# Response to RC1

(responses in blue)

Mutz et al. look to use paleoclimate GCMs to identify drivers of past geomorphic change. This is a topic for which I hold great interest, and I feel that the authors have crafted a very useful set of model results that they leave underutilized. As such, I feel inclined to accept the paper on the basis of the useful results, but to request major revisions such that they do their own work justice.

My major concerns, which will become clear in the line-by-line comments (please feel free to respond to similar comments en masse) are as follows: 1. The paper is motivated by denudation rates and landscape evolution, but really includes this as a speculative wrapper that is not substantiated. I suggest that instead you propose testable hypotheses surrounding your findings.

2. Related to #1, much of the text is a litany of "temperature was X here ...". I find such statements of results useful only insofar as they expand upon a figure (associated with a supplementary data set) that presents the results. While these sections are written clearly, I would suggest that the authors focus on a set of geomorphic questions (if this be their motivation) and how the model-data set informs those questions.

3. Many of the discussions of model results are of ice-covered regions, yet no consideration of direct glacial erosion is given. Furthermore, no reference to the changes in the statistics of discharge or catchment area in ice-covered regions is given. This seems a disservice to this relatively high-resolution paleoclimate AGCM: the geologic setting *must* be considered, otherwise it seems that the authors' pushing on the modeling end has not been matched by a simple geological history sanity check. I would suggest that either significantly glaciated regions and the catchments that they feed be masked out, or that glacial erosion and its associated processes be included in the discussion.

4. (Discussed only here): You have not compared your models against any data. I understand that this may be simply a modeling exercise that you do compare to other models. However, I think that such a comparison could assuage skepticism about your results and lend support to your case, especially if you include it as part of a local case study (see the third point below). The core of these three points is that, with a bit more care, I think your results could say something really useful to the geomorphic community. Currently, the paper seems to be more a statement of, "this is important to geomorphology", followed by a long list of the model results. I challenge the authors to demonstrate (rather than simply stating) the importance of their work to geomorphology in a way that includes how it may impact the way scientists view Quaternary landscape evolution. Ideas include:

• Changes in means (done)

• Changes in statistical distributions of temperature and precipitation – think extreme events, frost-cracking window, etc.

• A focus on a few iconic regions while *explicitly* ignoring significantly ice-covered domains (I think this would be easiest, though obviously would be thrilled if you decided to tackle glacial processes)

• Using this focus to build a template for how to use paleoclimate GCM outputs to advance the field of geomorphology. Currently, I think that the work is acceptable following changes for internal consistency and geological accuracy (see #1 and #3), but I think that you could be selling yourselves short if you don't dig just a tiny bit deeper to investigate your forcings and their impact on geomorphology. I hope that you find these comments helpful in continuing to craft an insightful piece of work out of what seems to be a strong modeling approach.

We thank Prof. Andrew Wickert for his highly valuable review of our manuscript. Many important points were raised in the review and we hope that our appreciation for the input is sufficiently reflected in the revisions we made in response to it. We also encourage him to see our response to the second reviewer, where we provide additional geologic relevance of this study by now including a comparison of available terrestrial proxy data to our model results. We also explain throughout our response to the 2nd reviewer why an application of the predicted climate change to predict denudation rate changes is a large undertaking that can not be meaningfully conducted in this paper, but warrants more detailed applications of the models to individual areas (a topic of ongoing work/application for us). We refrain from using the model predicted runoff in the global GCM (even though it's conducted at relatively high resolution compared to previous work) to calculate changes in fluvial incision. This would be better done by mapping the predicted precipitation changes onto higher resolution (<90 m) DEMs and solving the kinematic wave equation for each fluvial erosion in each catchment, for the changes in precipitation. However, as we repeatedly mention above, this is not possible to include in this manuscript without first characterising how the precipitation has changed in each region (the current manuscript goals). Work in progress we are conducting is trying to apply the kinematic wave equation and palaeoprecipitation to selected areas, but it's proving difficult to implement meaningfully without temporally continuous (e.g. LGM to present) simulations of precipitation change. We hope this brings to the reviewers attention the complications associated with doing full erosion history calculations based on these results. We have expanded the last paragraph in the instruction to convey the above perspective better, and more clearly articulate (and justify) the scope and limitations of the manuscript.

We appreciate the importance of addressing specific sets of geomorphic questions and hypotheses (1 & 2) and we are currently taking an in-depth look at quantifying the potential for erosion by a variety of processes. These include different methods of quantifying frost cracking intensity and extreme precipitation events and how these changed over time. However, in order to include those in this manuscript, we fear that would have to seriously compromise the thoroughness with which we investigate these questions at the moment. Instead, we hope that we can convey the usefulness of our consistently set up palaeoclimate simulations as a framework for addressing any of these particular questions in detail, and have modified sections of this manuscript accordingly. This includes, but is not restricted to, extensive compilations of proxy-based precipitation reconstructions for our two larger study sites (South Asia and western South America) and comparison of this data to our model output. With this, we hope that we were also able to address the concerns raised in point 4. In order to address the important point raised about glaciated areas (3), we added an ice cover layer on all of our difference plots, included global maps of ice extent (as used for our simulations) in the supplementary material, and discussed where the large differences in temperature and precipitation we highlight in the manuscript are accompanied by changes in ice cover. Thus we hope to prevent that interpretations of the implications of our results are made without consideration of changes in ice cover (and consequently shifts in the process domain).

Line-by-line:
23. US Pacific Northwest Pacific → drop second "Pacific"
This has been corrected. Thank you for catching that.

29. future observational studies interested in quantifying → future observational studies that quantify (studies can't be interested in things, strictly speaking)
That is right of course. It has been corrected as suggested.

53. orogen scale → orogen-scale
It has been corrected as suggested.

∼57. A couple of recent studies from the climate science community shed light on the impacts of the Andes (first ref below) and continents in general (second ref below). In case these are interesting to you, I'm pasting the bibliographic information here:
Maroon, E. A., D. M. W. Frierson, and D. S. Battisti (2015), The tropical precipitation response to Andes topography and ocean heat fluxes in an aquaplanet model, J. Clim., 28(1), 381–398, doi:10.1175/JCLI-D-14-00188.1.
Maroon, E. A., D. M. W. Frierson, S. M. Kang, and J. Scheff (2016), The precip itation response to an idealized subtropical continent, J. Clim., 29(12), 4543–4564, doi:10.1175/JCLI-D-15-0616.1.
Many thanks for the references. These are indeed of much interest to us and we included them in the revised manuscript.

73-75. "Furthermore, recent controversy exists concerning the spatial and temporal scales over which geologic and geochemical observations can record climate-driven changes in weathering and erosion [e.g. Whipple, 2009; von Blanckenburg et al., 2015; Braun, 2016].": I see that you do not return to this point later, so could you describe the controversy for those who are not familiar with it?
Thank you for pointing this out. We described the controversy briefly for those unfamiliar with it after the sentence quoted above.

81. I see that later you discuss a little about what an AOGCM may do, but I will be looking for justification

about how an AGCM may suffice. Is this in part because you prescribe the b.c.'s and you are running it for 17 years only? If so, could you discuss potential systematic variations between this and an AOGCM?
We prescribe sea surface temperature reconstructions (SSTs) boundary conditions, which allows us to bypass the computationally expensive coupled simulations. Because these are fixed climatologies (though with seasonality preserved), the simulation of fewer years suffices. As a consequence, however, we do not expect to see decadal scale variability as we would in case of coupled models or prescribed SSTs that vary from year to year (such as present day simulations using AMIP SST's). We discuss this in the revised manuscript.

89. "PLIO to the Last Glacial Maximum": as you include no time-slices between these, I suggest making these part of the list and dropping the "to the".
This has been corrected as suggested.

147. "This section describes the clustering method used in this study." You could drop this sentence – the section title should be enough for even an inattentive reader!
We followed this suggestion and dropped that sentence.

176-178. I was wondeirng how you picked the number of clusters: I am glad to see that you performed a thorough search.
Thank you. We added some text explaining this. We systematically increased the number of clusters from 3 to 10 and assessed the distinctiveness or similarities of resulting climate clusters. Once the increase in the number of cluster no longer resulted in the addition of another cluster that was distinctly different from the others, we used this as a cut off point and used the cluster number of the previous iteration as the optimal cluster number.

Section 3: Much of this is information that I find better communicated through figures than with text. It is clearly written, however, and I am reluctant to suggest a rewrite for brevity in a length-unconstrained journal so long as the text can be co-located with the figure.
190-192. I see you have another "This section describes..." sentence. If this is your preferred way to write, you may keep it; here, the second sentence is not such a good topic-sentence replacement.
Thank you. We kept the sentence in this instance as it also immediately draws attention to the relevant figures, which may also serve in addressing the previous point you raised.

197-198. i.e. over the ice sheets. (This applies to other regions as well, and should be important to point out if you are going to then discuss fluvial processes in orogens)
Many thanks for pointing this out. We are more mindful of this in the revised discussion.

203-paragraph: Also because of local ice loss, presumably. So I think that the two prior paragraphs could have a new summary that "The greatest changes in temperature is observed where the greatest change in local ice extent occurs."
Thank you. We followed your suggestion.

214-215. Have you considered discussions of the African Humid Period?
We had not considered discussion of precipitation changes in North Africa, since it lies outside the regions we focus on. However, we appreciate that Holocene precipitation changes in the region are important and may be of interest to many readers. We therefore included a short discussion of Holocene precipitation changes in Northern Africa in the revised manuscript.

373-374. If you are looking at the influence of temperature and precipitation on erosion, and you are not including subglacial erosion, then your preceding text must indicate where your changes really are indicative of ice extent – both as a separate process domain and as a driver of fluvial processes and potential changes in the statistics of river discharge.
Thank you. We are more mindful of this in our discussion.

Section 4.1. Your first paragraph (weathering) differs from the content (comparing your model results with those published). These should be in different subsections, and the weathering paragraph may need to be expanded. Your "weathering and erosion" paragraph also neglects direct effects of glaciers, ice caps, and ice sheets, which were globallly significant.

Thank you for this suggestion. We re-structured this section as suggested and took ice extent into consideration in the erosion section.

Section 4.2. Once again, your discussion is often of formerly (or currently) ice-covered regions without explicitly acknowledging that this is a different process domain. In addition, as with the previous section, the body paragraphs are mostly about model comparison and regional changes with sparse link to the landscape-evolution factors indicated in the topic paragraph.
As above, we restructured this section, took ice extent into account and chose a more fitting section title.

416-423. Please discuss the direct influence of glaciers on the erosion orogens in the context of changing precipitation (and therefore mass balance). Is it significant or not?
Although it is challenging to sufficiently quantify changes in glacier-related erosion due to differences in precipitation, we now include this point in our discussion.

433-434. "Coastal North America"? Doesn't look like it: seems to be most of NA south of the ice margin.
That is correct. We revised our descriptions accordingly.

Section 4.3. The authors describe the results here, but I find the connection to erosion rates to be insufficiently described compared to how they are highlighted in the topic sentence, as well as in the abstract. I would like you to go one step beyond "ought to be considered" and actually posit how you expect the erosion rates – and therefore, the balance between erosion and exhumation and perhaps the equilibrium shapes of the mountains and their rivers – to vary. Otherwise, you are suggesting future work rather than actually describing the possible geomorphic significance – and I think underutilizing your results in a paper that is clearly targeted towards geomorphologists.
As described in our response above, we believe that trying to address specific problems such as these or quantifying how differences would be expressed as erosion rates would be beyond the scope of this manuscript and come at the cost of not being able to address these as thoroughly as we are currently attempting in other ongoing work.

498. "which may favour frost driven weathering during glacial climate states" – the St. Elias range was covered by glaciers! Yes, there can be some frost-cracking around the ice, but don't you think this is important too? http://instaar.colorado.edu/groups/QGISL/ak_paleoglacier_atlas/gallery/index.html
Thank you for drawing our attention to this. We consider this in our discussion now and revised the manuscript accordingly.

508. "enhanced sediment production driven by frost processes" – same as above. Glaciers were there. Consider them.
As for the comment above, we also considered glaciers here in the revised manuscript.

Conclusions: Comparison to other models: is this match surprising or no? Did you (mostly) use the same inputs and simply increase the grid resolution? If so, could you comment on how the improved grid and possible variations in inputs and use of them ocean as a boundary condition may have affected (or not) your results as compared to those of earlier studies? This would be more useful to include in the discussion than a simple list of "Our temperature in place Y was T0 , and X et al. wrote that they found it was T1 , which is close to T0 . Think big-picture, in both process and numerics!
Due to model-specific parameterisation, deviation is possible. In the revised manuscript, we comment on this as well as on the model resolution and implications of using ocean as boundary conditions instead of an ocean model.

533. Did your 8-10 degC changes occur significantly over areas that would be affected by hillslope or fluvial processes? (i.e. unglaciated areas?)
Some unglaciated areas experience large differences in temperature, but the maxima of 8-10 degC geographically coincide with ice cover changes. We acknowledge and discuss this in the revised manuscript.

# Revised Manuscript

# Where is Late Cenozoic climate change most likely to impact denudation?

Sebastian G. Mutz[1], Todd A. Ehlers[1], Martin Werner[2], Gerrit Lohmann[2], Christian Stepanek[2], Jingmin Li[1,3]

[1] Department of Geosciences, University Tübingen, D-72074 Tübingen, Germany

[2] Department of Paleoclimate Dynamics, Alfred Wegener Institute, Helmholtz Centre for Polar and Marine Research, D-27570 Bremerhaven, Germany

[3] now at Institute for Geography and Geology, University of Würzburg, Würzburg, D-97074 Germany

*Correspondence to*: Sebastian G. Mutz (sebastian.mutz@uni-tuebingen.de)

**Abstract**

The denudation history of active orogens is often interpreted in the context of modern climate and vegetation gradients. Here we address the validity of this approach and ask the question: what are the spatial and temporal variations in palaeo~~-~~climate for a latitudinally diverse range of active orogens? We do this using high-resolution (T159, ca. 80 x 80 km at the equator) palaeo~~-~~climate 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 ~~are observed~~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 ~~Pacific~~. 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 changes in precipitation and temperature occur and offer the highest potential for future observational studies ~~interested in quantifying~~that quantify the impact of climate change on denudation and weathering rates.

**Keywords:** Cenozoic climate, ECHAM5, Last Glacial Maximum, Mid-Holocene, Pliocene, cluster analysis, Himalaya, Tibet, Andes, Alaska, Cascadia

## 1. Introduction

Interpretation of orogen denudation histories in the context of climate and tectonic interactions is often hampered by a paucity of terrestrial palaeo-climate proxy data needed to reconstruct spatial variations in palaeo-climate. While it is self-evident that palaeoclimate changes could influence palaeodenudation rates, it is not always self-evident what the magnitude of climate change over different geologic time scales is, or what geographic locations offer the greatest potential to investigate palaeoclimate impacts on denudation. Palaeoclimate reconstructions are particularly beneficial when denudation rates are determined using geo- and thermo-chronology techniques that integrate over timescales of $10^3$-$10^{6+}$ years (e.g. cosmogenic radionuclides or low-temperature thermochronology) [e.g., Kirchner et al., 2001; Schaller et al., 2002; Bookhagen et al., 2005; Moon et al., 2011; Thiede and Ehlers, 2013; Lease and Ehlers, 2013]. However, few studies using denudation rate determination methods that integrate over longer timescales have access to information about past climate conditions that could influence these palaeo-denudation rates. Palaeo-climate modelling offers an alternative approach to sparsely available proxy data for understanding the spatial and temporal variations in precipitation and temperature in response to changes in orography [e.g. Takahashi and Battisti, 2007a, b; Insel et al., 2010; Feng et al., 2013] and global climate change events [e.g. Salzmann, 2011; Jeffery et al., 2013]. In this study, we characteri~~z~~se the climate at different times in the Late Cenozoic, and the magnitude of climate change for a range of active orogens. Our emphasis is on identifying changes in climate parameters relevant to weathering and catchment denudation to illustrate the potential importance of various global climate change events on surface processes.

