Estimates of Late Cenozoic climate change relevant to Earth surface processes in tectonically active orogens

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Abstract

The denudation history of active orogens is often interpreted in the context of modern climate gradients. Here we address the validity of this approach and ask the question: what are the spatial and temporal variations in palaeoclimate for a latitudinally diverse range of active orogens? We do this using high-resolution (T159, ca. 80 x 80 km at the equator) palaeoclimate simulations from the ECHAM5 global Atmospheric General Circulation Model and a statistical cluster analysis of climate over different orogens (Andes, Himalaya, SE Alaska, Pacific NW USA). Time periods and boundary conditions considered include the Pliocene (PLIO, ~3 Ma), the Last Glacial Maximum (LGM, ~21 ka), Mid Holocene (MH, ~6 ka) and Pre-Industrial (PI, reference year 1850). The regional simulated climates of each orogen are described by means of cluster analyses based on the variability of precipitation, 2m air temperature, the intra-annual amplitude of these values, and monsoonal wind speeds where appropriate. Results indicate the largest differences to the PI climate existed for the LGM and PLIO climates in the form of widespread cooling and reduced precipitation in the LGM and warming and enhanced precipitation during the PLIO. The LGM climate shows the largest deviation in annual precipitation from the PI climate, and shows enhanced precipitation in the temperate Andes, and coastal regions for both SE Alaska and the US Pacific Northwest . Furthermore, LGM precipitation is reduced in the western Himalayas and enhanced in the eastern Himalayas, resulting in a shift of the wettest regional climates eastward along the orogen. The cluster-analysis results also suggest more climatic variability across latitudes east of the Andes in the PLIO climate than in other time-slice experiments conducted here. Taken together, these results highlight significant changes in Late Cenozoic regional climatology over the last ~3 Ma. Comparison of simulated climate with proxy-based reconstructions for the MH and LGM reveal satisfactory to good performance of the model in reproducing precipitation changes, although in some cases discrepancies between neighbouring proxy observations highlight contradictions between proxy observations themselves. Finally, we document regions where the largest magnitudes of Late Cenozoic

31 changes in precipitation and temperature occur and offer the highest potential for future observational studies that

32 quantify the impact of climate change on denudation and weathering rates.

33

36

1. Introduction

38   Interpretation of orogen denudation histories in the context of climate and tectonic interactions is often hampered

39 by a paucity of terrestrial palaeoclimate proxy data needed to reconstruct spatial variations in palaeoclimate. While it is

40 self-evident that palaeoclimate changes could influence palaeodenudation rates, it is not always self-evident what the

41 magnitude of climate change over different geologic time scales is, or what geographic locations offer the greatest

42 potential to investigate palaeoclimate impacts on denudation. Palaeoclimate reconstructions are particularly beneficial

43 when denudation rates are determined using geo- and thermo-chronology techniques that integrate over timescales of

44 10^3-10^{6+} years (e.g. cosmogenic radionuclides or low-temperature thermochronology) [e.g., Kirchner et al., 2001;

45 Schaller et al., 2002; Bookhagen et al., 2005; Moon et al., 2011; Thiede and Ehlers, 2013; Lease and Ehlers, 2013].

46 However, few studies using denudation rate determination methods that integrate over longer timescales have access to

47 information about past climate conditions that could influence these palaeo-denudation rates. Palaeoclimate modelling

48 offers an alternative approach to sparsely available proxy data for understanding the spatial and temporal variations in

49 precipitation and temperature in response to changes in orography [e.g. Takahashi and Battisti, 2007a, b; Insel et al.,

50 2010; Feng et al., 2013] and global climate change events [e.g. Salzmann, 2011; Jeffery et al., 2013]. In this study, we

51 characterise the climate at different times in the Late Cenozoic, and the magnitude of climate change for a range of

52 active orogens. Our emphasis is on identifying changes in climate parameters relevant to weathering and catchment

53 denudation to illustrate the potential importance of various global climate change events on surface processes.

54   Previous studies of orogen-scale climate change provide insight into how different tectonic or global climate

55 change events influence regional climate change. For example, sensitivity experiments demonstrated significant

56 changes in regional and global climate in response to landmass distribution and topography of the Andes, including

57 changes in moisture transport, the north-south asymmetry of the Intertropical Convergence Zone [e.g. Takahashi and

58 Battisti, 2007a, ; Insel et al., 2010] and (tropical) precipitation [Maroon et al., 2015, ; 2016]. Another example is the

59 regional and global climate changes induced by the Tibetan Plateau surface uplift due to its role as a physical obstacle to

60 circulation [Raymo and Ruddiman, 1992; Kutzbach et al., 1993; Thomas, 1997; Bohner, 2006; Molnar et al., 2010;

61 Boos and Kuang, 2010]. The role of tectonic uplift in long term regional and global climate change remains a focus of

62 research and continues to be assessed with geologic datasets [e.g. Dettman et al., 2003; Caves et al., 2017; Kent-

63 Corson et al., 2006; Lechler et al., 2013; Lechler and Niemi, 2011; Licht et al., 2016; Methner et al.,

64 2016; Mulch et al., 2015, 2008; Pingel et al., 2016] and climate modelling [e.g. Kutzbach et al., 1989; Kutzbach

65 et al., 1993; Zhisheng, 2001; Bohner, 2006; Takahashi and Battisti, 2007a; Ehlers and Poulsen, 2009; Insel et al., 2010;

66 Boos and Kuang, 2010]. Conversely, climate influences tectonic processes through erosion [e.g. Molnar and England,

67 1990; Whipple et al., 1999; Montgomery et al., 2001; Willett et al., 2006; Whipple, 2009]. Quaternary climate change

68 between glacial and interglacial conditions [e.g. Braconnot et al., 2007; Harrison et al., 2013] resulted in not only the

69 growth and decay of glaciers and glacial erosion [e.g. Yanites and Ehlers, 2012; Herman et al., 2013; Valla et al., 2011]

70 but also global changes in precipitation and temperature [e.g. Otto-Bliesner et al., 2006; Li et al., 2017] that could

71 influence catchment denudation in non-glaciated environments [e.g. Schaller and Ehlers, 2006; Glotzbach et al., 2013;

72 Marshall et al., 2015]. These dynamics highlight the importance of investigating how much climate has changed over

73 orogens that are the focus of studies of climate-tectonic interactions and their impact on erosion.

74   Despite recognition by previous studies that climate change events relevant to orogen denudation are prevalent

75 throughout the Late Cenozoic, few studies have critically evaluated how different climate change events may, or may

76 not, have affected the orogen climatology, weathering and erosion. Furthermore, recent controversy exists concerning

77 the spatial and temporal scales over which geologic and geochemical observations can record climate-driven changes in

78 weathering and erosion [e.g. Whipple, 2009; von Blanckenburg et al., 2015; Braun, 2016]. For example, the previous

79 studies highlight that although palaeoclimate impacts on denudation rates are evident in some regions and measurable

80 with some approaches, they are not always present (or detectable) and the spatial and temporal scale of climate change

81 influences our ability to record climate sensitive denudation histories. This study contributes to our understanding of

82 the interactions between climate, weathering, and erosion by bridging the gap between the palaeoclimatology and

83 surface processes communities by documenting the magnitude and distribution of climate change over tectonically

84 active orogens.

85   Motivated by the need to better understand climate impacts on Earth surface processes, especially the denudation

86 of orogens, we model palaeoclimate for four time slices in the Late Cenozoic, use descriptive statistics to identify the

87 extent of different regional climates, quantify changes in temperature and precipitation, and discuss the potential

88 impacts on fluvial and/or hillslope erosion. In this study, we employ the ECHAM5 global Atmospheric General

89 Circulation Model and document climate and climate change for time slices ranging between the Pliocene (PLIO, ~3

90 Ma) to pre-industrial (PI) times for the St. Elias Range of South East Alaska, the US Pacific Northwest (Olympic and

91 Cascade Range), western South America (Andes) and South Asia (incl. parts of Central- and East Asia). Our approach is

92 two-fold and includes:

93   1. An empirical characterisation of palaeo-climates in these regions based on the covariance and spatial

94   clustering of monthly precipitation and temperature, the monthly change in precipitation and temperature magnitude,

95   and wind speeds where appropriate.

96   2. Identification of changes in annual mean precipitation and temperature in selected regions for four time

97   periods: (PLIO, Last Glacial Maximum (LGM), the Mid-Holocene (MH) and PI) and subsequent validation of the

98   simulated precipitation changes for MH and LGM.

99   Our focus is on documenting climate and climate change in different locations with the intent of informing past and

100   ongoing palaeodenudation studies of these regions. The results presented here also provide a means for future work to

101   formulate testable hypotheses and investigations into whether or not regions of large palaeoclimate change produced a

102   measurable signal in denudation rates or other Earth surface processes. More specifically, different aspects of the

103   simulated palaeoclimate may be used as boundary conditions for vegetation and landscape evolution models, such as

104   LPJ-GUESS and Landlab, to bridge the gap between climate change and quantitative estimates for Earth surface system

105   responses. In this study, we intentionally refrain from applying predicted palaeoclimate changes to predict denudation

106   rate changes. Such a prediction is beyond the scope of this study because a convincing (and meaningful) calculation of

107   climate-driven transients in fluvial erosion (e.g. via the kinematic wave equation), variations in frost cracking intensity,

108   or changes in hillslope sediment production and transport at the large regional scales considered here is not tractable

109   within a single manuscript, and instead is the focus of our ongoing work. Merited discussion of climatically induced

110   changes in glacial erosion, as is important in the Cenozoic, is also beyond the scope of this study. Instead, our emphasis

111   lies on providing and describing a consistently setup GCM simulation framework for future investigations of Earth

112   surface processes, and to identify regions in which Late Cenozoic climate changes potentially have a significant impact

113   on fluvial and hillslope erosion.

114

115   **2. Methods: Climate modelling and cluster analyses for climate characterisation**

116

117   **2.1 ECHAM5 simulations**

118   The global Atmospheric General Circulation Model ECHAM5 [Roeckner et al., 2003] has been developed at the

119   Max Planck Institute for Meteorology and is based on the spectral weather forecast model of the ECMWF [Simmons et

120   al., 1989]. In the context of palaeoclimate applications, the model has been used mostly at lower resolution (T31, ca.

121   3.75°x3.75°; T63, ca. 1.9°x1.9° in case of Feng et al. [2016] and T106 in the case of Li et al. [2016] and Feng and

122   Poulsen [2016]) . The performed studies are not limited to the last millenium [e.g. Jungclaus et al., 2010] but also

123   include research in the field of both warmer and colder climates, at orbital [e.g. Gong et al., 2013; Lohmann et al., 2013;

124   Pfeiffer and Lohmann, 2016; Zhang et al., 2013a; Zhang et al., 2014; Wei and Lohmann, 2012] and tectonic time scales

125   [e.g. Knorr et al., 2011; Stepanek and Lohmann, 2012], and under anthropogenic influence [Gierz et al., 2015].

126   Here, the ECHAM5 simulations were conducted at a T159 spatial resolution (horizontal grid size ca. 80 km x 80

127   km at the equator) with 31 vertical levels (between the surface and 10hPa). This high model resolution is admittedly not

128   required for all of the climatological questions investigated in this study, and it should be noted that the skill of GCM's

129   in predicting orographic precipitation remains limited at this scale [e.g. Meehl et al. 2007]. However, simulations were

130   conducted at this resolution so that future work can apply the results in combination with different dynamical and

131   statistical downscaling methods to quantify changes at large catchment to orogen scales. The output frequency is

132   relatively high (1 day) to enhance the usefulness of our simulations as input for landscape evolution and other models

133   that may benefit from daily input. The simulations were conducted for five different time periods: present-day (PD), PI,

134   MH, LGM and PLIO.

135   A PD simulation (not shown here) was used to establish confidence in the model performance before conducting

136   palaeosimulations and has been compared with the following observation-based datasets: European Centre for Medium-

137   Range Weather Forecasts (ECMWF) re-analyses [ERA40, Uppala et al., 2005], National Centers for Environmental

138   Prediction and National Center for Atmospheric Research (NCEP/NCAR) re-analyses [Kalnay et al., 1996; Kistler et

139   al., 2001], NCEP Regional Reanalysis (NARR) [Mesinger et al., 2006], the Climate Research Unit (CRU) TS3.21

140   dataset [Harris et al., 2013], High Asia Refined Analysis (HAR30) [Maussion et al., 2014] and the University of

141   Delaware dataset (UDEL v3.01) [Legates et al., 1990]. (See Mutz et al. [2016] for a detailed comparison with a lower

142   resolution model).

143   The PI climate simulation is an ECHAM5 experiment with PI (reference year 1850) boundary conditions. Sea

144   Surface Temperatures (SST) and Sea Ice Concentration (SIC) are derived from transient coupled ocean-atmosphere

145   simulations [Lorenz and Lohmann, 2004; Dietrich et al., 2013]. Following Dietrich et al. [2013], greenhouse gas

146   (GHG) concentrations (CO2: 280 ppm) are taken from ice core based reconstructions of CO_2 [Etheridge et al., 1996],

147   CH_4 [Etheridge et al., 1998] and N2O [Sowers et al., 2003]. Sea surface boundary conditions for MH originate from a

148   transient, low-resolution, coupled atmosphere-ocean simulation of the mid (6 ka) Holocene [Wei and Lohmann, 2012;

149   Lohmann et al, 2013], where the GHG concentrations (CO2: 280 ppm) are taken from ice core reconstructions of

150   GHG's by Etheridge et al. [1996], Etheridge et al. [1998] and Sowers et al. [2003]. GHG's concentrations for the LGM

151   (CO2: 185 ppm) have been prescribed following Otto-Bliesner et al. [2006]. Orbital parameters for MH and LGM are

152   set according to Dietrich et al. [2013] and Otto-Bliesner et al. [2006], respectively. LGM land-sea distribution and ice

153   sheet extent and thickness are set based on the PMIP III (Palaeoclimate Modelling Intercomparison Project, phase 3)

154   guidelines (elaborated on by Abe-Ouchi et al [2015]). Following Schäfer-Neth and Paul [2003], SST and SIC for the

155   LGM are based on GLAMAP [Sarnthein et al. 2003] and CLIMAP [CLIMAP project members, 1981] reconstructions

156   for the Atlantic and Pacific/Indian Ocean, respectively. Global MH and LGM vegetation are based on maps of plant

157   functional types by the BIOME 6000 / Palaeovegetation Mapping Project [Prentice et al., 2000; Harrison et al., 2001;

158 Bigelow et al., 2003; Pickett et al., 2004] and model predictions by Arnold et al. [2009]. Boundary conditions for the

159 PLIO simulation, including GHG concentrations (CO2: 405), orbital parameters and surface conditions (SST, SIC, sea

160 land mask, topography and ice cover) are taken from the PRISM (Pliocene Research, Interpretation and Synoptic

161 Mapping) project [Haywood et al., 2010; Sohl et al., 2009; Dowsett et al., 2010], specifically PRISM3D. The PLIO

162 vegetation boundary condition was created by converting the PRISM vegetation reconstruction to the JSBACH plant

163 functional types as described by Stepanek and Lohmann [2012], but the built-in land surface scheme was used

164   SST reconstructions can be used as an interface between oceans and atmosphere [e.g. Li et al. 2016] instead of

165 conducting the computationally more expensive fully coupled Atmosphere-Ocean GCM experiments. While the use of

166 SST climatologies comes at the cost of capturing decadal-scale variability, and the results are ultimately biased towards

167 the SST reconstructions the model is forced with, the simulated climate more quickly reaches an equilibrium state and

168 the means of atmospheric variables used in this study do no change significantly after the relatively short spin-up

169 period. The palaeoclimate simulations (PI, MH, LGM, PLIO) using ECHAM5 are therefore carried out for 17 model

170 years, of which the first two years are used for model spin up. The monthly long-term averages (multi-year means for

171 individual months) for precipitation, temperature, as well as precipitation and temperature amplitude, i.e. the mean

172 difference between the hottest and coldest months, have been calculated from the following 15 model years for the

173 analysis presented below.

174   For further comparison between the simulations, the investigated regions were subdivided (Fig. 1). Western

175 South America was subdivided into four regions: parts of tropical South America (80°-60° W, 23.5-5° S), temperate

176 South America (80°-60° W, 50°-23.5° S), tropical Andes (80°-60° W, 23.5-5° S; high-pass filtered), i.e. most of the

177 Peruvian Andes, Bolivian Andes and northernmost Chilean Andes, and temperate Andes (80°-60° W, 50°-23.5° S, high-

178 pass filtered). South Asia was subdivided into three regions: tropical South Asia (40°-120°E, 0°-23.5°N), temperate

179 South Asia (40°-120°E, 23.5°-60°N), and high altitude South Asia (40°-120°E, 0°-60°N; high-pass filtered).

180   Our approach of using a single GCM (ECHAM5) for our analysis is motivated by, and differs from, previous

181 studies where inter-model variability exists from the use of different GCMs due to different parameterisations in each

182 model. The variability in previous inter-model GCM comparisons exists despite the use of the same forcings [e.g. see

183 results highlighted in IPCC AR5]. Similarities identified between these palaeoclimate simulations conducted with

184 different GCMs using similar boundary conditions can establish confidence in the models when in agreement with

185 proxy reconstructions. However, differences identified in inter-model GCM comparisons highlight biases by all or

186 specific GCMs, or reveal sensitivities to one changed parameter, such as model resolution. Given these limitations of

187 GCM modelling, we present in this study a comparison of a suite of ECHAM5 simulations to proxy-based

188 reconstructions (where possible) and, to a lesser degree, comment on general agreement or disagreement of our

189 ECHAM5 results with other modelling studies. A detailed inter-model comparison of our results with other GCMs is

190 beyond the scope of this study, and better suited for a different study in a journal with a different focus and audience.

191 Rather, by using the same GCM and identical resolution for the time slice experiments, we reduce the number of

192 parameters (or model parameterisations) varying between simulations and thereby remove potential sources of error or

193 uncertainty that would otherwise have to be considered when comparing output from different models with different

194 parameterisations of processes, model resolution, and in some cases model forcings (boundary conditions).

195 Nevertheless, the reader is advised to use these model results with the GCM's shortcoming and uncertainties in

196 boundary condition reconstructions in mind. For example, precipitation results may require dynamical or statistical

197 downscaling to increase accuracy where higher resolution precipitation fields are required. Furthermore, readers are

198 advised to familiarise themselves with the palaeogeography reconstruction initiatives and associated uncertainties. For

199 example, while Pliocene ice sheet volume can be estimated, big uncertainties pertaining to their locations remain

200 [Haywood et al. 2010].

201

202 **2.2 Cluster analysis to document temporal and spatial changes in climatology**

203 The aim of the clustering approach is to group climate model surface grid boxes together based on similarities in

204 climate. Cluster analyses are statistical tools that allow elements (i) to be grouped by similarities in the elements'

205 attributes. In this study, those elements are spatial units, the elements' attributes are values from different climatic

206 variables, and the measure of similarity is given by a statistical distance. The four basic variables used as climatic

207 attributes of these spatial elements are: near-surface (2m) air temperature, seasonal 2m air temperature amplitude,

208 precipitation rate, and seasonal precipitation rate amplitude. Since monsoonal winds are a dominant feature of the

209 climate in the South Asia region, near surface (10m) speeds of u-wind and v-wind (zonal and meridional wind

210 components, respectively) during the monsoon season (July) and outside the monsoon season (January) are included as

211 additional variables in our analysis of that region. Similarly, u-wind and v-wind speeds during (January) and outside

212 (July) the monsoon season in South America are added to the list of considered variables to take into account the South

213 American Monsoon System (SASM) in the cluster analysis for this region. The long-term monthly means of those

214 variables are used in a hierarchical clustering method, followed by a non-hierarchical k-means correction with

215 randomised re-groupment [Mutz et al., 2016; Wilks, 2011; Paeth, 2004; Bahrenberg et al., 1992].

216 The hierarchical part of the clustering procedure starts with as many clusters as there are elements (ni), then

217 iteratively combines the most similar clusters to form a new cluster using centroids for the linkage procedure for

218 clusters containing multiple elements. The procedure is continued until the desired number of clusters (k) is reached.

219 One disadvantage of a pure hierarchical approach is that elements cannot be re-categorised once they are assigned to a

220 cluster, even though the addition of new elements to existing clusters changes the clusters' defining attributes and could

221 warrant a re-categorisation of elements. We address this problem by implementation of a (non-hierarchical) k-means

222   clustering correction [e.g. Paeth, 2004]. Elements are re-categorizsed based on the multivariate centroids determined by

223   the hierarchical cluster analysis in order to minimisze the sum of deviations from the cluster centroids. The Mahalanobis

224   distance [e.g. Wilks, 2011] is used as a measure of similarity or distance between the cluster centroids, since it is a

225   statistical distance and thus not sensitive to different variable units. The Mahalanobis distance also accounts for possible

226   multi-collinearity between variables.

227   The end results of the cluster analyses are subdivisions of the climate in the investigated regions into k

228   subdomains or clusters based on multiple climate variables. The region-specific k has to be prescribed before the

229   analyses. A large k may result in redundant additional clusters describing very similar climates, thereby defeating the

230   purpose of the analysis to identify and describe the dominant, distinctly different climates in the region and their

231   geographical coverage. Since it is not possible to know a priori the ideal number of clusters, k was varied between 3 and

232   10 for each region and the results presented below identify the optimal number of visibly distinctly different clusters

233   from the analysis. Optimal k was determined by assessing the distinctiveness and similarities between the climate

234   clusters in the systematic process of increasing k from 3 to 10. Once an increase in k no longer resulted in the addition

235   of another cluster that was climatologically distinctly different from the others, and instead resulted in at least two

236   similar clusters, k of the previous iteration was chosen as the optimal k for the region.

