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Moisture amplification of the high-altitude deglacial warming

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Abstract

In response to anthropogenic warming, glaciers are shrinking almost everywhere, endangering water accessibility in areas located downstream. Glacier fluctuations are at first order controlled by local precipitation and temperature, but large uncertainties persist on the potential role of local moisture in amplifying or dampening temperature changes at high-elevation. Here, we combine glacier extents and Sea Surface Temperature (SST) during the Last Glacial Maximum (LGM) to quantify altitudinal thermal gradients (lapse rate) from 40°N to 40°S along the American Cordillera. We also constrain modern lapse rates based on present day temperature and SST database to explore how the lapse rate has changed since the LGM along this North South transect. Based on proxy-based quantitative paleo-precipitation estimations above 2000 m, we investigate how these lapse rate changes compare with moisture modifications around the Cordillera and discuss the mechanisms that potentially controlled lapse rate changes during the post-LGM deglacial warming.

We find that lapse rate changes are linearly related to changes in precipitation and derive a quantitative relationship between these two parameters. To further explore the processes involved in controlling lapse rate variations, we use the IPSL global climate model outputs, for the LGM and pre-industrial states in this region. The IPSL model also yields a shallower modern lapse rate in the wetter tropical region, confirming the observed correlation between precipitation changes and lapse rate variations. The IPSL model also supports a close coupling of continental relative moisture and mean annual precipitation in the studied area, indicating that moisture is involved in the precipitation – lapse rate relationship. Our results suggest that future warming may be enhanced in high altitude regions where precipitation is expected to increase. Using our most reliable relationship linking precipitation and lapse rate changes, we conclude that, assuming a 1°C warming at sea level, a mean annual precipitation increases of 500 mm.a⁻¹ could lead to a warming amplification of 4.1 ± 0.8°C at 4000 m asl (mean elevation of modern glaciers). In this case, a 2°C warming at sea level would yield >6°C degrees warming at 4000 m asl. This study therefore confirms that special attention should be given to the climate projections of glacier melting in tropical and mid latitude regions.

Main text

1 - Introduction

Global warming massively disrupts the cryosphere at low and mid-latitude, as alpine glaciers shrinking is now accelerating worldwide (Hugonnet et al., 2021; Thompson et al., 2011). Over the last two decades, melting of mountain glaciers (excluding ice cap) has contributed to 21±3% of the global sea level rise, i.e. more than Greenland and Antarctica considered separately (Hugonnet et al., 2021; Slater et al., 2020). In mountainous regions where precipitation is highly seasonal, alpine glaciers
represent the main water supply during the dry season (Kaser et al., 2010; Pritchard, 2019; Rabatel et al., 2013). It is therefore crucial to identify factors that modulate glacier melting in order to better understand and anticipate their stability, for the decades to come (Kraaijenbrink et al., 2017; Zekollari et al., 2019).

The annual mass balance of a glacier is primarily controlled by mean local temperature and precipitation (Condom et al., 2007; Ohmura, 1992). Globally, the temperature increase associated with anthropogenic fossil fuel combustion recently surpassed +1°C, at an average warming rate of +0.18 °C/decade (Intergovernmental Panel on Climate Change (IPCC), 2022). This warming rate is not spatially homogenous: it is notably amplified over the continent, reaching on average 1.4°C in the 40°S-40°N zone at low elevation (Seltzer et al., 2023a). At high elevation, in mountainous areas, anthropogenic warming has reached up to +0.7°C/decade (Intergovernmental Panel on Climate Change (IPCC), 2022). However, high elevation warming rates are very variable from one region to another, ranging between +0.1 and +0.7°C/decade above 2000 m ((Pepin, 2015). Many potential causes for high-altitude temperature amplification have been proposed, including surface albedo feedbacks, latent heat release, moisture-enhanced radiative changes, and aerosol feedbacks (Pepin, 2015). These mechanisms may at least partly explain the contrasting regional patterns of high-altitude warming, but the relative contributions of each of these forcing remain unknown.

Several studies have hypothesized that warming at high altitude should be greater than observed at sea level because warmer conditions at the regional scale will increase the atmospheric moisture content and yield a shallower lapse rate. The modern global annual average free atmospheric lapse rate (LR) is -6.5°C.km⁻¹, but it locally varies between the moist adiabat value around 4-6°C/km and the dry adiabat value of -9.8°C.km⁻¹ (Stone and Carlson, 1979). The temporal and spatial LR variations remains poorly known. Accurate and precise reconstructions of paleo LR under different global and regional paleo-climate conditions are useful to quantify and understand how the atmospheric moisture content may affect LR variations at the regional scale (Blard et al., 2007a; Loomis et al., 2017; Tripati et al., 2014a).

Recent compilations of oceanic and continental paleothermometers proposed that the global post-Last Glacial Maximum (LGM) warming was between 4-6°C, with large regional disparities (Annan et al., 2022; Seltzer et al., 2021; Tierney et al., 2020). Even if the exact amplitude of the post LGM warming could still be refined in future studies, the LGM climate is an ideal benchmark for studying the sensitivity of global temperatures to the net radiative forcing, and the specific responses of certain regions, partly driven by the polar amplification (Smith et al., 2019), the terrestrial amplification (Seltzer et al., 2023) or by the altitudinal dependent warming (Pepin et al., 2015). LGM temperature reconstructions at high elevations have yielded contrasting results: some tropical and mid latitude regions display an amplified post-LGM warming at high elevation (steeper LR during the LGM) (Blard et al., 2007a; Kuhlemann et al., 2008; Loomis et al., 2017), while others suggest no dependence of the post-LGM warming on elevation (i.e., no LR change since the LGM) (Tripati et al., 2014a). These
studies provide local reconstructions in different regions of the globe, with contrasting hydrological regimes. Thus, they may not be globally representative, hampering our ability to anticipate potentially upcoming LR changes under future climatic conditions.