Previous studies of orogen-scale climate change provide insight into how different tectonic or global climate change events influence regional climate change. For example, sensitivity experiments demonstrated significant changes in regional and global climate in response to landmass distribution and topography of the Andes, including changes in moisture transport, the north-south asymmetry of the Intertropical Convergence Zone ~~and the north-south asymmetry of the Inter Tropical Convergence Zone~~ [e.g. Takahashi and Battisti, 2007a, ; Insel et al., 2010] and (tropical) precipitation [Maroon et al., 2015, ; ~~Maroon et al.~~ 2016]. Another example is the regional and global climate changes induced by the Tibetan Plateau surface uplift due to its role as a ~~cold-temperature island and~~ physical obstacle to circulation [Raymo and Ruddiman, 1992; Kutzbach et al., 1993; Thomas, 1997; Bohner, 2006; Molnar et al., 2010; Boos and Kuang, 2010]. The role of tectonic uplift in long term regional and global climate change remains a focus of research and continues to be assessed with geologic datasets [e.g. ~~Zhisheng, 2001;~~ Dettman et al., 2003; Caves et al., 2017; Kent-Corson et al., 2006; Lechler et al., 2013; Lechler and Niemi, 2011; Licht et al., 2016; Methner et al., 2016; Mulch et al., 2015, 2008; Pingel et al., 2016] and climate modelling [e.g. Kutzbach et al., 1989; Kutzbach et al., 1993; Zhisheng, 2001;~~;~~ Bohner, 2006; Takahashi and Battisti, 2007a; Ehlers and Poulsen, 2009; Insel et al., 2010; Boos and Kuang, 2010]. Conversely, climate influences tectonic processes through erosion [e.g. Molnar and England, 1990; Whipple et al., 1999; Montgomery et al., 2001; Willett et al., 2006; Whipple, 2009]. Quaternary climate change between glacial and interglacial conditions [e.g. Braconnot et al., 2007; Harrison et al., 2013] resulted in not only the growth and decay of glaciers and glacial erosion [e.g. Yanites and Ehlers, 2012; Herman et al., 2013; Valla et al., 2011] but also global changes in precipitation and temperature [e.g. Otto-Bliesner et al., 2006; Li et al., 2017] that could influence catchment denudation in non-glaciated environments [e.g. Schaller and Ehlers, 2006; Glotzbach et al., 2013; Marshall et al., 2015]. These dynamics highlight the importance of investigating how much climate has changed over orogens that are the focus of studies of climate-tectonic interactions and their impact on erosion.

Despite recognition by previous studies that climate change events relevant to orogen denudation are prevalent throughout the Late Cenozoic, few studies have critically evaluated how different climate change events may, or may not, have affected the orogen climatology, weathering and erosion. Furthermore, recent controversy exists concerning the spatial and temporal scales over which geologic and geochemical observations can record climate-driven changes in weathering and erosion [e.g. Whipple, 2009; von Blanckenburg et al., 2015; Braun, 2016]. For example, the previous studies highlight that although palaeoclimate impacts on denudation rates are evident in some regions and measurable with some approaches, they are not always present (or detectable) and the spatial and temporal scale of climate change influences our ability to record climate sensitive denudation histories. This study contributes to our understanding of the interactions between climate, weathering, and erosion by bridging the gap between the palaeoclimatology and surface processes communities by documenting the magnitude and distribution of climate change over tectonically active orogens. ~~Our focus is on documenting the magnitude of paleoclimateclimate and climate change in different locations with the intent of informing past and ongoing paleodenudation studies of these regions. The application of these results to predicted changes in denudation rates is beyond the scope of this study and the focus of future work.~~

~~We~~ In this study, we employ the ECHAM5 global Atmospheric General Circulation Model and document climate and climate change for time slices ranging between the Pliocene (PLIO, ~3 Ma) to pre-industrial (PI) times for the St. Elias Range of South East Alaska, the US Pacific Northwest (Olympic and Cascade Range), western South America (Andes) and South Asia (incl. parts of Central- and East Asia). Our approach is two-fold and includes:

1. An empirical characteri~~z~~sation of palaeo-climates in these regions based on the covariance and spatial clustering of monthly precipitation and temperature, the monthly change in precipitation and temperature magnitude, and wind speeds where appropriate.

2. Identification of changes in annual mean precipitation and temperature in selected regions ~~over in the following time, specifically from the~~ for four time periods: (PLIO, ~~to the~~ Last Glacial Maximum (LGM), the Mid-Holocene (MH) and PI)~~.~~ and subsequent validation of the simulated precipitation changes for MH and LGM.

Our focus is on documenting climate and climate change in different locations with the intent of informing past and ongoing palaeodenudation studies of these regions. The results presented here also provide a means for future work to formulate testable hypotheses and investigations into whether or not regions of large palaeoclimate change produced a measurable signal in denudation rates. In this study, we intentionally refrain from applying predicted palaeoclimate changes to predict denudation rate changes. Such a prediction is beyond the scope of this study because a convincing (and meaningful) calculation of climate-driven transients in fluvial erosion (e.g. via the kinematic wave equation), variations in frost cracking intensity, or changes in hillslope sediment production and transport at the large regional scales considered here is not tractable within a single manuscript, and instead is the focus of our ongoing work. Instead, our emphasis lies on addressing the first question we are confronted with in our own research into denudation rate studies around world, namely - where is Late Cenozoic climate change most likely to impact denudation?

## 2. Methods: Climate modelling and cluster analyses for climate characteris~~z~~ation

### 2.1 ECHAM5 simulations

The global Atmospheric General Circulation Model ECHAM5 [Roeckner et al., 2003] has been developed at the Max Planck Institute for Meteorology and is based on the spectral weather forecast model of the ECMWF [Simmons et al., 1989]. In the context of palaeoclimate applications, the model has been used mostly at lower resolution (T31, ~~approximately~~ca. 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 Poulsen [2016]) . The performed studies are not limited to the last millenium [e.g. Jungclaus et al., 2010] but also include research in the field of both warmer and colder climates, at orbital [e.g. Gong et al., 2013; Lohmann et al., 2013; Pfeiffer and Lohmann, 2016; Zhang et al., 2013a; Zhang et al., 2014; Wei and Lohmann, 2012] and tectonic time scales [e.g. Knorr et al., 2011; Stepanek and Lohmann, 2012], and under anthropogenic influence [Gierz et al., 2015].

Here, the ECHAM5 simulations were conducted at a T159 spatial resolution (horizontal grid size ca. 80 km x 80 km at the equator) with 31 vertical levels (between the surface and 10hPa). This high model resolution is admittedly not required for all of the climatological questions investigated in this study, and it should be noted that the skill of GCM's in predicting orographic precipitation remains limited at this scale [e.g. Meehl et al. 2007]. However, simulations were conducted at this resolution so that future work can apply the results in combination with different dynamical and statistical downscaling methods to quantify changes at large catchment to orogen scales. The output frequency is relatively high (1 day) to enhance the usefulness of our simulations as input for landscape evolution and other models that may benefit from daily input. The simulations were conducted for five different time periods: present-day (PD), PI, MH, LGM and PLIO.

A PD simulation (not shown here) was used to establish confidence in the model performance before conducting palaeo-simulations and has been compared with the following observation-based datasets: European Centre for Medium-Range Weather Forecasts (ECMWF) re-analyses [ERA40, Uppala et al., 2005], National Centers for Environmental Prediction and National Center for Atmospheric Research (NCEP/NCAR) re-analyses [Kalnay et al., 1996; Kistler et al., 2001], NCEP Regional Reanalysis (NARR) [Mesinger et al., 2006], the Climate Research Unit (CRU) TS3.21 dataset [Harris et al., 2013], High Asia Refined Analysis (HAR30) [Maussion et al., 2014] and the University of Delaware dataset (UDEL v3.01) [Legates et al., 1990]. (See Mutz et al. [2016] for a detailed comparison with a lower resolution model).

The PI climate simulation is an ECHAM5 experiment with PI (reference year 1850) boundary conditions. Sea Surface Temperatures (SST) and Sea Ice Concentration (SIC) are derived from transient coupled ocean-atmosphere simulations [Lorenz and Lohmann, 2004; Dietrich et al., 2013]. Following Dietrich et al. [2013], greenhouse gas (GHG) concentrations (CO2: 280 ppm) are taken from ice core based reconstructions of $CO_2$ [Etheridge et al., 1996], $CH_4$ [Etheridge et al., 1998] and N2O [Sowers et al., 2003]. Sea surface boundary conditions for MH originate from a transient, low-resolution, coupled atmosphere-ocean simulation of the mid (6 ka) Holocene [Wei and Lohmann, 2012; Lohmann et al, 2013], where the GHG concentrations (CO2: 280 ppm) are taken from ice core reconstructions of GHG's by Etheridge et al. [1996], Etheridge et al. [1998] and Sowers et al. [2003]. GHG's concentrations for the LGM (CO2: 185 ppm) have been prescribed following Otto-Bliesner et al. [2006]. Orbital parameters for MH and LGM are set according to Dietrich et al. [2013] and Otto-Bliesner et al. [2006], respectively. LGM land-sea distribution and ice sheet extent and thickness are set based on the PMIP III (Palaeoclimate Modelling Intercomparison Project, phase 3) guidelines (elaborated on by Abe-Ouchi et al [2015]). Following Schäfer-Neth and Paul [2003], SST and SIC for the LGM are based on GLAMAP [Sarnthein et al. 2003] and CLIMAP [CLIMAP project members, 1981] reconstructions for the ~~for the~~ Atlantic and Pacific/Indian Ocean, respectively. Global MH and LGM vegetation are based on maps of plant functional types by the BIOME 6000 / Palaeovegetation Mapping Project [Prentice et al., 2000; Harrison et al., 2001; Bigelow et al., 2003; Pickett et al., 2004] and model predictions by Arnold et al. [2009]. Boundary conditions for the PLIO simulation, including GHG concentrations (CO2: 405), orbital parameters and surface conditions (SST, SIC, sea land mask, topography and ice cover) are taken from the PRISM (Pliocene Research, Interpretation and Synoptic Mapping) project [Haywood et al., 2010; Sohl et al., 2009; Dowsett et al., 2010], specifically PRISM3D. The PLIO vegetation boundary condition was created by converting the PRISM vegetation reconstruction to the JSBACH plant functional types as described by Stepanek and Lohmann [2012], but the built-in land surface scheme was used.

SST reconstructions can be used as an interface between oceans and atmosphere [e.g. Li et al. 2016] instead of conducting the computationally more expensive fully coupled Atmosphere-Ocean GCM experiments. While the use of SST climatologies comes at the cost of capturing decadal-scale variability, and the results are ultimately biased towards the SST reconstructions the model is forced with, the simulated climate more quickly reaches an equilibrium state and the means of atmospheric variables used in this study do no change significantly after the relatively short spin-up period. The palaeoclimate simulations (PI, MH, LGM, PLIO) using ECHAM5 are therefore carried out for 17 model years, of which the first two years are used for model spin up. The monthly long-term averages (multi-year means for individual months) for precipitation, temperature, as well as precipitation and temperature amplitude, i.e. the mean difference between the hottest and coldest months, have been calculated from the following 15 model years for the analysis presented below.

For further comparison between the simulations, the investigated regions were subdivided (Fig. 1). Western South America was subdivided into four regions: parts of tropical South America (80°-60° W, 23.5-5° S), temperate 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 Peruvian Andes, Bolivian Andes and northernmost Chilean Andes, and temperate Andes (80°-60° W, 50°-23.5° S, high-pass filtered). South Asia was subdivided into three regions: tropical South Asia (40°-120°E, 0°-23.5°N), temperate South Asia (40°-120°E, 23.5°-60°N), and high altitude South Asia (40°-120°E, 0°-60°N; high-pass filtered).

Our approach of using a single GCM (ECHAM5) for our analysis is motivated by, and differs from, previous studies where inter-model variability exists from the use of different GCMs due to different parameterisations in each model. The variability in previous inter-model GCM comparisons exists despite the use of the same forcings [e.g. see results highlighted in IPCC AR5]. Similarities identified between these palaeoclimate simulations conducted with different GCMs using similar boundary conditions can establish confidence in the models when in agreement with proxy reconstructions. However, differences identified in inter-model GCM comparisons highlight biases by all or specific GCMs, or reveal sensitivities to one changed parameter, such as model resolution. Given these limitations of GCM modelling, we present in this study a comparison of a suite of ECHAM5 simulations to proxy-based reconstructions (where possible) and, to a lesser degree, comment on general agreement or disagreement of our ECHAM5 results with other modelling studies. A detailed inter-model comparison of our results with other GCMs is beyond the scope of this study, and better suited for a different study in a journal with a different focus and audience.

Rather, by using the same GCM and identical resolution for the time slice experiments, we reduce the number of parameters (or model parameterisations) varying between simulations and thereby remove potential sources of error or uncertainty that would otherwise have to be considered when comparing output from different models with different parameterisations of processes, model resolution, and in some cases model forcings (boundary conditions). Nevertheless, the reader is advised to use these model results with the GCM's shortcoming and uncertainties in boundary condition reconstructions in mind. For example, precipitation results may require dynamical or statistical downscaling to increase accuracy where higher resolution precipitation fields are required. Furthermore, readers are advised to familiarise themselves with the palaeogeography reconstruction initiatives and associated uncertainties. For example, while Pliocene ice sheet volume can be estimated, big uncertainties pertaining to their locations remain [Haywood et al. 2010].

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

~~This section describes the clustering method used in this study.~~ The aim of the clustering approach is to group climate model surface grid boxes together based on similarities in climate. Cluster analyses are statistical tools that allow elements (i) to be grouped by similarities in the elements' attributes. In this study, those elements are spatial units, the elements' attributes are values from different climatic variables, and the measure of similarity is given by a statistical distance. The four basic variables used as climatic attributes of these spatial elements are: near-surface (2m) air temperature, seasonal 2m air temperature amplitude, precipitation rate, and seasonal precipitation rate amplitude. Since monsoonal winds are a dominant feature of the climate in the South Asia region, near surface (10m) speeds of u-wind and v-wind (zonal and meridional wind components, respectively) during the monsoon season (July) and outside the monsoon season (January) are included as additional variables in our analysis of that region. Similarly, u-wind and v-wind speeds during (January) and outside (July) the monsoon season in South America are added to the list of considered variables to take into account the South American Monsoon System (SASM) in the cluster analysis for this region. The long-term monthly means of those variables are used in a hierarchical clustering method, followed by a non-hierarchical k-means correction with randomis~~z~~ed re-groupment [Mutz et al., 2016; Wilks, 2011; Paeth, 2004; Bahrenberg et al., 1992].

