes from dated organic-rich sediments in the Hudson Bay Lowlands[68], which could not have formed under ice cover. Whilst different dating techniques have provided slightly different ages for these sediments, recent work[68] gives clusters of ages that suggest the area was ice-free during the middle of the LIG.

In Eurasia, it is generally accepted that there was an extensive MIS 6 ice sheet, but there is only limited and imprecise dating of the ice retreat prior to the LIG. Along the northern margin of Eurasia there is sedimentary evidence for a major marine transgression during the LIG and a complete disappearance of the Eurasian Ice Sheet (EIS)[69, 70]. Warm conditions and the absence of glaciation are also recorded by inter-morainic organic deposits on the Scandinavian Peninsula: the main inception area of the Fennoscandian sector of the EIS[71, 72]. As with the LIS, rapid EIS growth is documented following the MIS 5e-d transition[73, 74], but the EIS is an unlikely candidate for driving sea-level fluctuations within the LIG.

Evidence relating to smaller ice masses such as plateau icefields and valley glaciers in Alaska, Patagonia, Iceland, the European Alps, Siberia, the Himalaya and Southern Alps is sparse, with no direct evidence for LIG ice in these regions[75-78]. Moreover, although there is potential for LIG ice to have been present in mountainous regions covered by ice caps and glaciers today, their small net volume[7] (total modern mean sea-level equivalent 0.4 m) makes it unlikely that these small ice masses could drive multi-metre GMSL fluctuations during peak LIG warmth.

In summary, there is no empirical evidence to support the hypothesis of ice-sheet regrowth of sufficient magnitude to explain the putative GMSL fluctuation. The only potential sites where ice-sheet regrowth of a significant volume could have theoretically occurred during the LIG are Antarctica and Greenland, as other ice masses were either absent or too small to account for a multi-metre sea-level fall. In addition, the sea-level fluctuation cannot be explained by an out-of-phase melting of Greenland and Antarctica during the LIG, as there must be a period of net ice increase to cause a fall in GMSL. The absence of evidence of ice regrowth may also reflect the difficulty of constraining pre-LGM ice-sheet dynamics where subsequent glacial advances may have overprinted or erased evidence. However, the lack of evidence for regrowth is largely consistent with modelling studies (perhaps with the exception of Sutter et al.[61]), which are unable to model ice-sheet driven sea-level variations on millennial timescales during the LIG[79].

The role of thermal expansion

The absence of geomorphological or stratigraphical evidence for ice-sheet regrowth during the LIG leads us to consider the only other mechanism that might explain the GMSL fall as presented in Figure 1; thermosteric sea-level change. Kopp et al.[4] estimate a thermosteric component in their modelling, which varies with global ice volume (-1.6 ± 0.6 m per 100 m equivalent sea level ice-sheet growth). Based on a compilation of LIG SST data and a coupled atmosphere-ocean climate model, McKay et al.[80] suggest that it is unlikely that thermosteric sea-level rise exceeded 0.4 ± 0.3 m during the LIG. On millennial timescales, the equilibrium response of ocean thermal expansion to warming has been estimated as 0.2 to 0.6 m/°C [81]. If simply inverted for cooling and the consequent thermal contraction (without including potential changes in ocean stratification), a 4 m sea-level fall (Figure 1) would require a ~6-20^{o}C global temperature drop during the LIG, which is clearly not evident[62]. Regionally cooler SSTs, such as in the Nordic Seas[82], may have driven a small, localised thermosteric contraction, but not one which resulted in a multi-metre global sea-level fall.