In this study, we aim to fill this gap by investigating the main physical parameters and mechanisms controlling LR changes. For that, we compare reconstructed precipitation and LR data between the LGM and present. We focus our study on the American Cordillera, along a near-continuous, 8000 km-long latitudinal transect ranging from 40°N to 40°S. From Alaska to Patagonia, the American Cordillera is the sole mountain range that continuously covers both Hemispheres over a large latitudinal range (70°N – 50°S, Fig. 1), with a mean altitude of 4000 m and summits above 6000 m asl. The American Cordillera thus provides a unique collection of glaciers, encompassing both the tropics and mid-latitude regions. The Cordillera also have sub-regions with very contrasting precipitation regimes, which is ideal to test the potential impact of moisture variations on LR changes. It is also located near the Eastern Pacific Ocean, where records of SSTs permit constraining low altitude LGM temperatures. Local continental precipitation proxies (e.g., lake levels, speleothems, and pollen records) are also available in the vicinity of LGM paleoglaciars. Here, based on glacier paleo-extents, we reconstruct LGM equilibrium lines of the American cordillera glaciers and convert them into 0°C isotherm altitudes using independent estimates of paleoprecipitation, following the approach of Condom et al., (2007) and Legrain et al., (2022). We then combine these paleo-isotherm estimates with alkenone-based paleo SST data to compute LGM LR estimates along the latitudinal transect. These LR estimates are then compared with present-day LR values (computed with a comparable methodology) to identify where and how LR have changed during the last deglaciation. The mechanisms causing the observed spatial and temporal LR variability are then investigated using the IPSL global climate model outputs, for LGM and pre-industrial states.

2 – Materials and Methods

To reconstruct LR variations between the LGM and today, our approach includes four main steps (summarized on the synoptic flow diagram presented in Supplementary Fig. 1), which involve both already published and newly computed data:

1. Identifying sites with LGM glacial extents, well-dated by cosmogenic ^10^Be, ^36^Cl or radiocarbon (section 2.1.). The locations of the paleoclimatic data used in this study are presented in Supplementary Figs. 3-4).

2. Compiling /determining modern (section 2.2.) and LGM (section 2.3.) SST, 0°C isotherm and precipitation for each site considered. For the LGM, this step notably requires reconstructing LGM ELAs using the Accumulation-Area Ratio (AAR) method and
converting these into LGM 0°C isotherms using empirical relationship linking ELA, temperature, and paleoprecipitation data (Fox and Bloom, 1994; Legrain et al., 2022).

3. Converting SST data into equivalent continental sea-level temperatures at the sites where glacial landforms are studied using an empirical transfer function (section 2.4.).

4. Computing modern and LGM lapse rates (with LGM LR defined as the slope between SST – taken at -125 m below present-day sea level – and the glacier-derived isotherm 0°C elevation) to derive ΔLR values by subtracting the LGM LR from the modern LR (section 2.5.).

We then proceed to a model-data comparison (section 2.6.) using outputs of the IPSLCM5A2 model (Sepulchre et al., 2020) to investigate the underlying mechanism(s) controlling LR temporal and spatial variations.

2.1 - Studied LGM glacial landforms and their ages constraints

This study is based on the compilation of well-dated LGM landform extents dated by in situ cosmogenic nuclides (27/31 sites) and radiocarbon (4/31 sites) (Supplementary Tabs. 1 and 2). The inclusion criteria for these LGM glacial extents is to consider that every glacial morphology dated between 26-18 ka belongs to the LGM. This 26-18 ka range represents the broad duration of the global LGM (Clark et al., 2009). Considering this whole time window has the advantage to account for the inter-site age variability produced by the slight diachronism of the so-called “local” LGMs identified along the American Cordillera (Palacios et al., 2020).

10Be cosmogenic data used here comprise all the published LGM ages stored within the online ICE-D database http://alpine.ice-d.org/. All these 10Be exposure ages are updated and homogenized using the KNSTD07 standardization (Nishiizumi et al., 2007), and are recomputed with the online calculator Cosmic Ray Exposure Program (CREp) https://crep.otelo.univ-lorraine.fr/#/ (Martin et al., 2017) using the regional production rate of the Andes (Martin et al., 2015), as well as the world wide mean value, for comparison (Supplementary Tab. 2) (Martin et al., 2017). For comparison, we also compute these 10Be ages using two different scaling schemes: Lal stone time corrected (Lal, 1991) and Lifton-Sato-Dunai (Lifton et al., 2014) (Supplementary Tab. 2). In all cases, we use the ERA40 atmosphere model and the atmospheric 10Be based Virtual Dipolar Moment (VDM) (Muscheler et al., 2005) to account for time variations of the production rate. Our comparison shows that the choice of the scaling and production rates does not affect the final ages beyond uncertainties (Supplementary Tab. 2).

We retain the ages computed with the Lal stone time variable scaling with the regional Andes production rate (Martin et al., 2015). In the few cases where no regional production rate is available (southern Andes), we use the 10Be ages computed with the worldwide mean production rate (Supplementary Tab. 2). 36Cl ages of landforms (n=7) are considered without any additional processing (Supplementary Tab.
as there is no consensus on $^{36}$Cl elemental production rates (Marrero et al., 2016; Schimmelpfennig et al., 2009).

When no direct cosmogenic surface exposure age of a given glacial morphology is available (as is the case for 4 sites in Ecuador and Colombia; Supplementary Tab. 1), the age of the morphology is determined by $^{14}$C dating on paleosols bracketing the glacier moraines.

2.2 - Modern temperatures and precipitation at the studied sites

2.2.1. Modern Sea Surface Temperatures

We use the modern SST data from the Extended Reconstructed Sea Surface Temperature version 4 (ERSST_V4) (Huang et al., 2015; Smith and Reynolds, 2003), available at https://www.esrl.noaa.gov/psd/. This database is based on SST values extended from the Comprehensive Ocean–Atmosphere Data Set (COADS) database. Modern SSTs correspond to the mean annual value from the climatic period 1989-2018 (Supplementary Tab. 3). Data are compiled for each 0.5 latitudinal degree from 40°N to 40°S along the American Pacific coast and the Venezuelan and Colombian Caribbean coasts. We use the Pacific Ocean SSTs, except for the Santa Maria Colombian site and the Venezuelan sites, that are closest to the Caribbean Sea (Supplementary Tab. 3). At these two locations, we compute a mean SST between the Pacific and the Caribbean zone. It is noteworthy that SST values from the Pacific Ocean and the Caribbean Sea are similar ($\Delta$SST ≤ 1°C) at these latitudes (12-9°N). Thus, the potential bias induced by combining the Pacific Ocean and Caribbean Sea data is not considered as a significant source of uncertainty in our study.

2.2.2. Modern 0°C isotherms

Modern 0°C isotherms are directly compiled from literature data (Supplementary Tab. 5), except for the Sierra Nevada and San Bernardino sites, for which no previous estimate of the 0°C isotherm is available. For the sites from Sierra Nevada and San Bernardino, we estimate the 0°C isotherm using normal PRISM reanalysis data (1980-2011). We extract data for 118 sites of high-altitude glaciers and perform a regression of elevation as a function of mean annual temperature (Supplementary Fig. 6), yielding an averaged value of 0°C isotherm for the whole range of 3578 ± 22 m (2σ, standard deviation). Computed and compiled modern 0°C isotherms range from 3452 to 5100 m with an average of 4530 ±587 m (1σ standard deviation). We estimate an approximate uncertainty of 50m for modern 0°C isotherms directly compiled from literature.