The hierarchical part of the clustering procedure starts with as many clusters as there are elements (ni), then iteratively combines the most similar clusters to form a new cluster using centroids for the linkage procedure for clusters containing multiple elements. The procedure is continued until the desired number of clusters (k) is reached. One disadvantage of a pure hierarchical approach is that elements cannot be re-categoris~~z~~ed once they are assigned to a cluster, even though the addition of new elements to existing clusters changes the clusters' defining attributes and could warrant a re-categori~~z~~sation of elements. We address this problem by implementation of a (non-hierarchical) k-means clustering correction [e.g. Paeth, 2004]. Elements are re-categorizsed based on the multivariate centroids determined by the hierarchical cluster analysis in order to minimisze the sum of deviations from the cluster centroids. The Mahalanobis distance [e.g. Wilks, 2011] is used as a measure of similarity or distance between the cluster centroids, since it is a statistical distance and thus not sensitive to different variable units. The Mahalanobis distance also accounts for possible multi-collinearity between variables.

The end results of the cluster analyses are subdivisions of the climate in the investigated regions into *k* subdomains or clusters based on multiple climate variables. The region-specific *k* has to be prescribed before the analyses. A large *k* may result in redundant additional clusters describing very similar climates, thereby defeating the purpose of the analysis to identify and describe the dominant, distinctly different climates in the region and their geographical coverage. Since it is not possible to know a priori the ideal number of clusters, *k* was varied between 3 and 10 for each region and the results presented below identify the optimal number of visibly distinctly different clusters from the analysis. Optimal *k* was determined by assessing the distinctiveness and similarities between the climate clusters in the systematic process of increasing *k* from 3 to 10. Once an increase in *k* no longer resulted in the addition of another cluster that was climatologically distinctly different from the others, and instead resulted in at least two similar clusters, *k* of the previous iteration was chosen as the optimal *k* for the region.

The cluster analysis ultimately results in a description of the geographical extent of a climate (cluster) characterised by a certain combination of mean values for each of the variables associated with the climate. For example, climate cluster 1 may be the most tropical climate in a region and thus be characterised by a high precipitation values, high temperature values and low seasonal temperature amplitude. Each of the results (consisting of the geographical extent of climates and mean vectors describing the climate) can be viewed as an optimal classification for the specific region and time. It serves primarily as a means for providing an overview of the climate in each of the regions at different times, reduces dimensionality of the raw simulation output, and identify regions of climatic homogeneity that is difficult to notice by viewing simple maps of each climate variable. Its synoptic purpose is similar to that of the widely known Köppen-Geiger classification scheme [Peel et al., 2007], but we allow for optimal classification rather than prescribe classes, and our selection of variables is more restricted and made in accordance with the focus of this study.

## 3. Results

Results from our analysis are first presented for general changes in global temperature and precipitation for the different time slices (Fig. 2, 3), which is then followed by an analysis of changes in the climatology of selected orogens. A more detailed description of temperature and precipitation changes in our selected orogens is presented in subsequent subsections (Fig. 4 and following). All differences in climatology are expressed relative to the PI control run. Changes relative to the PI rather than PD conditions are presented to avoid interpreting an anthropogenic bias in the results and focusing instead on pre-anthropogenic variations in climate. For brevity, near-surface (2m) air temperature and total precipitation rate are referred to as temperature and precipitation.

### 3.1 Global differences in mean annual temperature

This section describes the differences between simulated MH, LGM, and PLIO annual mean temperature anomalies with respect to PI shown in Fig. 2b, and PI temperature absolute values shown in Fig. 2a. Most temperature differences between the PI and MH climate are within -1°C to 1°C. Exceptions to this are the Hudson Bay, Weddell Sea and Ross Sea regions which experience warming of 1-3°C, 1-5°C and 1-9°C respectively. Continental warming is mostly restricted to low-altitude South America, Finland, western Russia, the Arabian Peninsula (1-3°C) and subtropical north Africa (1-5°C). Simulation results show that LGM and PLIO annual mean temperature deviate from the PI means the most. The global PLIO warming and LGM cooling trends are mostly uniform in direction, but the magnitude varies regionally. The strongest LGM cooling is concentrated in regions where the greatest change in ice extent occurs (as indicated on Fig. 2), i.e. Canada, Greenland, the North Atlantic, Northern Europe and Antarctica. Central Alaska shows no temperature changes, whereas coastal South Alaska experiences cooling of ≤ 9°C. Cooling in the US Pacific northwest 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 the central Andes and ≤ 9°C in the south. Along the Himalayan orogen, LGM temperature values are 5-7°C below PI values. Much of central Asia and the Tibetan plateau cools by 3-5°C, and most of India, low-altitude China and southeast Asia by 1-3°C.

In the PLIO climate, parts of Antarctica, Greenland and the Greenland Sea experience the greatest temperature increase (≤ 19°C). Most of southern Alaska warms by 1-5°C and ≤ 9°C near McCarthy, Alaska. The US Pacific northwest warms by 1-5°C. The strongest warming in South America is concentrated at the Pacific west coast and the Andes (1-9°C), specifically between Lima and Chiclayo, and along the Chilean-Argentinian Andes south of Bolivia (≤ 9°C). Parts of low-altitude South America to the immediate east of the Andes experience cooling of 1-5°C. The Himalayan orogen warms by 3-9°C, whereas Myanmar, Bangladesh, Nepal, northern India and northeast Pakistan cool by 1-9°C.

### 3.2 Global differences in mean annual precipitation

Notable differences occur between simulated MH, LGM, PLIO annual mean precipitation anomalies with respect to PI shown in Fig. 3b, and the PI precipitation absolute values shown in Fig. 3a. Of these, MH precipitation deviates the least from PI values. The differences between MH and PI precipitation on land appear to be largest in northern tropical Africa (increase ≤1200 mm/a) and along the Himalayan orogen (increase ≤2000 mm/a) and in central Indian states (decrease) ≤500mm. The biggest differences in western South America are precipitation increases in central Chile between Santiago and Puerto Montt. The LGM climate shows the largest deviation in annual precipitation from the PI climate, and precipitation on land mostly decreases. Exceptions are increases in precipitation rates in North American coastal regions, especially in coastal South Alaska (≤2300 mm/a) and the US Pacific Northwest (≤1700 mm/a). Further exceptions are precipitation increases in low-altitude regions immediately east of the Peruvian Andes (≤1800 mm/a), central Bolivia (≤1000 mm/a), most of Chile (≤1000 mm/a) and northeast India (≤1900 mm/a). Regions of notable precipitation decrease are northern Brazil (≤1700 mm/a), southernmost Chile and Argentina (≤1900 mm/a), coastal south Peru (≤700 mm/a), central India (≤2300 mm/a) and Nepal (≤1600 mm/a).

Most of the precipitation on land in the PLIO climate is higher than those in the PI climate. Precipitation is enhanced by ca. 100-200 mm/a in most of the Atacama desert, by ≤1700 mm/a south of the Himalayan orogen and by ≤1400 mm/a in tropical South America. Precipitation significantly decreases in central Peru (≤2600mm), southernmost Chile (≤2600mm) and from eastern Nepal to northernmost northeast India (≤2500mm).

### 3.3 Palaeoclimate characteri~~z~~sation from the cluster analysis and changes in regional climatology

In addition to the above described global changes, the PLIO to PI regional climatology changes substantially in the four investigated regions of: South Asia (section 3.3.1), the Andes (section 3.3.2), South Alaska (section 3.3.3) and the Cascade Range (section 3.3.4). Each climate cluster defines separate distinct climate that is characterized by the mean values of the different climate variables used in the analysis. The clusters are calculated by taking the arithmetic means of all the values (climatic means) calculated for the grid boxes within each region. The regional climates are 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 for specific palaeo~~-~~climates are mentioned in the text as $C_{i(t)}$, where $i$ corresponds to the cluster number ($i$=1, …, $k$) and $t$ to the simulation time period ($t$=PI, MH, LGM, PLIO). The descriptions first highlight the similarities and then the differences in regional climate. The cluster means of seasonal near-surface temperature amplitude and seasonal precipitation amplitude are referred to as temperature and precipitation amplitude. The median, 25$^{th}$ percentile, 75$^{th}$ percentile, minimum and maximum values for annual mean precipitation are referred to as $P_{md}$, $P_{25}$, $P_{75}$, $P_{min}$ and $P_{max}$ respectively. Likewise, the same statistics for temperature are referred to as $T_{md,}$ $T_{25}$, $T_{75}$, $T_{min}$ and $T_{max}$. These are presented as boxplots of climate variables in different time periods. When the character of a climate cluster is described as “high”, “moderate” and “low”, the climatic attribute’s values are described relative to the value range of the specific region in time, thus high PLIO precipitation rates may be higher than high LGM precipitation rates. The character is presented a raster plots, to allow compact visual representation of it. The actual mean values for each variable in every time-slice and region-specific cluster are included in tables in the supplementary material.

#### 3.3.1 Climate change and palaeoclimate characteri~~z~~sation in South Asia, Central- and East Asia

This section describes the regional climatology of the four investigated Cenozoic time slices and how precipitation and temperature changes from PLIO to PI times in tropical, temperate and high altitude regions. LGM and PLIO simulations show the largest simulated temperature and precipitation deviations (Fig. 4b) from PI temperature and precipitation (Fig. 4a) in the South Asia region. LGM temperatures are 1-7°C below PI temperatures and the direction of deviation is uniform across the study region. PLIO temperature is mostly above PI temperatures by 1-7°C. The cooling of 3-5°C in the region immediately south of the Himalayan orogen represents one of the few exceptions. Deviations of MH precipitation from PI precipitation in the region are greatest along the eastern Himalayan orogeny, which experiences an increase in precipitation (≤2000 mm/a). The same region experiences a notable decrease in precipitation in the LGM simulation, which is consistent in direction with the prevailing precipitation trend on land during the LGM. PLIO precipitation on land is typically higher than PI precipitation.

Annual means of precipitation and temperature spatially averaged for the regional subdivisions and the different time slice simulations have been compared. The value range $P_{25}$ to $P_{75}$ of precipitation is higher for tropical South Asia 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 other time slice simulations, most visibly for tropical South Asia (ca. 100 mm/a). The temperature range (both $T_{75}$-$T_{25}$ 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 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 and 2°C below PI and MH temperatures in tropical, temperate and high altitude South Asia respectively, whereas the same temperature statistics for the PLIO simulation are ca. 1°C above PI and MH values in all regional subdivisions (Fig. 5 d-f). With respect to PI and MH values, precipitation and temperature are generally lower in the LGM and higher in the PLIO in tropical, temperate and high altitude South Asia.

In all time periods, the wettest climate cluster $C_1$ covers an area along the southeastern Himalayan orogen (Fig. 6 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 characterized by (dark blue, Fig. 6e-h) the highest temperatures, u-wind and v-wind speeds during the summer monsoon in their respective time periods, whereas $C_{4(PI)}$, $C_{5(MH)}$, and $C_{6(LGM)}$ are defined by low temperatures and highest temperature amplitude, u-wind and v-wind speeds outside the monsoon season (in January) in their respective time periods (Fig. 6 e-h). The latter 3 climate classes cover much of the more continental, northern landmass in their respective time periods and represents a cooler climate affected more by seasonal temperature fluctuations (Fig. 6 a-d). The two wettest climate clusters $C_1$ and $C_2$ are more restricted to the eastern end of the Himalayan orogen in the LGM than during other times, indicating that the LGM precipitation distribution over the South Asia landmass is more concentrated in this region than in other time slice experiments.

### 3.3.2 Climate change and palaeoclimate characteri~~z~~sation in the Andes, Western South America

This section describes the cluster analysis based regional climatology of the four investigated Late Cenozoic time slices and illustrates how precipitation and temperature changes from PLIO to PI in tropical and temperate low- and high altitude (i.e. Andes) regions in western South America (Fig. 7-9).

LGM and PLIO simulations show the largest simulated deviations (Fig. 7b) from PI temperature and precipitation (Fig. 7a) in western South America. The direction of LGM temperature deviations from PI temperatures is negative and uniform across the region. LGM temperatures are typically 1-3°C below PI temperatures across the region, and 1-7°C below PI values in the Peruvian Andes, which also experience the strongest and most widespread increase in precipitation during the LGM (≤1800 mm/a). Other regions, such as much of the northern Andes and tropical South America, experience a decrease of precipitation in the same experiment. PLIO temperature is mostly elevated above PI temperatures by 1-5°C. The Peruvian Andes experience a decrease in precipitation (≤2600mm), while the northern Andes are wetter in the PLIO simulation compared to the PI control simulation.

PI, MH, LGM and PLIO precipitation and temperature means for regional subdivisions have been compared. The $P_{25}$ to $P_{75}$ range is smallest for the relatively dry temperate Andes and largest for tropical South America and the 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 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 are ca. 50 mm/a lower in tropical South America and ca. 50 mm/a higher in the temperate Andes. Average PLIO temperatures are slightly warmer and LGM temperatures are slightly colder than PI and MH temperatures in tropical and temperate South America (Fig. 8 e and f). These differences are more pronounced in the Andes, however. $T_{md}$, $T_{25}$ and $T_{75}$ are ca. 5°C higher in the PLIO climate than in PI and MH climates in both temperate and tropical Andes, whereas the same temperatures for the LGM are ca. 2-4°C below PI and MH values (Fig. 8 g and h).

For the LGM, the model computes drier-than-PI conditions in tropical South America and tropical Andes, enhanced precipitation in the temperate Andes, and a decrease in temperature that is most pronounced in the Andes. For the PLIO, the model predicts precipitation similar to PI, but with lower precipitation maxima. PLIO temperatures generally increase from PI temperatures, and this increase is most pronounced in the Andes.

The climate variability in the region is described by six different clusters (Fig. 9 a-d), which have similar attributes in all time periods. The wettest climate $C_1$ is also defined by moderate to high precipitation amplitudes, low temperatures and moderate to high u-wind speeds in summer and winter in all time periods (dark blue, Fig. 9 e-h). $C_{2(PI)}$, $C_{2(MH)}$, $C_{3(LGM)}$ and $C_{2(PLIO)}$ are characterized by high temperatures and low seasonal temperature amplitude (dark blue, Fig. 9 e-h), geographically cover the north of the investigated region, and represent a more tropical climate. $C5_{(PI)}$, $C5_{(MH)}$, $C6_{(LGM)}$ and $C6_{(PLIO)}$ are defined by low precipitation and precipitation amplitude, high temperature amplitude and high u-wind speeds in winter (Fig. 9 e-h), cover the low-altitude south of the investigated region (Fig. 9 a-d) and represent dry, extra-tropical climates with more pronounced seasonality. In the PLIO simulation, the lower-altitude east of the region has four distinct climates, whereas the analysis for the other time slice experiments only yield three distinct climates for the same region.

### 3.3.3 Climate change and palaeoclimate characteris~~z~~ation in the St. Elias Range, Southeast Alaska

This section describes the changes in climate and the results from the cluster analysis for South Alaska (Fig. 10-12). As is the case for the other study areas, LGM and PLIO simulations show the largest simulated deviations (Fig. 10b) from PI temperature and precipitation (Fig. 10a). The sign of LGM temperature deviations from PI temperatures is negative and uniform across the region. LGM temperatures are typically 1-9°C below PI temperatures, with the east of the study area experiencing largest cooling. PLIO temperatures are typically 1-5°C above PI temperatures and the warming is uniform for the region. In comparison to the PI simulation, LGM precipitation is lower on land, but higher (≤2300mm) in much of the coastal regions of South Alaska. Annual PLIO precipitation is mostly higher (≤800mm) than for PI.