Last Interglacial relative sea-level data

The absence of direct evidence for ice-sheet regrowth or thermosteric cooling sufficient to drive a >4 m sea-level fall during the LIG, leads us to review the data, which underpins the GMSL curve in Figure 1. Reconstructions of GMSL must be supported by local RSL data which includes a location, age, elevation (both the measured elevation of the sample and the modern relationship to the tide level at which such an indicator would form today), and ideally a summary of whether the indicator describes an increase or decrease in marine influence[83, 84]. The Kopp et al.[4] dataset comprises 108 RSL observations from 47 sites. The highly-fluctuating Red Sea record[8] (based on stable-oxygen isotopes of planktonic foraminifera) provides 29 of the sea-level observations in the dataset and plays an important role in anchoring the timescale (although the timing of the highstand has since been revised by 6300 years[85]). Kopp et al.[5] show that their reconstructed GMSL fall is not precluded by the inclusion of the Red Sea data. Outside of the Red Sea basin, the most widespread LIG sea-level archives (both within the Kopp et al. database and in records published since) are preserved in low-latitude regions. They are therefore typically based on fossil corals and reef terraces[10, 86, 87], some of which provide potential evidence for one or more local RSL falls[86, 88, 89]. A recent comprehensive analysis of modern coral distributions[90] has shown that there is no direct relationship between coral growth and water-depth *per se,* and that coral-depth distributions are clearly variable. This suggests notable uncertainties in the elevation of former sea levels reconstructed by some records[86, 88, 89, 91]. Reworked corals are also common in Barbados, which can result in erroneous interpretations of past sea level[92]. Furthermore, using coral archives as LIG sea-level records is complicated by uncertainty surrounding the use of open- versus closed-system U-Th ages[93], with a compilation of closed-system U-Th ages from corals in West Australia, Bahamas, Bermuda and Yucatan providing no evidence for a local sea-level fall within the interglacial[10].

These issues raise an important question: after accounting for solid Earth processes, is there any clear empirical evidence for a global mean sea-level fall of >4 m during the LIG? In the Seychelles - a site predicted to approximately record GMSL[94] - there appears to be a regionally-consistent interruption to LIG reef growth, potentially indicative of a stable or briefly falling RSL, or alternatively caused by changes in local accommodation space or storm bleaching[87, 95]. A mid-interglacial falling RSL is suggested by a survey of palaeo-shorelines in Western Australia, but the authors explain this by regional solid Earth processes[13]. In Southern Australia, sedimentary evidence points to a single phase of sea-level rise[96], while on the Yucatán peninsula, Mexico, LIG RSL was stable until a rapid late-interglacial sea-level rise[97]. Such records could be produced if falling GMSL was matched by local land subsidence. However, given that GMSL fall is estimated to be ~3-8 m in 1000 years, subsidence would need to be very rapid to prevent a hiatus being recorded in the stratigraphy at these sites. A Mediterranean facies succession, previously interpreted to record a double LIG highstand[88], has recently been reinterpreted as providing no evidence of a LIG RSL fall, instead recording highstands from two separate interstadials[98]. In temperate-latitude locations there is no evidence for local LIG RSL oscillations[2, 99]. Without unequivocal empirical evidence for locally falling sea level[96, 98], and considering the uncertainties in coral-growth distributions[90], the local RSL data provides limited support for >4 m fluctuations in GMSL during the middle of the LIG.

Summary

In conclusion, reconstructions of GMSL during the LIG[4, 5] have raised the intriguing possibility that fluctuations in ice-sheet volume occurred within the interglacial, that is, ice sheets regrew and then decayed. We have considered several possible driving mechanisms, acting alone or in combination, for multi-metre changes in GMSL during the LIG. We find that the current understanding of ice-sheet histories during MIS 6 is not adequate enough to rule out the possibility that limitations in the modelling of the solid Earth response could be contributing to the appearance of a GMSL fall during the LIG[3, 11]. However, if the GMSL fall was driven by changes in ice sheet mass balance, it would require 1.15 – 3.45 million km^3 of ice to form in less than 1000 years; we found little geomorphological or sedimentary evidence for such substantial ice-sheet regrowth during the LIG. It is also clear that there remain large uncertainties associated with the interpretation of some local RSL data that underpin the reconstructed GMSL curve. Taken together, our analysis leads us to question the occurrence of a rapid GMSL fall within the LIG, which also raises important questions about the reconstructed very high rates of GMSL rise following the lowstand; reported to be approximately 3 to 7 m/kyr [5].

We conclude that it is critical that future reconstructions of GMSL during the LIG include a range of realistic ice-sheet scenarios from the preceding glacial (MIS 6); take into account the impact of dynamic topography on the reconstructed elevations of former RSLs; and assemble a geographically and temporally widespread dataset of local RSL, with careful interpretation of fossil sea-level indicators with respect to tidal datums and accurate chronologies. Until these issues are better resolved, we would urge caution in using rates of GMSL rise from the LIG to project future sea-level changes.