2.2.3. Modern precipitation

The values of modern precipitation at the LGM precipitation sites are PRISM product, that is compiled from high-resolution radar measurements (Western Regional Climate Center, https://wrcc.dri.edu/), from global modelling of precipitation (Climate Change Knowledge Portal,
https://climateknowledgeportal.worldbank.org), and from climatic station measurements (Supplementary Tab. 6).

### 2.3 – LGM temperatures and precipitation at the studied sites

#### 2.3.1. LGM Sea Surface Temperatures

To reconstruct SSTs during LGM, we use alkenone data from 16 marine cores located along the American cordillera at a restricted distance (<1000 km) from the coastline (Fig. 1; Supplementary Fig. 2). We use a single SST proxy approach to avoid potential biases due to interproxy disparities (Bova et al., 2021; Leduc et al., 2017, 2010). Moreover, alkenone is the only SST proxy available from marine cores along the whole American Cordillera (Fig. 1) and comparison of SST proxies suggest that alkenones provide accurate SST estimates in the Tropics (Lea et al., 2014). The LGM alkenone-derived SSTs are averaged from 26 to 18 ka (Supplementary Tab. 8). To estimate LGM SSTs corresponding to the sites studied for glacial landforms, we first consider a simple linear extrapolation between the reconstructed SSTs for the two nearest LGM SST cores.

Importantly, this simple approach may suffer biases due to the heterogeneous spatial distribution of marine cores. To circumvent this issue and smooth local discrepancies, we consider time variations of the SST and how they may relate to latitude. Because the Holocene variability of temperature is low, we assume in this calculation that the modern SSTs are equivalent to Holocene average SSTs derived from alkenone data. For the Holocene, we average the alkenone-derived SST for the time period 11.7 to 0 ka. We consider the Holocene as a whole because global temperature changes during this period were marginal (1σ standard deviation <1°C) and the averages should therefore represent robust and representative SST values. Then we consider the LGM SST as the mean of the 26-18 ka period. The ΔSST (Holocene-LGM) are then calculated as the difference between the mean Holocene SST and the mean LGM SST (Supplementary Tab. 8).

As expected, the minimal ΔSST values are found around the equator and increase exponentially towards higher latitudes, suggesting that alkenone data provide LGM SST estimates that are precise enough to capture the polar temperature amplification (Fig. 2). At ~34°N, we observe a cluster of LGM SST anomalies that are much lower than surrounding estimates from other marine sites. These LGM SST anomalies of ~1°C compared to modern SST are caused by an early warming in the area, more than 10 ka before the end of the LGM (Herbert et al., 2001). This early SST warming was likely caused by the collapse of the California Current during glacial maximum, in response to a local climate change occurring on land (Herbert et al., 2001). Using the local LGM SST defined for ~25 ka from these 3 cores would yield LGM anomalies of about -4°C, a value in good agreement with other SST estimates at those latitudes (Herbert et al., 2001, Fig. 2). As this LGM SST anomaly is very localized and is not
representative of the large-scale latitudinal trend, we remove these three sites from our LGM SST database (labelled in red in Fig. 2). As the ΔSST are driven at first order by latitude, we can use it to compute the LGM SST at each latitude location from the modern SST. From these data, we then define a relationship between ΔSST at any given latitude and the latitude (Fig. 2):

\[
\Delta SST \text{ (latitude)} = A \times \exp(| \text{Latitude} | \times B) \quad (S1)
\]

where Latitude is expressed in decimal degrees (DD) and A and B are empirical parameters specific to the eastern Pacific SST latitudinal gradient. This LGM SST at any given latitude, especially at studied sites for glacier paleoextents, can be calculated by subtracting the latitude-specific ΔSST from the modern SST at the same latitude.

2.3.2. LGM 0°C isotherms

ELA, 0°C isotherm and mean annual precipitation are linked with robust empirical relationships (Condom et al., 2007; Fox and Bloom, 1994; Legrain et al., 2022). Knowing this relationship and LGM precipitation (see section 3.3.3 below, Supplementary Tabs. 5 and 6) allows using LGM ELAs and LGM precipitation to compute LGM 0°C Isotherm elevations. To estimate paleo ELAs from paleo glacial extents we use the Accumulation Ablation Ratio (AAR) (Meier and Post, 1962) method, that is a robust and easy-to-use technique (Benn and Lehmkuhl, 2000). The AAR defines the ELA as the elevation at which the ratio between (i) the glacier area above this ELA and (ii) the total surface of the glacier, equals the AAR (Meier and Post, 1962) (i.e., the ratio of the accumulation area to the area of glacier). Thus, the LGM ELA can be computed if both the total surface of the LGM glacier and the AAR are known.

The LGM glacial extent and total surface of each studied glacier are determined by glacial morphology analysis and identification of lateral and frontal moraines mapped using high-resolution Landsat imagery (Supplementary Fig. 7). Then, we assume that the AAR remained constant through time and estimate paleo-ELA based on the modern AAR. Since AAR may vary spatially, we use the most accurate AAR estimates available (from the World Glacier Monitoring Service (WGMS) mass balance measurements) to calibrate regional AAR values (Supplementary Fig. 8). Following Martin et al. (2020), we compile interannual AAR and annual mass balances to estimate the AAR value (Supplementary Fig. 8) of each glacier, at equilibrium with the climate state. Data from modern glaciers available in WGMS permitted to calibrate AARs in the following five regions: the Sierra Nevada (n=1 glacier), Mexico (n=1), the Central Andes (n=1), the Bolivian and the Peruvian Andes (n=2) (Supplementary Tab. 7). In the case of Venezuela and Central America, for which modern data are absent from the WGMS database, we use AAR regional values directly compiled from the literature (Stansell et al. (2007) (Venezuela) and Lachniet and Vazquez-Selem (2005) (Central America) (Supplementary Tab. 7).
Local LGM ELAs for each glacial valley are given in Supplementary Tab. 7. The final uncertainties associated with these ELAs are computed using the quadratic sum of the standard deviation associated with each regional AAR coefficient (ranging from 0.03 to 0.06) and a spatial uncertainty of 50 m to account for the lack of precision of paleo glacier mapping.