$P_{md}$, $P_{25}$, $P_{75}$, $P_{min}$ and $P_{max}$ for South Alaskan mean annual precipitation do not differ much between PI, MH and 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). The Alaskan PLIO climate is distinguished from the PI and MH climates by its higher (ca. 2°C) regional temperature 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 other considered time period (Fig. 11b), and about 3-5°C lower than during the PI and MH.

Distinct climates are present in the PLIO to PI simulations for Southeast Alaska. Climate cluster $C_1$ is always geographically restricted to coastal southeast Alaska (Fig. 12 a-d) and characterized by the highest precipitation, precipitation amplitude, temperature, and by relatively low temperature amplitude (dark blue, Fig. 12 e-h). Climate $C_2$ is characterized by moderate to low precipitation, precipitation amplitude, temperature, and by low temperature amplitude. $C_2$ is either restricted to coastal southeast Alaska (in MH and LGM climates) or coastal southern Alaska (in PI and PLIO climates). Climate $C_3$ is described by low precipitation, precipitation amplitude, temperature, and moderate temperature amplitude in all simulations. It covers coastal western Alaska and separates climate $C_1$ and $C_2$ from the northern $C_4$ climate. Climate $C_4$ is distinguished by the highest mean temperature amplitude, by low temperature and precipitation amplitude, and by lowest precipitation.

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 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 coastal southwest of Alaska. The main difference in characterization between PI and PLIO climates $C_1$- $C_4$ lies in the greater difference (towards lower values) in precipitation, precipitation amplitude and temperature from $C_{1(PLIO)}$ to $C_{2(PLIO)}$ compared to the relatively moderate decrease in those means from $C_{1(PI)}$ to $C_{2(PI)}$.

### 3.3.4 Climate change and palaeoclimate characteri~~z~~sation in the Cascade Range, US Pacific Northwest

This section describes the character of regional climatology in the US Pacific Northwest and its change over time (Fig. 13-15). The region experiences cooling of typically 9-11°C on land during the LGM, and warming of 1-5°C during the PLIO (Fig. 13b) when compared to PI temperatures (Fig. 13a). LGM precipitation increases over water, 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 Vancouver Island, around Portland and the Olympic Mountains, whereas PLIO precipitation does not deviate much from PI values over water and varies in the direction of deviation on land. MH temperature and precipitation deviation from PI values are negligible.

$P_{md}$, $P_{25}$, $P_{75}$, $P_{min}$ and $P_{max}$ for the Cascade Range do not notably differ between the four time periods (Fig. 14a). The LGM range of precipitation values is slightly larger than that of the PI and MH with slightly increased $P_{md}$, while 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 higher than those values for PI and MH (Fig. 14b). All temperature statistics for the LGM are notably (ca. 13°C) below their analogues in the other time periods (Fig. 14b).

PI, LGM and PLIO clusters are similar in both their geographical patterns (Fig. 15 a, c, d) and their characterization by mean values (Fig. 15 e, g, h). $C_1$ is the wettest cluster and shows the highest amplitude in precipitation. The common characteristics of the $C_2$ cluster are moderate to high precipitation and precipitation amplitude. $C_4$ is characterized by the lowest precipitation and precipitation amplitudes, and the highest temperature amplitudes. Regions assigned to clusters $C_1$ and $C_2$ are in proximity to the coast, whereas $C_4$ is geographically restricted to more continental settings.

In the PI and LGM climates, the wettest cluster $C_1$ is also characterized by high temperatures (Fig 10 e, g). 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 also described by moderate to high near surface temperature and temperature amplitude (Fig 10 f), and in that it is geographically less restricted and, covering much of Vancouver Island and the continental coastline north of it (Fig 10 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 10 f). $C_{2(MH)}$ is also geographically more restricted than $C_2$ clusters in PI, LGM and PLIO climates (Fig 10 a-d). $C_{2(PI)}$, $C_{2(MH)}$ and $C_{2(LGM)}$ have a low temperature amplitude (Fig 10 e-g), whereas $C_{2(PLIO)}$ is characterized by a moderate temperature amplitude (Fig 10 h).

## 4. Discussion

In the following, we synthesis~~z~~e our results and compare to previous studies that investigate the effects of temperature and precipitation change on erosion. Since our results do not warrant merited discussion of subglacial processes without additional work that is beyond the scope of this study, we instead advise caution in interpreting the presented precipitation and temperature results in an erosional context where the regions are covered with ice. For convenience, ice cover is indicated on figures 2,3,47,10 and 13, and a summary of ice cover used as boundary conditions for the different time slice experiments is included in the supplemental material. Where~~n~~ possible, we relate the magnitude of climate change predicted in each geographical study area with terrestrial proxy data.

### 4.1 Synthesis of temperature changes ~~and implications for weathering and erosion~~

#### 4.1.1 Temperature changes and implications for weathering and erosion

Changes in temperature can affect physical weathering due to temperature-induced changes in periglacial processes and promote frost cracking and frost creep [e.g., Matsuoka, 2001; Schaller et al., 2002; Matsuoka and Murton, 2008; Delunel et al., 2010; Andersen et al., 2015; Marshall et al., 2015], and also biotic weathering and erosion [e.g. Moulton et al., 1998; Banfield et al., 1999; Dietrich and Perron, 2006]. Quantifying and understanding past changes in temperature is thus vital for our understanding of denudation histories. In the following, we highlight regions in the world where future observational studies might be able to document significant warming or cooling that would influence temperature related changes in physical and chemical weathering over the last ~3 Ma.

Simulated MH temperatures show little deviation (typically < 1°C) from PI temperatures in the investigated regions (Fig. 2b), suggesting little difference in MH temperature-related weathering. The LGM experiences widespread cooling, which is accentuated at the poles. ~~LGM cooling is accentuated at the poles, in general agreement with studies such as Otto-Bliesner et al. [2006] and Braconnot et al. [2007], and~~ , increas~~es~~in the equator-to-pole pressure gradient and consequently strengthens global atmospheric circulation. Despite this global trend, cooling in coastal South Alaska is higher (≤ 9°C) than in central Alaska (0±1°C). The larger temperature difference in South Alaska geographically coincides with ice cover (Fig. 10b), and should thus be interpreted in context of a different erosional regime. Cooling in most of the lower-latitude regions in South America and central to southeast Asia is relatively mild. The greatest temperature differences in South America are observed for western Patagonia, which was mostly covered by glaciers. The Tibetan plateau experiences more cooling (3-5°C) than adjacent low-altitude regions (1-3°C) during the LGM.

The PLIO simulation is generally warmer, and temperature differences are ~~shows little to no warming in the tropics and~~ accentuated warming at the poles. ~~, as do findings of Salzmann et al. [2011] and Robinson [2009] and Ballantyne [2010] respectively. This would reduce the equator-to-pole sea and land surface temperature gradient, as also reported by Dowsett et al. [2010], and also weaken global atmospheric circulation. Agreement with proxy-based reconstructions, as is the case of the relatively little warming in lower latitudes, is not surprising given that sea surface temperature reconstructions are prescribed in this uncoupled atmosphere simulation. It should be noted that coupled ocean-atmosphere simulations do predict more low-latitude warming [e.g. Stepanek and Lohmann 2012; Zhang et al. 2013b].~~ Warming in simulation PLIO is ~~present~~greatest in parts of Canada, ~~and~~ Greenland and Antarcti~~t~~ca (up to

19°C), which geographically coincides with the presence of ice in the PI reference simulation and thus may be attributed to differences in ice cover. It should therefore also be regarded as areas in which process domain shifted from glacial to non-glacial. ~~and consistent with values based on multi-proxy studies [Ballantyne et al., 2010]. Due to a scarcity of paleo-botanical proxies in Antarctica, reconstruction-based temperature and ice-sheet extent estimates for a PLIO climate have high uncertainties [Salzmann et al., 2011], making model validation difficult. Furthermore, controversy about relatively little warming in the south polar regions compared to the north polar regions remains [e.g. Hillenbrand and Fütterer, 2002; Wilson et al., 2002]. Mid-latitude PLIO warming is mostly in the 1-3°C range with notable exceptions of cooling in the northern tropics of Africa and on the Indian subcontinent, especially south of the Himalayan orogen.~~ The warming in simulation PLIO in South Alaska and the US Pacific northwest is mostly uniform and in the range of 1-5°C. As before, changes in ice cover reveal that the greatest warming may be associated with the absence of glaciers relative to the PI simulation. ~~whereas w~~Warming in South America is concentrated at the Pacific west coast and the Andes between Lima and Chiclayo, and along the Chilean-Argentinian Andes south of Bolivia (≤ 9°C).

Overall, annual mean temperatures in the MH simulation show little deviation from PI values.  The more significant temperature deviations of the colder LGM and of the warmer PLIO simulations are accentuated at the poles leading to higher and lower equator-to-pole temperature gradients respectively. The largest temperature-related changes (relative to PI conditions) in weathering and subsequent erosion, in many cases through a shift in the process domain from glacial to non-glacial or vice versa, are therefore to be expected in the LGM and PLIO climates.

**4.1.2 Temperature comparison to other studies**

LGM cooling is accentuated at the poles, thus increases the equator-to-pole pressure gradient and consequently strengthens global atmospheric circulation, and is in general agreement with studies such as Otto-Bliesner et al. [2006] and Braconnot et al. [2007]. The PLIO simulation shows little to no warming in the tropics and accentuated warming at the poles, as do findings of Salzmann et al. [2011] and Robinson [2009] and Ballantyne [2010] respectively. This would reduce the equator-to-pole sea and land surface temperature gradient, as also reported by Dowsett et al. [2010], and also weaken global atmospheric circulation. Agreement with proxy-based reconstructions, as is the case of the relatively little warming in lower latitudes, is not surprising given that sea surface temperature reconstructions (derived from previous coarse resolution coupled ocean-atmosphere models) are prescribed in this uncoupled atmosphere simulation. It should be noted that coupled ocean-atmosphere simulations do predict more low-latitude warming [e.g. Stepanek and Lohmann 2012; Zhang et al. 2013b]. The PLIO warming in parts of Canada and Greenland (up to 19°C) and consistent with values based on multi-proxy studies [Ballantyne et al., 2010]. Due to a scarcity of palaeobotanical proxies in Antarctica, reconstruction-based temperature and ice-sheet extent estimates for a PLIO climate have high uncertainties

[Salzmann et al., 2011], making model validation difficult. Furthermore, controversy about relatively little warming in the south polar regions compared to the north polar regions remains [e.g. Hillenbrand and Fütterer, 2002; Wilson et al., 2002]. Mid-latitude PLIO warming is mostly in the 1-3°C range with notable exceptions of cooling in the northern tropics of Africa and on the Indian subcontinent, especially south of the Himalayan orogen.

## 4.2 Synthesis of precipitation changes ~~and implications for orogen denudation~~

### 4.2.1 Precipitation and implications for weathering and erosion

Changes in precipitation affects erosion through river incision, sediment transport, and erosion due to extreme precipitation events and storms [e.g. Whipple and Tucker, 1999; Hobley et al., 2010]. Furthermore, vegetation type and cover also co-evolve with variations in precipitation and with changes in geomorphology [e.g. Marston 2010; Roering et al., 2010]. These vegetation changes in turn modify hillslope erosion by increasing root mass and canopy cover, and decreasing water-induced erosion via surface runoff [e.g. Gyssels et al., 2005]. Therefore, understanding and quantifying changes in precipitation in different palaeo~~-~~climates is necessary for a more complete reconstruction of orogen denudation histories. A synthesis of predicted precipitation changes is provided below, and highlights regions where changes in river discharge and hillslope processes might be impacted by climate change over the last ~3 Ma.

Most of North Africa is notably wetter during the MH, which is characteristic of the African Humid Period [Sarnthein 1978]. This pluvial regional expression of the Holocene Climatic Optimum is attributed to sudden changes in the strength of the African monsoon caused by orbital-induced changes in summer insolation [e.g. deMenocal et al. 2000]. Southern Africa is characterised by a wetter climate to the east and drier climate to the west of the approximate location of the Congo Air Boundary (CAB), the migration of which has previously been cited as a cause for precipitation changes in East Africa [e.g. Juninger et al. 2014]. In contrast, simulated MH precipitation rates show little deviation from the PI in most of the investigated regions, suggesting little difference in MH precipitation-related erosion. The Himalayan orogen is an exception and shows a precipitation increase of ~~≤~~ up to 2000 mm/a. The climate's enhanced erosion potential, that could result from such a climatic change, should be taken into consideration when palaeo-erosion rates estimated from the geological record in this area are interpreted [e.g. Bookhagen et al., 2005]. Specifically, higher precipitation rates (along with differences in other rainfall-event parameters) could increase the probability of mass movement events on hillslopes, especially where hillslopes are close to the angle of failure [e.g. Montgomery, 2001], and modify fluxes to increase shear stresses exerted on river beds and increase stream capacity to enhance erosion on river beds (e.g. by abrasion).

Most precipitation on land is decreased during the LGM due to large-scale cooling and decreased evaporation over the tropics, resulting in an overall decrease in inland moisture transport [e.g., Braconnot et al. 2007]. ~~Coastal~~ North

America, south of the continental ice sheets, is an exception and experiences increases in precipitation. ~~, t~~For example, the investigated US Pacific Northwest and the southeastern coast of Alaska ~~are exceptions in that there is~~experience experience strongly enhanced precipitation of ≤1700 mm/a and ≤2300 mm/a, respectively. These changes geographically coincide with differences in ice extent. An increase in precipitation in these regions may have had direct consequences on the glaciers' mass balance and equilibrium line altitudes, where the glaciers' effectiveness in erosion is highest [e.g. Egholm et al., 2009; Yanites and Ehlers, 2012]. The differences in the direction of precipitation changes, and accompanying changes in ice cover ~~Reduced precipitation in other parts of southern Alaska result in a stronger south-to-north drying gradient than in the PI simulation. This could~~ would likely result in more regionally differentiated variations in precipitation-specific erosional processes in the St. Elias Range rather than causing systematic offsets for the LGM. Although precipitation is significantly reduced along much of the Himalayan orogen (≤1600 mm/a), ~~which is consistent with findings by, e.g., Braconnot et al. [2007]~~, northeast India experiences strongly enhanced precipitation (≤1900 mm/a). This could have large implications for studies of uplift and erosion at orogen syntaxes, where highly localized and extreme denudation has been documented [e.g.  Koons et al., 2013; Bendick and Ehlers, 2014].

Overall, the PLIO climate is wetter than the PI climate, in particular in the (northern) mid-latitudes, and possibly related to a northward shift of the northern Hadley cell boundary that is ultimately the result of a reduced equator-to-pole temperature gradient [e.g. Haywood et al. 2000, 2013; 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).~~ Most of the PLIO precipitation over land increases ~~during the PLIO. This finding agrees well with simulations performed at a lower spatial model resolution [cf. Stepanek and Lohmann, 2012]. PLIO precipitation significantly 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. Cohen 1990], the simulation warrants careful consideration of possible, geographically non-uniform offsets in precipitation in investigations of denudation and uplift histories.