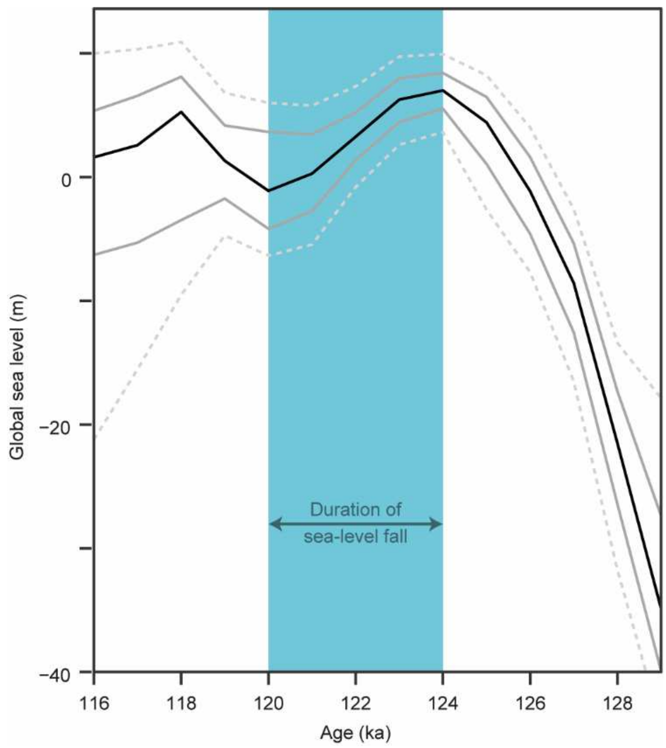

Figure 1: Probabilistic reconstruction of global mean sea level by Kopp et al.[4] during the Last Interglacial. The solid black line = median value; grey lines = 16th and 84th percentiles; dotted grey lines = 2.5th and 97.5th percentiles. Duration of the sea-level fall depends on the uncertainties. It is *very likely* (95% probability) that the fastest rate of sea-level fall, prior to the final LIG sea-level decline, was 2.8 to 8.4 m/kyr (Kopp, pers. comm). The rates of GMSL rise that are inferred to have followed the lowstand are *likely* (i.e. 67% probability) between 3 and 7 m/kyr [5].