237   The cluster analysis ultimately results in a description of the geographical extent of a climate (cluster)

238   characterised by a certain combination of mean values for each of the variables associated with the climate. For

239   example, climate cluster 1 may be the most tropical climate in a region and thus be characterised by a high precipitation

240   values, high temperature values and low seasonal temperature amplitude. Each of the results (consisting of the

241   geographical extent of climates and mean vectors describing the climate) can be viewed as an optimal classification for

242   the specific region and time. It serves primarily as a means for providing an overview of the climate in each of the

243   regions at different times, reduces dimensionality of the raw simulation output, and identify regions of climatic

244   homogeneity that is difficult to notice by viewing simple maps of each climate variable. Its synoptic purpose is similar

245   to that of the widely known Köppen-Geiger classification scheme [Peel et al., 2007], but we allow for optimal

246   classification rather than prescribe classes, and our selection of variables is more restricted and made in accordance with

247   the focus of this study.

248

249   **3. Results**

250   Results from our analysis are first presented for general changes in global temperature and precipitation for the

251   different time slices (Fig. 2, 3), which is then followed by an analysis of changes in the climatology of selected orogens.

252   A more detailed description of temperature and precipitation changes in our selected orogens is presented in subsequent

253   subsections (Fig. 4 and following). All differences in climatology are expressed relative to the PI control run. Changes

254 relative to the PI rather than PD conditions are presented to avoid interpreting an anthropogenic bias in the results and

255 focusing instead on pre-anthropogenic variations in climate. For brevity, near-surface (2m) air temperature and total

256 precipitation rate are referred to as temperature and precipitation.

257

258 **3.1 Global differences in mean annual temperature**

259   This section describes the differences between simulated MH, LGM, and PLIO annual mean temperature anomalies

260 alies with respect to PI shown in Fig. 2b, and PI temperature absolute values shown in Fig. 2a. Most temperature differ-

261 ences between the PI and MH climate are within -1°C to 1°C. Exceptions to this are the Hudson Bay, Weddell Sea and

262 Ross Sea regions which experience warming of 1-3°C, 1-5°C and 1-9°C respectively. Continental warming is mostly re-

263 stricted to low-altitude South America, Finland, western Russia, the Arabian Peninsula (1-3°C) and subtropical north

264 Africa (1-5°C). Simulation results show that LGM and PLIO annual mean temperature deviate from the PI means the

265 most. The global PLIO warming and LGM cooling trends are mostly uniform in direction, but the magnitude varies re-

266 gionally. The strongest LGM cooling is concentrated in regions where the greatest change in ice extent occurs (as indic-

267 ated on Fig. 2), i.e. Canada, Greenland, the North Atlantic, Northern Europe and Antarctica. Central Alaska shows no

268 temperature changes, whereas coastal South Alaska experiences cooling of ≤ 9°C. Cooling in the US Pacific northwest

269 is uniform and between 11 and 13°C. Most of high-altitude South America experiences mild cooling of 1-3°C, 3-5°C in

270 the central Andes and ≤ 9°C in the south. Along the Himalayan orogen, LGM temperature values are 5-7°C below PI

271 values. Much of central Asia and the Tibetan plateau cools by 3-5°C, and most of India, low-altitude China and south-

272 east Asia by 1-3°C.

273   In the PLIO climate, parts of Antarctica, Greenland and the Greenland Sea experience the greatest temperature

274 increase (≤ 19°C). Most of southern Alaska warms by 1-5°C and ≤ 9°C near McCarthy, Alaska. The US Pacific northw-

275 est warms by 1-5°C. The strongest warming in South America is concentrated at the Pacific west coast and the Andes

276 (1-9°C), specifically between Lima and Chiclayo, and along the Chilean-Argentinian Andes south of Bolivia (≤ 9°C).

277 Parts of low-altitude South America to the immediate east of the Andes experience cooling of 1-5°C. The Himalayan

278 orogen warms by 3-9°C, whereas Myanmar, Bangladesh, Nepal, northern India and northeast Pakistan cool by 1-9°C.

279

280 **3.2 Global differences in mean annual precipitation**

281   Notable differences occur between simulated MH, LGM, PLIO annual mean precipitation anomalies with re-

282 spect to PI shown in Fig. 3b, and the PI precipitation absolute values shown in Fig. 3a. Of these, MH precipitation devi-

283 ates the least from PI values. The differences between MH and PI precipitation on land appear to be largest in northern

284 tropical Africa (increase ≤1200 mm/a) and along the Himalayan orogen (increase ≤2000 mm/a) and in central Indian

285 states (decrease) ≤500mm. The biggest differences in western South America are precipitation increases in central Chile

286 between Santiago and Puerto Montt. The LGM climate shows the largest deviation in annual precipitation from the PI

287 climate, and precipitation on land mostly decreases. Exceptions are increases in precipitation rates in North American

288 coastal regions, especially in coastal South Alaska ($\leq$2300 mm/a) and the US Pacific Northwest ($\leq$1700 mm/a). Further

289 exceptions are precipitation increases in low-altitude regions immediately east of the Peruvian Andes ($\leq$1800 mm/a),

290 central Bolivia ($\leq$1000 mm/a), most of Chile ($\leq$1000 mm/a) and northeast India ($\leq$1900 mm/a). Regions of notable pre-

291 cipitation decrease are northern Brazil ($\leq$1700 mm/a), southernmost Chile and Argentina ($\leq$1900 mm/a), coastal south

292 Peru ($\leq$700 mm/a), central India ($\leq$2300 mm/a) and Nepal ($\leq$1600 mm/a).

293   Most of the precipitation on land in the PLIO climate is higher than those in the PI climate. Precipitation is en-

294 hanced by ca. 100-200 mm/a in most of the Atacama desert, by $\leq$1700 mm/a south of the Himalayan orogen and by

295 $\leq$1400 mm/a in tropical South America. Precipitation significantly decreases in central Peru ($\leq$2600mm), southernmost

296 Chile ($\leq$2600mm) and from eastern Nepal to northernmost northeast India ($\leq$2500mm).

297

3.3 Palaeoclimate characterisation from the cluster analysis and changes in regional climatology

299   In addition to the above described global changes, the PLIO to PI regional climatology changes substantially in

300 the four investigated regions of: South Asia (section 3.3.1), the Andes (section 3.3.2), South Alaska (section 3.3.3) and

301 the Cascade Range (section 3.3.4). Each climate cluster defines separate distinct climate that is characterized by the

302 mean values of the different climate variables used in the analysis. The clusters are calculated by taking the arithmetic

303 means of all the values (climatic means) calculated for the grid boxes within each region. The regional climates are

304 referred to by their cluster number C_1, C_2, …, C_k, where k is the number of clusters specified for the region. The clusters

305 for specific palaeoclimates are mentioned in the text as $C_{i(t)}$, where i corresponds to the cluster number (i=1, …, k) and t

306 to the simulation time period (t=PI, MH, LGM, PLIO). The descriptions first highlight the similarities and then the

307 differences in regional climate. The cluster means of seasonal near-surface temperature amplitude and seasonal

308 precipitation amplitude are referred to as temperature and precipitation amplitude. The median, 25th percentile, 75th

309 percentile, minimum and maximum values for annual mean precipitation are referred to as P_{md}, P_{25}, P_{75}, P_{min} and P_{max}

310 respectively. Likewise, the same statistics for temperature are referred to as T_{md}, T_{25}, T_{75}, T_{min} and T_{max}. These are

311 presented as boxplots of climate variables in different time periods. When the character of a climate cluster is described

312 as "high", "moderate" and "low", the climatic attribute's values are described relative to the value range of the specific

313 region in time, thus high PLIO precipitation rates may be higher than high LGM precipitation rates. The character is

314 presented a raster plots, to allow compact visual representation of it. The actual mean values for each variable in every

315 time-slice and region-specific cluster are included in tables in the supplementary material.

316

3.3.1 Climate change and palaeoclimate characterisation in South Asia, Central- and East Asia

318   This section describes the regional climatology of the four investigated Cenozoic time slices and how

319   precipitation and temperature changes from PLIO to PI times in tropical, temperate and high altitude regions. LGM and

320   PLIO simulations show the largest simulated temperature and precipitation deviations (Fig. 4b) from PI temperature and

321   precipitation (Fig. 4a) in the South Asia region. LGM temperatures are 1-7°C below PI temperatures and the direction

322   of deviation is uniform across the study region. PLIO temperature is mostly above PI temperatures by 1-7°C. The

323   cooling of 3-5°C in the region immediately south of the Himalayan orogen represents one of the few exceptions.

324   Deviations of MH precipitation from PI precipitation in the region are greatest along the eastern Himalayan orogeny,

325   which experiences an increase in precipitation ($\leq$2000 mm/a). The same region experiences a notable decrease in

326   precipitation in the LGM simulation, which is consistent in direction with the prevailing precipitation trend on land

327   during the LGM. PLIO precipitation on land is typically higher than PI precipitation.

328   Annual means of precipitation and temperature spatially averaged for the regional subdivisions and the different

329   time slice simulations have been compared. The value range P_{25} to P_{75} of precipitation is higher for tropical South Asia

330   than for temperate and high altitude South Asia (Fig. 5 a-c). The LGM values for P_{25}, P_{md} and P_{75} are lower than for the

331   other time slice simulations, most visibly for tropical South Asia (ca. 100 mm/a). The temperature range (both T_{75}-T_{25}

332   and T_{max}- T_{min}) is smallest in the hot (ca. 21°C) tropical South Asia, wider in the high altitude (ca. -8°C) South Asia, and

333   widest in the temperate (ca. 2°C) South Asia region (Fig. 5 d-f). T_{md}, T_{25} and T_{75} values for the LGM are ca. 1°C, 1-2°C

334   and 2°C below PI and MH temperatures in tropical, temperate and high altitude South Asia respectively, whereas the

335   same temperature statistics for the PLIO simulation are ca. 1°C above PI and MH values in all regional subdivisions

336   (Fig. 5 d-f). With respect to PI and MH values, precipitation and temperature are generally lower in the LGM and higher

337   in the PLIO in tropical, temperate and high altitude South Asia.

338   In all time periods, the wettest climate cluster C_1 covers an area along the southeastern Himalayan orogen (Fig. 6

339   a-d) and is defined by the highest precipitation amplitude (dark blue, Fig. 6 e-h). $C_{5(PI)}$, $C_{3(MH)}$, $C_{4(LGM)}$ and $C_{5(PLIO)}$ are

340   characterized by (dark blue, Fig. 6e-h) the highest temperatures, u-wind and v-wind speeds during the summer monsoon

341   in their respective time periods, whereas $C_{4(PI)}$, $C_{5(MH)}$, and $C_{6(LGM)}$ are defined by low temperatures and highest

342   temperature amplitude, u-wind and v-wind speeds outside the monsoon season (in January) in their respective time

343   periods (Fig. 6 e-h). The latter 3 climate classes cover much of the more continental, northern landmass in their

344   respective time periods and represents a cooler climate affected more by seasonal temperature fluctuations (Fig. 6 a-d).

345   The two wettest climate clusters C_1 and C_2 are more restricted to the eastern end of the Himalayan orogen in the LGM

346   than during other times, indicating that the LGM precipitation distribution over the South Asia landmass is more

347   concentrated in this region than in other time slice experiments.

348

349   **3.3.2 Climate change and palaeoclimate characterisation in the Andes, Western South America**

350   This section describes the cluster analysis based regional climatology of the four investigated Late Cenozoic

351   time slices and illustrates how precipitation and temperature changes from PLIO to PI in tropical and temperate low-

352   and high altitude (i.e. Andes) regions in western South America (Fig. 7-9).

353   LGM and PLIO simulations show the largest simulated deviations (Fig. 7b) from PI temperature and

354   precipitation (Fig. 7a) in western South America. The direction of LGM temperature deviations from PI temperatures is

355   negative and uniform across the region. LGM temperatures are typically 1-3°C below PI temperatures across the region,

356   and 1-7°C below PI values in the Peruvian Andes, which also experience the strongest and most widespread increase in

357   precipitation during the LGM ($\leq$1800 mm/a). Other regions, such as much of the northern Andes and tropical South

358   America, experience a decrease of precipitation in the same experiment. PLIO temperature is mostly elevated above PI

359   temperatures by 1-5°C. The Peruvian Andes experience a decrease in precipitation ($\leq$2600mm), while the northern

360   Andes are wetter in the PLIO simulation compared to the PI control simulation.

361   PI, MH, LGM and PLIO precipitation and temperature means for regional subdivisions have been compared.

362   The P_{25} to P_{75} range is smallest for the relatively dry temperate Andes and largest for tropical South America and the

363   tropical Andes (Fig. 8 a-d). P_{max} is lowest in the PLIO in all four regional subdivisions even though P_{md}, P_{25} and P_{75} in

364   the PLIO simulation are similar to the same statistics calculated for PI and MH time slices. P_{md}, P_{25} and P_{75} for the LGM

365   are ca. 50 mm/a lower in tropical South America and ca. 50 mm/a higher in the temperate Andes. Average PLIO

366   temperatures are slightly warmer and LGM temperatures are slightly colder than PI and MH temperatures in tropical

367   and temperate South America (Fig. 8 e and f). These differences are more pronounced in the Andes, however. T_{md}, T_{25}

368   and T_{75} are ca. 5°C higher in the PLIO climate than in PI and MH climates in both temperate and tropical Andes,

369   whereas the same temperatures for the LGM are ca. 2-4°C below PI and MH values (Fig. 8 g and h).

370   For the LGM, the model computes drier-than-PI conditions in tropical South America and tropical Andes,

371   enhanced precipitation in the temperate Andes, and a decrease in temperature that is most pronounced in the Andes. For

372   the PLIO, the model predicts precipitation similar to PI, but with lower precipitation maxima. PLIO temperatures

373   generally increase from PI temperatures, and this increase is most pronounced in the Andes.

374   The climate variability in the region is described by six different clusters (Fig. 9 a-d), which have similar

375   attributes in all time periods. The wettest climate C_1 is also defined by moderate to high precipitation amplitudes, low

376   temperatures and moderate to high u-wind speeds in summer and winter in all time periods (dark blue, Fig. 9 e-h). $C_{2(PI)}$,

377    $C_{2(MH)}$, $C_{3(LGM)}$ and $C_{2(PLIO)}$ are characterized by high temperatures and low seasonal temperature amplitude (dark blue,

378   Fig. 9 e-h), geographically cover the north of the investigated region, and represent a more tropical climate. $C5_{(PI)}$,

379    $C5_{(MH)}$, $C6_{(LGM)}$ and $C6_{(PLIO)}$ are defined by low precipitation and precipitation amplitude, high temperature amplitude and

380   high u-wind speeds in winter (Fig. 9 e-h), cover the low-altitude south of the investigated region (Fig. 9 a-d) and

381   represent dry, extra-tropical climates with more pronounced seasonality. In the PLIO simulation, the lower-altitude east

382 of the region has four distinct climates, whereas the analysis for the other time slice experiments only yield three

383 distinct climates for the same region.

384

385 ### 3.3.3 Climate change and palaeoclimate characterisation in the St. Elias Range, Southeast Alaska

386 This section describes the changes in climate and the results from the cluster analysis for South Alaska (Fig. 10-12). As

387 is the case for the other study areas, LGM and PLIO simulations show the largest simulated deviations (Fig. 10b) from

388 PI temperature and precipitation (Fig. 10a). The sign of LGM temperature deviations from PI temperatures is negative

389 and uniform across the region. LGM temperatures are typically 1-9°C below PI temperatures, with the east of the study

390 area experiencing largest cooling. PLIO temperatures are typically 1-5°C above PI temperatures and the warming is

391 uniform for the region. In comparison to the PI simulation, LGM precipitation is lower on land, but higher ($\leq$2300mm)

392 in much of the coastal regions of South Alaska. Annual PLIO precipitation is mostly higher ($\leq$800mm) than for PI.

393 P_{md}, P_{25}, P_{75}, P_{min} and P_{max} for South Alaskan mean annual precipitation do not differ much between PI, MH and

394 PLIO climates, while P_{md}, P_{25}, P_{75} and P_{min} decrease by ca. 20-40 mm/a and P_{max} increases during the LGM (Fig. 11a).

395 The Alaskan PLIO climate is distinguished from the PI and MH climates by its higher (ca. 2°C) regional temperature

396 means, T_{25}, T_{75} and T_{md} (Fig. 11b). Mean annual temperatures, T_{25}, T_{75}, T_{min} and T_{max} are lower in the LGM than in any

397 other considered time period (Fig. 11b), and about 3-5°C lower than during the PI and MH.

398 Distinct climates are present in the PLIO to PI simulations for Southeast Alaska. Climate cluster C_1 is always

399 geographically restricted to coastal southeast Alaska (Fig. 12 a-d) and characterized by the highest precipitation,

400 precipitation amplitude, temperature, and by relatively low temperature amplitude (dark blue, Fig. 12 e-h). Climate C_2 is

401 characterized by moderate to low precipitation, precipitation amplitude, temperature, and by low temperature amplitude.

402 C_2 is either restricted to coastal southeast Alaska (in MH and LGM climates) or coastal southern Alaska (in PI and PLIO

403 climates). Climate C_3 is described by low precipitation, precipitation amplitude, temperature, and moderate temperature

404 amplitude in all simulations. It covers coastal western Alaska and separates climate C_1 and C_2 from the northern C_4

405 climate. Climate C_4 is distinguished by the highest mean temperature amplitude, by low temperature and precipitation

406 amplitude, and by lowest precipitation.

407 The geographical ranges of PI climates C_1- C_4 and PLIO climates C_1- C_4 are similar. $C_{1(PI/PLIO)}$ and $C_{2(PI/PLIO)}$ spread

408 over a larger area than $C_{1(MH/LGM)}$ and $C_{2(MH/LGM)}$. $C_{2(PI/PLIO)}$ are not restricted to coastal southeast Alaska, but also cover the

409 coastal southwest of Alaska. The main difference in characterization between PI and PLIO climates C_1- C_4 lies in the

410 greater difference (towards lower values) in precipitation, precipitation amplitude and temperature from $C_{1(PLIO)}$ to

411 $C_{2(PLIO)}$ compared to the relatively moderate decrease in those means from $C_{1(PI)}$ to $C_{2(PI)}$.

412

413 ### 3.3.4 Climate change and palaeoclimate characterisation in the Cascade Range, US Pacific Northwest

414   This section describes the character of regional climatology in the US Pacific Northwest and its change over time

415   (Fig. 13-15). The region experiences cooling of typically 9-11°C on land during the LGM, and warming of 1-5°C

416   during the PLIO (Fig. 13b) when compared to PI temperatures (Fig. 13a). LGM precipitation increases over water,

417   decreases on land by ≤800 mm/a in the North and in the vicinity of Seattle and increases on land by ≤1400 mm/a on

418   Vancouver Island, around Portland and the Olympic Mountains, whereas PLIO precipitation does not deviate much

419   from PI values over water and varies in the direction of deviation on land. MH temperature and precipitation deviation

420   from PI values are negligible.

421   P_{md}, P_{25}, P_{75}, P_{min} and P_{max} for the Cascade Range do not notably differ between the four time periods (Fig. 14a).

422   The LGM range of precipitation values is slightly larger than that of the PI and MH with slightly increased P_{md}, while

423   the respective range is smaller for simulation PLIO. The T_{md}, T_{25}, T_{75} and T_{max} values for the PLIO climate are ca. 2°C

424   higher than those values for PI and MH (Fig. 14b). All temperature statistics for the LGM are notably (ca. 13°C) below

425   their analogues in the other time periods (Fig. 14b).

426   PI, LGM and PLIO clusters are similar in both their geographical patterns (Fig. 15 a, c, d) and their

427   characterization by mean values (Fig. 15 e, g, h). C_1 is the wettest cluster and shows the highest amplitude in

428   precipitation. The common characteristics of the C_2 cluster are moderate to high precipitation and precipitation

429   amplitude. C_4 is characterized by the lowest precipitation and precipitation amplitudes, and the highest temperature

430   amplitudes. Regions assigned to clusters C_1 and C_2 are in proximity to the coast, whereas C_4 is geographically restricted

431   to more continental settings.

432   In the PI and LGM climates, the wettest cluster C_1 is also characterized by high temperatures (Fig 10 e, g).

433   However, virtually no grid boxes were assigned to $C_{1(LGM)}$. $C_{1(MH)}$ differs from other climate state's C_1 clusters in that it is

434   also described by moderate to high near surface temperature and temperature amplitude (Fig 10 f), and in that it is

435   geographically less restricted and, covering much of Vancouver Island and the continental coastline north of it (Fig 10

436   b). Near surface temperatures are highest for C_2 in PI, LGM and PLIO climates (Fig 10 e, g, h) and low for $C_{2(MH)}$ (Fig

437   10 f). $C_{2(MH)}$ is also geographically more restricted than C_2 clusters in PI, LGM and PLIO climates (Fig 10 a-d). $C_{2(PI)}$,

438    $C_{2(MH)}$ and $C_{2(LGM)}$ have a low temperature amplitude (Fig 10 e-g), whereas $C_{2(PLIO)}$ is characterized by a moderate

439   temperature amplitude (Fig 10 h).

440

441   **4. Discussion**

442   In the following, we synthesise our results and compare to previous studies that investigate the effects of

443   temperature and precipitation change on erosion. Since our results do not warrant merited discussion of subglacial

444   processes without additional work that is beyond the scope of this study, we instead advise caution in interpreting the

445   presented precipitation and temperature results in an erosional context where the regions are covered with ice. For

446 convenience, ice cover is indicated on figures 2,3,47,10 and 13, and a summary of ice cover used as boundary

447 conditions for the different time slice experiments is included in the supplemental material. Where possible, we relate

448 the magnitude of climate change predicted in each geographical study area with terrestrial proxy data.