MH precipitation and temperature in tropical, temperate and high-altitude South Asia is similar to PI precipitation and temperature, whereas LGM precipitation and temperatures are generally lower (by ca. 100 mm/a and 1-2°C respectively), possibly reducing precipitation-driven erosion and enhancing frost-driven erosion in areas pushed into a near-zero temperature range during the LGM.

*Andes, South America*

Clusters in South America (Fig. 9), which are somewhat reminiscent of the Köppen and Geiger classification [Kraus, 2001], remain mostly the same over the last 3 Ma. In the PLIO simulation, the lower-altitude east of the region is characterized by four distinct climates, which suggests enhanced latitudinal variability in the PLIO climate compared to PI with respect temperature and precipitation.

The largest temperature deviations from PI values are derived for the PLIO simulation in the (tropical and temperate) Andes, where temperatures exceed PI values by 5°C. On the other hand, LGM temperatures in the Andes are ca. 2-4°C below PI values in the same region (Fig 7 g and h). In the LGM simulation, tropical South America experiences ca. 50 mm/a less precipitation, the temperate Andes receive ca. 50 mm/a more precipitation than in PI and MH simulations. These latitude-specific differences in precipitation changes ought to be considered in attempts to reconstruct precipitation-specific palaeo~~-~~erosion rates in the Andes on top of longitudinal climate gradients highlighted by, e.g., Montgomery et al. [2001].

*St. Elias Range, South Alaska*

South Alaska is subdivided into two wetter and warmer clusters in the south, and two drier, colder clusters in the north. The latter are characteris~~z~~ed by increased seasonal temperature variability due to being located at higher latitudes (Fig. 12). The different equator-to-pole temperature gradients for LGM and PLIO may affect the intensity of the Pacific North American Teleconnection (PNA) [Barnston and Livzey, 1987], which has significant influence on temperatures and precipitation, especially in southeast Alaska, and may in turn result in changes in regional precipitation and temperature patterns and thus on glacier mass balance. Changes in the Pacific Decadal Oscillation, which is related to the PNA pattern, has previously been connected to differences in Late Holocene precipitation [Barron and Anderson, 2011]. While this climate cluster pattern appears to be a robust feature for the considered climate states, and hence over the recent geologic history, the LGM sets itself apart from PI and MH climates by generally lower precipitation (20-40 mm) and lower temperatures (3-5°C, Fig. 10, 11), which may favour frost driven weathering during glacial climate states [e.g. Andersen et al., 2015; Marshall et al. 2015] in unglaciated areas, whereas glacial processes would have dominated most of this region as it was covered by ice. Simulation PLIO is distinguished by temperatures that exceed PI and MH conditions by ca. 2°C, and by larger temperature and precipitation value ranges, possibly modifying temperature- and precipitation-dependent erosional processes in the region of South Alaska.

*Cascade Range, US Pacific Northwest*

In all time slices, the geographic climate patterns, based on the cluster analysis (Fig. 15), represents an increase in the degree of continentality from the wetter coastal climates to the further inland located climates with greater seasonal temperature amplitude and lower precipitation and precipitation amplitude (Fig 15 e-h). The most notable difference between the time slices is the strong cooling during the LGM, when temperatures are ca. 13°C (Fig. 13, 14) below those of other time periods~~,~~. Given that the entire investigated region was covered by ice (Fig 13), we can assume a shift to glacially dominated processes. ~~possibly leading to enhanced sediment production driven by frost processes, as proposed for parts of the Pacific Northwest by Marshall et al. [2015].~~

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

The predicted precipitation differences reported in this study were compared with observed (proxy record) palaeoprecipitation change. Proxy based precipitation reconstructions for the MH and LGM are presented for South Asia and South America for the purpose of assessing ECHAM5 model performance, and for identifying inconsistencies between neighbouring proxy data. Due to the repeated glaciations, detailed terrestrial proxy records for the time slices investigated here are not available, to the best of our knowledge, for the Alaskan and Pacific NW USA studies. Although marine records and records of glacier extent are available in these regions, the results from them do not explicitly provide estimates of wetter/drier, or colder/warmer conditions that can be spatially compared to the simulation estimates. For these two areas with no available records, the ECHAM5 predicted results therefore provide predictions from which future studies can formulate testable hypotheses to evaluate.

The palaeoclimate changes in terrestrial proxy records compiled here are reported as "wetter than today", "drier than today" or "the same as today" for each of the study locations, and plotted on top of the simulation-based difference maps as upward facing blue triangles, downward facing red triangles and grey circles respectively (Fig. 16, 17). The numbers listed next to those indicators are the ID numbers assigned to the studies compiled for this comparison and are associated with a citation provided in the figure captions.

In South Asia, 14/26 results from local studies agree with the model predicted precipitation changes for the MH. The model seems able to reproduce the predominantly wetter conditions on much of the Tibetan plateau, but predicts slightly drier conditions north of Chengdu, which is not reflected in local reconstructions. The modest mismatch between ECHAM5 predicted and proxy-based MH climate change in south Asia was also documented by Li et al., [2017], whose simulations were conducted at a coarser (T106) resolution. Despite these model-proxy differences, we note that there are significant discrepancies between the proxy data themselves in neighbouring locations in the MH, highlighting caution in relying solely upon these data for regional palaeoclimate reconstructions. These differences could result from either poor age-constraints in the reported values, or systematic errors in the transfer functions used to convert proxy measurements to palaeoclimate conditions. The widespread drier conditions on the Tibetan Plateau and immediately north of Laos are confirmed by 7/7 of the palaeoprecipitation reconstructions. 23/39 of the reconstructed precipitation changes agree with model predictions for South America during the MH. The model predicted wetter conditions in the central Atacama desert, as well as the drier conditions northwest of Santiago are confirmed by most of the reconstructions. The wetter conditions in southernmost Peru and the border to Bolivia and Chile cannot be confirmed by local studies. 11/17 of the precipitation reconstructions for the LGM are in agreement with model predictions. These include wetter conditions in most of Chile. The most notable disagreement can be seen in northeast Chile at the border to Argentina and Bolivia, where model predicted wetter conditions are not confirmed by reported reconstructions from local sites.

Model performance is, in general, higher for the LGM than for the MH and overall satisfactory given that it cannot be expected to resolve sub-grid scale differences in reported palaeoprecipitation reconstructions. However, as mentioned above, it should be noted that some locations (MH of south Asia, and MH of norther Chile) discrepancies exist between neighbouring proxy samples and highlight the need for caution in how these data are interpreted. Other potential sources of error resulting in disagreement of simulated and proxy-based precipitation estimates are the model's shortcomings in simulating orographic precipitation at higher resolutions, and uncertainties in palaeoclimate reconstructions at the local sites. In summary, although some differences are evident in both the model-proxy data comparison and between neighbouring proxy data themselves, the above comparison highlights an overall good agreement between the model and data for the south Asia and South American study areas. Thus, although future advances in GCM model parameterisations and new or improved palaeoclimate proxy techniques are likely, the palaeoclimate changes documented here are found to be in general robust and provide a useful framework for future studies investigating how these predicted changes in palaeoclimate impact denudation.

## 4.45 Conclusions

We present a statistical cluster-analysis-based description of the geographic coverage of possible distinct regional expressions of climates from four different time slices (Fig. 6, 9, 12, 15). These are determined with respect to a selection of variables that characterize the climate of the region and may be relevant to weathering and erosional processes. While the geographic distribution of climate ~~patterns~~ remains similar throughout time (as indicated by results of four different climate states representative for the climate of the last 3 Ma), results for the PLIO simulation suggests more climatic variability east of the Andes (with respect to near-surface temperature, seasonal temperature amplitude, precipitation, seasonal precipitation amplitude and seasonal u-wind and v-wind speeds). Furthermore, the wetter climates in the South Asia region retreat eastward along the Himalayan orogen for the LGM simulation, this is due to decreased precipitation along the western part of the orogen and enhanced precipitation on the eastern end, possibly signifying more localised high erosion rates.

Most global trends of the high-resolution LGM and PLIO simulations conducted here are in general agreement with previous studies [Otto-Bliesner et al., 2006; Braconnot et al., 2007; Wei and Lohmann, 2012; Lohmann et al., 2013; Zhang et al., 2013b, 2014; Stepanek and Lohmann, 2012]. The MH does not deviate notably from the PI, the LGM is relatively dry and cool, while the PLIO is comparably wet and warm. While the simulated regional changes in temperature and precipitation usually agree with the sign (or direction) of the simulated global changes, there are region-specific differences in the magnitude and direction. For example, the LGM precipitation of the Tropical Andes does not deviate significantly from PI precipitation, whereas LGM precipitation in the Temperate Andes is enhanced.

Comparisons to local, proxy-based reconstructions of MH and LGM precipitation in South Asia and South America reveal satisfactory performance of the model in simulating the reported differences. The model performs better for the LGM than the MH. We note however that compilations of proxy data such as we present here, also identify inconsistences between neighbouring proxy data themselves, warranting caution in the extent to which both proxy data and palaeoclimate models are interpreted for MH climate change in south Asia, and western South America.

The changes in regional climatology presented here are manifested, in part, by small to large magnitude changes in fluvial and hillslope relevant parameters such as precipitation and temperature. For the regions investigated here we find that precipitation differences between the PI, MH, LGM, and PLIO are in many areas around +/- 200-600 mm/yr, and locally can reach maximums of +/- 1000-2000 mm/yr (Figs. 4, 7, 10, 13). In areas where significant precipitation increases are accompanied by changes in ice extent, such as parts of southern Alaska during the LGM, we would expect a shift in the erosional regime to glacier dominated processes. Temperature differences between these same time periods are around 1-4 °C in many places, but reach maximum values of 8-10 °C. Many of these maxima in the temperature differences geographically coincide with changes in ice sheet extent and must therefore be interpreted as part of a different erosional process domains. However, we also observe large temperature differences (~5°C) in unglaciated areas that would be affected by hillslope, frost cracking, and fluvial processes. The magnitude of these differences are not trival, and will likely impact fluvial and hillslope erosion and sediment transport, as well as biotic and abiotic weathering. The regions of large magnitude changes in precipitation and temperature documented here (Figs. 4, 7, 10, 13) offer the highest potential for future observational studies interested in quantifying the impact of climate change on denudation and weathering rates.

## Acknowledgements

The model simulations presented in this study are freely available to interested persons by contacting S. Mutz or T. Ehlers. We note however that the data files are very large (~4 TB, and too large to archive in journal supplementary material) and require familiarity in reading/plotting NetCDF formatted files. ~~Support from~~ European Research Council (ERC) Consolidator Grant number 615703 provided support for S. Mutz. Additional support is acknowledged from the German science foundation (DFG) priority research program 1803 (EarthShape: Earth Surface Shaping by Biota; grants EH329/14-1 and EH329/17-1). We thank B. Adams and J. Starke for constructive discussions. The DKRZ is thanked for computer time used for some of the simulations presented here. C. Stepanek, M. Werner, and G. Lohmann acknowledge funding by the Helmholtz Climate Initiative Reklim and the Alfred Wegener Institute's research programme Marine, Coastal and Polar Systems.

**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 Cascadia.

**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-surface 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) temperate South Asia, and (f) high-altitude South Asia. For each time slice, the minimum, lower 25$^{th}$ percentile, median, upper 75$^{th}$ percentile and maximum are plotted.

**Figure 6** Geographical coverage and characterization of climate classes $C_1$- $C_6$ based on cluster-analysis of 8 variables (near surface temperature, seasonal near surface temperature amplitude, total precipitation, seasonal precipitation amplitude, u-wind in January and July, v-wind in January and July) in the South Asia region. The geographical coverage 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-slice specific characterization of $C_1$- $C_6$ for PI (e), MH (f), LGM (g) and PLIO (h) is shown on the right.

**Figure 7** PI annual mean near-surface temperatures (a), and deviations of MH, LGM and PLIO annual mean near-surface temperatures from PI values (b) for western South America. Insignificant ($p < 99\%$) differences (as determined by a t-test) are greyed out.

**Figure 8** PI, MH, LGM and PLIO annual mean precipitation in (a) tropical South America, (b) temperate South America, (c) tropical Andes, and (d) temperate Andes; PI, MH, LGM and PLIO annual mean temperatures in (e) tropical

South America, (f) temperate South America, (g) tropical Andes, and (h) temperate Andes. For each time slice, the minimum, lower 25th percentile, median, upper 75th percentile and maximum are plotted.

**Figure 9** Geographical coverage and characterization of climate classes $C_1$- $C_6$ based on cluster-analysis of 8 variables (near surface temperature, seasonal near surface temperature amplitude, precipitation, seasonal precipitation amplitude, u-wind in January and July, v-wind in January and July) in western South America. The geographical coverage 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-slice specific characterization of $C_1$- $C_6$ for PI (e), MH (f), LGM (g) and PLIO (h) is shown on the right.

**Figure 10** PI annual mean near-surface temperatures (a), and deviations of MH, LGM and PLIO annual mean near-surface temperatures from PI values (b) for the South Alaska region. Insignificant ($p < 99\%$) differences (as determined by a t-test) are greyed out.

**Figure 11** PI, MH, LGM and PLIO annual mean precipitation (a), and mean annual temperatures (b) in South Alaska. For each time slice, the minimum, lower 25th percentile, median, upper 75th percentile and maximum are plotted.

**Figure 12** Geographical coverage of climate classes $C_1$- $C_4$ based on cluster-analysis of 4 variables (near surface temperature, seasonal near surface temperature amplitude, total precipitation, seasonal total precipitation amplitude) in southern Alaska. The geographical coverage of the climates $C_1$- $C_4$ is shown on the left for PI (a,), MH (b), LGM (c) and PLIO (d); the complementary, time-slice specific characterization of $C_1$- $C_6$ for PI (e), MH (f), LGM (g) and PLIO (h) is shown on the right.

**Figure 13** PI annual mean near-surface temperatures (a), and deviations of MH, LGM and PLIO annual mean near-surface temperatures from PI values (b) for the US Pacific Northwest. Insignificant ($p < 99\%$) differences (as determined by a t-test) are greyed out.

**Figure 14** PI, MH, LGM and PLIO annual mean precipitation (a), and annual mean temperatures (b) in the Cascades, US Pacific Northwest. For each time slice, the minimum, lower 25th percentile, median, upper 75th percentile and maximum are plotted.