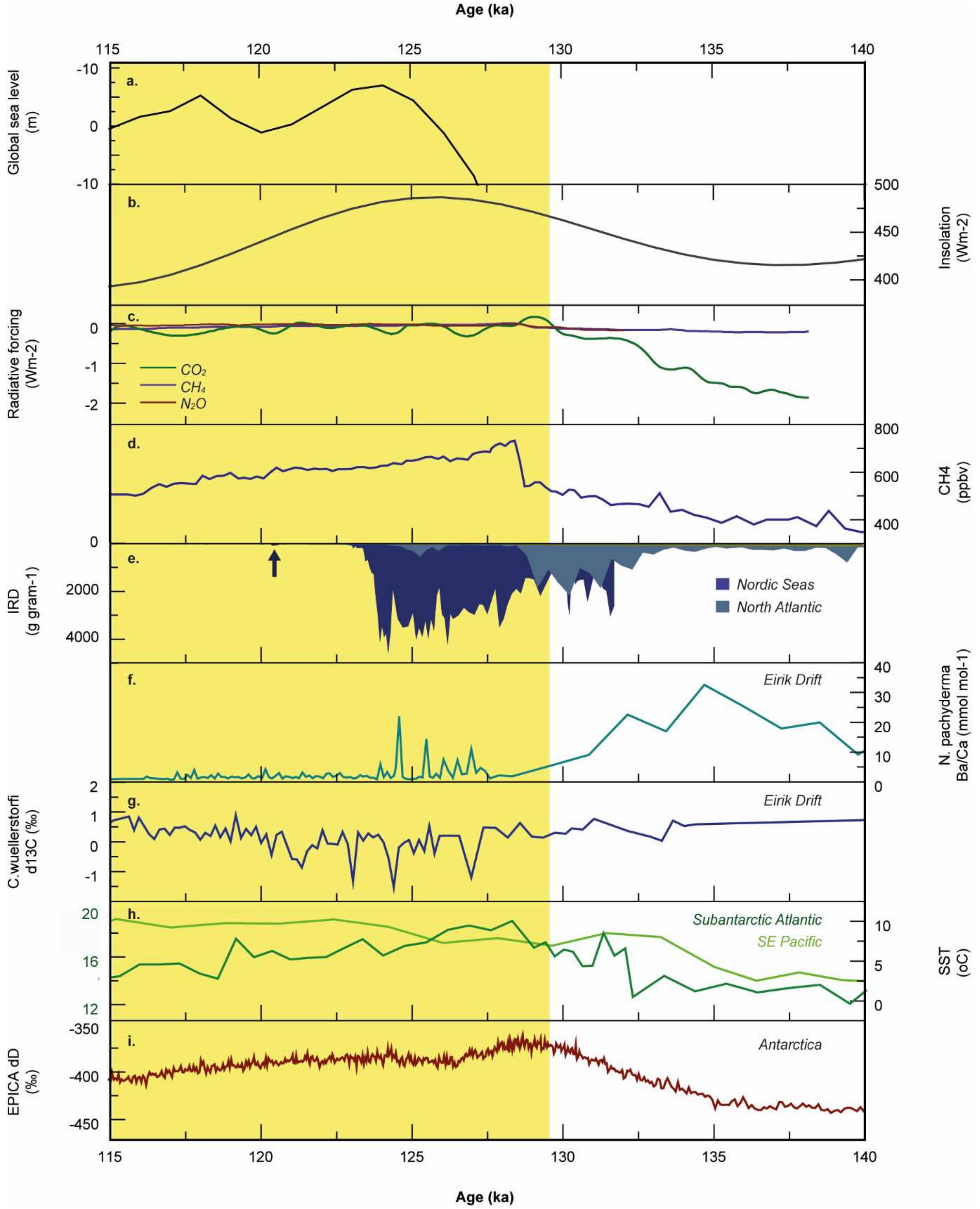

Figure 2: Selected records indicating climate changes and/or ice-sheet changes through the LIG. MIS 5e is shaded yellow. (a) global sea-level reconstruction (median value) [4] **;** (b) July insolation at 65°N[100]; (c) radiative forcing variations driven by the major greenhouse gases[101]; (d) Antarctic methane concentrations[102], which reflect Greenland temperature [62]; (e) IRD from the Nordic Seas[82] and North Atlantic[103]. Bauch et al.[82] highlight a small increase in IRD at 120-121 ka (marked by arrow); (f) planktonic foraminifera Ba/Ca ratios from Eirik drift, a proxy for Greenland ice-sheet melt[44]; (g) Eirik Drift benthic $\delta^{13}C$ (3442 m), a proxy for AMOC strength[44]; (h) SE Pacific SSTs, offshore of Patagonia[104], and Subantarctic Atlantic SSTs[105]; (i) EPICA δD as a proxy for Antarctic air temperature[106].