449

450 **4.1 Synthesis of temperature changes**

451

452 **4.1.1 Temperature changes and implications for weathering and erosion**

453   Changes in temperature can affect physical weathering due to temperature-induced changes in periglacial

454 processes and promote frost cracking and frost creep [e.g., Matsuoka, 2001; Schaller et al., 2002; Matsuoka and

455 Murton, 2008; Delunel et al., 2010; Andersen et al., 2015; Marshall et al., 2015], and also biotic weathering and erosion

456 [e.g. Moulton et al., 1998; Banfield et al., 1999; Dietrich and Perron, 2006]. Quantifying and understanding past

457 changes in temperature is thus vital for our understanding of denudation histories. In the following, we highlight regions

458 in the world where future observational studies might be able to document significant warming or cooling that would

459 influence temperature related changes in physical and chemical weathering over the last ~3 Ma.

460   Simulated MH temperatures show little deviation (typically < 1°C) from PI temperatures in the investigated

461 regions (Fig. 2b), suggesting little difference in MH temperature-related weathering. The LGM experiences widespread

462 cooling, which is accentuated at the poles. , increasin the equator-to-pole pressure gradient and consequently

463 strengthens global atmospheric circulation. Despite this global trend, cooling in coastal South Alaska is higher (≤ 9°C)

464 than in central Alaska (0±1°C). The larger temperature difference in South Alaska geographically coincides with ice

465 cover (Fig. 10b), and should thus be interpreted in context of a different erosional regime. Cooling in most of the lower-

466 latitude regions in South America and central to southeast Asia is relatively mild. The greatest temperature differences

467 in South America are observed for western Patagonia, which was mostly covered by glaciers. The Tibetan plateau

468 experiences more cooling (3-5°C) than adjacent low-altitude regions (1-3°C) during the LGM.

469   The PLIO simulation is generally warmer, and temperature differences are accentuated warming at the poles.

470 Warming in simulation PLIO is greatest in parts of Canada, Greenland and Antarctica (up to 19°C), which

471 geographically coincides with the presence of ice in the PI reference simulation and thus may be attributed to

472 differences in ice cover. It should therefore also be regarded as areas in which process domain shifted from glacial to

473 non-glacial. The warming in simulation PLIO in South Alaska and the US Pacific northwest is mostly uniform and in

474 the range of 1-5°C. As before, changes in ice cover reveal that the greatest warming may be associated with the absence

475 of glaciers relative to the PI simulation. Warming in South America is concentrated at the Pacific west coast and the

476 Andes between Lima and Chiclayo, and along the Chilean-Argentinian Andes south of Bolivia (≤ 9°C).

477   Overall, annual mean temperatures in the MH simulation show little deviation from PI values. The more

478   significant temperature deviations of the colder LGM and of the warmer PLIO simulations are accentuated at the poles

479   leading to higher and lower equator-to-pole temperature gradients respectively. The largest temperature-related changes

480   (relative to PI conditions) in weathering and subsequent erosion, in many cases through a shift in the process domain

481   from glacial to non-glacial or vice versa, are therefore to be expected in the LGM and PLIO climates.

482

483   **4.1.2 Temperature comparison to other studies**

484   LGM cooling is accentuated at the poles, thus increases the equator-to-pole pressure gradient and consequently

485   strengthens global atmospheric circulation, and is in general agreement with studies such as Otto-Bliesner et al. [2006]

486   and Braconnot et al. [2007]. The PLIO simulation shows little to no warming in the tropics and accentuated warming at

487   the poles, as do findings of Salzmann et al. [2011] and Robinson [2009] and Ballantyne [2010] respectively. This would

488   reduce the equator-to-pole sea and land surface temperature gradient, as also reported by Dowsett et al. [2010], and also

489   weaken global atmospheric circulation. Agreement with proxy-based reconstructions, as is the case of the relatively

490   little warming in lower latitudes, is not surprising given that sea surface temperature reconstructions (derived from

491   previous coarse resolution coupled ocean-atmosphere models) are prescribed in this uncoupled atmosphere simulation.

492   It should be noted that coupled ocean-atmosphere simulations do predict more low-latitude warming [e.g. Stepanek and

493   Lohmann 2012; Zhang et al. 2013b]. The PLIO warming in parts of Canada and Greenland (up to 19°C) and consistent

494   with values based on multi-proxy studies [Ballantyne et al., 2010]. Due to a scarcity of palaeobotanical proxies in

495   Antarctica, reconstruction-based temperature and ice-sheet extent estimates for a PLIO climate have high uncertainties

496   [Salzmann et al., 2011], making model validation difficult. Furthermore, controversy about relatively little warming in

497   the south polar regions compared to the north polar regions remains [e.g. Hillenbrand and Fütterer, 2002; Wilson et al.,

498   2002]. Mid-latitude PLIO warming is mostly in the 1-3°C range with notable exceptions of cooling in the northern

499   tropics of Africa and on the Indian subcontinent, especially south of the Himalayan orogen.

500

501   **4.2 Synthesis of precipitation changes**

502

503   **4.2.1 Precipitation and implications for weathering and erosion**

504   Changes in precipitation affects erosion through river incision, sediment transport, and erosion due to extreme

505   precipitation events and storms [e.g. Whipple and Tucker, 1999; Hobley et al., 2010]. Furthermore, vegetation type and

506   cover also co-evolve with variations in precipitation and with changes in geomorphology [e.g. Marston 2010; Roering

507   et al., 2010]. These vegetation changes in turn modify hillslope erosion by increasing root mass and canopy cover, and

508   decreasing water-induced erosion via surface runoff [e.g. Gyssels et al., 2005]. Therefore, understanding and

509   quantifying changes in precipitation in different palaeoclimates is necessary for a more complete reconstruction of

510   orogen denudation histories. A synthesis of predicted precipitation changes is provided below, and highlights regions

511   where changes in river discharge and hillslope processes might be impacted by climate change over the last ~3 Ma.

512   Most of North Africa is notably wetter during the MH, which is characteristic of the African Humid Period

513   [Sarnthein 1978]. This pluvial regional expression of the Holocene Climatic Optimum is attributed to sudden changes in

514   the strength of the African monsoon caused by orbital-induced changes in summer insolation [e.g. deMenocal et al.

515   2000]. Southern Africa is characterised by a wetter climate to the east and drier climate to the west of the approximate

516   location of the Congo Air Boundary (CAB), the migration of which has previously been cited as a cause for

517   precipitation changes in East Africa [e.g. Juninger et al. 2014]. In contrast, simulated MH precipitation rates show little

518   deviation from the PI in most of the investigated regions, suggesting little difference in MH precipitation-related

519   erosion. The Himalayan orogen is an exception and shows a precipitation increase of up to 2000 mm/a. The climate's

520   enhanced erosion potential, that could result from such a climatic change, should be taken into consideration when

521   palaeo-erosion rates estimated from the geological record in this area are interpreted [e.g. Bookhagen et al., 2005].

522   Specifically, higher precipitation rates (along with differences in other rainfall-event parameters) could increase the

523   probability of mass movement events on hillslopes, especially where hillslopes are close to the angle of failure [e.g.

524   Montgomery, 2001], and modify fluxes to increase shear stresses exerted on river beds and increase stream capacity to

525   enhance erosion on river beds (e.g. by abrasion).

526   Most precipitation on land is decreased during the LGM due to large-scale cooling and decreased evaporation

527   over the tropics, resulting in an overall decrease in inland moisture transport [e.g., Braconnot et al. 2007]. North

528   America, south of the continental ice sheets, is an exception and experiences increases in precipitation. For example, the

529   investigated US Pacific Northwest and the southeastern coast of Alaska experience experience strongly enhanced

530   precipitation of ≤1700 mm/a and ≤2300 mm/a, respectively. These changes geographically coincide with differences in

531   ice extent. An increase in precipitation in these regions may have had direct consequences on the glaciers' mass balance

532   and equilibrium line altitudes, where the glaciers' effectiveness in erosion is highest [e.g. Egholm et al., 2009; Yanites

533   and Ehlers, 2012]. The differences in the direction of precipitation changes, and accompanying changes in ice cover

534   would likely result in more regionally differentiated variations in precipitation-specific erosional processes in the St.

535   Elias Range rather than causing systematic offsets for the LGM. Although precipitation is significantly reduced along

536   much of the Himalayan orogen (≤1600 mm/a), , northeast India experiences strongly enhanced precipitation (≤1900

537   mm/a). This could have large implications for studies of uplift and erosion at orogen syntaxes, where highly localized

538   and extreme denudation has been documented [e.g. Koons et al., 2013; Bendick and Ehlers, 2014].

539   Overall, the PLIO climate is wetter than the PI climate, in particular in the (northern) mid-latitudes, and possibly

540   related to a northward shift of the northern Hadley cell boundary that is ultimately the result of a reduced equator-to-

541   pole temperature gradient [e.g. Haywood et al. 2000, 2013; Dowsett et al. 2010]. Most of the PLIO precipitation over

land increases , esp. at the Himalayan orogen by ≤1400 mm/a, and decreases from eastern Nepal to Namcha Barwa

(≤2500 mm/a). Most of the Atacama Desert experiences an increase in precipitation by 100-200 mm/a, which may have

to be considered in erosion and uplift history reconstructions for the Andes. A significant increase (~2000 mm/a) in

precipitation from simulation PLIO to modern conditions is simulated for the eastern margin of the Andean Plateau in

Peru and for northern Bolivia. This is consistent with recent findings of a pulse of canyon incision in these locations in

the last ~3 Ma [Lease and Ehlers, 2013].

Overall, the simulated MH precipitation varies least from PI precipitation. The LGM is generally drier than the

PI simulation, even though pockets of a wetter-than-PI climate do exist, such as much of coastal North America. Extra-

tropical increased precipitation of the PLIO simulation and decreased precipitation of the LGM climate may be the

result of decreased and increased equator-to-pole temperature gradients, respectively.

4.2.2 Precipitation comparison to other studies

The large scale LGM precipitation decrease on land, related to cooling and decreased evaporation over the

tropics, and greatly reduced precipitation along much of the Himalayan orogeny, is consistent with previous studies by,

(for example) Braconnot et al. [2007]. The large scale PLIO precipitation increase due to a reduced equator-to-pole

temperature gradient, has previously been pointed out by e.g. Haywood et al. [2000, 2013] and Dowsett et al. [2010]. A

reduction of this gradient by ca. 5°C is indeed present in the PLIO simulation of this study (Fig. 2b). This precipitation

increase over land agrees well with simulations performed at a lower spatial model resolution [cf. Stepanek and

Lohmann, 2012]. Section 4.4 includes a more in-depth discussion of how simulated MH and LGM precipitation

differences compare with proxy-based reconstructions in South Asia and South America.

4.3 Trends in Late Cenozoic changes in regional climatology

This section describes the major changes in regional climatology and highlights their possible implications on

erosion rates.

Himalaya-Tibet, South Asia

In South Asia, cluster-analysis based categorization and description of climates (Fig. 6) remains similar

throughout time. However, the two wettest climates (C_1 and C2) are geographically more restricted to the eastern

Himalayan orogen in the LGM simulation. Even though precipitation over the South Asia region is generally lower, this

shift indicates that rainfall on land is more concentrated in this region and that the westward drying gradient along the

orogen is more accentuated than during other time periods investigated here. While there is limited confidence in the

global Atmospheric General Circulation Model's abilities to accurately represent meso-scale precipitation patterns [e.g.

574 Cohen 1990], the simulation warrants careful consideration of possible, geographically non-uniform offsets in

575 precipitation in investigations of denudation and uplift histories.

576   MH precipitation and temperature in tropical, temperate and high-altitude South Asia is similar to PI

577 precipitation and temperature, whereas LGM precipitation and temperatures are generally lower (by ca. 100 mm/a and

578 1-2°C respectively), possibly reducing precipitation-driven erosion and enhancing frost-driven erosion in areas pushed

579 into a near-zero temperature range during the LGM.

580

581 *Andes, South America*

582   Clusters in South America (Fig. 9), which are somewhat reminiscent of the Köppen and Geiger classification

583 [Kraus, 2001], remain mostly the same over the last 3 Ma. In the PLIO simulation, the lower-altitude east of the region

584 is characterized by four distinct climates, which suggests enhanced latitudinal variability in the PLIO climate compared

585 to PI with respect temperature and precipitation.

586   The largest temperature deviations from PI values are derived for the PLIO simulation in the (tropical and

587 temperate) Andes, where temperatures exceed PI values by 5°C. On the other hand, LGM temperatures in the Andes are

588 ca. 2-4°C below PI values in the same region (Fig 7 g and h). In the LGM simulation, tropical South America

589 experiences ca. 50 mm/a less precipitation, the temperate Andes receive ca. 50 mm/a more precipitation than in PI and

590 MH simulations. These latitude-specific differences in precipitation changes ought to be considered in attempts to

591 reconstruct precipitation-specific palaeoerosion rates in the Andes on top of longitudinal climate gradients highlighted

592 by, e.g., Montgomery et al. [2001].

593

594 *St. Elias Range, South Alaska*

595   South Alaska is subdivided into two wetter and warmer clusters in the south, and two drier, colder clusters in the

596 north. The latter are characterised by increased seasonal temperature variability due to being located at higher latitudes

597 (Fig. 12). The different equator-to-pole temperature gradients for LGM and PLIO may affect the intensity of the Pacific

598 North American Teleconnection (PNA) [Barnston and Livzey, 1987], which has significant influence on temperatures

599 and precipitation, especially in southeast Alaska, and may in turn result in changes in regional precipitation and

600 temperature patterns and thus on glacier mass balance. Changes in the Pacific Decadal Oscillation, which is related to

601 the PNA pattern, has previously been connected to differences in Late Holocene precipitation [Barron and Anderson,

602 2011]. While this climate cluster pattern appears to be a robust feature for the considered climate states, and hence over

603 the recent geologic history, the LGM sets itself apart from PI and MH climates by generally lower precipitation (20-40

604 mm) and lower temperatures (3-5°C, Fig. 10, 11), which may favour frost driven weathering during glacial climate

605 states [e.g. Andersen et al., 2015; Marshall et al. 2015] in unglaciated areas, whereas glacial processes would have

606 dominated most of this region as it was covered by ice. Simulation PLIO is distinguished by temperatures that exceed

607 PI and MH conditions by ca. 2°C, and by larger temperature and precipitation value ranges, possibly modifying

608 temperature- and precipitation-dependent erosional processes in the region of South Alaska.

609

610 *Cascade Range, US Pacific Northwest*

611 In all time slices, the geographic climate patterns, based on the cluster analysis (Fig. 15), represents an increase

612 in the degree of continentality from the wetter coastal climates to the further inland located climates with greater

613 seasonal temperature amplitude and lower precipitation and precipitation amplitude (Fig 15 e-h). The most notable

614 difference between the time slices is the strong cooling during the LGM, when temperatures are ca. 13°C (Fig. 13, 14)

615 below those of other time periods. Given that the entire investigated region was covered by ice (Fig 13), we can assume

616 a shift to glacially dominated processes.

617

618 **4.4 Comparison of simulated and observed precipitation differences**

619 The predicted precipitation differences reported in this study were compared with observed (proxy record)

620 palaeoprecipitation change. Proxy based precipitation reconstructions for the MH and LGM are presented for South

621 Asia and South America for the purpose of assessing ECHAM5 model performance, and for identifying inconsistencies

622 between neighbouring proxy data. Due to the repeated glaciations, detailed terrestrial proxy records for the time slices

623 investigated here are not available, to the best of our knowledge, for the Alaskan and Pacific NW USA studies.

624 Although marine records and records of glacier extent are available in these regions, the results from them do not

625 explicitly provide estimates of wetter/drier, or colder/warmer conditions that can be spatially compared to the

626 simulation estimates. For these two areas with no available records, the ECHAM5 predicted results therefore provide

627 predictions from which future studies can formulate testable hypotheses to evaluate.

628 The palaeoclimate changes in terrestrial proxy records compiled here are reported as "wetter than today", "drier

629 than today" or "the same as today" for each of the study locations, and plotted on top of the simulation-based difference

630 maps as upward facing blue triangles, downward facing red triangles and grey circles respectively (Fig. 16, 17). The

631 numbers listed next to those indicators are the ID numbers assigned to the studies compiled for this comparison and are

632 associated with a citation provided in the figure captions.

633 In South Asia, 14/26 results from local studies agree with the model predicted precipitation changes for the MH.

634 The model seems able to reproduce the predominantly wetter conditions on much of the Tibetan plateau, but predicts

635 slightly drier conditions north of Chengdu, which is not reflected in local reconstructions. The modest mismatch

636 between ECHAM5 predicted and proxy-based MH climate change in south Asia was also documented by Li et al.,

637 [2017], whose simulations were conducted at a coarser (T106) resolution. Despite these model-proxy differences, we

638 note that there are significant discrepancies between the proxy data themselves in neighbouring locations in the MH,

639 highlighting caution in relying solely upon these data for regional palaeoclimate reconstructions. These differences

640 could result from either poor age-constraints in the reported values, or systematic errors in the transfer functions used to

641 convert proxy measurements to palaeoclimate conditions. The widespread drier conditions on the Tibetan Plateau and

642 immediately north of Laos are confirmed by 7/7 of the palaeoprecipitation reconstructions. 23/39 of the reconstructed

643 precipitation changes agree with model predictions for South America during the MH. The model predicted wetter

644 conditions in the central Atacama desert, as well as the drier conditions northwest of Santiago are confirmed by most of

645 the reconstructions. The wetter conditions in southernmost Peru and the border to Bolivia and Chile cannot be

646 confirmed by local studies. 11/17 of the precipitation reconstructions for the LGM are in agreement with model

647 predictions. These include wetter conditions in most of Chile. The most notable disagreement can be seen in northeast

648 Chile at the border to Argentina and Bolivia, where model predicted wetter conditions are not confirmed by reported

649 reconstructions from local sites.

650 Model performance is, in general, higher for the LGM than for the MH and overall satisfactory given that it

651 cannot be expected to resolve sub-grid scale differences in reported palaeoprecipitation reconstructions. However, as

652 mentioned above, it should be noted that some locations (MH of south Asia, and MH of norther Chile) discrepancies

653 exist between neighbouring proxy samples and highlight the need for caution in how these data are interpreted. Other

654 potential sources of error resulting in disagreement of simulated and proxy-based precipitation estimates are the model's

655 shortcomings in simulating orographic precipitation at higher resolutions, and uncertainties in palaeoclimate

656 reconstructions at the local sites. In summary, although some differences are evident in both the model-proxy data

657 comparison and between neighbouring proxy data themselves, the above comparison highlights an overall good

658 agreement between the model and data for the south Asia and South American study areas. Thus, although future

659 advances in GCM model parameterisations and new or improved palaeoclimate proxy techniques are likely, the

660 palaeoclimate changes documented here are found to be in general robust and provide a useful framework for future

661 studies investigating how these predicted changes in palaeoclimate impact denudation.

662

663 **4.5 Conclusions**

664 We present a statistical cluster-analysis-based description of the geographic coverage of possible distinct

665 regional expressions of climates from four different time slices (Fig. 6, 9, 12, 15). These are determined with respect to

666 a selection of variables that characterize the climate of the region and may be relevant to weathering and erosional

667 processes. While the geographic distribution of climate remains similar throughout time (as indicated by results of four

668 different climate states representative for the climate of the last 3 Ma), results for the PLIO simulation suggests more

669 climatic variability east of the Andes (with respect to near-surface temperature, seasonal temperature amplitude,

670 precipitation, seasonal precipitation amplitude and seasonal u-wind and v-wind speeds). Furthermore, the wetter

671 climates in the South Asia region retreat eastward along the Himalayan orogen for the LGM simulation, this is due to

672 decreased precipitation along the western part of the orogen and enhanced precipitation on the eastern end, possibly

673 signifying more localised high erosion rates.

674   Most global trends of the high-resolution LGM and PLIO simulations conducted here are in general agreement

675 with previous studies [Otto-Bliesner et al., 2006; Braconnot et al., 2007; Wei and Lohmann, 2012; Lohmann et al.,

676 2013; Zhang et al., 2013b, 2014; Stepanek and Lohmann, 2012]. The MH does not deviate notably from the PI, the

677 LGM is relatively dry and cool, while the PLIO is comparably wet and warm. While the simulated regional changes in

678 temperature and precipitation usually agree with the sign (or direction) of the simulated global changes, there are

679 region-specific differences in the magnitude and direction. For example, the LGM precipitation of the Tropical Andes

680 does not deviate significantly from PI precipitation, whereas LGM precipitation in the Temperate Andes is enhanced.

681   Comparisons to local, proxy-based reconstructions of MH and LGM precipitation in South Asia and South

682 America reveal satisfactory performance of the model in simulating the reported differences. The model performs better

683 for the LGM than the MH. We note however that compilations of proxy data such as we present here, also identify

684 inconsistences between neighbouring proxy data themselves, warranting caution in the extent to which both proxy data

685 and palaeoclimate models are interpreted for MH climate change in south Asia, and western South America.

686   The changes in regional climatology presented here are manifested, in part, by small to large magnitude changes

687 in fluvial and hillslope relevant parameters such as precipitation and temperature. For the regions investigated here we

688 find that precipitation differences between the PI, MH, LGM, and PLIO are in many areas around +/- 200-600 mm/yr,

689 and locally can reach maximums of +/- 1000-2000 mm/yr (Figs. 4, 7, 10, 13). In areas where significant precipitation

690 increases are accompanied by changes in ice extent, such as parts of southern Alaska during the LGM, we would expect

691 a shift in the erosional regime to glacier dominated processes. Temperature differences between these same time periods

692 are around 1-4 °C in many places, but reach maximum values of 8-10 °C. Many of these maxima in the temperature

693 differences geographically coincide with changes in ice sheet extent and must therefore be interpreted as part of a

694 different erosional process domains. However, we also observe large temperature differences (~5°C) in unglaciated

695 areas that would be affected by hillslope, frost cracking, and fluvial processes. The magnitude of these differences are

696 not trival, and will likely impact fluvial and hillslope erosion and sediment transport, as well as biotic and abiotic

697 weathering. The regions of large magnitude changes in precipitation and temperature documented here (Figs. 4, 7, 10,

698 13) offer the highest potential for future observational studies interested in quantifying the impact of climate change on

699 denudation and weathering rates.