**Figure 15** Geographical coverage and characterization of climate classes $C_1$- $C_4$ based on cluster-analysis of 4 variables (near surface temperature, seasonal near surface temperature amplitude, total precipitation, seasonal total precipitation amplitude) in the Cascades, US Pacific Northwest. The geographical coverage of the climates $C_1$- $C_4$ is shown on the left for PI (a), MH (b), LGM (c) and PLIO (d); the complementary, time-slice specific characterization of $C_1$-

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

**Figure 16** Simulated annual mean precipitation deviations of MH (left) and LGM (right) from PI values in South Asia, and temporally corresponding proxy-based reconstructions, indicating wetter (upward facing blue triangles), drier (downward facing red triangles) or similar (grey circles) conditions in comparison with modern climate. MH proxy-based precipitation differences are taken from Mügler et al. (2010) (**66**), Wischnewski et al. (2011) (**67**), Mischke et al. (2008), Wischnewski et al. (2011), Herzschuh et al. (2009) (**68**), Yanhong et al. (2006) (**69**), Morrill et al. (2006) (**70**), Wang et al. (2002) (**71**), Wuennemann et al. (2006) (**72**), Zhang et al. (2011), Morinaga et al. (1993), Kashiwaya et al. (1995) (**73**), Shen et al. (2005) (**74**), Liu et al. (2014) (**75**), Herzschuh et al. (2006) (**76**), Zhang and Mischke (2009)  (**77**), Nishimura et al. (2014) (**78**), Yu and Lai (2014) (**79**), Gasse et al. (1991) (**80**), Van Campo et al. (1996) (**81**), Demske et al. (2009) (**82**), Kramer et al. (2010) (**83**), Herzschuh et al. (2006) (**84**), Hodell et al. (1999)(**85**), Hodell et al. (1999) (**86**), Shen et al. (2006) (**87**), Tang et al. (2000) (**88**), Tang et al. (2000) (**89**), Zhou et al. (2002) (**90**), Liu et al. (1998) (**91**), Asashi (2010)(**92**), Kotila et al. (2009) (**93**), Kotila et al. (2000) (**94**), Wang et al. (2002) (**95**), Hu et al. (2014) (**96**), Hodell et al. (1999) (**97**), Hodell et al. (1999) (**98**).

**Figure 17** Simulated annual mean precipitation deviations of MH (left) and LGM (right) from PI values in South America, and temporally corresponding proxy-based reconstructions, indicating wetter (upward facing blue triangles), drier (downward facing red triangles) or similar (grey circles) conditions in comparison with modern climate. MH proxy-based precipitation differences are taken from Bird et al. (2011) (**1**), Hansen et al (1994) (**2**), Hansen et al (1994) (**3**), Hansen et al (1994) (**4**), Hansen et al (1994) (**5**), Hansen et al (1994) (**6**), Hillyer et al. (2009) (**7**), D'Agostino et al. (2002) (**8**), Baker et al. (2001) (**9**), Schwalb et al (1999) (**10**), Schwalb et al (1999) (**11**), Schwalb et al (1999) (**12**), Schwalb et al (1999) (**13**), Moreno et al (2009) (**14**), Pueyo et al (2011) (**15**), Mujica et al (2015) (**16**), Fritz et al. (2004) (**17**), Gayo et al. (2012) (**18**), Latorre et al. (2006) (**19**), Latorre et al. (2003) (**20**), Quade et al (2008) (**21**), Bobst et al. (2001) (**22**), Grosjean et al. (2001) (**23**), Betancourt et al. (2000) (**24**), Latorre et al. (2002) (**25**), Rech et al. (2003) (**26**), Diaz et al. (2012) (**27**), Maldonado et al (2005) (**28**), Diaz et al. (2012) (**29**), Lamy et al. (2000) (**30**), Kaiser et al. (2008) (**31**), Maldonado et al. (2010) (**32**), Villagrán et al. (1990) (**33**), Méndez et al. (2015) (**34**), Maldonado et al. (2006) (**35**), Lamy et al. (1999) (**36**), Jenny et al. (2002) (**37**), Jenny et al. (2002b) (**38**), Villa-Martínez et al. (2003) (**39**), Bertrand et al. (2008) (**40**), De Basti et al. (2008) (**41**), Lamy et al. (2009) (**42**), Lamy et al. (2002) (**43**), Szeicz et al. (2003) (**44**), de Porras et al. (2012) (**45**), de Porras et al. (2014) (**46**), Markgraf et al. (2007) (**47**), Siani et al. (2010) (**48**), Gilli et al. (2001) (**49**), Markgraf et al. (2003) (**50**), Stine et al. (1990) (**51**).

**References**


Abe-Ouchi, A., Saito, F., Kageyama, M., Bracannot, P., Harrison, S. P., Lambeck, K., Otto-Bliesner, B. L., Peltier, W.R., Tarasov, L., Peterschmitt, J.-Y., Takahashi, K.: Ice-sheet configuration in the CMUP5/PMIP3 Last Glacial Maximum experiments. *Geosci. Model Dev.*, 8, 3621-3637, doi:10.5194/gmd-8-3621-2015, 2015

Andersen, J., Egholm, D.L., Knudsen, M.F., Jansen, J., Nielsen, S.B., The periglacial engine of mountain erosion – Part 1: Rates of frost cracking and frost creep., *Earth Surface Dynamics*, 3(4), 447-462, 2015.

Arnold, L., Breon, F.M., and Brewer, S: The earth as an extrasolar planet: the vegetation spectral signature today and during the last Quaternary climatic extrema, *Int. J. Astrobiol*., 8, 81–94. http://dx.doi.org/10.1017/ S1473550409004406, 2009.

Asahi, K.: Equilibrium-line altitudes of the present and Last Glacial Maximum in the eastern Nepal Himalayas and their implications for SW monsoon climate, Quatern. Int., 212, 26-34, doi:10.1016/j.quaint.2008.08.004, 2010.

Bahrenberg , G., Giese, E., and Nipper, J.: Multivariate Statistik. *Statistische Methoden in der Geographie 2*, Stuttgart, 1992.

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. P.: The History of South America Tropical Precipitation for the Past 25,000 Years, Earth and Atmospheric Sciences, 291, 640-643, doi:10.1126/science.291.5504.640, 2001.

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

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

Diaz, F.P., Latorre, C., Maldonado, A., Quade, J., and Betancourt, J.L.: Rodent middens reveal episodic, long-distance plant colonizations across the hyperarid Atacama Desert over the last 34,000 years, J. Biogeogr., 39, 510-525, doi:10.1111/j.1365-2699.2011.02617.x, 2012.

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

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

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 record from the NE margin of the Tibetan plateau. *Earth and Planetary Science Letters*, 214(1-2), 267-277, 2003.

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

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

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

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

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

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

Feng, R., Poulsen, C.J., Werner, M., 2016. Tropical circulation intensification and tectonic extension recorded by Neogene terrestrial d18O records of the western United States. Geology 44. doi:10.1130/G38212.1

Feng, R., Poulsen, C.J., 2016. Refinement of Eocene lapse rates, fossil-leaf altimetry, and North American Cordilleran surface elevation estimates. Earth Planet. Sci. Lett. doi:10.1016/j.epsl.2015.12.022

Fritz, S. C., Baker, P. A., Lowenstein, T. K., Seltzer, G. O., Rigsby, C. A., Dwyer, G. S., Tapia, P. M., Arnold, K. K., Ku, T. L., and Luo, S: Hydrologic variation during the last 170,000 years in the southern hemisphere tropics of South America, Quaternary Res., 61, 95 – 104, doi:10.1016/j.yqres.2003.08.007, 2004.Gasse, F., Arnold, M., Fontes, J. C., Fort, M., Gibert, E., Huc, A., Li, B., Li, Y., Liu, Q., Melleres, F., Van Campo, E., Wang, F., and Zhang, Q.: A 13,000-year climate record from western Tibet, Nature, 353, 742-745, doi:10.1016/j.quaint.2006.02.001, 1991.

Gayo E. M., Latorre, C., Santoro, C. M., Maldonado, A., and De Pol-Holz, R.: Hydroclimate variability in the low-elevation Atacama Desert over the last 2500 yr, Clim. Past, 8, 287-306, doi:10.5194/cp-8-287-2012, 2012.

Gierz, P., Lohmann, G., and Wei, W.: Response of Atlantic Overturning to future warming in a coupled atmosphere-ocean-ice sheet model, *Geophysical Research Letters*, 42, 6811-6818, doi:10.1002/2015GL065276, 2015.

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

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

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

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

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

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

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

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

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

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

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

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

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

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

Herzschuh, U., Winter, K., Wuennemann, B., and Li, S.: A general cooling trend on the central Tibetan Plateau throughout the Holocene recorded by the lake Zigetang pollen spectra, Quatern. Int., 154, 113-121, doi:10.1016/j.quaint.2006.02.005, 2006b.

Herzschuh, U., Kramer A., Mischke S., and Zhang C.: Quantitative climate and vegetation trends since the late glacial on the northeastern Tibetan Plateau deduced from Koucha lake pollen spectra, Quaternary Res., 71, 162-171, doi:10.1016/j.yqres.2008.09.003, 2009.

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

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

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

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

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

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

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

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

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

Jungclaus, J. H., Lorenz, S. J., Timmreck, C., Reick, C. H., Brovkin, V., Six, K., Segschneider, J., Giorgetta, M.A., Crowley, T.J., Pongratz, J., Krivova, N.A., Vieira, L.E., Solanski, S.K., Klocke, D., Botzet, M., Esch, M., Gayler, V., Haak, H., Raddatz, T.J., Roeckner, E., Schnur, R., Widmann, H., Claussen, M., Stevens, B., and Marotzke, J.: Climate and carbon-cycle variability over the last millennium. *Climate of the Past*, *6*, 723-737. doi:10.5194/cp-6-723-2010, 2010.

Junginger, A., Roller, S., Olaka, L. A., and Trauth, M. H.:The effects of solar irradiation changes on the migration of the Congo Air Boundary and water levels of paleo-Lake Suguta, Northern Kenya Rift, during the African Humid Period (15–5 ka BP), Palaeogeogr. Palaeocl., 396, 1-16, doi:10.1016/ j.palaeo.2013.12.007, 2014.

Kaiser, J., Schefuss, E., Lamy, F., Mohtadi, M., and Hebbeln, D.: Glacial to Holocene changes in sea surface temperature and coastal vegetation in north central Chile : high *versus* low latitude forcing, Quaternary

Sci. Rev., 27, 2064–2075, doi:10.1016/j.quascirev.2008.08.025, 2008.

Kalnay, E., Kanamitsu, M., Kistler, R., Collins, W., Deaven, D., Gandin, L., Iredell, M., Saha, S., White, G., Woollen, J., Zhu, Y., Chelliah, M., Ebisuzaki, W., Higgins, W., Janowiak, J., Mo, K.C., Ropelewski, C., Wang, J., Leetmaa, A., Reynolds, R., Jenne, R., and Joseph, D.: The NCEP/NCAR 40-year reanalysis project. *Bulletin of the American Meteorological Society*, 77(3), 437-471, 1996.

Kashiwaya, K., Masuzawa, T., Morinaga, H., Yaskawa, K., Yuan, B. Y., Liu, J. Q., and Gu, Z.: Changes in hydrological conditions in the central Qing-Zang (Tibetan) Plateau inferred from lake bottom sediments, Earth Planet. Sc. Lett., 135, 31-39, doi:10.1016/0012-821X(95)00136-Z, 1995.

Kent-Corson, M., Sherman, L., Mulch, A. and Chamberlain, C.: Cenozoic topographic and climatic response to changing tectonic boundary conditions in Western North America, Earth Planet. Sc. Lett., 252(3-4), 453-466, doi:10.1016/j.epsl.2006.09.049, 2006.

Kistler, R., Collins, W., Saha, S., White, G., Woollen, J., Kalnay, E., Chelliah, M., Ebisuzaki, W., Kanamitsu, M., Kousky, V., Van den Dool, H., Jenne, R., and Fiorino, M.: The NCEP−NCAR 50−Year Reanalysis: Monthly Means CD−ROM and Documentation. *Bulletin of the American Meteorological Society*, 82(2), 247–267, 2001.

Kirchner, J.W., Finkel, R.C., Riebe, C.S., Granger, D.E., Clayton, J.L., King, J.G., and Megahan, W.F.: Mountain erosion over 10 yr, 10 k.y., and 10 m.y. time scales, *Geology*, 29 (7), 591-594, 2001.

Knorr, G., Butzin, M., Micheels, A., and Lohmann, G.: A Warm Miocene Climate at Low Atmospheric CO2 levels. Geophysical Research Letters, 38, L20701, doi:10.1029/2011GL048873, 2011.

Koons, P.O., Zeitler, P.K., and Hallet, B.: 5.14 Tectonic Aneurysms and Mountain Building, in *Treatise on Geomorphology*, pp. 318–349, Elsevier, 2013.

Kotlia, B. S., Sharma, C., Bhalla, M. S., Rajagopalan, G.,. Subrahmanyam, K., Bhattacharya, A., and Valdiya, K. S.: Paleoclimatic conditions in the late Pleistocene Wadda Lake, eastern Kumaun Himalaya (India), Palaeogeogr. Palaeocl., 162, 105-118, doi:10.1016/S0031-0182(00)00107-3, 2000.

Kramer, A., Herzschuh, U., Mischke, S., and Zhang, C.: Late glacial vegetation and climate oscillations on the southeastern Tibetan Plateau inferred from the lake Naleng pollen profile, Quaternary Res., 73, 324-335, doi:10.1016/j.yqres.2009.12.003, 2010.

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

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

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

Lamy, F. and Kaiser. J.: Past Climate Variability in South America and Surrounding Regions, Chapter 6:

Glacial to Holocene Paleoceanographic and Continental Paleoclimate Reconstructions Based on ODP Site 1233/GeoB 3313 Off southern Chile, 2009.

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

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

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

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

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

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

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

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

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

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

Li, J., Ehlers, T.A., Werner, M., Mutz, S.G., Steger, C., Paeth, H.: Late quarternary climate, precipitation $\delta^{18}O$, and Indian monsoon variations over the Tibetan Plateau. *Earth and Planetary Science Letters* 457, 412-422, 2017.

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 climate threshold at the eastern edge of the Tibetan plateau, Geophys. Res. Lett., doi: 10.1002/2014GL060833, 2014.

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

Lohmann, G., Pfeiffer, M., Laepple, T., Leduc, G., and Kim, J.-H.: A model-data comparison of the Holocene global sea surface temperature evolution. Clim. Past, 9, 1807-1839, doi:10.5194/cp-9-1807-2013, 2013.

Lorenz, S.J., and Lohmann, G.: Acceleration technique for Milankovitch type forcing in a coupled atmosphere-ocean circulation model: method and application for the Holocene. Climate Dynamics (2004) 23: 727. doi:10.1007/ s00382-004-0469-y, 2004.

Maldonado, A. and Villagrán, C.: Climate variability over the last 9900 cal yr BP from a swamp forest pollen record along the semiardi coast of Chile, Quaternary Res., 66, 146-258, doi:10.1016/j.yqres.2006.04.003, 2006.

Maldonado, A. J., Betancourt, J. L., Latorre, C., and Villagrán, C : Pollen analyses from a 50000-yr rodent midden series in the southern Atacama Desert (25°30' S), J. Quaternary Sci., 20, 493-507, doi:10.1002/jqs.936, 2005.

Maldonado, A., Méndez, C., Ugalde, P., Jackson, D., Seguel, R., and Latorre, C.: Early Holocene climate change and human occupation along the semiarid coast of north-central Chile, J. Quaternary Sci., 25, 985–988, doi:10.1002/jqs.1385, 2010.

Markgraf, V., Bradbury, J. P., Schwalb, A., Burns, S., Stern, C., Ariztegui, D., Gilli, D., Anselmetti, F. S., Stine, S., and Maidana, N.: Holocene palaeoclimates of southern Patagonia: limnological and environmental history of Lago Cardiel, Argentina, Holocene, 13, 581–591, doi:10.1191/0959683603hl648rp, 2003.