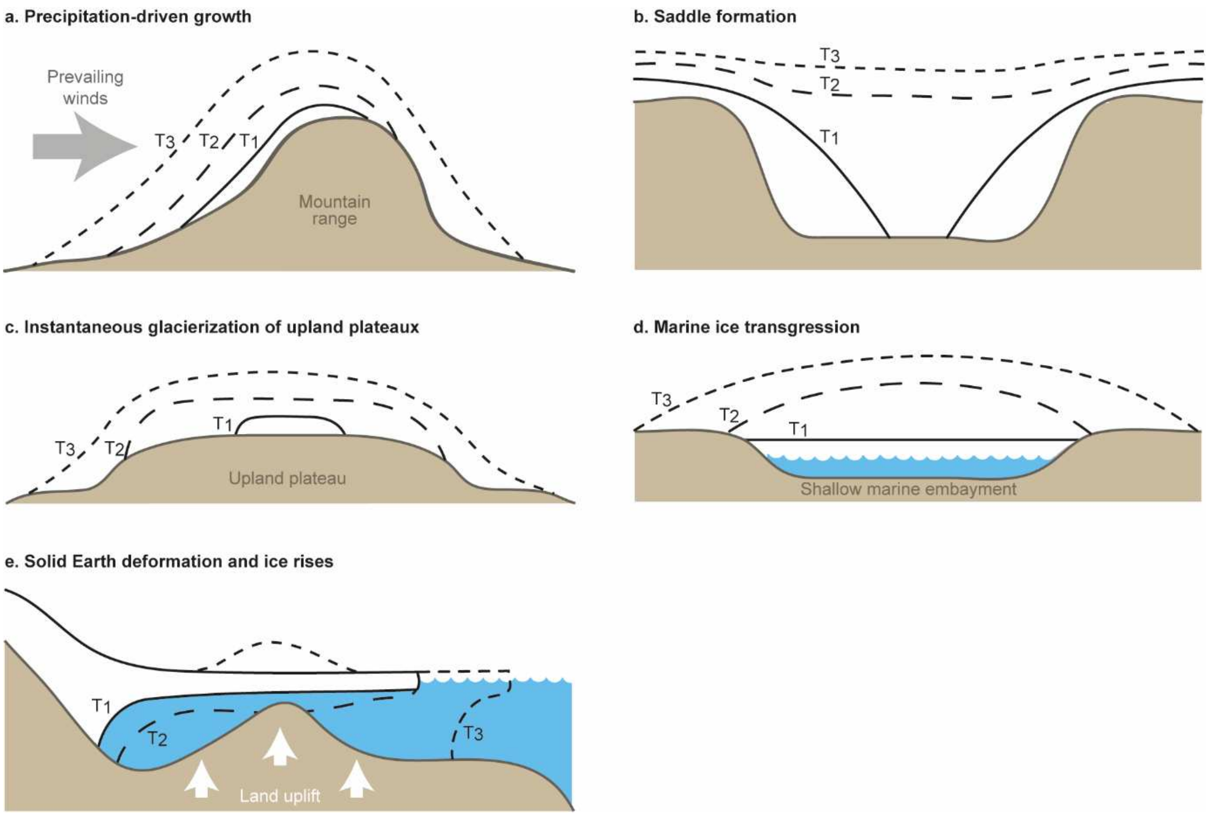

Figure 3: Theoretical mechanisms of ice-sheet growth and inception at three successive time steps (T_1, T_2, T_3). Corresponding author

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Author contribution statement

NLMB and ELM conceived and led the study. PLW conducted the GIA modelling. All authors contributed ideas and to the development and writing of the paper.

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Supplementary information

Within sections 1-3 of the supplementary information we set out the climatic context within the Last Interglacial (LIG) in which sea-level oscillations and ice sheet regrowth (as outlined in the main paper and in section 4 here) may have occurred. A number of recent syntheses have focussed on LIG climate, either overall or for discrete time slices[1-3], and we provide a key summary in Figure 2 in the main paper.

1. Radiative forcing within the Last Interglacial

Overall, the LIG had a warmer than modern climate, consistent with higher insolation forcing[2] (Figure 2b). Radiative forcing associated with varying greenhouse gases through the last glacial-interglacial cycle, shown in Figure 2, was calculated by Schilt et al.[4]. The results demonstrate that during the last interglacial (LIG) the overall radiative forcing was comparable to modern, driven largely by CO_2 concentrations which were similar to the Pre-Industrial (~280 ppmv). Millennial-scale oscillations in CO_2 contribute to small-amplitude (<0.5 Wm^{-2}) changes in radiative forcing during the LIG. A rapid increase in CH_4 occurs during the latest stages of the deglaciation, reaching a peak at 128 ka before declining, but the impact on radiative forcing is minimal[4]. However, this pattern of CH_4 concentrations through the LIG has been linked to northern hemisphere temperature changes: during millennial-scale climate events during the glacial stage, high CH_4 concentrations align with increases in Greenland temperatures[5]. Thus, it has been proposed that the CH_4 pattern over the LIG represents the northern hemisphere warming occurring later and more abruptly than in the south (as shown by Antarctic ice core δD), then gradually cooling towards MIS 5d[1].