700

701

Figure Captions

Figure 1 Topography for regions (a) tropical South Asia, (b) temperate South Asia, (c) high altitude South Asia, (d) temperate South America, (e) tropical South America, (f) temperate Andes, (g) tropical Andes, SE Alaska and Casca-dia.

Figure 2 Global PI annual mean near-surface temperatures (a), and deviations of MH, LGM and PLIO annual mean near-surface temperatures from PI values (b). Units are °C and insignificant (p < 99%) differences (as determined by a t-test) are greyed out.

Figure 3 Global PI annual mean precipitation (a), and deviations of MH, LGM and PLIO annual mean near-surface temperatures from PI values (b). Units are mm/yr.

Figure 4 PI annual mean near-surface temperatures (a), and deviations of MH, LGM and PLIO annual mean near-sur-face temperatures from PI values (b) for the South Asia region. Insignificant (p < 99%) differences (as determined by a t-test) are greyed out.

Figure 5 PI, MH, LGM and PLIO annual mean precipitation in (a) tropical South Asia, (b) temperate South Asia, and (c) high-altitude South Asia; PI, MH, LGM and PLIO annual mean temperatures in (d) tropical South Asia, (e) tem-

730   perate South Asia, and (f) high-altitude South Asia. For each time slice, the minimum, lower 25th percentile, median,

731   upper 75th percentile and maximum are plotted.

732   **Figure 6** Geographical coverage and characterization of climate classes C_1- C_6 based on cluster-analysis of 8 variables

733   (near surface temperature, seasonal near surface temperature amplitude, total precipitation, seasonal precipitation

734   amplitude, u-wind in January and July, v-wind in January and July) in the South Asia region. The geographical cov-

735   erage of the climates C_1- C_6 is shown on the left for PI (a), MH (b), LGM (c) and PLIO (d); the complementary,

736   time-slice specific characterization of C_1- C_6 for PI (e), MH (f), LGM (g) and PLIO (h) is shown on the right.

737   **Figure 7** PI annual mean near-surface temperatures (a), and deviations of MH, LGM and PLIO annual mean near-sur-

738   face temperatures from PI values (b) for western South America. Insignificant (p < 99%) differences (as determined

739   by a t-test) are greyed out.

740   **Figure 8** PI, MH, LGM and PLIO annual mean precipitation in (a) tropical South America, (b) temperate South Amer-

741   ica, (c) tropical Andes, and (d) temperate Andes; PI, MH, LGM and PLIO annual mean temperatures in (e) tropical

742   South America, (f) temperate South America, (g) tropical Andes, and (h) temperate Andes. For each time slice, the

743   minimum, lower 25th percentile, median, upper 75th percentile and maximum are plotted.

744   **Figure 9** Geographical coverage and characterization of climate classes C_1- C_6 based on cluster-analysis of 8 variables

745   (near surface temperature, seasonal near surface temperature amplitude, precipitation, seasonal precipitation amp-

746   litude, u-wind in January and July, v-wind in January and July) in western South America. The geographical cover-

747   age of the climates C_1- C_6 is shown on the left for PI (a), MH (b), LGM (c) and PLIO (d); the complementary, time-

748   slice specific characterization of C_1- C_6 for PI (e), MH (f), LGM (g) and PLIO (h) is shown on the right.

749   **Figure 10** PI annual mean near-surface temperatures (a), and deviations of MH, LGM and PLIO annual mean near-sur-

750   face temperatures from PI values (b) for the South Alaska region. Insignificant (p < 99%) differences (as determined

751   by a t-test) are greyed out.

752   **Figure 11** PI, MH, LGM and PLIO annual mean precipitation (a), and mean annual temperatures (b) in South Alaska.

753   For each time slice, the minimum, lower 25th percentile, median, upper 75th percentile and maximum are plotted.

754   **Figure 12** Geographical coverage of climate classes C_1- C_4 based on cluster-analysis of 4 variables (near surface tem-

755   perature, seasonal near surface temperature amplitude, total precipitation, seasonal total precipitation amplitude) in

756   southern Alaska. The geographical coverage of the climates C_1- C_4 is shown on the left for PI (a,), MH (b), LGM (c)

757   and PLIO (d); the complementary, time-slice specific characterization of C_1- C_6 for PI (e), MH (f), LGM (g) and

758   PLIO (h) is shown on the right.

759   **Figure 13** PI annual mean near-surface temperatures (a), and deviations of MH, LGM and PLIO annual mean near-sur-

760   face temperatures from PI values (b) for the US Pacific Northwest. Insignificant (p < 99%) differences (as determ-

761   ined by a t-test) are greyed out.

762   **Figure 14** PI, MH, LGM and PLIO annual mean precipitation (a), and annual mean temperatures (b) in the Cascades,

763   US Pacific Northwest. For each time slice, the minimum, lower 25th percentile, median, upper 75th percentile and

764   maximum are plotted.

765   **Figure 15** Geographical coverage and characterization of climate classes C_1- C_4 based on cluster-analysis of 4 variables

766   (near surface temperature, seasonal near surface temperature amplitude, total precipitation, seasonal total precipita-

767   tion amplitude) in the Cascades, US Pacific Northwest. The geographical coverage of the climates C_1- C_4 is shown

768   on the left for PI (a), MH (b), LGM (c) and PLIO (d); the complementary, time-slice specific characterization of C_1-

769   C_6 for PI (e), MH (f), LGM (g) and PLIO (h) is shown on the right.

770   **Figure 16** Simulated annual mean precipitation deviations of MH (left) and LGM (right) from PI values in South Asia,

771   and temporally corresponding proxy-based reconstructions, indicating wetter (upward facing blue triangles), drier

772   (downward facing red triangles) or similar (grey circles) conditions in comparison with modern climate. MH proxy-

773   based precipitation differences are taken from Mügler et al. (2010) (**66**), Wischnewski et al. (2011) (**67**), Mischke et

774   al. (2008), Wischnewski et al. (2011), Herzschuh et al. (2009) (**68**), Yanhong et al. (2006) (**69**), Morrill et al. (2006)

775   (**70**), Wang et al. (2002) (**71**), Wuennemann et al. (2006) (**72**), Zhang et al. (2011), Morinaga et al. (1993),

776   Kashiwaya et al. (1995) (**73**), Shen et al. (2005) (**74**), Liu et al. (2014) (**75**), Herzschuh et al. (2006) (**76**), Zhang and

777   Mischke (2009) (**77**), Nishimura et al. (2014) (**78**), Yu and Lai (2014) (**79**), Gasse et al. (1991) (**80**), Van Campo et

778   al. (1996) (**81**), Demske et al. (2009) (**82**), Kramer et al. (2010) (**83**), Herzschuh et al. (2006) (**84**), Hodell et al.

779   (1999)(**85**), Hodell et al. (1999) (**86**), Shen et al. (2006) (**87**), Tang et al. (2000) (**88**), Tang et al. (2000) (**89**), Zhou et

780   al. (2002) (**90**), Liu et al. (1998) (**91**), Asashi (2010)(**92**), Kotila et al. (2009) (**93**), Kotila et al. (2000) (**94**), Wang et

781   al. (2002) (**95**), Hu et al. (2014) (**96**), Hodell et al. (1999) (**97**), Hodell et al. (1999) (**98**).

782   **Figure 17** Simulated annual mean precipitation deviations of MH (left) and LGM (right) from PI values in South Amer-

783   ica, and temporally corresponding proxy-based reconstructions, indicating wetter (upward facing blue triangles),

784   drier (downward facing red triangles) or similar (grey circles) conditions in comparison with modern climate. MH

785   proxy-based precipitation differences are taken from Bird et al. (2011) (**1**), Hansen et al (1994) (**2**), Hansen et al

786   (1994) (**3**), Hansen et al (1994) (**4**), Hansen et al (1994) (**5**), Hansen et al (1994) (**6**), Hillyer et al. (2009) (**7**),

787   D'Agostino et al. (2002) (**8**), Baker et al. (2001) (**9**), Schwalb et al (1999) (**10**), Schwalb et al (1999) (**11**), Schwalb

788   et al (1999) (**12**), Schwalb et al (1999) (**13**), Moreno et al (2009) (**14**), Pueyo et al (2011) (**15**), Mujica et al (2015)

789   (**16**), Fritz et al. (2004) (**17**), Gayo et al. (2012) (**18**), Latorre et al. (2006) (**19**), Latorre et al. (2003) (**20**), Quade et al

790   (2008) (**21**), Bobst et al. (2001) (**22**), Grosjean et al. (2001) (**23**), Betancourt et al. (2000) (**24**), Latorre et al. (2002)

791   (**25**), Rech et al. (2003) (**26**), Diaz et al. (2012) (**27**), Maldonado et al (2005) (**28**), Diaz et al. (2012) (**29**), Lamy et

792   al. (2000) (**30**), Kaiser et al. (2008) (**31**), Maldonado et al. (2010) (**32**), Villagrán et al. (1990) (**33**), Méndez et al.

793   (2015) (**34**), Maldonado et al. (2006) (**35**), Lamy et al. (1999) (**36**), Jenny et al. (2002) (**37**), Jenny et al. (2002b)

794   (**38**), Villa-Martínez et al. (2003) (**39**), Bertrand et al. (2008) (**40**), De Basti et al. (2008) (**41**), Lamy et al. (2009)

795   (**42**), Lamy et al. (2002) (**43**), Szeicz et al. (2003) (**44**), de Porras et al. (2012) (**45**), de Porras et al. (2014) (**46**),

796   Markgraf et al. (2007) (**47**), Siani et al. (2010) (**48**), Gilli et al. (2001) (**49**), Markgraf et al. (2003) (**50**), Stine et al.

797   (1990) (**51**).

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800   **References**

801

802   Abe-Ouchi, A., Saito, F., Kageyama, M., Bracannot, P., Harrison, S. P., Lambeck, K., Otto-Bliesner, B. L., Peltier, W.R.,

803   Tarasov, L., Peterschmitt, J.-Y., Takahashi, K.: Ice-sheet configuration in the CMUP5/PMIP3 Last Glacial Maximum

804   experiments. *Geosci. Model Dev.*, 8, 3621-3637, doi:10.5194/gmd-8-3621-2015, 2015

805   Andersen, J., Egholm, D.L., Knudsen, M.F., Jansen, J., Nielsen, S.B., The periglacial engine of mountain erosion – Part

806   1: Rates of frost cracking and frost creep., *Earth Surface Dynamics,* 3(4), 447-462, 2015.

807   Arnold, L., Breon, F.M., and Brewer, S: The earth as an extrasolar planet: the vegetation spectral signature today and

808   during the last Quaternary climatic extrema, *Int. J. Astrobiol*., 8, 81–94.

809   http://dx.doi.org/10.1017/S1473550409004406, 2009.

810   Asahi, K.: Equilibrium-line altitudes of the present and Last Glacial Maximum in the eastern Nepal Himalayas and

811   their implications for SW monsoon climate, Quatern. Int., 212, 26-34, doi:10.1016/j.quaint.2008.08.004, 2010.

812   Bahrenberg , G., Giese, E., and Nipper, J.: Multivariate Statistik. *Statistische Methoden in der Geographie 2*, Stuttgart,

813   1992.

814   Baker, P. A., Seltzer, G. O., Fritz, S. C., Dunbar, R. B., Grove, M. J., Tapia, P. M., Cross L., Rowe H. D., and Baroda, J.

815   P.: The History of South America Tropical Precipitation for the Past 25,000 Years, Earth and Atmospheric Sciences,

816   291, 640-643, doi:10.1126/science.291.5504.640, 2001.

817   Ballantyne, A.P., Greenwood, D.R., Sinninghe Damste, J.S., Csank, A.Z., Eberle, J.J., and Rybczynski, N.: Significantly
818   warmer Arctic surface temperatures during the Pliocene indicated by multiple independent proxies, *Geology* 38 (7),
819   603–606, 2010.

820   Banfield, J.F., Barker, W.W., Welch, S.A., and Taunton, A.: Biological impact on mineral dissolution: application of the
821   lichen model to understanding mineral weathering in the rhizophere, *Proceedings of the National Academy of*
822   *Sciences*, 96, 3404-3411, 1999.

823   Barnston, A. G. and Livezey, R. E.: Classification, Seasonality and Persistence of Low-Frequency Atmospheric
824   Circulation Patterns, Mon. Weather Rev., 115, 1083-1126, doi:10.1175/1520-
825   0493(1987)1151083:CSAPOL2.0.CO;2, 1987.

826   Barron, J.A., and Anderson, L.: Enhanced Late Holocene ENSO/PDO expression along the margins of the eastern North
827   Pacific. *Quaternary International*, 235, 3-12, 2011.

828   Bendick, R., and Ehlers, T.A.: Extreme localized exhumation at syntaxes initiated by subduction geometry, *Geophys.*
829   *Res. Lett.*, *41*(16), 2014GL061026, doi:10.1002/2014GL061026, 2014.

830   Bertrand, S., Charlet F., Charlier B., Renson V., and Fagel N.: Climate variability of southern Chile since the Last
831   Glacial Maximum: a continuous sedimentlogical records from Lago Puyehue (40°S), J. Paleolimnol., 39, 179-195,
832   doi:10.1007/s10933-007-9117-y, 2008.

833   Betancourt, J. L., Latorre C., Rech J. A., Quade J., and Rylander K,A.: A 22,000 Year Record of Monsoonal Precipita-
834   tion from Northern Chile's Atacama Desert, Science, 289, 1542-1546, doi: 10.1126/science.289.5484.1542, 2000.

835   Bigelow, N. H., Brubaker, L. B., Edwards, M. E., Harrison, S. P., Prentice, I. C., Anderson, P. M., Andreev, A. A.,
836   Bartlein, P. J., Christensen, T. R., Cramer, W., Kaplan, J. O., Lozhkin, A. V., Matveyeva, N. V., Murray, D. V.,
837   McGuire, A. D., Razzhivin, V. Y., Ritchie, J. C., Smith, B., Walker, D. A., Gajewski, K., Wolf, V., Holmqvist, B. H.,
838   Igarashi, Y., Kremenetskii, K., Paus, A., Pisaric, M. F. J., and Vokova, V. S.: Climate change and Arctic ecosystems
839   I. Vegetation changes north of 55°N between the last glacial maximum, mid-Holocene and present. Journal of Geo-
840   physical Research - Atmospheres, 108(D19), doi: 10.1029/2002JD002558, 2003.

841   Bird, B. W., Abbott M .B., Rodbell D. T,. and Vuille M.: Holocene tropical South American hydroclimate revealed from
842   a decadally resolved lake sediment δ18O record, Earth Planet. Sc. Lett., 310, 192-202,
843   doi:10.1016/j.epsl.2011.08.040, 2011.

844   Bobst, A. L., Lowenstein, T. K., Jordan, T. E., Godfrey, L. V., Ku, T.-L., and Luo, S.: A 106 ka paleoclimate record from
845   drill core of the Salar de Atacama, northern Chile, Palaeogeogr. Palaeocl., 173, 21-42, doi:10.1016/S0031-
846   0182(01)00308, 2001.

847
848   Braconnot, P., Otto-Bleisner, B., Harrison, S.P., Joussaume, S., Peterschmitt, J.-Y., Abe-Ouchi, A., Crucifix, M.,
849   Driesschaert, E., Fichefet, T., Hewitt, C.D., Kagayama, M., Kitoh, A., Loutre, M.-F., Marti, O., Merkel, U.,
850   Ramstein, G., Valdes, P., Weber, L., Yu, Y., and Zhao, Y.: Results of PMIP2 coupled simulations of the mid-
851   Holocene and Last Glacial maximum, part 1: experiments and large-scale features, Clim Past 3:261–277, 2007.

852 Bohner, J.: General climatic controls and topoclimatic variations in Central and High Asia, *Boreas*, 35(2), 279-295,
853   2006.

854 Bookhagen, B., Thiede, R.C., and Strecker, M.R.: Late Quaternary intensified monsoon phases control landscape
855   evolution in the northwest Himalaya, *Geology*, 33(2), 149-152. doi: 10.1130/G20982.1, 2005.

856 Boos, W. R., and Kuang, Z.: Dominant control of the South Asian monsoon by orographic insulation versus plateau
857   heating, *Nature*, 463, 218-222, 2010.

858 Braun, J.: A simple model for regolith formation by chemical weathering: Regolith Formation, *Journal of Geophysical*
859   *Research Earth's Surface,* DOI: 10.1002/2016JF003914, 2016.

860 Caves, J.: Late Miocene Uplift of the Tian Shan and Altai and Reorganization of Central Asia Climate, GSA Today,
861   doi:10.1130/gsatg305a.1, 2017.

862 CLIMAP Project Members: Seasonal Reconstruction of the Earth's Surface at the Last Glacial Maximum.
863 *Map and Chart Series*, Vol. 36, Geological Society of America, 18 pp., 1981.
864

865 Cohen, S.J.: Bringing the global warming issue close to home: The challenge of regional impact studies. *Bulletin of the*
866   *American Meteorological Society,* 71: 520 – 526., 1990.

867 D'Agostino, K ., Seltzer, G., Baker, P., Fritz, S., and Dunbar, R.: Late-Quaternary lowstands of Lake Titicaca: evidence
868   from high-resolution seismic data, Palaeogeogr. Palaeocl., 179, 97-111, doi:10.1016/S0031-0182(01)00411-4, 2002.

869 De Batist, M., Fagel N., Loutre M. F., and Chapron E.: A 17,900-year multi-proxy lacustrine record of Lago Puyehue
870   (Chilean Lake District): introduction, J. Paleolimnol., 39, 151-161, doi:10.1007/s10933-007-9113-2, 2007.

871 de Porras, M. E., Maldonado, A., Abarzua, A. M., Cardenas, M. L., Francois, J. P., Martel-Cea, A., Stern, C. R.,
872   Mendez, C., and Reyes, O.: Postglacial vegetation, fire and climate dynamics at Central Chilean Patagonia (Lake
873   Shaman, 44°S), Quaternary Sci. Rev., 50, 71-85, doi:10.1016/j.quascirev.2012.06.015; 2012.

874 de Porras, M. E., Maldonado, A., Quintana, F. A., Martel-Cea, A., Reyes, O., and Méndez, C.: Environmental and cli-
875   matic changes in central Chilean Patagonia since the Late Glacial (Mallín El Embudo, 44° S), Clim. Past,10, 1063–
876   1078, doi:10.5194/cp-10-1063-2014, 2014.

877 Delunel, R., van der Beek, P. A., Carcaillet, J., Bourlès, D. L., and Valla, P.G.: Frost-cracking control on catchment de-
878   nudation rates: Insights from in situ produced 10Be concentrations in stream sediments (Ecrins–Pelvoux massif,
879   French Western Alps), *Earth and Planetary Science Letters*, *293*(1–2), 72–83, doi:10.1016/j.epsl.2010.02.020, 2010.

880 DeMenocal, P., Ortiz, J., Guilderson, T., Adkins, J., Samthein, M., Baker, L., and Yarusinsky, M.: Abrupt onset and ter-
881   mination of the African Humid Period:: rapid climate responses to gradual insolation forcing, Quaternary Sci. Rev.,
882   19, 347-361, doi:10.1016/S0277-3791(99)00081-5, 2000.

883 Demske, D., Tarasov, P. E., Wünnemann, B., and Riedel, F.: Late glacial and Holocene vegetation, Indian monsoon and
884   westerly circulation in the Trans-Himalaya recorded in the lacustrine pollen sequence from Tso Kar, Ladakh, NW
885   India, Palaeogeogr. Palaeocl., 279, 172-185, doi:10.1016/j.palaeo.2009.05.008, 2009.

886 Diaz, F.P., Latorre, C., Maldonado, A., Quade, J., and Betancourt, J.L.: Rodent middens reveal episodic, long-distance

887   plant colonizations across the hyperarid Atacama Desert over the last 34,000 years, J. Biogeogr., 39, 510-525,
888   doi:10.1111/j.1365-2699.2011.02617.x, 2012.

889   Dietrich, W. E., and Perron, J.T.: The search for a topographic signature of life, *Nature*, *439*(7075), 411–418,
890   doi:10.1038/nature04452, 2006.

891   Dietrich, S., Werner, M., Spangehl, T., and Lohmann, G.: Influence of orbital forcing and solar activity on water iso-
892   topes in precipitation during the mid and late Holocene, *Clim. Past*, 9, 13-26. doi:10.5194/cp-9-13-2013, 2013.
893

894   Dettman, D.L., Fang, X.M., Garzione, C.N., and Li, J.J.: Uplift-driven climate change at 12 Ma: a long delta O-18 re-
895   cord from the NE margin of the Tibetan plateau. *Earth and Planetary Science Letters*, 214(1-2), 267-277, 2003.

896   Dowsett, H.J., Robinson, M., Haywood, A., Salzmann, U., Hill, D., Sohl, L., Chandler, M., Williams, M., Foley, K.,
897   and Stoll, D.: The PRISM3D paleoenvironmental reconstruction. *Stratigraphy*, 7, 123–139, 2010.

898   Egholm, D.L., Nielsen, S.B., Pedersen, V.K., Lesemann, J.,: Glacial efects limiting moutain height, *Nature,* 460(7257),
899   884-887, 2009.

900   Ehlers, T.A., and Poulsen, C.J.: Influence of Andean uplift on climate and paleoaltimetry estimates, *Earth and Planet-*
901   *ary Science Letters*, 281(3-4), 238-248, 2009.