For South and Central America (20°N – 36°S), we use the relationship calibrated by Fox and Bloom, (1994) to calculate the elevation of the LGM 0°C isotherm, \(\text{Iso}_{0\text{LGM}}\) (m asl), as:

\[
\text{Iso}_{0\text{LGM}} = \text{ELA}_{\text{LGM}} - 3427 + (1148 \times \log_{10} P_{\text{LGM}})
\]

where \(P_{\text{LGM}}\) is the mean annual precipitation (mm.a\(^{-1}\)) and \(\text{ELA}_{\text{LGM}}\) is the Equilibrium Line Altitude (m asl), both during the LGM. This equation is established for the latitudinal range of 5-17°S but was proven to yield robust and accurate results between 10°N and 50°S (Condom et al., 2007). In Central America (20°N – 10°N), climatic conditions are fairly similar to the calibration area (Tropical Andes), implying that this equation is likely to also provide robust results for this region.

For the Sierra Nevada and San Bernardino Mountains (40°N – 34°N), we use the relationship established by Legrain et al. (2022) who used data from Northern America glaciers to define a regional version of the Fox and Bloom, (1994) equation:

\[
\text{Iso}_{0\text{LGM}} = \text{ELA}_{\text{LGM}} - 5150 + (1640 \times \log_{10} P_{\text{LGM}})
\]

Total uncertainties on LGM 0°C Isotherms are computed by propagating uncertainties on LGM precipitation, LGM ELA, as well as the calibration relationship.

### 2.3.3. LGM precipitation

A few LGM precipitation databases exist (Bartlein et al., 2011; Cleator et al., 2019) and were used in climate-model comparison exercises (e.g. Kageyama et al., 2021). Unfortunately, only very few quantitative LGM precipitation estimates are included in these databases for western America. A few isotopic records based on speleothems (Lachniet et al., 2013) and leaf wax (Bhattacharya et al., 2018) are available, but as the isotopic composition of rainfall records mixed source signals, precipitation temperatures, and extents of Rayleigh-type distillation, comparing isotopic results from the LGM and Holocene is not straightforward. In particular, the deglacial warming and continental ice sheet melting invariably obscure precipitation signals and eventually compromise quantitative precipitation rate estimates. For this reason, we ignore rainfall estimates based on stable isotopes. Instead, we derive LGM precipitation from other proxies (pollen assemblage, paleolake shoreline, packrat midden) (Supplementary Tab. 6) that provide relative precipitation change between the LGM and modern conditions. In practice, we compute LGM precipitation by correcting modern precipitation from these changes derived from these proxies (Supplementary Tab. 6, Fig. 3). When uncertainties are not provided
in the original publication, we arbitrarily attribute a relative uncertainty of 20% for LGM precipitation estimates.

2.4 – Continental effect: translation of SST into continental-equivalent temperature

We restrict our compilation to the American Cordillera in order to minimize the potential bias introduced by the effect of continentality between SST and 2m air temperatures above continents. Despite the relative proximity (≤ 500km) between the glaciated sites in the Andes Cordillera and the Pacific Ocean (or the Caribbean Sea), a correction must be applied to convert modern and LGM coastal SST into equivalent continental sea-level temperatures. This correction accounts for the fact that continents have a different radiative budget and climatic inertia compared to the ocean (Seltzer et al., 2023).

In order to compute this continental-ocean correction for the present day, we compile weather station data at variable altitudes along the American cordillera, and defined 9 regions (Supplementary Fig. 5-6) whereby the LR appears to be distinct from those of the adjacent areas (spatial LR variability > 1°C/km). Each region is also characterized by a specific climate. For instance, the distinction between San Bernardino Mountains and Sierra Nevada corresponds to an abrupt change in precipitation rates (Pandey et al., 1999) associated with an orographic depression (<1000 m asl) between these two regions. For Mexico, we separate an Eastern and a Western Transmexican region, since available climatic data show strong differences in the atmosphere thermal structures of these two regions (spatial LR variability ≥ 1.5°C.km⁻¹). These differences are probably due to the influences of Caribbean Sea and Pacific Ocean on the corresponding coastal sites. In this case, the boundary between the two regions at a given latitude is defined as the equidistant point between the two coasts. The Central America region conforms to the specific climate of Central America, with a geographical setting characterized by high moisture transport across the Central America isthmus (Leduc et al., 2007). Equatorial Andes are divided into two distinct regions: the Northern and the Southern Equatorial Andes, representing the Colombian Venezuelan Andes (Caribbean Sea influence and discontinuous Andes) and the Peruvian Bolivian Andes (Pacific coast and high-altitude plateau), respectively. The Northern Central Andes correspond to the climatic region of the Chilean Dry Andes, characterized by a near absence of glaciers due to extremely dry conditions (Ammann et al., 2001). The Southern Central Andes correspond to the glaciated region South of the North Central Andes.

For each region, we compile mean annual temperature data from weather stations (wrcc.dri.edu; explorador.cr2.cl; http://berkeleyearth.lbl.gov). Combining the elevation and temperature of these station, we obtain the temperature of the theoretical 0 m asl altitude within the mountain range, for each of the 9 sub-climatic regions (Supplementary Fig. 9). This theoretical temperature is referred to as the CNET (Continental Null Elevation Temperature). Then, we compare this value with the modern SST of
the East Pacific at the same latitude and expressed the Correction Factor of Continental Effect (CFCE) as:

\[
\text{CFCE} (i) = \text{CNET}(i) - \text{SST}(i) \quad (S4)
\]

where (i) corresponds to the index of a given climatic region. The mean CFCE value is +4.3 ± 5.2 °C, ranging from +14.8°C (Bolivian and Peruvian Andes) down to -2.3°C (Central America) (Supplementary Tab. 4). Then, we compute the modern SST_{corr} for each of the 31 studied sites:

\[
\text{SST}_{\text{corr}}(\text{modern})(i) = \text{SST}_{\text{Modern}} - \text{CFCE}(i) \quad (S5)
\]

where CFCE is given in °C and varies for each calibrated region. Modern SSTs data corrected for the continental effect using the regional computed CFCE value are reported in Supplementary Tab. 3.

To compute the continental-ocean correction for the LGM, we use the same approach as for modern data but we apply an additional correction to account for continental amplification. Recently, it has been shown that during a major climatic change, the temperature change over the continent is about 37% larger than that of the ocean (Seltzer et al., 2023a). To compute the SST during the LGM period we thus account for the terrestrial amplification of the post-deglacial warming. Hence, the SST_{corr} (LGM) is:

\[
\text{SST}_{\text{corr}}(\text{LGM}) = \text{SST}_{\text{Modern}} + \text{CFCE} - 1.37 \times (\Delta\text{SST}) \quad (S6)
\]

with \(\Delta\text{SST}_{\text{LGM}}\) defined by equation (S1).