Markgraf, V., Whitlock, C., and Haberle, S.: Vegetation and fire history during the last 18,000 cal yr B.P. in Southern Patagonia: Mallín Pollux, Coyhaique, Province Aisén (45°41'30'' S, 71°50'30'' W, 640 m elevation), Palaeogeogr. Palaeocl , 254, 492–507, doi:10.1016/j.palaeo.2007.07.008, 2007.

Maroon, E. A., Frierson, D. M. W., and Battisti, D. S.: The tropical precipitation response to Andes topography and ocean heat fluxes in an aquaplanet model, J. Climate, 28, 381–398, doi:10.1175/JCLI-D-14-00188.1, 2015.

Maroon, E. A., Frierson, D. M. W., Kang, S. M., and Scheff, J.: The precipitation response to an idealized subtropical continent, J. Climate, 29, 4543–4564, doi:10.1175/JCLI-D-15-0616.1, 2016.

Marshall, J.A., Roering, J.J., Bartlein, P.J., Gavin, D.G., Granger, D.E., Rempel, A.W., Praskievicz, S.J., and Hales, T.C.: Frost for the trees: Did climate increase erosion in unglaciated landscapes during the late Pleistocene? *Science Advances*, 1, 1-10, 2015.

Marston, R.A.: Geomorphology and vegetation on hillslopes: Interactions, dependencies, and feedback loops, *Geomorphology*, 116(3–4), 206–217. doi:http://doi.org/10.1016/j.geomorph.2009.09.028, 2010.

Matsuoka, N., and Murton, J.: Frost weathering: Recent advances and future directions. *Permafr. Periglac. Process.* 19, 195–210, 2008.

Matsuoka, N.: Solifluction rates, processes and landforms: A global review. *Earth Science Reviews* 55, 107–134, 2001.

Maussion, F., Scherer, D., Mölg, T., Collier, E., Curio, J., and Finkelnburg, R.: Precipitation seasonality and variability over the Tibetan Plateau as resolved by the High Asia Reanalysis, *J. Climate, 27, 1910-1927, doi:10.1175/JCLI-D-13-00282.1*, 2014

Méndez, C., Gil, A., Neme, G., Nuevo Delaunay, A., Cortegoso, V., Huidobro, C., Durán, and Maldano, A.: Mid Holocene radiocarbon ages in the Subtropical Andes (~29° - 35° S), climatic change an implicaton for human space organization, Quatern. Int., 356, 15-26, doi:10.1016/j.quaint.2014.06.059, 2015.

Mesinger, F., DiMego, G., Kalnay, E., Mitchell, K., Shafran, P.C., Ebisuzaki, W., Jovic, D., Woollen, J., Rogers, E., Berbery, E.H., Ek, M.B., Fan, Y., Grumbine, R., Higgins, W., Li, H., Lin, Y., Manikin, G., Parrish, D., and Shi, W.: North American Regional Reanalysis, *Bulletin of the American Meteorological Society, 87, 343–360.doi:10.1175/BAMS-87-3-343, 2006.*

Methner, K., Fiebig, J., Wacker, U., Umhoefer, P., Chamberlain, C. and Mulch, A.: Eocene-Oligocene proto-Cascades topography revealed by clumped (Δ47) and oxygen isotope (δ18O) geochemistry (Chumstick Basin, WA, USA), Tectonics, 35(3), 546-564, doi:10.1002/2015tc003984, 2016.

Mischke S., Kramer M., Herzschuh U., Shang H., Erzinger J., and Zhang C.: Reduced early holocene moisture availability in the Bayan Har Mountains, northeastern Tibetan Plateau, inferred from a multi-proxy lake record, Palaeogeogr. Palaeocl, 267, 59-76, doi:10.1016/j.palaeo.2008.06.002, 2008.

Molnar, P., Boos, W.R., and Battisti, D.S.: Orographic Controls on Climate and Paleoclimate of Asia: Thermal and Mechanical Roles for the Tibetan Plateau, *Annual Review of Earth and Planetary Sciences*, 38, 77-102, 2010.

Molnar, P., and England, P.: Late Cenozoic uplift of mountain ranges and global climate change: chicken or egg?, *Nature*, 346, 29-34, 1990.

Montgomery, D.R.: Slope distributions, threshold hillslopes, and steady-state topography, *American Journal of Science*, 301, 432-454, 2001.

Montgomery, D.R., Balco, G., and Willett, S.D.: Climate, tectonics, and the morphology of the Andes, *Geology*, 29(7), 579-582, 2001.

Moon, S., Chamberlain, C.P., Blisniuk, K., Levine, N., Rood, D.H., Hilley, G.E.: Climatic control of denudation in the deglaciated landscape of the Washington Cascades, *Nature Geoscience*, 4, 469-473, 2011.

Moreno, A., Santoro, C. M., and Latorre, C.: Climate change and human occupation in the northernmost Chilean Altiplano over the last ca. 11500 cal. a BP, Quaternary Sci., 24, 373–382, doi:10.1002/jqs.1240, 2009.

Morinaga, H., Itota, C., Isezaki, N., Goto, H., Yaskawa, K., Kusakabe, M., Liu, J., Gu, Z., Yuan, B., and Cong, S.: Oxygen-18 and carbon-13 records for the last 14,000 years from lacustrine carbonates of Siling-Co (lake) in the Qinghai-Tibetan Plateau, Geophys. Res. Lett., 20, 2909-2912, doi:10.1029/93GL02982, 1993.

Morrill, C., Overpeck, J. T., Cole, J. E., Liu, K., Shen, C., and Tang, L.: Holocene variations in theAsian mon-soon inferred from the geochemistry of lake sediments in central Tibet, Quaternary Res., 65, 232-243, doi:10.1016/j.yqres.2005.02.014, 2006.

Moulton, K.L., and Berner, R.A.: Quantification of the effect of plants on weathering: studies in Iceland, *Geology*, 26, 895-898, 1998.

Muegler, I., Gleixner, G., Guenther, F., Maeusbacher, R., Daut, G., Schuett, B., Berking, J., Schwalb, A., Schwark, L., Xu, B., Yao, T., Zhu, L., and Yi, C.: A multi-proxy approach to reconstruct hydrological changes and Holocene climate development of Nam Co, Central Tibet, J. Paleolimnol., 43, 625-648, doi:10.1007/s10933-009-9357-0, 2010.

Mujica, M. I., Latorre, C., Maldonado, A., González-Silvestre, L., Pinto, R., De Pol-Holz, R., and Santoro, C. M.: Late Quaternary climate change, relict populations and present-day refugia in the northern Atacama Desert: a case study from Quebrada La Higuera (18° S), J. Biogeogr. , 42, 76–88, doi:10.1111/jbi.12383, 2015.

Mulch, A., Sarna-Wojcicki, A., Perkins, M. and Chamberlain, C.: A Miocene to Pleistocene climate and elevation record of the Sierra Nevada (California), P. Natl. Acad. Sci. USA, 105(19), 6819-6824, doi:10.1073/pnas.0708811105, 2008.

Mulch, A., Chamberlain, C., Cosca, M., Teyssier, C., Methner, K., Hren, M. and Graham, S.: Rapid change in high-elevation precipitation patterns of western North America during the Middle Eocene Climatic Optimum (MECO), Am. J. Scie, 315(4), 317-336, doi:10.2475/04.2015.02, 2015.

Mutz, S.G., Ehlers, T.A., Li, J., Steger, C., Peath, H., Werner, M., Poulsen, C.J.: Precipitation $\delta^{18}O$ over the South Asia Orogen from ECHAM5-wiso Simulation: Statistical Analysis of Temperature, Topography and Precipitation. *Journal of Geophysical Research-Atmospheres*, 121(16),9278-9300, doi: 10.1002/2016JD024856, 2016.

Nishimura, M., Matsunaka, T., Morita, Y., Watanabe, T., Nakamura, T., Zhu, L., Nara, W. F., Imai, A., Izutsu, Y., and Hasuike, N.: Paleoclimatic changes on the southern Tibetan Plateau over the past 19,000 years recorded in lake Pumoyum Co, and their implications for the southwest monsoon evolution, Palaeogeogr. Palaeocl., 396, 75-92, doi:10.1016/j.palaeo.2013.12.015, 2014.

Otto-Bliesner, B.L., Brady, C.B., Clauzet, G., Tomas, R., Levis, S., and Kothavala, Z.: Last Glacial Maximum and Holocene Climate in CCSM3. Journal of Climate, 19, 2526-2544, 2006.

Paeth, H.: Key Factors in African Climate Change Evaluated by a Regional Climate Model, *Erdkunde*, 58, 290-315, 2004.

Peel, M.C., Finlayson, B.L., McMahon, T.A.: Updated world map of the Koppen- Geiger climate classification. Hydrol. Earth Syst. Sci. 11, 1633–1644, 2007.

Pfeiffer, M., and Lohmann, G.: Greenland Ice Sheet influence on Last Interglacial climate: global sensitivity studies performed with an atmosphere-ocean general circulation model. *Climate of the Past*, 12, pp. 1313-1338, doi:10.5194/cp-12-1313-2016, 2016.

Pickett, E.J., Harrison, S.P., Flenley, J., Grindrod, J., Haberle, S., Hassell, C., Kenyon, C., MacPhail, M., Martin, H., Martin, A.H., McKenzie, M., Newsome, J.C., Penny, D., Powell, J., Raine, J.I., Southern, W., Stevenson, J., Sutra, J.-P., Thomas, I., van der Kaars, S., Ward, J.: Pollen-based reconstructions of biome distributions for Australia, South-East Asia and the Pacific (SEAPAC region) at 0, 6000 and 18,000 14C years B.P.. Journal of Biogeography, 31, 1381–1444, doi: 10.1111/j.1365-2699.2004.01001.x, 2004.

Pingel, H., Mulch, A., Alonso, R., Cottle, J., Hynek, S., Poletti, J., Rohrmann, A., Schmitt, A., Stockli, D. and Strecker, M.: Surface uplift and convective rainfall along the southern Central Andes (Angastaco Basin, NW Argentina), 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 Asia~~a~~n plateau since Late Miocene times. *Nature*, 411(6833), 62-66, 2001.

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 climate change in the Zoigê Plateau and AMS radiocarbon dating, Chinese Sci. Bull., 47, 66-70, doi:10.1360/02tb9013, 2002.

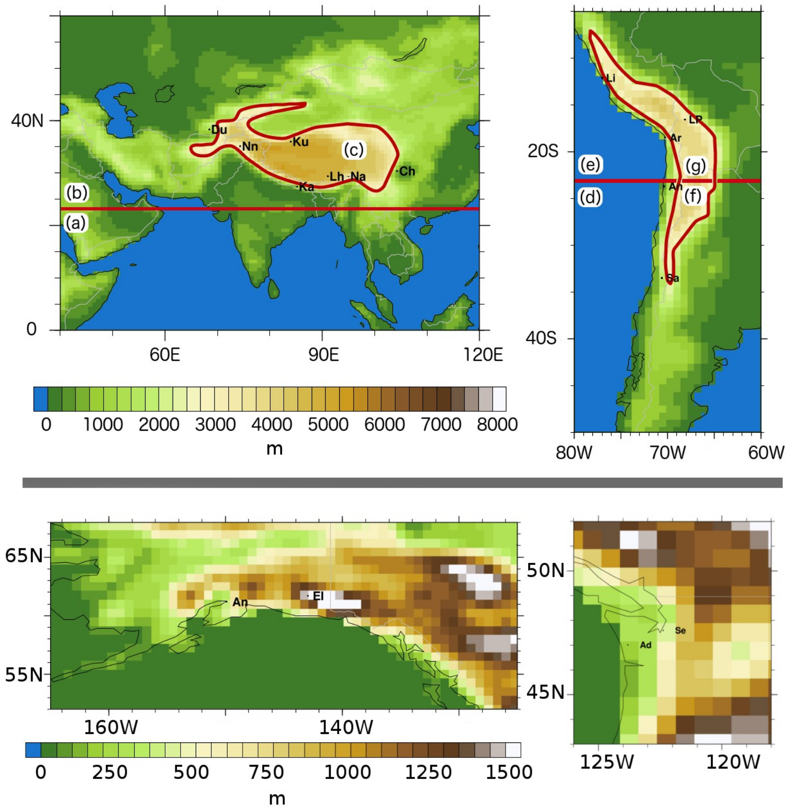

40N
0
60E
90E
120E
Du
Nn
Ku
(c)
Ch
Lh
Na
Ka
(b)
(a)
0 1000 2000 3000 4000 5000 6000 7000 8000
m
Li
LP
Ar
An
Sa
(e)
(g)
(d)
(f)
20S
40S
80W
70W
60W
65N
55N
An
El
160W
140W
0 250 500 750 1000 1250 1500
m
50N
45N
Se
Ad
125W
120W

Figure 1

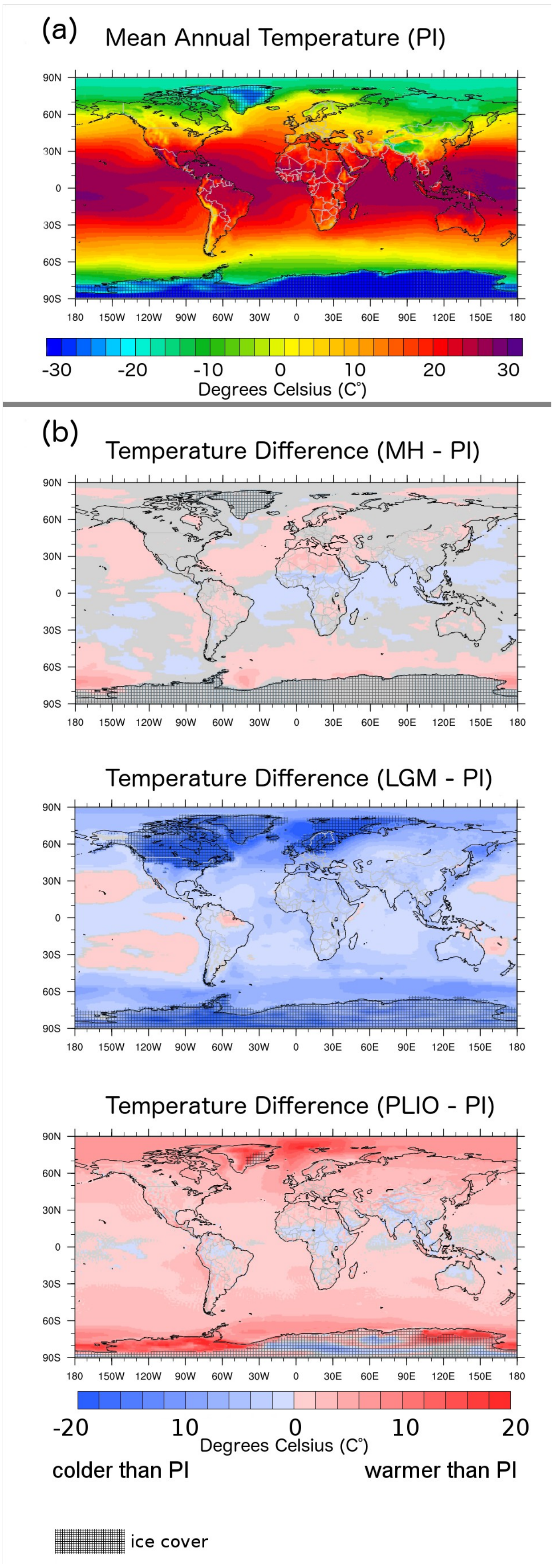

(a)
Mean Annual Temperature (PI)
90N
60N
30N
0
30S
60S
90S
180
150W
120W
90W
60W
30W
0
30E
60E
90E
120E
150E
180
-30
-20
-10
0
10
20
30
Degrees Celsius (C°)
(b)
Temperature Difference (MH - PI)
Temperature Difference (LGM - PI)
Temperature Difference (PLIO - PI)
-20
10
0
10
20
Degrees Celsius (C°)
colder than PI
warmer than PI
ice cover