902   Etheridge, D.M., Steele, L., Langenfelds, R., Francey, R., Barnola, J., and Morgan, V.:, Natural and anthropogenic
903   changes in atmospheric CO_2 over the last 1000 years from air in Antarctic ice and firn, *J Geophys Res* 101:4115–
904   4128, 1996.
905

906   Etheridge, D.M., L. Steele, R. Francey, and R. Langenfelds (1998), Atmospheric methane between 1000 a.d. and
907   present: evidence of anthropogenic emissions and climatic variability. *J Geophys Res,* 103:15979–15993
908

909   Feng, R., Poulsen, C.J., Werner, M., Chamberlain, C.P., Mix, H.T., and Mulch, A.: Early Cenozoic evolution of topo-
910   graphy, climate, and stable isotopes in precipitation in the North American Cordillera, *American Journal of Science*,
911   *313*(7), 613–648, 2013.

912   Feng, R., Poulsen, C.J., Werner, M., 2016. Tropical circulation intensification and tectonic extension recorded
913   by Neogene terrestrial d18O records of the western United States. Geology 44. doi:10.1130/G38212.1
914
915   Feng, R., Poulsen, C.J., 2016. Refinement of Eocene lapse rates, fossil-leaf altimetry, and North American
916   Cordilleran surface elevation estimates. Earth Planet. Sci. Lett. doi:10.1016/j.epsl.2015.12.022
917
918   Fritz, S. C., Baker, P. A., Lowenstein, T. K., Seltzer, G. O., Rigsby, C. A., Dwyer, G. S., Tapia, P. M., Arnold, K.
919   K., Ku, T. L., and Luo, S: Hydrologic variation during the last 170,000 years in the southern hemisphere
920   tropics of South America, Quaternary Res., 61, 95 – 104, doi:10.1016/j.yqres.2003.08.007, 2004.Gasse, F.,
921   Arnold, M., Fontes, J. C., Fort, M., Gibert, E., Huc, A., Li, B., Li, Y., Liu, Q., Melleres, F., Van Campo, E., Wang, F.,
922   and Zhang, Q.: A 13,000-year climate record from western Tibet, Nature, 353, 742-745,
923   doi:10.1016/j.quaint.2006.02.001, 1991.

924   Gayo E. M., Latorre, C., Santoro, C. M., Maldonado, A., and De Pol-Holz, R.: Hydroclimate variability in the low-elev-
925   ation Atacama Desert over the last 2500 yr, Clim. Past, 8, 287-306, doi:10.5194/cp-8-287-2012, 2012. Gierz, P.,
926   Lohmann, G., and Wei, W.: Response of Atlantic Overturning to future warming in a coupled atmosphere-ocean-ice

927   sheet model, *Geophysical Research Letters*, 42, 6811-6818, doi:10.1002/2015GL065276, 2015.

928   Gilli, A., Ariztegui, D., Bradbury, J. P., Kelts, K. R., Markgraf, V., and McKenzie, J. A.: Tracking abrupt climate change
929   in the Southern Hemisphere: a seismic stratigraphic study of Lago Cardiel, Argentina (49°S), Terra Nova, 13, 443-
930   448, doi:10.1046/j.1365-3121.2001.00377.x , 2001.

931   Glotzbach, C., van der Beek, P., Carcaillet, J., and Delunel, R.: Deciphering the driving forces of erosion rates on mil-
932   lennial to million−year timescales in glacially impacted landscapes: An example from the Western Alps, *Journal of*
933   *Geophysical Research: Earth Surface,* 118, 1491-1515, 2013.

934   Gong, X., Knorr, G., Lohmann, G., and Zhang, X.: Dependence of abrupt Atlantic meridional ocean circulation changes
935   on climate background states, *Geophysical Research Letters*, 40 (14), 3698-3704, doi:10.1002/grl.50701, 2013.

936   Grosjean, M., Van Leeuwen, J., Van der Knaap, W., Geyh, M., Ammann, B., Tanner, W., Messerli, B., Núñez, L.,
937   Valero-Garcés, B., and Veit, H. : A 22,000 14C year BP sediment and pollen record of climate change from Laguna
938   Miscanti (23°S), northern Chile, Global Planet. Change, 28, 35–51, doi:10.1016/S0921-8181(00)00063-1, 2001.

939   Gyssels, G., Poesen, J., Bochet, E., and Li, Y.: Impact of plant roots on the resistance of soils to erosion by water: a re-
940   view, *Progress in Physical Geography*, 29, 189–217. http://doi.org/10.1191/0309133305pp443ra, 2005.

941   Hansen, B. C. S., Seltzer, G. O., and Wright Jr., H.E.: Late Quaternary vegetational change in the central Peruvian
942   Andes, Palaeogeogr. Palaeocl., 109, 263-285, doi:10.1016/0031-0182(94)90179-1, 1994.

943   Harris, I., Jones, P.D., Osborn, T.J., and Lister, D.H.: Updated high-resolution grids of monthly climatic observations -
944   the CRU TS3.10 Dataset, *International Journal of Climatology*, doi:10.1002/joc.3711, 2013.

945   Harrison, S. P., Yu, G., Takahara, H., and Prentice, I. C.: Palaeovegetation - Diversity of temperate plants in east Asia.
946   Nature 413, 129-130, 2001.
947

948   Harrison, S.P., Bartlein, P.J., Brewer, S., Prentice, I.C., Boyd, M., Hessler, I., Holmgren, K., Izumi, K., and Willis, K.:
949   Climate model benchmarking with glacial and mid-Holocene climates, Climate Dynamics, 43, 671-688. doi
950   10.1007/s00382-013-1922-6, 2013.

951   Haywood, A.M., Valdes, P.J., and Sellwood, B.W.: Global scale palaeoclimate reconstruction of the middle Pliocene cli-
952   mate using the UKMO GCM: initial results. *Global and Planetary Change,* 25 (3–4), 239–256, 2000.

953   Haywood, A.M., Dowsett, H.J., Otto-Bliesner, B., Chandler, M.A., Dolan, A.M., Hill, D.J., Lunt, D.J., Robinson, M.M.,
954   Rosenbloom, N., Salzmann, U., and Sohl, L.E.: Pliocene Model Intercomparison Project (PlioMIP): experimental
955   design and boundary conditions (Experiment 1), *Geoscientific Model Development* (3), 227-242, 2010.
956

957   Haywood, A.M., Hill, D.J., Dolan, A.M., Otto-Bliesner, B., Bragg, F., Chan, W.-L., Chandler, M.A., Contoux, C., Jost,
958   A., Kamae, Y., Lohmann, G., Lunt, D.J., Abe-Ouchi, A., Pickering, S.J., Ramstein, G., Rosenbloom, N.A., Sohl, L.,
959   Stepanek, C., Yan, Q., Ueda, H., and Zhang, Z.: Large-scale features of Pliocene climate: results from the Pliocene
960   Model Intercomparison Project, *Clim. Past*, 9, 191-209. doi:10.5194/cp-9-191-2013, 2013.
961

962 Herman, F., Seward, D., Valla, P.G., Carter, A., Kohn, B., Willett, S.D., and Ehlers, T.A.: Worldwide acceleration of
963   mountain erosion under a cooling climate, *Nature,* 504, 423–426. doi:10.1038/nature12877, 2013.

964 Herzschuh, U., Kuerschner, H., and Mischke, S.: Temperature variability and vertical vegetation belt shifts during the
965   last ~50,000 yr in the Qilian Mountains (NE margin of the Tibetan Plateau, China), Quaternary Res., 66, 133-146,
966   doi:10.1016/j.yqres.2006.03.001, 2006a.

967 Herzschuh, U., Winter, K., Wuennemann, B., and Li, S.: A general cooling trend on the central Tibetan Plateau
968   throughout the Holocene recorded by the lake Zigetang pollen spectra, Quatern. Int., 154, 113-121,
969   doi:10.1016/j.quaint.2006.02.005, 2006b.
970
971 Herzschuh, U., Kramer A., Mischke S., and Zhang C.: Quantitative climate and vegetation trends since the late glacial
972   on the northeastern Tibetan Plateau deduced from Koucha lake pollen spectra, Quaternary Res., 71, 162-171,
973   doi:10.1016/j.yqres.2008.09.003, 2009.

974 Hillenbrand, C.-D., and Fütterer, D.K.: Neogene to Quaternary deposition of opal on the continental rise west of the
975   Antarctic Peninsula, ODP Leg 178, Sites 1095, 1096, and 1101. In: Barker, P.F., Camerlenghi, A., Acton, G.D.,
976   Ramsay, A.T.S. (Eds.), Proceedings of the Ocean Drilling Programme, Scientific Results, 178. Texas A and M Uni-
977   versity, College Station, Texas, pp. 1–40 (CD-ROM), 2002.

978 Hillyer, R., Valencia, B. G., Bush, M.B., Silman, M.R., and Steinitz-Kannan, M.: A 24,700-yr paleolimnological history
979   from the Peruvian Andes, Quaternary Res., 71, 71-82, doi:10.1016/j.yqres.2008.06.006, 2009.

980 Hobley, D.E., Sinclair, H.D., and Cowie, P.A.: Processes, rates, and time scales of fluvial response in an ancient postgla-
981   cial landscape of the northwest Indian Himalaya. *Geological Society of America Bulletin,* 122, 1569-1584, 2010.

982 Hodell, D. A., Brenner, M., Kanfoush, S. L., Curtis, J. H., Stoner, J. S., Song, X., Wu, Y., and Whitmore, T. J.: Paleocli-
983   mate of southwestern China for the past 50,000 yr inferred from lake sediment records, Quaternary Res., 52, 369-
984   380, doi:10.1006/ qres.1999.2072, 1999.

985 Hu, G., Yi, C.-L., Zhang, J.-F., Liu, J.-L., Jiang, T., and Qin, X.: Optically stimulated luminescence dating of a moraine
986   and a terrace in Laohugou valley, western Qilian Shan, northeastern Tibet, Quaternary International 321, 37-49,
987   doi:10.1016/j.quaint.2013.12.019, 2014.

988 Insel, N., Poulsen, C.J., and Ehlers, T.A.: Influence of the Andes Mountains on South American moisture transport, con-
989   vection, and precipitation, *Climate Dynamics*, 35 (7-8), 1477-1492, 2010.

990 Jeffery, M.L., Ehlers, T.A., Yanites, B.J., and Poulsen, C.J.: Quantifying the role of paleoclimate and Andean Plateau
991   uplift on river incision: PALEOCLIMATE ROLE IN RIVER INCISION, *Journal of Geophysical Research: Earth*
992   *Surface*, *118*(2), 852–871, doi:10.1002/jgrf.20055, 2013.

993 Jenny, B., Valero-Garces, B. L., Urrutia, R., Kelts, K., Veit, H., and Geyh, M.: Moisture changes and fluctuations of the
994   Westerlies in Mediterranean Central Chile durin the last 2000 years: The Laguna Aculeo record (33°50' S), Quatern.
995   Int., 87, 3-18, doi:10.1016/S1040-6182(01)00058-1, 2002a.

996 Jenny, B., Valero-Garcés, B.L., Villa-Martínez, R., Urrutia, R., Geyh, M. A., and Veit, H.: Early to Mid-Holocene Arid-
997   ity in Central Chile and the Southern Westerlies: The Laguna Aculeo Record (34°S), Quaternary Res., 58, 160–170,
998   doi:10.1006/qres.2002.2370, 2002b.

999   Jungclaus, J. H., Lorenz, S. J., Timmreck, C., Reick, C. H., Brovkin, V., Six, K., Segschneider, J., Giorgetta, M.A.,

1000   Crowley, T.J., Pongratz, J., Krivova, N.A., Vieira, L.E., Solanski, S.K., Klocke, D., Botzet, M., Esch, M., Gayler, V.,

1001   Haak, H., Raddatz, T.J., Roeckner, E., Schnur, R., Widmann, H., Claussen, M., Stevens, B., and Marotzke, J.:

1002   Climate and carbon-cycle variability over the last millennium. *Climate of the Past, 6*, 723-737. doi:10.5194/cp-6-723-

1003   2010, 2010.

1004   Junginger, A., Roller, S., Olaka, L. A., and Trauth, M. H.:The effects of solar irradiation changes on the migration of the

1005   Congo Air Boundary and water levels of paleo-Lake Suguta, Northern Kenya Rift, during the African Humid Period

1006   (15–5 ka BP), Palaeogeogr. Palaeocl., 396, 1-16, doi:10.1016/j.palaeo.2013.12.007, 2014.

1007   Kaiser, J., Schefuss, E., Lamy, F., Mohtadi, M., and Hebbeln, D.: Glacial to Holocene changes in sea surface temperat-

1008   ure and coastal vegetation in north central Chile : high *versus* low latitude forcing, Quaternary Sci. Rev., 27, 2064–

1009   2075, doi:10.1016/j.quascirev.2008.08.025, 2008.

1010   Kalnay, E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D., Gandin, L., Iredell, M., Saha, S., White, G., Woollen, J.,

1011   Zhu, Y., Chelliah, M., Ebisuzaki, W., Higgins, W., Janowiak, J., Mo, K.C., Ropelewski, C., Wang, J., Leetmaa, A.,

1012   Reynolds, R., Jenne, R., and Joseph, D.: The NCEP/NCAR 40-year reanalysis project. *Bulletin of the American*

1013   *Meteorological Society*, 77(3), 437-471, 1996.

1014   Kashiwaya, K., Masuzawa, T., Morinaga, H., Yaskawa, K., Yuan, B. Y., Liu, J. Q., and Gu, Z.: Changes in hydrological

1015   conditions in the central Qing-Zang (Tibetan) Plateau inferred from lake bottom sediments, Earth Planet. Sc. Lett.,

1016   135, 31-39, doi:10.1016/0012-821X(95)00136-Z, 1995.

1017   Kent-Corson, M., Sherman, L., Mulch, A. and Chamberlain, C.: Cenozoic topographic and climatic response to chan-

1018   ging tectonic boundary conditions in Western North America, Earth Planet. Sc. Lett., 252(3-4), 453-466,

1019   doi:10.1016/j.epsl.2006.09.049, 2006.

1020   Kistler, R., Collins, W., Saha, S., White, G., Woollen, J., Kalnay, E., Chelliah, M., Ebisuzaki, W., Kanamitsu, M.,

1021   Kousky, V., Van den Dool, H., Jenne, R., and Fiorino, M.: The NCEP−NCAR 50−Year Reanalysis: Monthly Means

1022   CD−ROM and Documentation. *Bulletin of the American Meteorological Society*, 82(2), 247–267, 2001.

1023   Kirchner, J.W., Finkel, R.C., Riebe, C.S., Granger, D.E., Clayton, J.L., King, J.G., and Megahan, W.F.: Mountain

1024   erosion over 10 yr, 10 k.y., and 10 m.y. time scales, *Geology,* 29 (7), 591-594, 2001.

1025   Knorr, G., Butzin, M., Micheels, A., and Lohmann, G.: A Warm Miocene Climate at Low Atmospheric CO2 levels.

1026   Geophysical Research Letters, 38, L20701, doi:10.1029/2011GL048873, 2011.

1027   Koons, P.O., Zeitler, P.K., and Hallet, B.: 5.14 Tectonic Aneurysms and Mountain Building, in *Treatise on Geomorpho-*

1028   *logy*, pp. 318–349, Elsevier, 2013.

1029   Kotlia, B. S., Sharma, C., Bhalla, M. S., Rajagopalan, G.,. Subrahmanyam, K., Bhattacharya, A., and Valdiya, K. S.: Pa-

1030   leoclimatic conditions in the late Pleistocene Wadda Lake, eastern Kumaun Himalaya (India), Palaeogeogr. Pa-

1031   laeocl., 162, 105-118, doi:10.1016/S0031-0182(00)00107-3, 2000.

1032   Kramer, A., Herzschuh, U., Mischke, S., and Zhang, C.: Late glacial vegetation and climate oscillations on the south-

1033   eastern Tibetan Plateau inferred from the lake Naleng pollen profile, Quaternary Res., 73, 324-335,

1034   doi:10.1016/j.yqres.2009.12.003, 2010.

1035   Kraus, H.: Die Atmosphäre der Erde. Eine Einführung in die Meteorologie. Berlin, 2001.

1036   Kutzbach, J.E., Guetter, P.J., Ruddiman, W.F., and Prell, W.L.: Sensitivity of Climate to Late Cenozoic Uplift in
1037   Southern Asia and the American West - Numerical Experiments, *Journal of Geophysical Research-Atmospheres*,
1038   94(D15), 18393-18407, 1989.

1039   Kutzbach, J.E., Prell, W.L., and Ruddiman, W.F.: Sensitivity of Eurasian Climate to Surface Uplift of the Tibetan Plat-
1040   eau, *Journal of Geology*, 101(2), 177-190, 1993.

1041   Lamy, F. and Kaiser. J.: Past Climate Variability in South America and Surrounding Regions, Chapter 6: Glacial to
1042   Holocene Paleoceanographic and Continental Paleoclimate Reconstructions Based on ODP Site 1233/GeoB 3313
1043   Off southern Chile, 2009.

1044   Lamy, F., Hebbeln, D., and Wefer, G.: High Resolution Marine Record of Climatic Change in Mid-latitude Chile during
1045   the Last 28,000 Years Based on Terrigenous Sediment Parameters, Quaternary Res., 51, 83-93,
1046   doi:10.1006/qres.1998.2010, 1999.

1047   Lamy, F., Klump, J., Hebbeln, D., and Wefer, G.: Late Quaternary rapid climate change in northern Chile, Terra Nova,
1048   12, 8-13, doi:10.1046/j.1365-3121.2000.00265.x, 2000.

1049   Lamy, F., Rühlemann, C., Hebbeln, D. and Wefer, G.: High- and low-latitude climate control on the position of the
1050   southern Peru-Cihle Current during the Holocene, Paleoceanography, 17, 16-1–16-10, doi:10.1029/2001PA000727,
1051   2002.

1052   Latorre, C., Betancourt, J., Rylander, K., and Quade, J.: Vegetation invasions into absolute desert: A 45 000 yr rodent
1053   midden record from the Calama-Salar de Atacama basins, northern Chile (lat 22° - 24°S), Geol. Soc. Am. Bull., 114,
1054   349–366, doi:10.1130/0016-7606(2002)114<0349:VIIADA>2.0.CO;2, 2002.

1055   Latorre, C., Betancourt, J.L., Rylander, K.A., Quade, J., and Matthei, O.: A vegetation history from the arid prepuna of
1056   northern Chile (22-23°S) over the last 13500 years, Palaeogeogr. Palaeocl.,194, 223-246, doi:10.1016/S0031-
1057   0182(03)00279-7, 2003.

1058   Latorre, C., Betancourt, J.L., and Arroyo, M.T.K.: Late Quaternary vegetation and climate history of a perennial river
1059   canyon in the Río Salado basin (22°S) of northern Chile, Quaternary Res., 65, 450-466,
1060   doi:10.1016/j.yqres.2006.02.002, 2006.

1061   Lechler, A. and Niemi, N.: Sedimentologic and isotopic constraints on the Paleogene paleogeography and
1062   paleotopography of the southern Sierra Nevada, California, Geology, 39(4), 379-382, doi:10.1130/g31535.1, 2011.

1063   Lechler, A., Niemi, N., Hren, M. and Lohmann, K.: Paleoelevation estimates for the northern and central proto-Basin
1064   and Range from carbonate clumped isotope thermometry, Tectonics, 32(3), 295-316, doi:10.1002/tect.20016, 2013.

1065   Lease, R.O., and Ehlers, T.A.: Incision into the Eastern Andean Plateau During Pliocene Cooling, *Science*, *341*(6147),
1066   774–776, doi:10.1126/science.1239132, 2013.

1067   Legates, D.R., and Willmott, C.J.: Mean Seasonal and Spatial Variability in Gauge-Corrected, Global Precipitation, *In-*
1068   *ternational Journal of Climatology 10(2), 111-127. doi: 10.1002/joc.3370100202,* 1990.

1069   Li, J., Ehlers, T.A., Werner, M., Mutz, S.G., Steger, C., Paeth, H.: Late quarternary climate, precipitation δ^{18}O, and In-

1070   dian monsoon variations over the Tibetan Plateau. *Earth and Planetary Science Letters* 457, 412-422, 2017.

1071   Liu, X., Colman, S. M., Brown, E. T., An, Z., Zhou, W., Jull, A. J. T., Huang, Cheng, Y., P., Liu, W., and Xu, H.: A
1072   climate threshold at the eastern edge of the Tibetan plateau, Geophys. Res. Lett., doi: 10.1002/2014GL060833,
1073   2014.

1074   Licht, A., Quade, J., Kowler, A., de los Santos, M., Hudson, A., Schauer, A., Huntington, K., Copeland, P. and Lawton,
1075   T.: Impact of the North American monsoon on isotope paleoaltimeters: Implications for the paleoaltimetry of the
1076   American southwest, Am. J. Scie, 317(1), 1-33, doi:10.2475/01.2017.01, 2017.

1077   Lohmann, G., Pfeiffer, M., Laepple, T., Leduc, G., and Kim, J.-H.: A model-data comparison of the Holocene global sea
1078   surface temperature evolution. Clim. Past, 9, 1807-1839, doi:10.5194/cp-9-1807-2013, 2013.

1079

1080   Lorenz, S.J., and Lohmann, G.: Acceleration technique for Milankovitch type forcing in a coupled atmosphere-ocean
1081   circulation model: method and application for the Holocene. Climate Dynamics (2004) 23: 727.
1082   doi:10.1007/s00382-004-0469-y, 2004.

1083
1084   Maldonado, A. and Villagrán, C.: Climate variability over the last 9900 cal yr BP from a swamp forest pollen record
1085   along the semiardi coast of Chile, Quaternary Res., 66, 146-258, doi:10.1016/j.yqres.2006.04.003, 2006.