SST_{corr} (LGM) values are reported in Supplementary Tab. 3.

### 2.5 - Lapse rate and ΔLR computations

We compute both modern and LGM LR using a linear interpolation between a high-altitude temperature (glacier data) and a zero-altitude temperature (continental temperature reconstruction derived from coastal SSTs), following:

\[
\text{LR}(t) = \frac{-\text{SST}_{\text{corr}}(t)}{\text{Iso0}(t) - Z(\text{SST}_{\text{corr}}(t))} \quad (S7)
\]

Where \((t)\) refers to either the LGM or modern time, \(\text{Iso0}(t)\) is the 0°C isotherm elevation in m asl. at the considered time \(t\), \(\text{SST}_{\text{corr}}(t)\) and \(Z(\text{SST}_{\text{corr}}(t))\) are the SST corrected from the CFCE (°C) and the altitude of the \(\text{SST}_{\text{corr}}(t)\) compared to the modern value, respectively, both also at the considered time \(t\). In practice, we use \(Z(\text{SST})\) values of 0 and -125 m for the modern and LGM sea level (Dutton et al., 2015; Fleming et al., 1998), respectively. The last step of our procedure is to compute the difference between the two LR in order to get the ΔLR (in °C.km\(^{-1}\)) for the 31 glacial sites along the American cordillera (Supplementary Tab. 9).
2.6 IPSLCM5A2 climate model

The model results used in this study are based on the IPSLCM5A2 model (Sepulchre et al., 2020). The pre-industrial simulation is described in the same reference. The LGM simulations follow the PMIP4 protocol (Kageyama et al., 2017) and is further described in the first overview of PMIP4 LGM simulations (Kageyama et al., 2021). The main changes in boundary conditions to obtain this LGM simulation are those due to the LGM ice sheets: the altitude, land ice extent and coastlines are set up following the PMIP3 ice sheet reconstruction (Abe-Ouchi et al., 2015). Greenhouse gases and insolation parameters are set up to their LGM values (Kageyama et al., 2017). The Pre-industrial simulation has been run for 2800 years up to equilibrium (Sepulchre et al., 2020). The LGM simulation has been run for 1200 years, starting from pre-industrial conditions. The final 100 years of the pre-industrial and LGM experiments are used to define the averages used in the present work. We examine the 3D (longitude, latitude, altitude) annual mean temperatures and relative humidity simulated by the model for both periods, over the Americas and adjacent oceans, as well as the annual mean precipitation.

We compute the free atmospheric lapse rate for each grid point as the slope of the regression of the vertically dependent temperature vs altitude of the grid point, computed by using the mean geopotential height. Grid points with altitude below 6000m are retained.

3 – Results

3.1 – Low altitude vs high altitude temperature changes

Fig. 4a displays the post-LGM warming on the continent at low elevation and at high elevation (computed from the 0° isotherm shift), respectively, from -40° to 40° latitude. This shows that low elevation T changes are comparable or slightly higher than those at high elevation, from 40°S to 10°S of latitude and beyond 35°N of latitude, with warming ranging from 1 to 6°C. A contrasting pattern is visible in the inner Tropics, between, 8°S to 20°N, where high elevation post LGM warming are significantly larger (8 to 14°C) than those observed at low elevation (1 to 6°C). Consequently, the lapse rate changes consecutive to the deglacial warming are much larger in this Inner Tropic region than those observed at higher latitudes and are locally driven by the large temperature changes that occurred at high elevation (Fig. 4a).

3.2 - Comparing LGM and modern lapse rates

Reconstructed LR values for LGM and modern time stand between -9.5 ±0.8 and -3.8 ±0.3°C.km⁻¹ (average LR of -6.7 ± 1.6 °C.km⁻¹; 1σ) and between -7.3 and -5.2°C.km⁻¹ (average LR of -
6.1 ± 0.7 °C.km⁻¹; 1σ), respectively (Table S9; Fig. 4b, Fig. 5). The interquartile range of LGM LR is markedly larger than the one for modern LR (Fig. 5). The ΔLR ranges from -1.5 to +3.7 °C.km⁻¹ (Supplementary Tab. 9; Fig. 4c). Positive ΔLR values, indicative of shallower modern lapse rates relative to LGM, are observed in the Colombian Andes (Inner Northern Tropics), whereas slightly negative ΔLRs are recorded in Northern and Southern Mid-latitudes. Shallower modern lapse rates indicate amplified post LGM warming at high elevation, compared to sea level. In the search for a spatial relationship between ΔLR and latitude along the American Cordillera, we identify three distinct regions (Fig. 4c): (i) between -40 and 20°N (Sierra Nevada and San Bernardino mountains), where the modern LR is slightly steeper than the LGM LR (mean ΔLR of -1.1 ± 0.4°C.km⁻¹), (ii) between 20°N and 9°S, where the modern LR is shallower than the LGM LR (mean ΔLR of 1.9 ± 0.9°C.km⁻¹), and (iii) between 14°S and 35°S, where the modern LR is similar to the LGM LR (ΔLR average of -0.2 ± 0.5°C.km⁻¹). In summary, a high-altitude amplification of post LGM warming is observed at low latitudes (20°N to 9°S), in the central and the northern South American Cordillera (Fig. 4).

The ΔLR simulated by IPSL-CM5A2 (Fig. 6g) is positive over the tropical Americas, the maximum value being reached over northern South America. On the other hand, negative values are simulated over mid-latitude North America. Even if the simulated ΔLR are much smaller than the reconstructed values, these model results suggest that changes in lapse rates are indeed spatially heterogeneous.

4 – Discussions

4.1 Robustness of the reconstructed ΔLR

The robustness of our ΔLR could be questioned because our approach includes several computation steps with inherent uncertainties, assumptions and corrections. However, several independent arguments and observations support our results and suggest our approach is robust.

First, as a sensitivity test, we also calculated the modern LR directly from weather station data (Supplementary Fig. 5). The average difference between these modern LR estimates and our modern SST versus 0°C isotherm-based LR is only 0.2°C.km⁻¹ on average, with a maximum of 0.94 °C.km⁻¹ for Costa Rican sites (Supplementary Fig. 10). Such a small difference between two independent methods appears negligible and within the uncertainty of our regional ΔLR computation. It is much lower than the modern versus LGM variations of the LR we report here for the Tropics (Fig. 4c). In other words, using the alternative stations-based methodology would not significantly affect the ΔLR changes computed in this study between 20°N – 9°S (Supplementary Fig. 10). Nonetheless, we prefer to keep the SST and 0°C isotherm-based approach to compute modern LR, for two main reasons: (i) we consider it preferable to use the same methodology for LGM and modern LR computation, and (ii) the weather
station-based methodology does not permit taking into account the local LR variability due to the large geographical range covered by a weather-station region.