2. Marine archives of Last Interglacial climate

In the oceans there is asynchrony in the timing of peak warmth: the Southern Hemisphere oceans warmed earlier (130 ka) than the Northern Hemisphere (125 ka)[1]. Several northern high-latitude ocean records indicate pronounced (>4°C) but short (~1-2 kyr) sea surface cooling episodes, e.g. at 120 ka [6] and between 129-116 ka [7] in the Nordic Seas (Figure 2e), and at 125-124 ka and between 122-120 ka on the Eirik Drift[8]. However, the overall temporal resolution of the marine records tends to limit our ability to identify millennial-scale variability in sea surface conditions (Figure 2).

There is evidence for LIG variability in the strength of Atlantic Meridional Overturning Circulation (AMOC), which transports heat and moisture to the ice sheet source regions around the North Atlantic and Nordic Seas. By driving the bipolar seesaw[9], changes in AMOC could explain the different temperature trends observed in the northern and southern polar regions during the LIG[1]. Centennial-scale fluctuations in AMOC during the LIG are indicated by benthic foraminifera $\delta^{13}C$ oscillations, which are either of comparable magnitude to the Holocene (in the northeast Atlantic and Nordic Seas[6, 7, 10, 11]), or larger than the Holocene and equivalent to glacial-stage oscillations (e.g. on the Eirik Drift[8, 10], Figure 2g). On the Eirik Drift, the centennial-scale reductions in North Atlantic Deep Water (NADW) formation between 126 and 123.5 ka correlate with periods of ice rafting, meltwater input, and a southward migration of the Polar Front[10], highlighting both marine and ice-sheet (likely Greenland) variability within the LIG.

During peak LIG warmth in the Atlantic Sector of the Southern Ocean (128-125 ka) (Figure 2i), there is evidence for both cyclical cold periods of 200-300 year duration, interpreted as solar-driven, meltwater-induced coolings of 1.0-1.5°C, as well as non-cyclic, 1-2°C amplitude sea-surface temperature (SST) fluctuations on 2-4 kyr timescales[12]. From 125 ka, the amplitude of the SST variability decreased and Southern Ocean temperatures began to fall in a stepwise manner to temperatures slightly warmer than today, in tandem with apparent strengthening of AMOC[13]. In considering the links between offshore and onshore conditions, it is important to note that reconstructions of Antarctic ice core terrestrial air temperature anomalies relative to modern for the LIG (+3.5°C) exceed those recorded in the Southern Ocean (+1.2°C on average)[1].

3. Terrestrial archives of Last Interglacial climate

Acquiring terrestrial palaeoenvironmental reconstructions that span the LIG is problematic. Suitable sites are rare, and it is difficult to produce robust, independent and accurate chronological control to enable record comparisons. With this in mind, a short review on well constrained, continuous records from regions that could potentially experience ice sheet regrowth are described below. We do not seek to provide a comprehensive review of LIG climate, which others have undertaken[1, 2, 14], but target records which reflect either potential climate forcing of ice-sheet regrowth and/or environmental signals of ice-sheet behaviour.

The onset of MIS 5e in the terrestrial record appears globally near-synchronous, initiating between 129-127 ka BP and generally marked by a switch from a dominance of cold-climate species and open vegetation to increasing abundances of thermophilous species. Climatic optimum is reached by 125-124 ka[15] and in all records the species assemblage suggests a climate that was warmer than the Holocene[16]. Two records from Europe[17] and Siberia[18] suggest the initial warming trend is interrupted by a short-lived (up to 400 years long) reversal occurring ca. 127-126 ka BP. However, more widespread was the occurrence of the Intra-Eemian Cold Event (IECE) that occurred sometime between ca. 124-120 ka, although the exact timing is record dependant[15, 18-20]. In most records this was a short-lived (~400 years long), low amplitude (1-2°C) cooling event with temperatures falling to values similar to modern. In pollen records from Europe this is often marked by an increase in the species *Carpinus*[21]. Two temperature reconstructions based on pollen and chironomid records from Sokli in northern Finland have suggested that a period of abrupt cooling occurred at ~120 ka BP. This event lasted between 500-1000 years with summer temperatures cooling by 2-4 °C. Even taking the temperature uncertainties into account, this would make local climate cooler than that at present[22]. However, evidence of the IECE is far from conclusive in Europe, with some records also suggesting climate stability through the LIG[23]. This discrepancy is partly explained by the limitations of quantitative temperature reconstructions, including the lack of appropriate modern analogues to reconstruct temperature from pollen assemblages, and disagreements over the climatic interpretation of the species *Carpinus*[21]. In all of the records outlined above that contain evidence of a climate oscillation, the post-cool event climate did not return to the optimum conditions that were experienced in the earlier part of the interglacial.