1086   Maldonado, A. J., Betancourt, J. L., Latorre, C., and Villagrán, C : Pollen analyses from a 50000-yr rodent midden
1087   series in the southern Atacama Desert (25°30' S), J. Quaternary Sci., 20, 493-507, doi:10.1002/jqs.936, 2005.

1088   Maldonado, A., Méndez, C., Ugalde, P., Jackson, D., Seguel, R., and Latorre, C.: Early Holocene climate change and
1089   human occupation along the semiarid coast of north-central Chile, J. Quaternary Sci., 25, 985–988,
1090   doi:10.1002/jqs.1385, 2010.

1091   Markgraf, V., Bradbury, J. P., Schwalb, A., Burns, S., Stern, C., Ariztegui, D., Gilli, D., Anselmetti, F. S., Stine, S., and
1092   Maidana, N.: Holocene palaeoclimates of southern Patagonia: limnological and environmental history of Lago
1093   Cardiel, Argentina, Holocene, 13, 581–591, doi:10.1191/0959683603hl648rp, 2003.

1094   Markgraf, V., Whitlock, C., and Haberle, S.: Vegetation and fire history during the last 18,000 cal yr B.P. in Southern
1095   Patagonia: Mallín Pollux, Coyhaique, Province Aisén (45°41'30" S, 71°50'30" W, 640 m elevation), Palaeogeogr.
1096   Palaeocl , 254, 492–507, doi:10.1016/j.palaeo.2007.07.008, 2007.

1097   Maroon, E. A., Frierson, D. M. W., and Battisti, D. S.: The tropical precipitation response to Andes topography and
1098   ocean heat fluxes in an aquaplanet model, J. Climate, 28, 381–398, doi:10.1175/JCLI-D-14-00188.1, 2015.

1099   Maroon, E. A., Frierson, D. M. W., Kang, S. M., and Scheff, J.: The precipitation response to an idealized subtropical
1100   continent, J. Climate, 29, 4543–4564, doi:10.1175/JCLI-D-15-0616.1, 2016.

1101   Marshall, J.A., Roering, J.J., Bartlein, P.J., Gavin, D.G., Granger, D.E., Rempel, A.W., Praskievicz, S.J., and Hales,
1102   T.C.: Frost for the trees: Did climate increase erosion in unglaciated landscapes during the late Pleistocene? *Science*
1103   *Advances*, 1, 1-10, 2015.

1104   Marston, R.A.: Geomorphology and vegetation on hillslopes: Interactions, dependencies, and feedback loops, *Geomor-*
1105   *phology*, 116(3–4), 206–217. doi:http://doi.org/10.1016/j.geomorph.2009.09.028, 2010.

1106   Matsuoka, N., and Murton, J.: Frost weathering: Recent advances and future directions. *Permafr. Periglac. Process.* 19,
1107   195–210, 2008.

1108   Matsuoka, N.: Solifluction rates, processes and landforms: A global review. *Earth Science Reviews* 55, 107–134, 2001.

1109 Maussion, F., Scherer, D., Mölg, T., Collier, E., Curio, J., and Finkelnburg, R.: Precipitation seasonality and variability
1110   over the Tibetan Plateau as resolved by the High Asia Reanalysis, *J. Climate, 27, 1910-1927, doi:10.1175/JCLI-D-*
1111   *13-00282.1*, 2014

1112 Méndez, C., Gil, A., Neme, G., Nuevo Delaunay, A., Cortegoso, V., Huidobro, C., Durán, and Maldano, A.: Mid
1113   Holocene radiocarbon ages in the Subtropical Andes (~29° - 35° S), climatic change an implicaton for human space
1114   organization, Quatern. Int., 356, 15-26, doi:10.1016/j.quaint.2014.06.059, 2015.

1115 Mesinger, F., DiMego, G., Kalnay, E., Mitchell, K., Shafran, P.C., Ebisuzaki, W., Jovic, D., Woollen, J., Rogers, E.,
1116   Berbery, E.H., Ek, M.B., Fan, Y., Grumbine, R., Higgins, W., Li, H., Lin, Y., Manikin, G., Parrish, D., and Shi,
1117   W.:North American Regional Reanalysis, *Bulletin of the American Meteorological Society, 87, 343–*
1118   *360.doi:10.1175/BAMS-87-3-343, 2006.*

1119 Methner, K., Fiebig, J., Wacker, U., Umhoefer, P., Chamberlain, C. and Mulch, A.: Eocene-Oligocene proto-Cascades
1120   topography revealed by clumped ($\Delta47$) and oxygen isotope ($\delta18O$) geochemistry (Chumstick Basin, WA, USA),
1121   Tectonics, 35(3), 546-564, doi:10.1002/2015tc003984, 2016.

1122 Mischke S., Kramer M., Herzschuh U., Shang H., Erzinger J., and Zhang C.: Reduced early holocene moisture
1123   availability in the Bayan Har Mountains, northeastern Tibetan Plateau, inferred from a multi-proxy lake record,
1124   Palaeogeogr. Palaeocl, 267, 59-76, doi:10.1016/j.palaeo.2008.06.002, 2008.

1125 Molnar, P., Boos, W.R., and Battisti, D.S.: Orographic Controls on Climate and Paleoclimate of Asia: Thermal and
1126   Mechanical Roles for the Tibetan Plateau, *Annual Review of Earth and Planetary Sciences*, 38, 77-102, 2010.

1127 Molnar, P., and England, P.: Late Cenozoic uplift of mountain ranges and global climate change: chicken or egg?,
1128   *Nature*, 346, 29-34, 1990.

1129 Montgomery, D.R.: Slope distributions, threshold hillslopes, and steady-state topography, *American Journal of Science,*
1130   301, 432-454, 2001.

1131 Montgomery, D.R., Balco, G., and Willett, S.D.: Climate, tectonics, and the morphology of the Andes, *Geology*, 29(7),
1132   579-582, 2001.

1133 Moon, S., Chamberlain, C.P., Blisniuk, K., Levine, N., Rood, D.H., Hilley, G.E.: Climatic control of denudation in the
1134   deglaciated landscape of the Washington Cascades, *Nature Geoscience,* 4, 469-473, 2011.

1135 Moreno, A., Santoro, C. M., and Latorre, C.: Climate change and human occupation in the northernmost Chilean
1136   Altiplano over the last ca. 11500 cal. a BP, Quaternary Sci., 24, 373–382, doi:10.1002/jqs.1240, 2009.

1137 Morinaga, H., Itota, C., Isezaki, N., Goto, H., Yaskawa, K., Kusakabe, M., Liu, J., Gu, Z., Yuan, B., and Cong, S.:
1138   Oxygen-18 and carbon-13 records for the last 14,000 years from lacustrine carbonates of Siling-Co (lake) in the
1139   Qinghai-Tibetan Plateau, Geophys. Res. Lett., 20, 2909-2912, doi:10.1029/93GL02982, 1993.

1140 Morrill, C., Overpeck, J. T., Cole, J. E., Liu, K., Shen, C., and Tang, L.: Holocene variations in theAsian monsoon in-
1141   ferred from the geochemistry of lake sediments in central Tibet, Quaternary Res., 65, 232-243,
1142   doi:10.1016/j.yqres.2005.02.014, 2006.

1143 Moulton, K.L., and Berner, R.A.: Quantification of the effect of plants on weathering: studies in Iceland, *Geology*, 26,

1144   895-898, 1998.

1145   Muegler, I., Gleixner, G., Guenther, F., Maeusbacher, R., Daut, G., Schuett, B., Berking, J., Schwalb, A., Schwark, L.,,
1146   Xu, B., Yao, T., Zhu, L., and Yi, C.: A multi-proxy approach to reconstruct hydrological changes and Holocene
1147   climate development of Nam Co, Central Tibet, J. Paleolimnol., 43, 625-648, doi:10.1007/s10933-009-9357-0, 2010.

1148   Mujica, M. I., Latorre, C., Maldonado, A., González-Silvestre, L., Pinto, R., De Pol-Holz, R., and Santoro, C. M.: Late
1149   Quaternary climate change, relict populations and present-day refugia in the northern Atacama Desert: a case study
1150   from Quebrada La Higuera (18° S), J. Biogeogr. , 42, 76–88, doi:10.1111/jbi.12383, 2015.

1151   Mulch, A., Sarna-Wojcicki, A., Perkins, M. and Chamberlain, C.: A Miocene to Pleistocene climate and elevation record
1152   of the Sierra Nevada (California), P. Natl. Acad. Sci. USA, 105(19), 6819-6824, doi:10.1073/pnas.0708811105,
1153   2008.
1154
1155   Mulch, A., Chamberlain, C., Cosca, M., Teyssier, C., Methner, K., Hren, M. and Graham, S.: Rapid change in high-elev-
1156   ation precipitation patterns of western North America during the Middle Eocene Climatic Optimum (MECO), Am. J.
1157   Scie, 315(4), 317-336, doi:10.2475/04.2015.02, 2015.

1158   Mutz, S.G., Ehlers, T.A., Li, J., Steger, C., Peath, H., Werner, M., Poulsen, C.J.: Precipitation δ^{18}O over the South Asia
1159   Orogen from ECHAM5-wiso Simulation: Statistical Analysis of Temperature, Topography and Precipitation.
1160   *Journal of Geophysical Research-Atmospheres*, 121(16),9278-9300, doi: 10.1002/2016JD024856, 2016.

1161   Nishimura, M., Matsunaka, T., Morita, Y., Watanabe, T., Nakamura, T., Zhu, L., Nara, W. F., Imai, A., Izutsu, Y., and
1162   Hasuike, N.: Paleoclimatic changes on the southern Tibetan Plateau over the past 19,000 years recorded in lake
1163   Pumoyum Co, and their implications for the southwest monsoon evolution, Palaeogeogr. Palaeocl., 396, 75-92,
1164   doi:10.1016/j.palaeo.2013.12.015, 2014.

1165   Otto-Bliesner, B.L., Brady, C.B., Clauzet, G., Tomas, R., Levis, S., and Kothavala, Z.: Last Glacial Maximum and
1166   Holocene Climate in CCSM3. Journal of Climate, 19, 2526-2544, 2006.

1167   Paeth, H.: Key Factors in African Climate Change Evaluated by a Regional Climate Model, *Erdkunde*, 58, 290-315,
1168   2004.

1169   Peel, M.C., Finlayson, B.L., McMahon, T.A.: Updated world map of the Koppen- Geiger climate classification.
1170   Hydrol. Earth Syst. Sci. 11, 1633–1644, 2007.
1171
1172   Pfeiffer, M., and Lohmann, G.: Greenland Ice Sheet influence on Last Interglacial climate: global sensitivity studies
1173   performed with an atmosphere-ocean general circulation model. *Climate of the Past*, 12, pp. 1313-1338,
1174   doi:10.5194/cp-12-1313-2016, 2016.

1175   Pickett, E.J., Harrison, S.P., Flenley, J., Grindrod, J., Haberle, S., Hassell, C., Kenyon, C., MacPhail, M., Martin, H.,
1176   Martin, A.H., McKenzie, M., Newsome, J.C., Penny, D., Powell, J., Raine, J.I., Southern, W., Stevenson, J., Sutra,
1177   J.-P., Thomas, I., van der Kaars, S., Ward, J.: Pollen-based reconstructions of biome distributions for Australia,
1178   South-East Asia and the Pacific (SEAPAC region) at 0, 6000 and 18,000 14C years B.P.. Journal of Biogeography,
1179   31, 1381–1444, doi: 10.1111/j.1365-2699.2004.01001.x, 2004.
1180
1181   Pingel, H., Mulch, A., Alonso, R., Cottle, J., Hynek, S., Poletti, J., Rohrmann, A., Schmitt, A., Stockli, D. and Strecker,
1182   M.: Surface uplift and convective rainfall along the southern Central Andes (Angastaco Basin, NW Argentina),
1183   Earth Planet. Sc. Lett., 440, 33-42, doi:10.1016/j.epsl.2016.02.009, 2016.

Prentice, I. C., Jolly, D., and BIOME 6000 Participants. (2000). Mid-Holocene and glacial-maximum vegetation geography of the northern continents and Africa. Journal of Biogeography 27, 507-519.

Pueyo, J. J., Sáez, A., Giralt, S., Valero-Garcés, B.L., Moreno, A., Bao, R., Schwalb, A., Herrera, C., Klosowska, B., and Taberner, C.: Carbonate and organic matter sedimentation and isotopic signatures in Lake Chungará, Chilean Altiplano, during the last 12.3 kyr, Palaeogeogr. Palaeocl., 307, 339-355, doi:10.1016/j.palaeo.2011.05.036, 2011.

Quade, J., Rech, J. A., Betancourt, J. L., Latorre, C., Quade, B., Rylander, K. A., and Fisher, T.: Paleowetlands and regional climate change in the central Atacama Desert, northern Chile, Quaternary Res., 69, 343-360, doi:10.1016/j.yqres.2008.01.003, 2008.

Raymo, M.E., and Ruddiman, W.F.: Tectonic forcing of late cenozoic climate. *Nature*, 359(6391), 117-122, 1992.

Rech, J. A., Pigati, J. S., Quade, J., and Betancourt, J. L.: Re-evaluation of mid-Holocene deposits at Quebrada Puripica, northern Chile, Palaeogeogr. Palaeocl., 194, 207-222, doi:0.1016/S0031-0182(03)00278-5, 2003.

Robinson, M.M.: New quantitative evidence of extreme warmth in the Pliocene Arctic, Stratigraphy 6, 265–275, 2009.

Roeckner, E., Bäuml, G., Bonaventura, L., Brokoph, R., Esch, M., Giorgetta, M., Hagemann, S., Kirchner, I., Kornblueh, L., Manzini, E., Rhodin, A., Schlese, U., Schulzweida, U., and Tompkins, A.: The atmospheric general circulation model ECHAM5. Part I: Model description. Rep. 349Rep., 127 pp, Max Planck Institute for Meteorology, Hamburg, 2003.

Roering, J.J., Marshall, J., Booth, A.M., Mort, M., and Jin, Q.: Evidence for biotic controls on topography and soil production, *Earth and Planetary Science Letters*, *298*(1–2), 183–190, doi:10.1016/j.epsl.2010.07.040, 2010.

Salzmann, U., Williams, M., Haywood, A.M., Johnson, A.L.A., Kender, S., and Zalasiewicz, J.: Climate and environment of a Pliocene warm world. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 309 (1-8), 2011.

Sarnthein, M.: Sand deserts during glacial maximum and climatic optimum, Nature, 272, 43–46, doi:10.1038/272043a0, 1978.

Sarnthein, M., Gersonde, R., Niebler, S., Pflaumann, U., Spielhagen, R., Thiede, J., Wefer, G., Weinelt, M.: Overview of Glacial Atlantic Ocean Mapping (GLAMAP 2000), *Paleoceanography*, 18(2), doi:10.1029/2002PA000769, 2003.

Schäfer-Neth C., and Paul, A.: The Atlantic Ocean at the last glacial maximum: objective mapping of the GLAMAP sea-surface conditions. In: Wefer G, Mulitza S, Ratmeyer V (eds) The South Atlantic in the late quaternary: reconstruction of material budgets and current systems. Springer, Berlin, pp 531–548, 2003.

Schaller, M., von Blanckenburg, F., Veldkamp, A., Tebbens, L.A., Hovius, N., and Kubik, P.W.: A 30 000 yr record of erosion rates from cosmogenic 10 Be in Middle European river terraces, *Earth and Planetary Science Letters*, *204*(1), 307–320, 2002.

Schaller, M., and Ehlers T.A.: Limits to quantifying climate driven changes in denudation rates with cosmogenic radionuclides, *Earth and Planetary Science Letters*, v. 248, pp. 153-167. doi:10.1016/j.epsl.2006.05.027, 2006.

Schwalb, A., Burns, S., and Kelts, K.: Holocene environments from stable isotope stratigraphy of ostracods and authigenic carbonate in Chilean Altiplano Lakes, Palaeogeogr. Palaeocl.,148, 153-168, doi:10.1016/S0031-0182(98)00181-3, 1999.

Shen J., Liu X. Q., Wang S. M., and Matsumoto R.: Palaeoclimatic changes in the Qinghai, Lake area during the last 18,000 years, Quatern. Int., 136, 131-140, doi:10.1016/j.quaint.2004.11.014, 2005.

Shen C., Liu K., Tang L., and Overpeck J. T.: Quantitative relationships between modern pollen rain and climate in the Tibetan Plateau. Rev. Palaeobot. Palyno., 140, 61–77, doi:10.1016/j.revpalbo.2006.03.001, 2006.

Siani, G., Colin, C., Michel, E., Carel, M., Richter, T., Kissel, C., and Dewilde, F.: Late Glacial to Holocene terrigenous sediment record in the Northern Patagonian margin: Paleoclimate implications, Palaeogeogr. Palaeocl., 297, 26-36, doi:10.1016/j.palaeo.2010.07.011, 2010.

Simmons, A.J., Burridge, D.M., Jarraud, M., Girard, C., and Wergen, W.: The ECMWF Medium-Range prediction models development of the numerical formulations and the impact of increased resolution, Meteorology and Atmospheric Physics, 40(1-3), 28-60, 1989.

Sohl, L.E., Chandler, M.A., Schmunk, R.B., Mankoff, K., Jonas, J.A., Foley, K.M., and Dowsett, H.J.: PRISM3/GISS topographic reconstruction, *U.S. Geological Survey Data Series,* 419, 6p., 2009.

Sowers T., Alley, R.B., and Jubenville, J.: Ice core records of atmospheric N_2O covering the last 106,000 years. *Science*, 301:945–948, 2003.

Stepanek, C., and Lohmann, G.: Modelling mid-Pliocene climate with COSMOS , *Geosci. Model Dev.* , 5 , pp. 1221-1243 . doi:10.5194/gmd-5-1221-2012, 2012.

Stine, S. and Stine M.: A record from Lake Cardiel of climate change in southern South America, Nature, 345, 705-708, doi:10.1038/345705a0,1990.

Szeicz, J. M., Haberle, S. G., and Bennett, K. D.: Dynamics of North Patagonian rainforests from fine-resolution pollen, charcoal and tree-ring analysis, Chonos Archipelago, Southern Chile, Austral Ecol., 28, 413–422, doi:10.1046/j.1442-9993.2003.01299.x , 2003.

Takahashi, K., and Battisti, D.: Processes controlling the mean tropical pacific precipitation pattern. Part I: The Andes and the eastern Pacific ITCZ. *Journal of Climate*, 20(14), 3434-3451, 2007a.

Takahashi, K., and Battisti, D.: Processes controlling the mean tropical pacific precipitation pattern. Part II: The SPCZ and the southeast pacific dry zone. Journal of Climate, 20(23), 5696-5706, 2007b.

Tang, L., Shen, S., Liu, K., and Overpeck, J. T.: Changes in south Asian monsoon: new high-resolution paleoclimatic records from Tibet, China, Chinese Sci. Bull., 45, 87-91, doi:10.1007/BF02884911, 2000.

Thiede, R.C., and Ehlers, T.A.: Large spatial and temporal variations in Himalayan denudation. *Earth and Planetary Science Letters*, *374, 256-257. doi:10.1016/j.epsl.2013.03.004*, 2013.

Thomas, A.: The climate of the Gongga Shan range, Sichuan Province, PR China. Arctic and Alpine Research, 29(2), 226-232, 1997.

Uppala, S.M., Kållberg, P.W., Simmons, A.J., Andrae, U., da Costa Bechtold, V., Fiorino, M., Gibson, J.K., Haseler, J., Hernandez, A., Kelly, G.A., Li, X., Onogi, K., Saarinen, S., Sokka, N., Allan, R.P., Andersson, E., Arpe, K., Balmaseda, M.A., Beljaars, A.C.M., van de Berg, L., Bidlot, J., Bormann, N., Caires, S., Chevallier, F., Dethof, A., Dragosavac, M., Fisher, M. Fuentes, M., Hagemann, S., Hólm, E., Hoskins, B.J., Isaksen, L., Janssen, P.A.E.M., Jenne, R., McNally, A.P., Mahfouf, J.-F., Morcrette, J.-J , Rayner, N.A., Saunders, R.W., Simon, P., Sterl, A., Trenberth, K.E., Untch, A., Vasiljevic, D., Viterbo, P., and Woollen, J.: The ERA-40 re-analysis. Quart. *J. R. Meteorol. Soc.*, 131, 2961-3012, 2005.

von Blanckenburg, F., Bouchez, J., Ibarra, D.E., and Maher, K.: Stable runoff and weathering fluxes into the oceans over Quaternary climate cycles, *Nature Geoscience*, *8*(7), 538–542, doi:10.1038/ngeo2452, 2015.

Valla, P.G., Shuster, D.L., and van der Beek, P.A.: Significance increase in relief of European Alps during mid-Pleistocene glaciations. *Nature Geoscience*, 4, 688–692. doi:10.1038/ngeo1242, 2011.

Van Campo, E., Cour, P., and Huang, S.: Holocene environmental changes in Bangong Co basin (Western Tibet). Part 2: The pollen record, Palaeogeogr. Palaeocl., 120, 49-63, doi:10.1016/0031-0182(95)00033-X, 1996.

Villagrán, C. and Varela, J.: Palynological Evidence for Increased Aridity on the Cenral Chilean Coast during the Holocene, Quaternary Res., 34, 198–207, doi:10.1016/0033-5894(90)90031-F, 1990.

Villa-Martínez, R. Villagrán, C., and Jenny, B.: The last 7500 cal yr B.P. sf westerly rainfall in Central Chile inferred from a high-resolution pollen record from Laguna Aculeo (34°S), Quaternary Res., 60, 284-293, doi:https://doi.org/10.1016/j.yqres.2003.07.007, 2003.