Second, we show that the final ΔLR values are not very sensitive to the choice of the method used to estimate the LGM SST at the studied site (either latitudinal correction to account for polar amplification or linear interpolation) (Fig. 7).

Third, we also performed a sensitivity analysis to assess how each parameter involved in the lapse rate reconstructions contribute to our ΔLR estimates, and hence identify potential bias. For this, we computed the ΔLR by iteratively suppressing each of the following corrections (Fig. 7): (i) the conversion of LGM glacier paleoELAs into 0°C isotherms using independent precipitation estimates, (ii) the latitudinal interpolation of ΔSST, and (iii) the conversion of modern and LGM SST into “equivalent” continental air temperatures with the CFCE. This sensitivity analysis indicates that the computed regional ΔLR pattern remains similar, within uncertainty, when any of these three corrections is not accounted for (Fig. 7). In the extreme scenario where the ΔLR are computed without any of these three corrections being applied, positive ΔLRs values are still present beyond uncertainties in the equatorial region (Fig. 7). These sensitivity tests therefore demonstrate that the patterns of our ΔLR calculations are robust.

4.2 The role of precipitation and humidity on LR changes

4.2.1 Correlation between LR and precipitation variations

For both LGM and modern values, the steepest LR values (≤ -6.5°C/km) are only observed in the driest regions (P ≤ 1000mm/yr; Fig. 5). However, the relationship between absolute LR and annual precipitation remains unclear for both LGM and modern data (Fig. 5). The modern LR displays a lower spread than LGM LR and the relationship between LR and annual precipitation is less significant under the warmer climatic state of the modern compared to the colder LGM (Fig. 5). This different LR vs precipitation sensitivity suggests that boundary conditions matter, but the main drivers (average temperature, atmospheric CO₂ concentrations, differences in atmospheric dynamics) still need further research to be identified. To better understand the mechanism(s) that control(s) spatial and temporal variations of LR, we evaluate the correlation between the computed ΔLR and absolute precipitation changes (ΔP = modern precipitation – LGM precipitation) (Fig. 8). Precipitation proxies indicate that rainfall significantly increased between the LGM and the Holocene in the inner tropics (20°N – 9°S: ΔP range from 0 to +1400 mm yr⁻¹), while it slightly decreased in the outer tropics and at mid-latitudes (10°S – 35°S and 34°N – 40°N: ΔP range from -15 to -770 mm yr⁻¹) (Fig. 3, Supplementary Tab. 6, Supplementary Fig. 4). Notably, pollen records and lake shoreline levels in Central America (20°N-9°S) indicate a significant precipitation increase from the LGM to the Holocene (Supplementary Tab.
Although these absolute precipitation estimates are based on different proxies with their own characteristics and potential biases, their good overall agreement depicts a consistent picture. In detail, differences between the absolute precipitation estimates for the inner and the outer tropics display a precipitation pattern which potentially reflects a global southward shift of the Intertropical Convergence Zone (ITCZ) during the LGM, from Central America to South America, and/or an intensification of the South American Summer Monsoon (SASM). This scenario is corroborated by several precipitation records in Southern America (Blard et al., 2011; Martin et al., 2020, 2018; McGee, 2020; Woods et al., 2020).

Importantly, ΔLR values display a linear relationship with ΔP ($R^2 = 0.81$; p-value = 3.2x10$^{-11}$) (Fig. 8). Shallower modern lapse rates are associated with higher modern precipitation rates compared to the LGM. Conversely, areas characterized by a limited precipitation change (or a slight aridification) in the modern period (still relative to the LGM) yield unchanged or only slightly negative ΔLR values (Fig. 8).

4.2.2 Robustness of the ΔLR versus ΔP relationship

The precipitation values are used in the computation of the LR values to convert ELAs into 0°C isotherms. One could hence wonder whether the ΔLR versus ΔP relationship (Fig. 8) could result from a bias inherent to the method. To evaluate the influence of the precipitation-correction on the ΔLR versus ΔP relationship, we compute a similar relationship but using ΔLR values that have not been corrected for the effect of precipitation (Supplementary Fig. 12). This methodological choice represents an extreme case where ELAs are considered as equivalent to 0°C isotherm, which is not physically realistic (Condom et al., 2007; Legrain et al., 2022). However, this approach allows us to evaluate the extent to which the observed correlation between ΔP and ΔLR is due to the intervention of precipitation in the computation of ΔLR values. Using this approach, we find that the correlation between ΔP and ΔLR qualitatively stands qualitatively similar, (i.e., wetter sites still are associated with shallower lapse rate; Supplementary Fig. 12). This sensitivity test thus reinforces the robustness of the observed correlation between ΔLR and ΔP. Nevertheless, the slope of the linear relationship is modified, implying that quantitative estimates is affected by the choice of using the precipitation correction. Since the glacier ELAs are sensitive to absolute precipitation, we still consider the relationship using precipitation values in the LR computation as the most reliable approach (Fig. 8).

4.2.3 Role of moisture in the ΔLR variations

Based on the relationship results, we propose that shallower lapse rates may result from a post-LGM increase of the atmospheric water content. When the atmospheric water content increases, the capacity of the atmosphere to transfer heat from low to high elevations by convective and radiative transfer is enhanced, yielding less steep lapse rates (Kattel et al., 2013; Stone and Carlson, 1979). Our
suggestion of a shallower modern LR relative to the LGM in the inner tropics (20°N – 9°S) corroborates local studies of Blard et al. (2007) and Loomis et al. (2017), who established less steep modern LR compared to LGM in Hawaii and Kenya, respectively. Alternatively, constant modern LR could have prevailed in some tropical regions (e.g., between 11°S and 30°S in this study, but also in Papua New Guinea, in the Western Pacific (Tripati et al. 2014)). The main difference between these two regions is that the summits of New Guinea are subject to much higher precipitation than those of Hawaii and Mount Kenya (> 1000 mm yr\(^{-1}\) versus < 500 mm yr\(^{-1}\)), and the local LR is hence close to the moist adiabatic typical range (around 4-6°C/km). While our results demonstrate a linear relationship between ΔLR and ΔP, some of the noise in this correlation may be attributed to differences in the absolute climatic conditions of the glaciers. Specifically, mountainous areas experiencing very high precipitation levels (> 1000 mm/yr) are likely to be less influenced by changes in precipitation, making them less susceptible to recording a change in lapse rates. Moreover, paleo-precipitation records suggest that the modern climate conditions in Papua New Guinea are not significantly wetter than during the LGM (Denham and Haberle, 2008; Hope, 2009). The absence of precipitation changes between the LGM and the modern would explain the absence of significant variation of the LR computed in these areas.