Outside of Europe, in North America[16, 24-27], South America[28-31], New Zealand[32-34], and Tasmania[35] a relatively stable LIG climate is recorded, with temperatures warmer than present, although the magnitude of LIG warmth is variable. In the Southern Hemisphere temperatures are thought to have been, at most, 1-2 °C above modern temperatures. However, evidence in northern North America point toward an extreme warming that possible reached 6-7°C above present during the LIG. The available terrestrial evidence thus indicates that climatic oscillations within the LIG were largely restricted to Europe and Siberia. However, it must be noted there is a relative scarcity of terrestrial records spanning the LIG in areas of likely ice-sheet growth beyond Europe and Siberia. Speleothems can also provide archives of climate continental conditions, but providing a quantitative estimate of temperature changes is a challenge, with records dominated by changes in water availability[36]. A composite speleothem record from New Zealand suggests peaks in $\delta^{18}O$ at ca. 128, 125 and 121 ka[34], and an Austrian alpine stalagmite records higher $\delta^{18}O$ at 130.7-130.0 ka and 125.7-118.2 ka[37], which may potentially be evidence of locally higher temperatures. There is a general consensus that a return to colder environments towards the end of the LIG began between 119-117 ka BP, particularly marked by a replacement of forest species by open vegetation[16, 17, 26, 38].

4. Calculating rates of mass balance change

Modelling by Kopp et al.[39] suggests that it is *very likely* (within 95% confidence limits) that the fastest rate of relative sea level fall during the LIG, prior to the final sea level decline, was between 2.8 and 8.4 m/kyr (Robert Kopp, personal communication). To place these rates into context, we have calculated the ice thickness increase across the Antarctic and/or Greenland Ice Sheets that would be required to produce the observed rates of LIG sea-level fall (Table S1). The extent of the Antarctic and Greenland ice-sheets during LIG is poorly constrained (as discussed in the main text), and therefore we simply use the area of the present-day ice sheets to convert ice volume to ice thickness. We do, however, account for the difference in the area of the ocean at 125 ka BP (total ocean area: 366 million km^2), as determined from a small suite of GIA models. The ICE-5G[40] deglaciation model was used to simulate two full glacial cycles. The length of the LIG was varied between 7 and 13 ka to reflect uncertainty in its duration. In addition, three different rheological models were used to reflect uncertainty in Earth properties. The range of predicted values for ocean area has a negligible effect on the results (differences in ice thickness change are <1 mm/yr), and therefore mean values are given in Table S1.

Ice-sheet thickening required	**2.8 m/kyr sea level fall**	**8.4 m/kyr sea level fall**
Across Antarctic Ice Sheet only	0.082 m/yr	0.246 m/yr
Across Greenland Ice Sheet only	0.676 m/yr	2.027 m/yr
Across both Antarctic and Greenland ice sheets	0.073 m/yr	0.220 m/yr

Table S1: Estimates of the rate of ice sheet thickening over 1000 years needed across the Antarctic and/or Greenland ice sheets to explain rates of mean sea-level fall of 2.8 and 8.4 m/kyr during the LIG.

On a global scale, considering that mass balance changes in ice sheets beyond Antarctica and Greenland may contribute to the recorded sea-level lowstand, a 2.8 m/kyr sea level fall requires a sustained increase in global ice volume of ~1150 km^3/yr for 1000 years (total global ice volume increase over 1000 years: ~1,150,000 km^3). At the maximum *very likely* range[39], an 8.4 m/kyr sea level fall requires a sustained increase in global ice volume of ~3,450 km^3/yr for 1000 years (total global ice volume increase over 1000 years: ~3,450,000 km^3 i.e. greater than the size of the modern Greenland ice sheet).

Supplementary Information References

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