Wei, W., and Lohmann, G.: Simulated Atlantic Multidecadal Oscillation during the Holocene. *Journal of Climate*, 25, 6989–7002. doi: http://dx.doi.org/10.1175/JCLI-D-11-00667.1, 2012.

Wang, R.L., Scarpitta, S. C., Zhang, S. C., and Zheng, M. P.: Later Pleistocene/Holocene climate conditions of Qinghai-Xizhang Plateau (Tibet) based on carbon and oxygen stable isotopes of Zabuye Lake sediments, Earth Planet. Sc. Lett., 203, 461-477, doi:10.1016/S0012-821X(02)00829-4, 2002.

Whipple, K.X, Kirby, E., and Broecklehurst, S.H.: Geomorphic limits to climate-induced increases in topographic relief. *Nature*, 401, 39-43. doi:10.1038/43375, 1999.

Whipple, K.X., and Tucker, G.E.: Dynamics of the stream-power river incision model: Implications for height limits of mountain ranges, landscape response timescales, and research needs, *Journal of Geophysical Research-Solid Earth,* 104, 17661-17674, 1999.

Whipple, K. X.: The influence of climate on the tectonic evolution of mountain belts, *Nat. Geosci.*, 2, doi:10.1038/ngeo413, 2009.

Wilks, D.S.: Statistical methods in the atmospheric sciences - 3rd ed. Academic Press, Oxford, 2011.

Willett, S.D., Schlunegger, F., and Picotti, V.: Messinian climate change and erosional destruction of the central

European Alps. *Geology*, 34(8), 613-616, 2006.

Wilson, G.S., Barron, J.A., Ashworth, A.C., Askin, R.A., Carter, J.A., Curren, M.G., Dalhuisen, D.H., Friedmann, E.I., Fyodorov-Davidov, D.G., Gilichinsky, D.A., Harper, M.A., Harwood, D.M., Hiemstra, J.F., Janecek, T.R., Licht, K.J., Ostroumov, V.E., Powell, R.D., Rivkina, E.M., Rose, S.A., Stroeven, A.P., Stroeven, P., van der Meer, J.J.M., Wizevich, M.C.: The Mount Feather Diamicton of the Sirius Group: an accumulation of indicators of Neogene Antarctic glacial and climatic history, *Palaeogeography, Palaeoclimatology, Palaeoecology,* 182 (1–2), 117–131, 2002.

Wischnewski, J., Mischke, S., Wang, Y., and Herzschuh, U.: Reconstructing climate variability on the northeastern Tibetan Plateau since the last lateglacial – a multi-proxy, dual-site approach comparing terrestrial and aquatic signals, Quaternary Science Reviews, 30, 82-97, doi:10.1016/j.quascirev.2010.10.001, 2011.

Wünnemann, B., Mischke, S., and Chen, F.: A Holocene sedimentary record from Bosten lake, China, Palaeogeogr. Palaeocl., 234, 223-238, doi:10.1016/j.palaeo.2005.10.016, 2006.

Yanhong, W., Lücke, A., Zhangdong, J., Sumin, W., Schleser, G.H., Battarbee, R.W., and Weilan, X.: Holocene climate development on the central Tibetan Plateau: A sedimentary record from Cuoe Lake, Palaeogeogr. Palaeocl., 234, 328-340, doi:10.1016/j.palaeo.2005.09.017, 2006

Yanites, B.J., and Ehlers, T.A.: Global climate and tectonic controls on the denudation of glaciated mountains. *Earth and Planetary Science Letters,* 325, 63-75, 2012.

Yu, L. and Lai, Z.: Holocene climate change inferred from stratigraphy and OSL chronology of Aeolian sediments in the Qaidam Basin, northeastern Qinghai-Tibetan Plateau, Quaternary Res., 81, 488-499, doi:10.1016/j.yqres.2013.09.006, 2014.

Zhang, C. and Mischke, S.: A lateglacial and Holocene lake record from the Nianbaoyeze Mountains and inferences of lake, glacier and climate evolution on the eastern Tibetan Plateau, Quaternary Sci. Rev., 28, 1970-1983, doi:10.1016/j.quascirev.2009.03.007, 2009.

Zhang, J., Chen, F., Holmes, J. A., Li, H., Guao, X., Wang, J., Li, S., Lu, Y., Zhao, Y., and Qiang, M.: Holocene monsoon climate documented by oxygen and carbon isotopes from lake sediments and peat bogs in China: a review and synthesis, Quaternary Sci. Rev., 30, 1973-1987, doi:10.1016/j.quascirev.2011.04.023, 2011.

Zhang, X., Lohmann, G., Knorr, G., and Xu X.: Different ocean states and transient characteristics in Last Glacial Maximum simulations and implications for deglaciation. *Clim. Past*, 9, 2319-2333, doi:10.5194/cp-9-2319-2013, 2013a.

Zhang, R., Yan, Q., Zhang, Z.S., Jiang, D., Otto-Bliesner, B.L., Haywood, A.M., Hill, D.J., Dolan, A.M., Stepanek, C., Lohmann, G., Contoux, C., Bragg, F., Chan, W.-L., Chandler, M.A., Jost, A., Kamae, Y., Abe-Ouchi, A., Ramstein, G., Rosenbloom, N.A., Sohl, L., and Ueda, H.: East Asian monsoon climate simulated in the PlioMIP. Clim. Past, 9, 2085-2099, doi:10.5194/cp-9-2085-2013, 2013b.

Zhang, X., Lohmann, G., Knorr, G., and Purcell C.: Abrupt glacial climate shifts controlled by ice sheet changes. *Nature*, 512 (7514), 290-294, doi:10.1038/nature13592, 2014.

Zhisheng, A., Kutzbach, J.E., Prell, W.L., and Porter, S.C.: Evolution of Asian monsoons and phased uplift of the South

1323   Asian plateau since Late Miocene times. *Nature*, 411(6833), 62-66, 2001.

1324 Zhou, W. J., Lu, X. F., Wu, Z. K., Deng, L., Jull, A. J. T., Donahue, D. J., and Beck, W.: Peat record reflecting Holocene

1325   climate change in the Zoigê Plateau and AMS radiocarbon dating, Chinese Sci. Bull., 47, 66-70,

1326   doi:10.1360/02tb9013, 2002.

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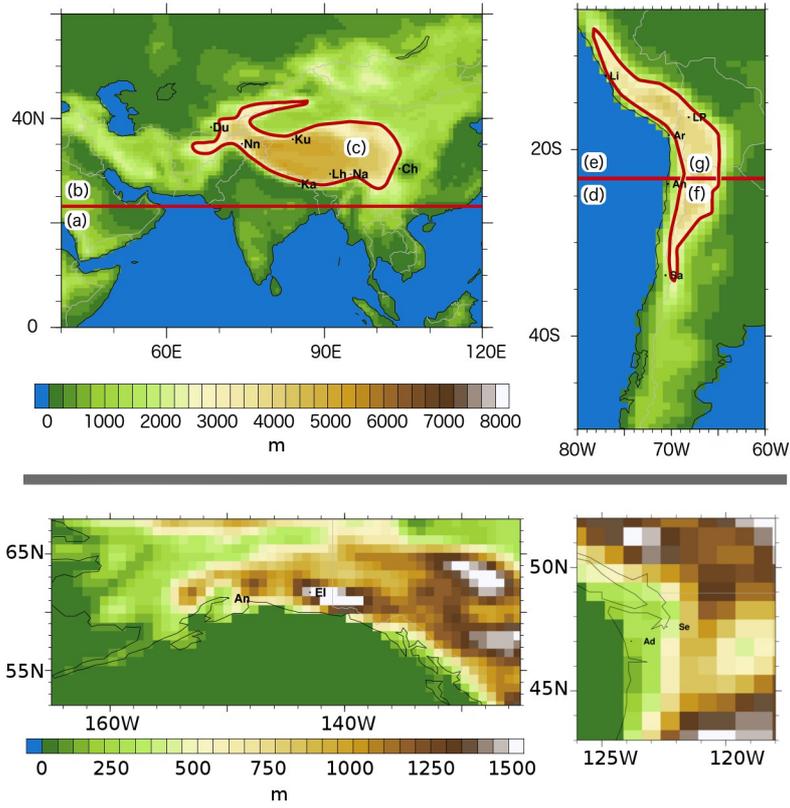
Figure 1

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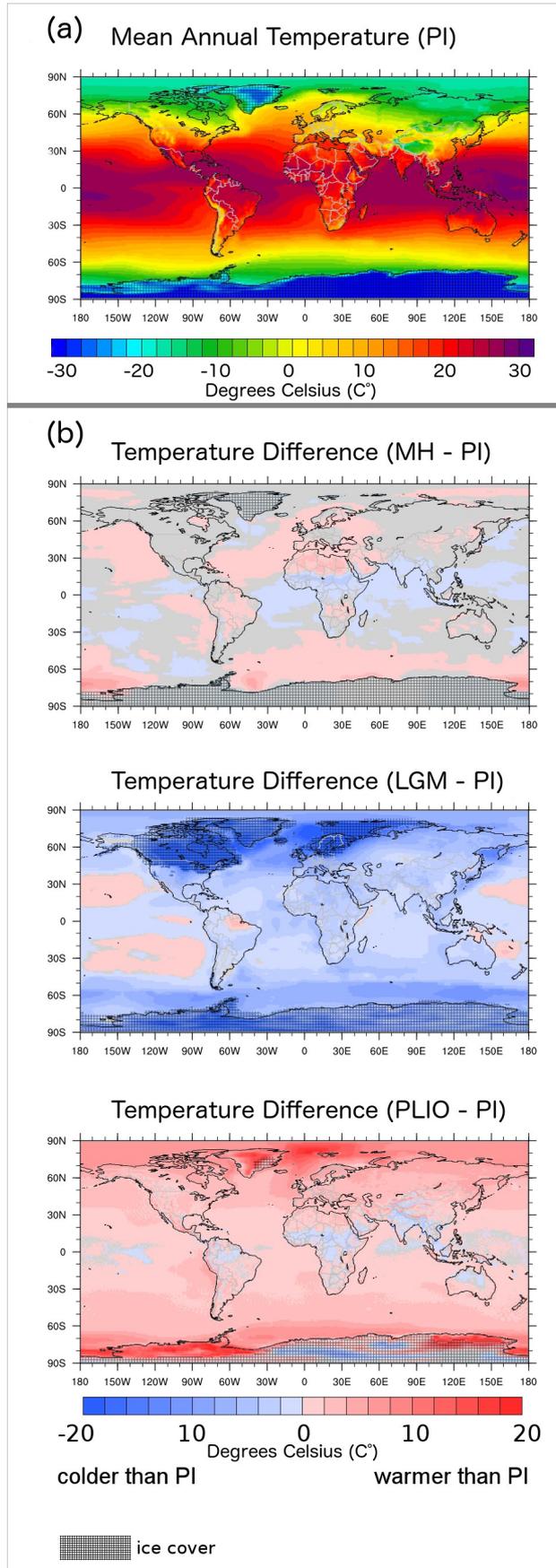
Figure 2

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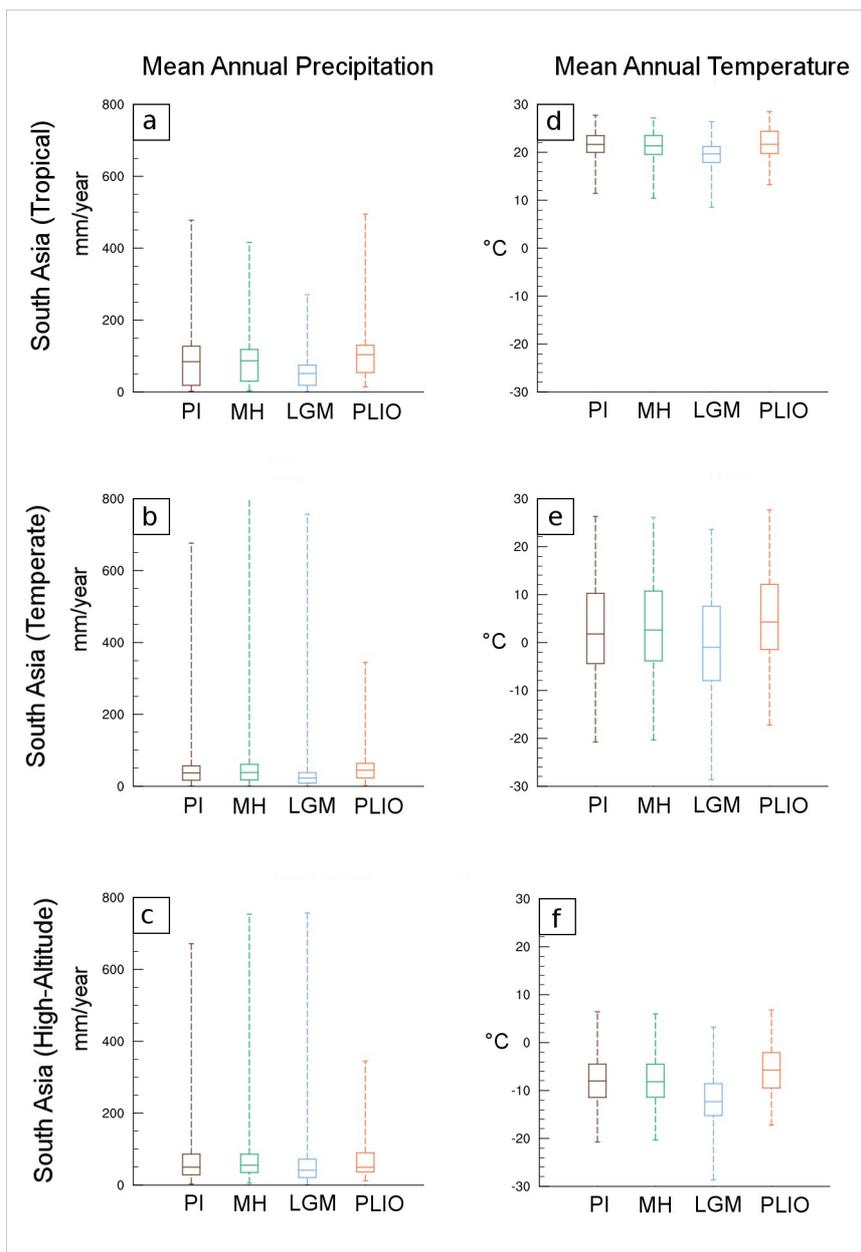
Figure 6

Figure 7

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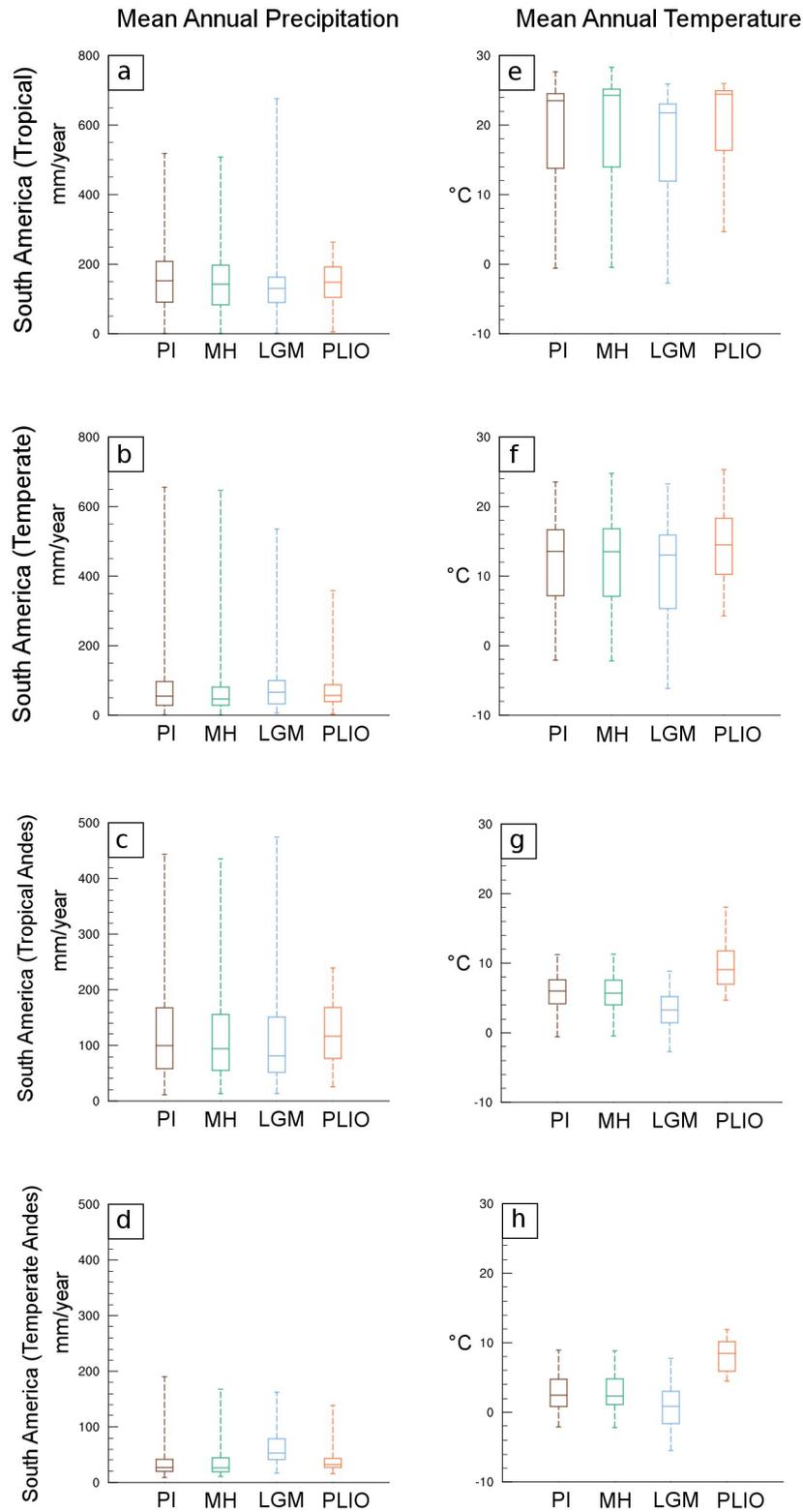
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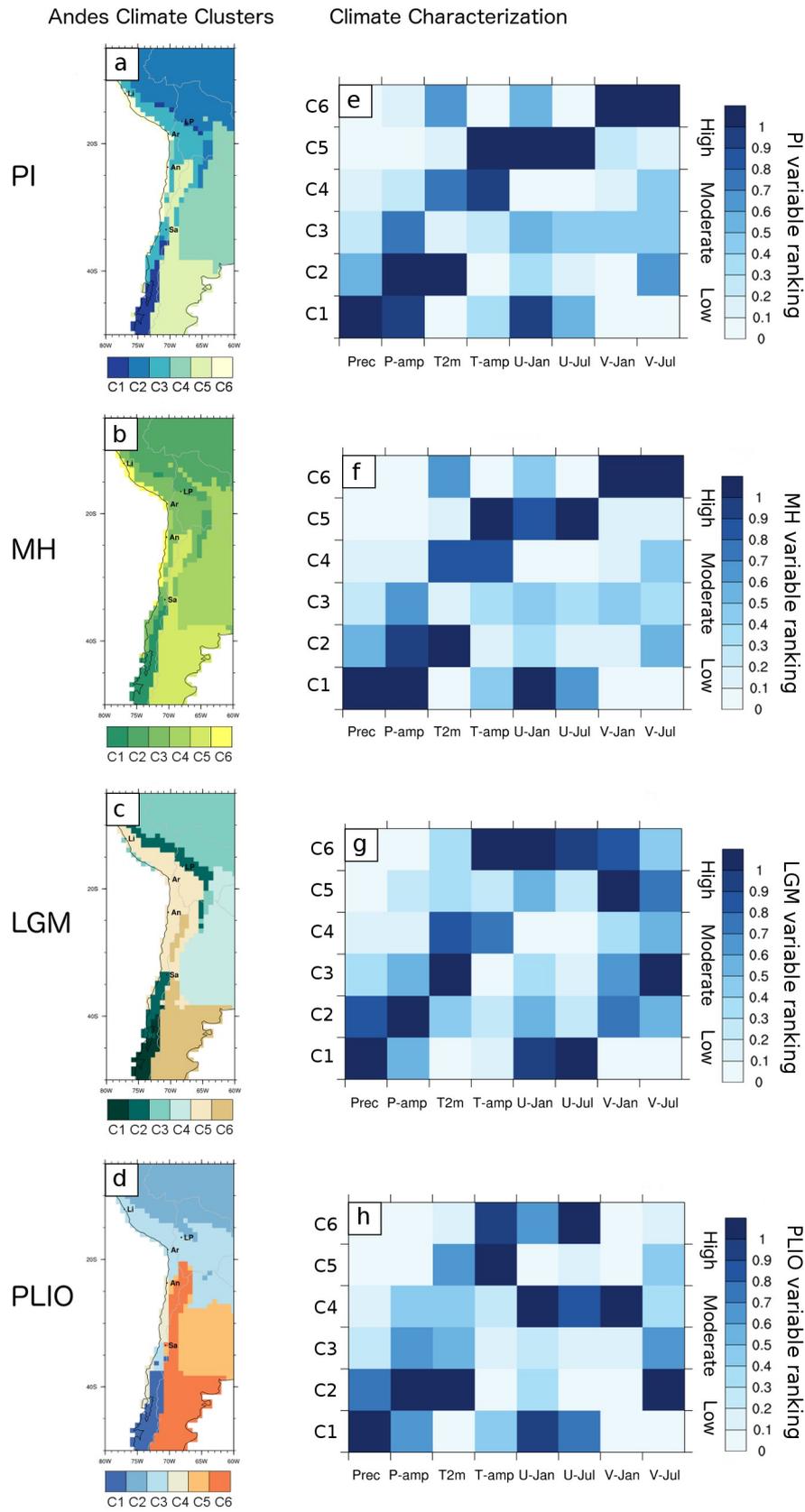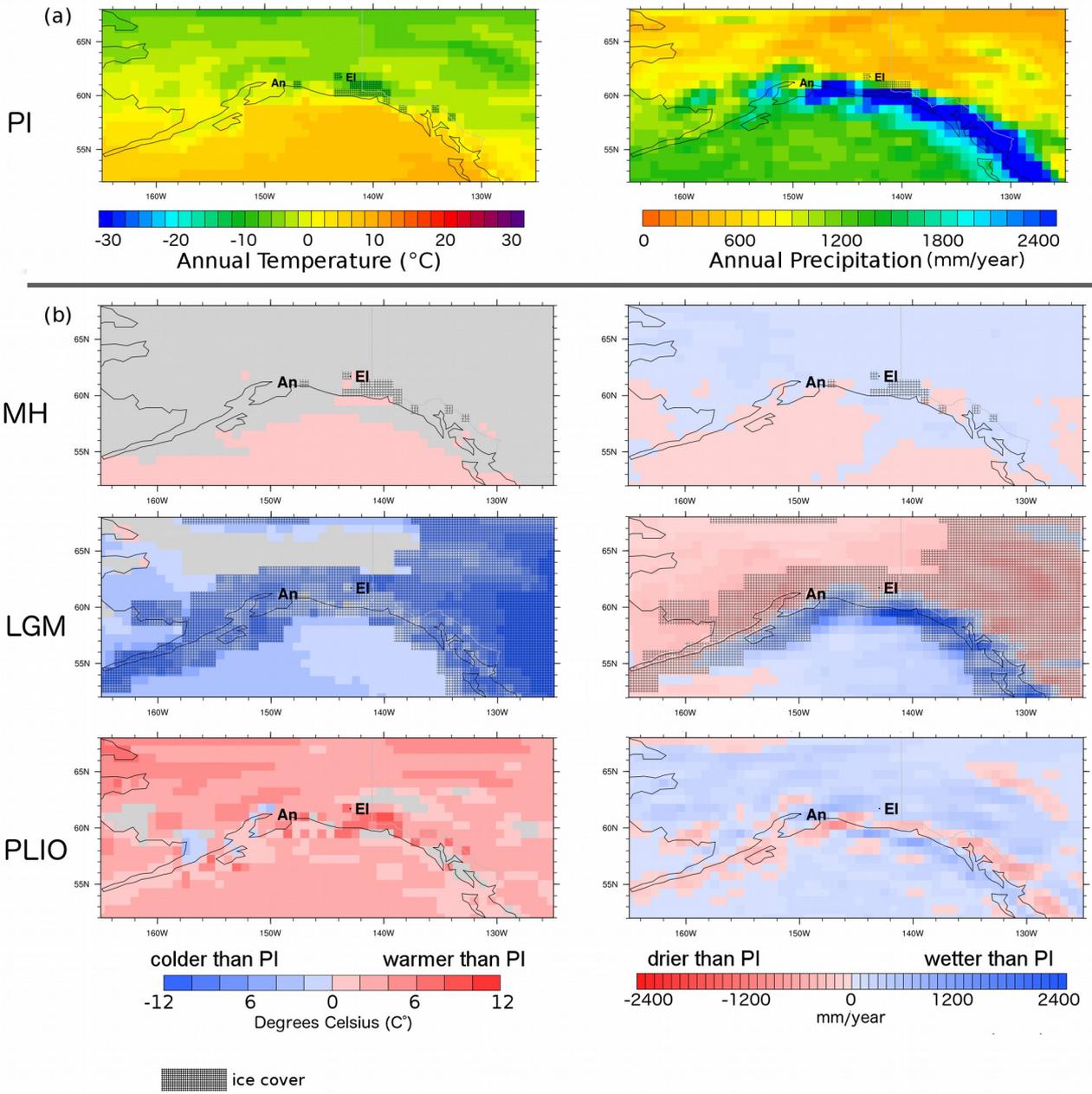
Figure 9