4.3 What can we learn by comparing the paleoLR derived from paleoELA and those from the IPSLCM5A2 climate model?

The GCM modeled ΔLR and ΔP (Pre-industrial – LGM) map indicates a first order spatial correlation that we derive from our reconstruction based on high altitude continental proxies between ΔLR and ΔP (Fig. 8). The GCM outputs also indicate a tight spatial correlation between ΔP and the difference of relative humidity (noted ΔRH) (Fig. 6). This correlation between RH and precipitation changes is supported by the scatter plot (Fig. 9b) between these individual outputs retrieved from the GCM grid (\(R^2 = 0.65\)). The observed correlation in the paleoclimate record between ΔP and ΔLR could hence reflect a close control of ΔRH on ΔLR. The fact that RH changes may be a major driver of LR changes is also confirmed by plotting these metrics extracted from the IPSLCM5A2 model output over and close to the American cordillera from 40°N to 40°S (Fig. 9c) (\(R^2 = 0.80\)). The same plot realized between the modelled specific humidity and LR changes display a poorer correlation mainly due to the sites located north of 20°N where ΔLR is negative (Supplementary Fig. 11; \(R^2 = 0.27\)). These observations provide a strong support for the involvement of a moisture control on LR changes, increased relative humidity conditions implying shallower lapse rates (or drier conditions implying steeper lapse rates) (Fig. 9).
However, the agreement between the IPSL model outputs and our proxy-based reconstruction does not hold for regions that experienced a post-LGM precipitation increase. In such cases, the model shows a saturation effect for ΔLR, hence causing the overall ΔLR response to be underestimated (Fig. 8). A comparison between present day precipitation rates and those modeled for PI conditions (Fig. 10) shows that the IPSL model underestimates precipitation rates over the Amazonian basin, while overestimating precipitation rates in the Northern part of the Andes (Ecuador, Columbia). This could result from the low resolution of the topography, which causes a poor representation of the real altitude of the range, and, hence, an inaccurate representation of the orographic effect in this narrow part of the Cordillera (Fig. 10). There are also other long-recognized biases of the IPSL models, as discussed by (Sepulchre et al., 2020). This low resolution could cause post-LGM LR changes to be underestimated in Ecuador and Columbia, where GCM-modeled precipitation are overestimated at the location of glaciers (> 1500 mm/a, compared to < 1000 mm/a), implying that the modeled LRs are already close to a moist adiabatic value (around 5°C/km), thus hampering any post-LGM LR change to be observed in the simulations. Finally, it is important to acknowledge that lapse rates measured along mountain slope may be decoupled from those observed in the adjacent free atmosphere (Pepin et al., 2015), especially in valleys. Since the IPSL model is run at ~300 km resolution, it only has an incomplete description of the real topography, and therefore of the processes occurring within the mountain atmospheric boundary layer, well below the subgrid scale. This observation may partly explain some of the discrepancies between modelled LR and those derived from paleoELA reconstruction. Future studies could test this by performing high-resolution GCM simulations with a much higher resolution topography (i.e., < 10 km grid cell).

4.4 Amplification of high-altitude temperature changes: implications for anthropogenic warming

During the past decade, several studies have developed the concept of an elevation-dependant warming (Pepin, 2015; Wang et al., 2016, 2014). Recent observations have shown an amplification of the high-altitude warming, albeit this pattern seems to yield contrasting outcomes (Ohmura, 2012). A recent study based on noble gas reconstruction of continental temperature evidenced a larger post-LGM warming over the continent relative to the oceans, even at elevations < 1000 m (Seltzer et al., 2021). In order to test the robustness of the amplification of temperature changes that we obtain here for high elevations, we compute the post-LGM warming values derived from paleo-glaciers and plotted them against the altitude of observation (Fig. 11). Our data correspond to high elevation (2500 m to 5000 m) post-LGM warming ranging between 6 and 14°C in locations that were drier during the LGM, and 0 to 5°C in locations that were wetter. For Brazil, noble gas paleotemperatures indicate a low altitude LGM cooling of 5.4±0.6°C, at an average elevation of 600 m (Seltzer et al., 2021; Stute et al., 1995). This value is at the low end of the range of post-LGM warming computed in our study for higher altitudes.
This observation thus implies that the amplification of the post LGM warming is significant at high elevation, and exceeds the effect of pure continental amplification (Seltzer et al., 2023) (Fig. 11).

Many factors have been proposed to explain the peculiar trends of anthropogenic warming, including surface-based land use feedbacks (Zeng et al., 2021), as well as atmospheric radiative forcings and convective process (Keil et al., 2021; Mountain Research Initiative EDW Working Group, 2015). A recent study also mentioned the potential role of seasonality in modulating short term warming amplification at high-altitude (Qixiang et al., 2018).

Several studies had already identified correlations between lapse rate and humidity, both for present day (Comand and Soden, 2021; Li et al., 2015) and the LGM (Blard et al., 2007; Kuhlemann et al., 2008; Loomis et al., 2017). Previous reconstructions of LGM lapse rates yielded contrasting results from different regions (Blard et al., 2007; Kuhlemann et al., 2008; Loomis et al., 2017; Tripati et al., 2014) but the scarcity of these LGM reconstructions and their lack of quantitative precipitation estimate did not permit to demonstrate a clear forcing mechanism. Our synthesis identifies for the first time a robust quantitative relationship between changes in lapse rate and changes in local precipitation (and probably relative humidity) during the post LGM warming. This finding permits to better understand and constrain the roles of precipitation and moisture on lapse rate modifications, temporal and spatial changes in moisture being able to drive variable interregional lapse rate changes at different elevations. Consequently, as evidenced by our dataset (Fig. 4a), the magnitude of the LGM cooling at high elevation has a high probability to have been spatially heterogenous, which represents a significant caveat to keep in mind when deriving a uniform and universal temperature change on land (Seltzer et al., 2023). Using the obtained relationship between precipitation rates and lapse rate changes (Fig. 8), we estimate the high-altitude amplification of the future warming, assuming mean annual precipitation changes modelled by 2100 projection scenarios. For instance, a precipitation increase of 500 mm.a⁻¹ would lead to a -1.1°C.km⁻¹. Hence, based on our most reliable ΔLR–ΔP relationship, we find that a sea level +1°C warming could induce an additional warming of 4.1 ±0.8°C at 4000 m asl (1σ standard deviation). Following this reasoning, we speculate that regions becoming increasingly wet as a result of climate change may coincide with regions where the low altitude anthropogenic warming will be amplified at high-altitude and potentially accelerate mountainous glaciers melting here despite the increase in precipitation. Reversely, in the regions that will become drier in the future, global warming could be attenuated at high elevation. Future modelling studies should help to quantify these amplification/attenuation mechanisms and better address future high-altitude climatic changes.