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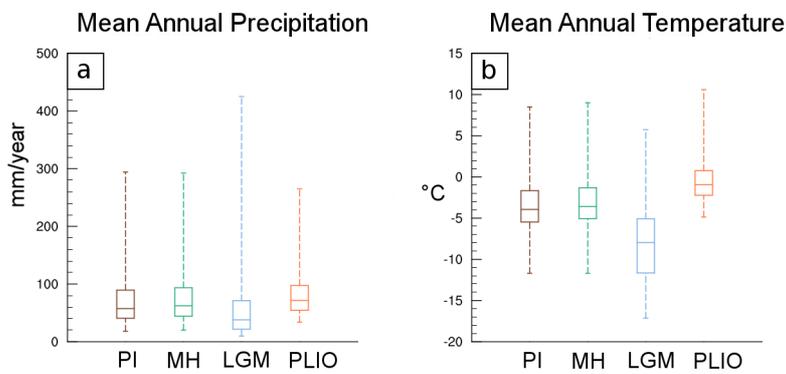
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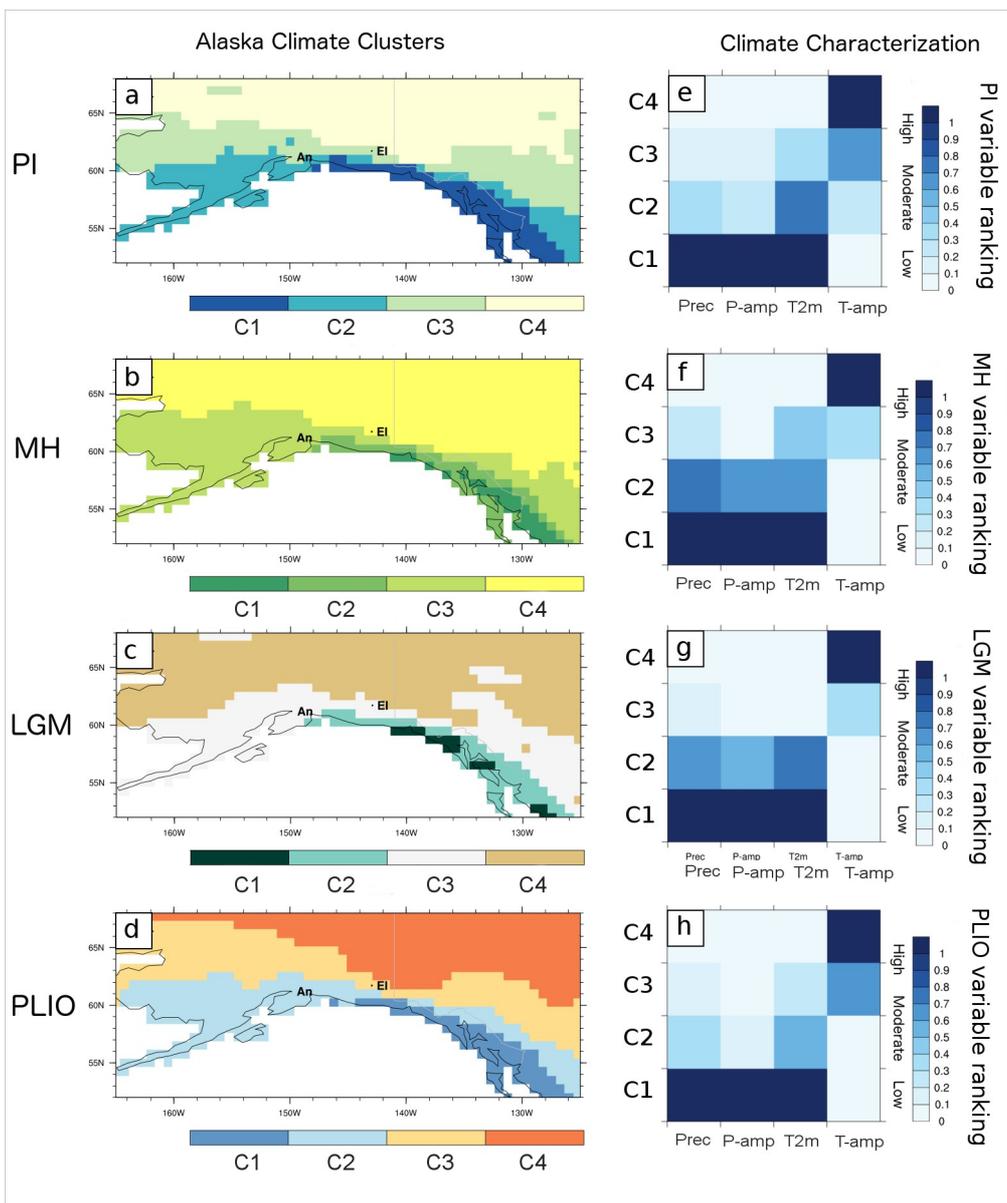
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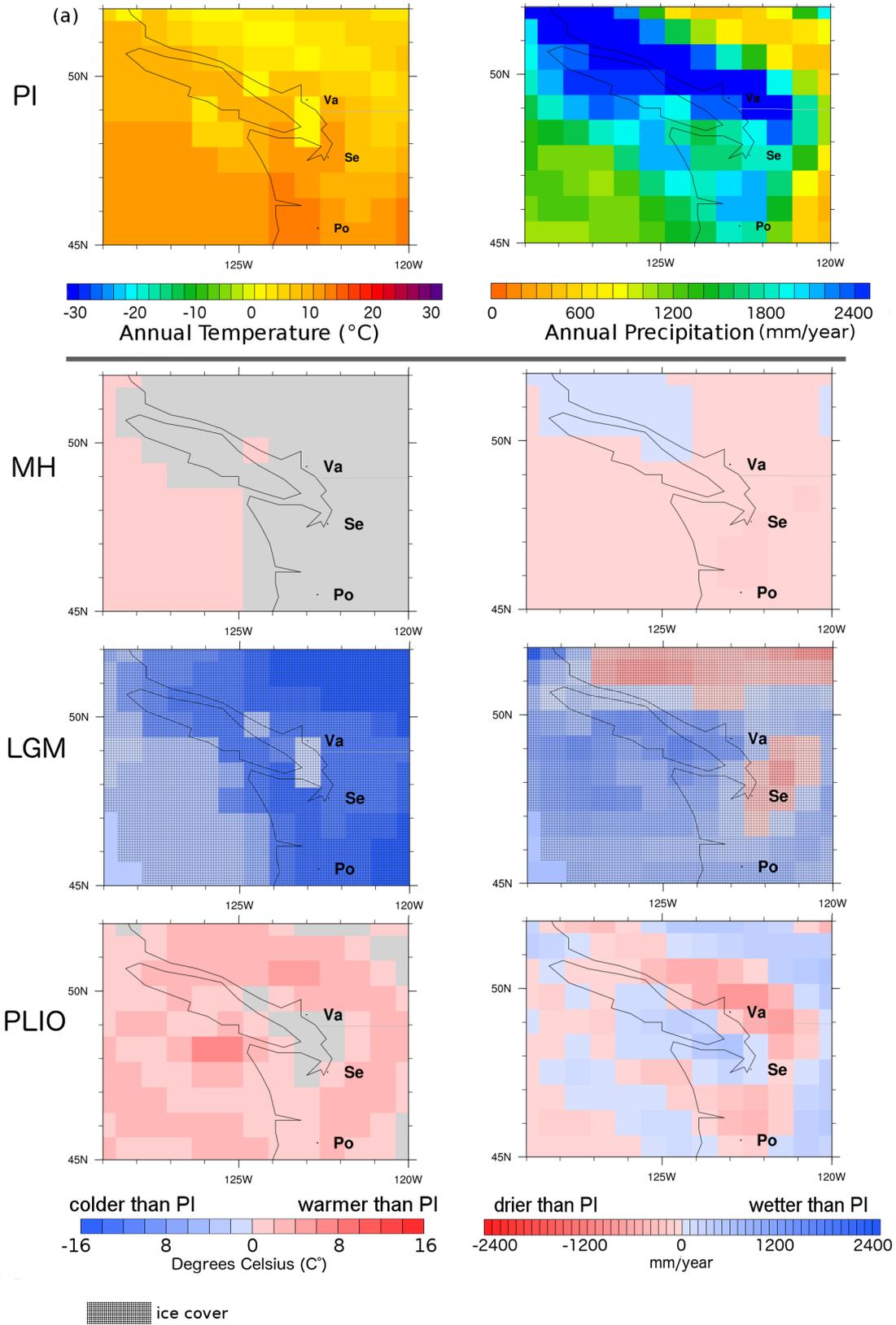
Figure 13

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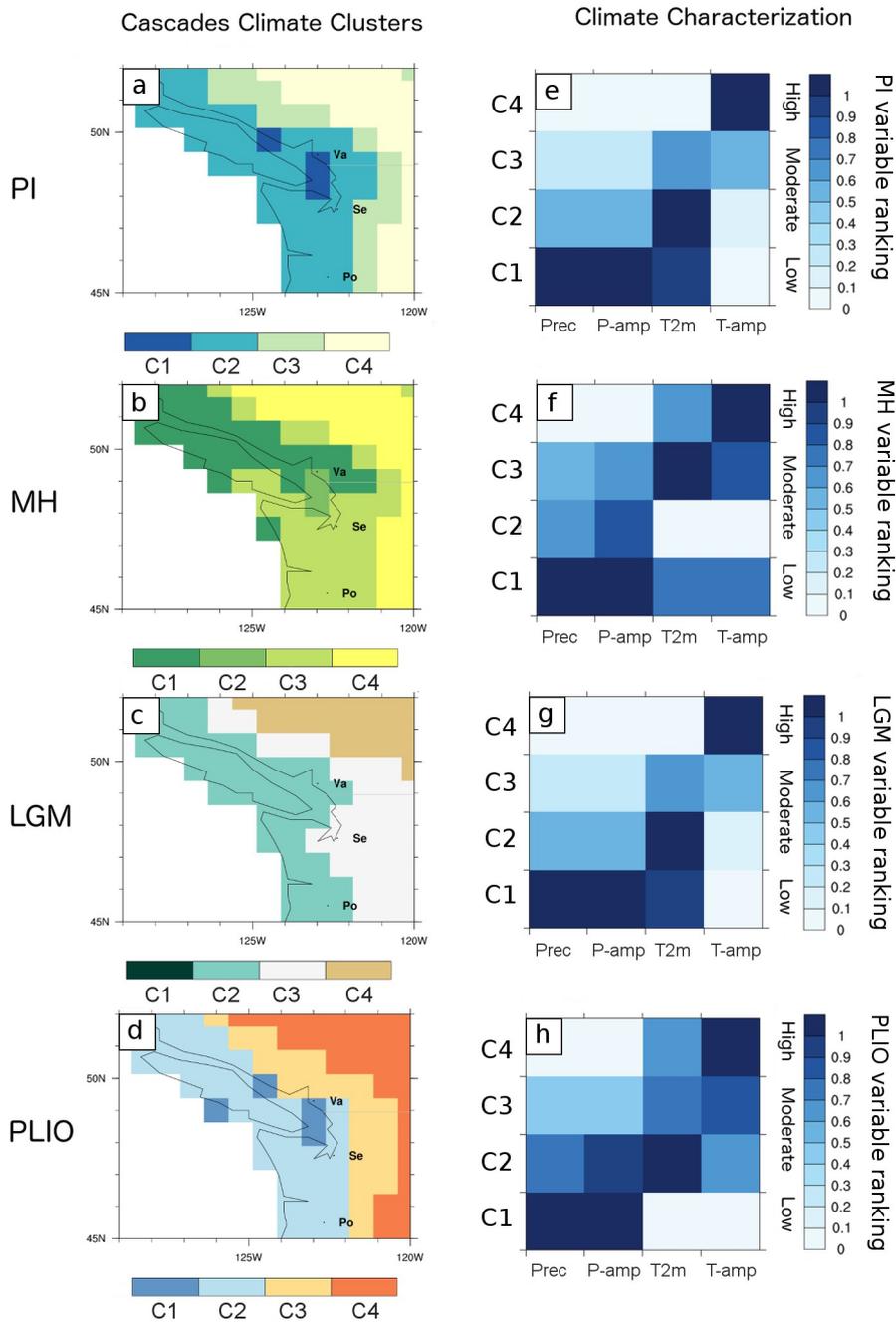2000
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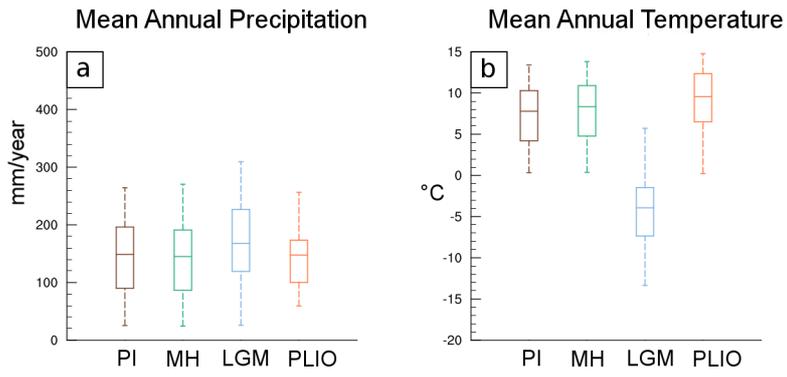
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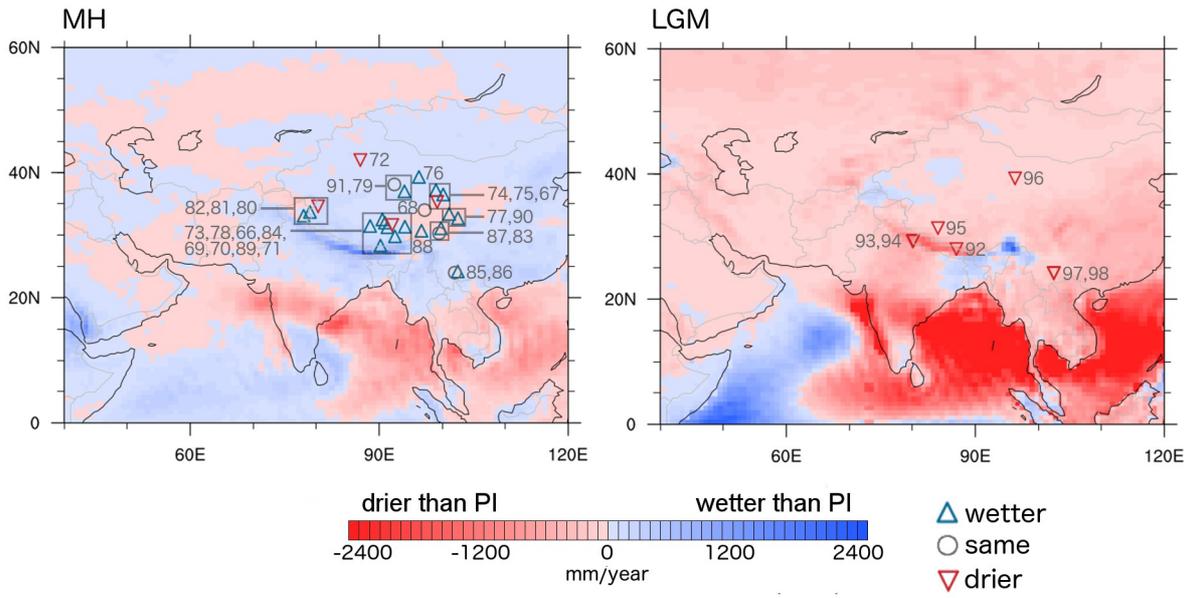
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Figure 16

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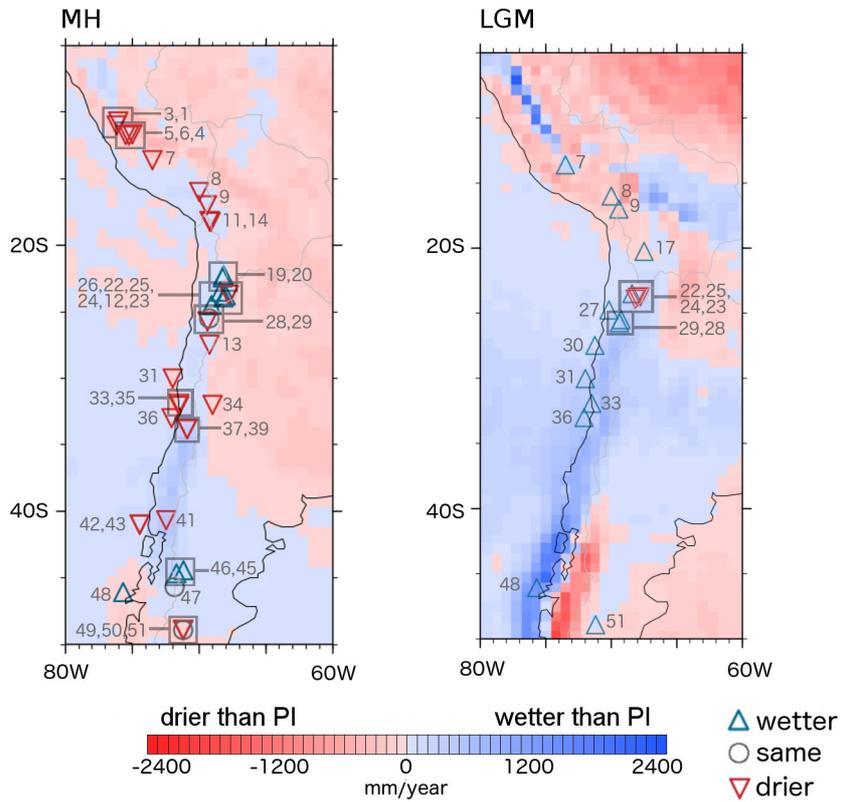2041
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Figure 17