5 – Conclusions
We summarize the main outcomes of our study in these four key findings:

- **Quantitative relationship between lapse rate and precipitation changes**: by using LGM 0°C isotherm from glacier paleo-equilibrium line, SST and paleoprecipitation proxies, we establish a relationship linking post LGM changes in precipitation (ΔP) and changes in lapse rates (ΔLR).

- **Amplified post-LGM warming in wetter areas**: the deglacial warming amplitude at high-altitude is correlated with changes in mean annual precipitation: in areas that have become wetter, the lapse rate has become shallower, and reversely. This mechanism is also encountered in the IPSL-GCM model outputs, although some discrepancies exist in regions that experienced the largest post LGM moisture increase.

- **Spatial heterogeneities in post-LGM lapse rate variations**: by linking moisture and lapse rate changes, our study provides key insights to explain the heterogeneities in post-LGM lapse rate variations inferred from previous studies.

- **Implications for anthropogenic-induced high-elevation warming**: our results suggest that the current high-altitude warming is underestimated in areas that will become wetter. Future modelled projections need to ensure that this moisture amplification mechanism is well captured by numerical models.
Fig. 1: Location of the studied sites. Grey diamonds identify marine core locations used for the SST calibration (Fig. 4). White squares are the 31 \( \Delta LR \) computation sites. Dots are sites with LGM paleoprecipitation estimates. Blue, red and grey dots respectively indicate sites where the modern period is wetter, drier, and unchanged than LGM.
Fig. 2: Calibration of the $\Delta SST_{\text{Modern-LGM}}$. a. White and red dots are SST estimates from marine cores included in, and excluded from, the calibration, respectively (see Methods). All SST estimates are based on alkenone reconstructions. The U-shaped relationship between the $\Delta SST_{\text{Modern-LGM}}$ and latitude highlights the polar amplification. b. $\Delta SST_{\text{Modern-LGM}}$ is plotted as a function of absolute latitude used to estimate the $\Delta SST_{\text{Modern-LGM}}$ at each latitude (blue line). Dashed lines represent the 1σ envelope ($R^2 = 0.89$; MSWD = 1.3; $p$ value = $2 \times 10^{-12}$).
Fig. 3: Modern and LGM 0°C Isotherm elevation and associated precipitation changes. Red and grey circles are modern and LGM 0°C isotherm values, respectively. Dark blue sticks are LGM precipitation at the Isotherm 0°C computation sites. Light blue and yellow sticks are modern precipitation when they are higher and lower than during LGM, respectively.
Fig. 4 - A) Low elevation vs high elevation post LGM warming. Low elevation T changes are calculated from nearby SST, applying the continental correction factor (CFCE), while high elevation T changes are computed from the glacier-derived 0° isotherm shift, using modern local lapse rates. B) Modern and LGM lapse rates. C) ΔLR_{modern-LGM} for the 31 studied sites. Central tropics (20°N - 9°S) present the highest value of ΔLR_{modern-LGM} and thus a steeper lapse rate during the LGM. Post-deglacial warming is thus amplified at high-altitude, between 20°N – 9°S.
Fig. 5: Annual precipitation (mm) plotted against lapse rates, for both modern (orange circles) and LGM (blue circles) lapse rates (Lower panel). Box plot showing the distribution of modern and LGM lapse rates (Upper panel).
**Fig. 6:** IPSLCM5A2 coupled model results for the LGM climate (top row) and the PI – LGM anomaly (bottom row). Surface air temperature (a: LGM, f: PI – LGM), lapse rate (b: LGM, g: PI – LGM), annual precipitation (c: LGM, h: PI – LGM), relative humidity (d: LGM, i: PI – LGM), and specific humidity (e: LGM, j: PI – LGM). All fields are annual means. Reconstructions compiled in this study for the lapse rate and precipitation are superimposed on panels e and f, based on the same colour scale as for the maps.
Fig. 7: Testing the model dependence of ΔLR results. Dark diamonds are the computed ΔLR of the study (see methods, Fig. 5). Coloured dots show ΔLR computed removing iteratively each correction: a) ΔLR computed without precipitation correction (yellow dots), b) ΔLR computed without CFCE correction (red dots), c) ΔLR computed assuming that corresponding LGM SSTs are interpolated from the 2 nearest cores, without using the SST calibration relationship (blue dots). d) ΔLR computed without performing any of the 3 corrections (white dots). Note that ΔLR between 20°N and 9°S is significantly superior to zero independently of the modelling choices used.
Fig. 8: Correlation between ΔLR and ΔP. (a) Data, blue dots: LGM precipitation has been calculated directly from modern precipitation and independent paleoprecipitation proxy (See Methods). The 31 sites of calculation are shown as blue dots. Note that steeper and shallower modern LR are systemically associated with drier and wetter areas than during LGM, respectively. IPSL model, orange dots: precipitation and lapse rate variations have been extracted at each data site location to perform the comparison.
Fig. 9: Relationships between variables simulated by IPSLCM5A2 over and close to the American Cordillera, on the grid points highlighted on the map. The American Cordillera is separated into three main domains highlighted in green, red and blue (a), respectively. Δ indicate PI - LGM anomalies. RH: relative humidity at first level, P: precipitation, LR: lapse rate.
Fig. 10: Comparison of observed (GPCC dataset, https://opendata.dwd.de/climate_environment/GPCC/html/gpcc_normals_v2022_doi_download.html) and modeled (GCM IPSLCM5A2) modern precipitation field in the American Cordillera.
Fig. 11: Correlation between $\Delta T_{\text{Modern-LGM}}$ and LGM ELA (m). $\Delta T_{\text{Modern-LGM}}$ is computed from the ELA depression value using the modern lapse rate value. Blue and red dots represent sites that were drier and wetter during the LGM relative to present-day, respectively. LGM precipitation was calculated directly from modern precipitation and independent paleoprecipitation proxies (see Methods). Note that the highest values of $\Delta T$ are higher than the land $\Delta T$ from the study of Seltzer et al. (2021) (green shaded area).
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