Figure 2

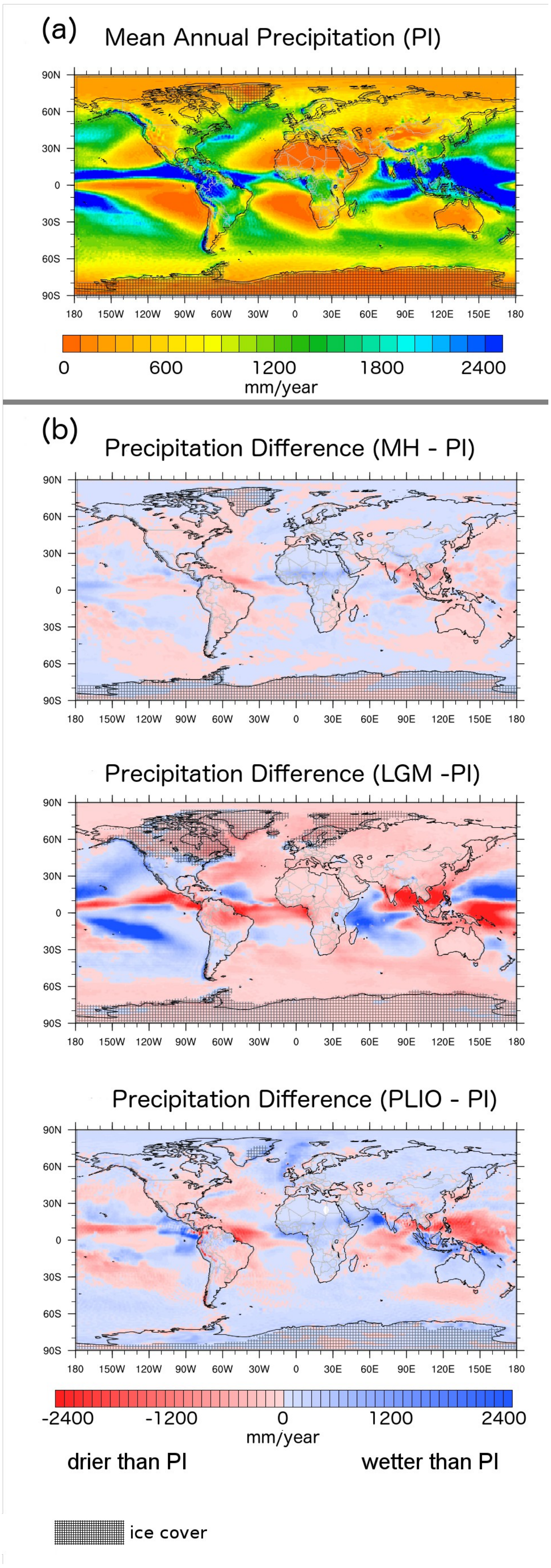

(a)
Mean Annual Precipitation (PI)
90N
60N
30N
0
30S
60S
90S
180 150W 120W 90W 60W 30W 0 30E 60E 90E 120E 150E 180
0 600 1200 1800 2400
mm/year
(b)
Precipitation Difference (MH - PI)
Precipitation Difference (LGM -PI)
Precipitation Difference (PLIO - PI)
-2400 -1200 0 1200 2400
mm/year
drier than PI
wetter than PI
ice cover

Figure 3

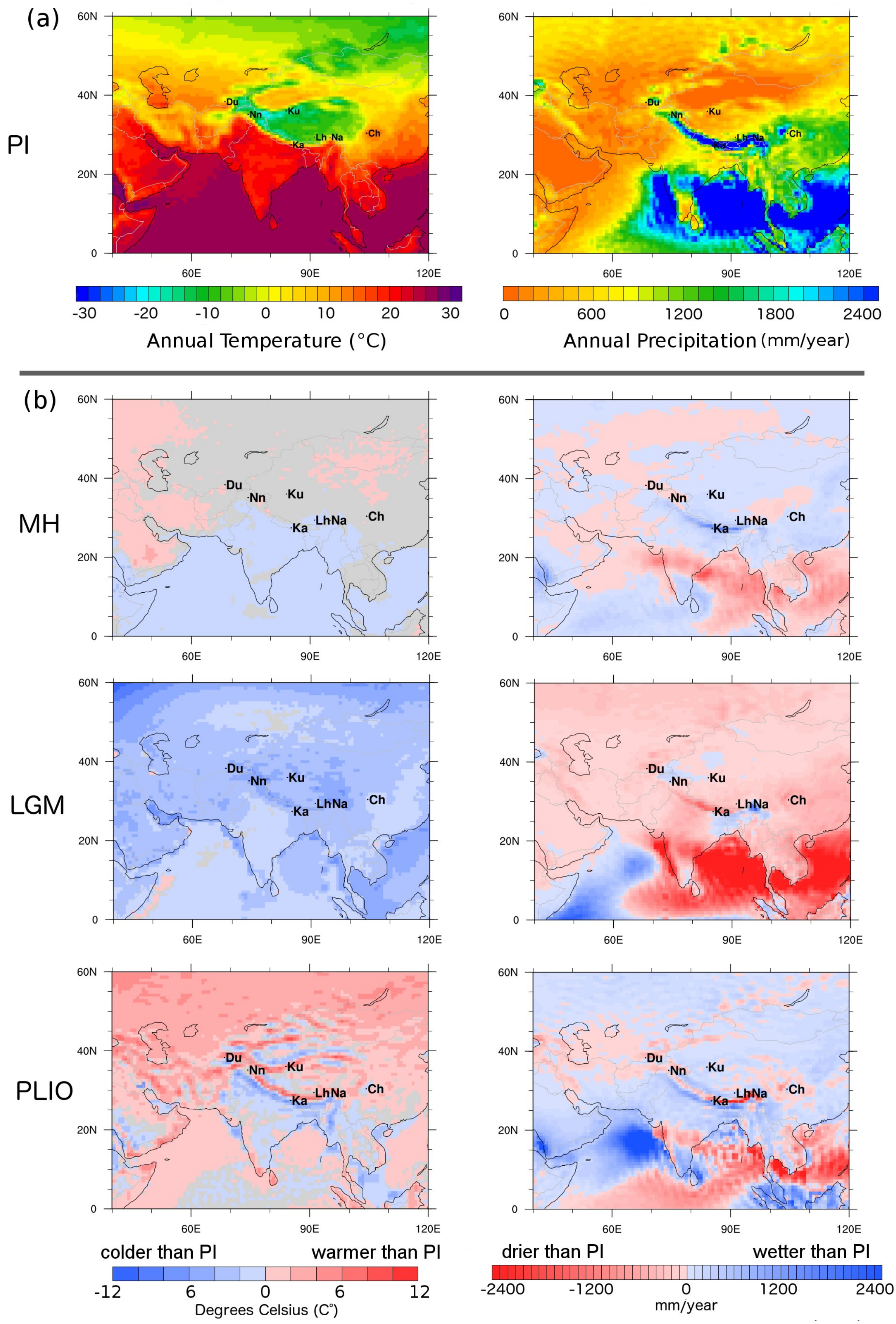

(a)
PI
60N
40N
20N
0
60E
90E
120E
Du
Nn
Ku
Ka
Lh Na
Ch
-30 -20 -10 0 10 20 30
Annual Temperature (°C)
0 600 1200 1800 2400
Annual Precipitation (mm/year)
(b)
MH
LGM
PLIO
colder than PI
warmer than PI
-12 6 0 6 12
Degrees Celsius (C°)
drier than PI
wetter than PI
-2400 -1200 0 1200 2400
mm/year



Figure 4

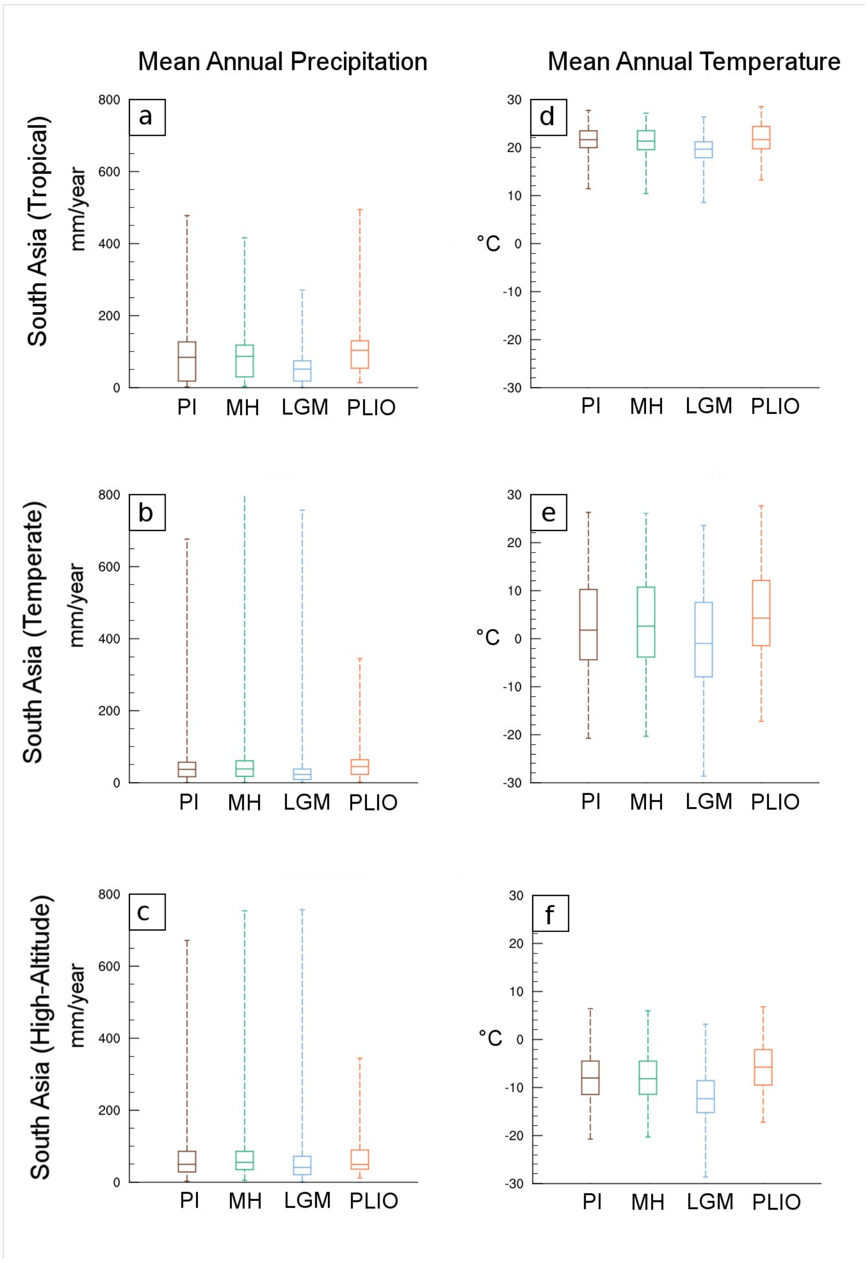
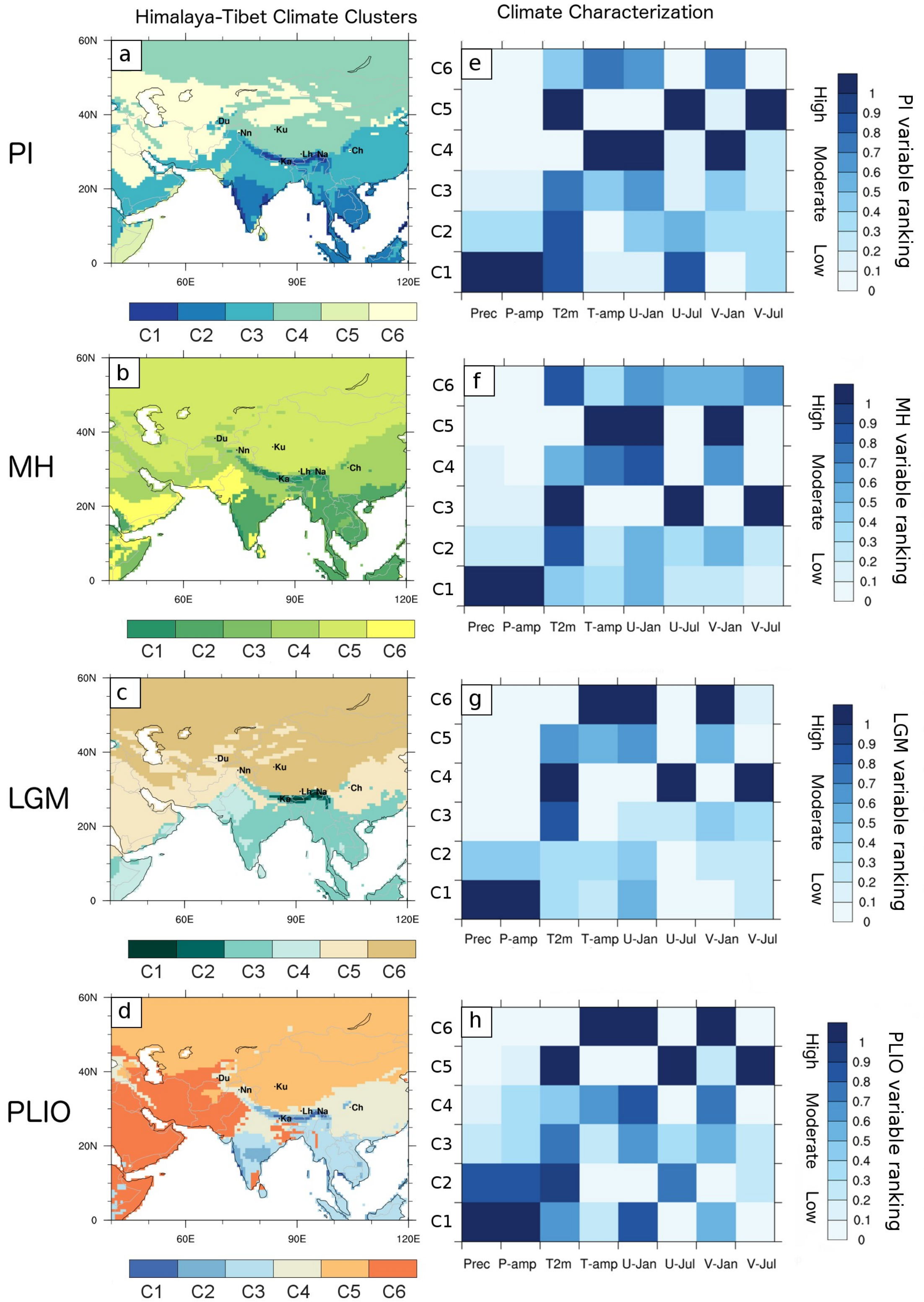
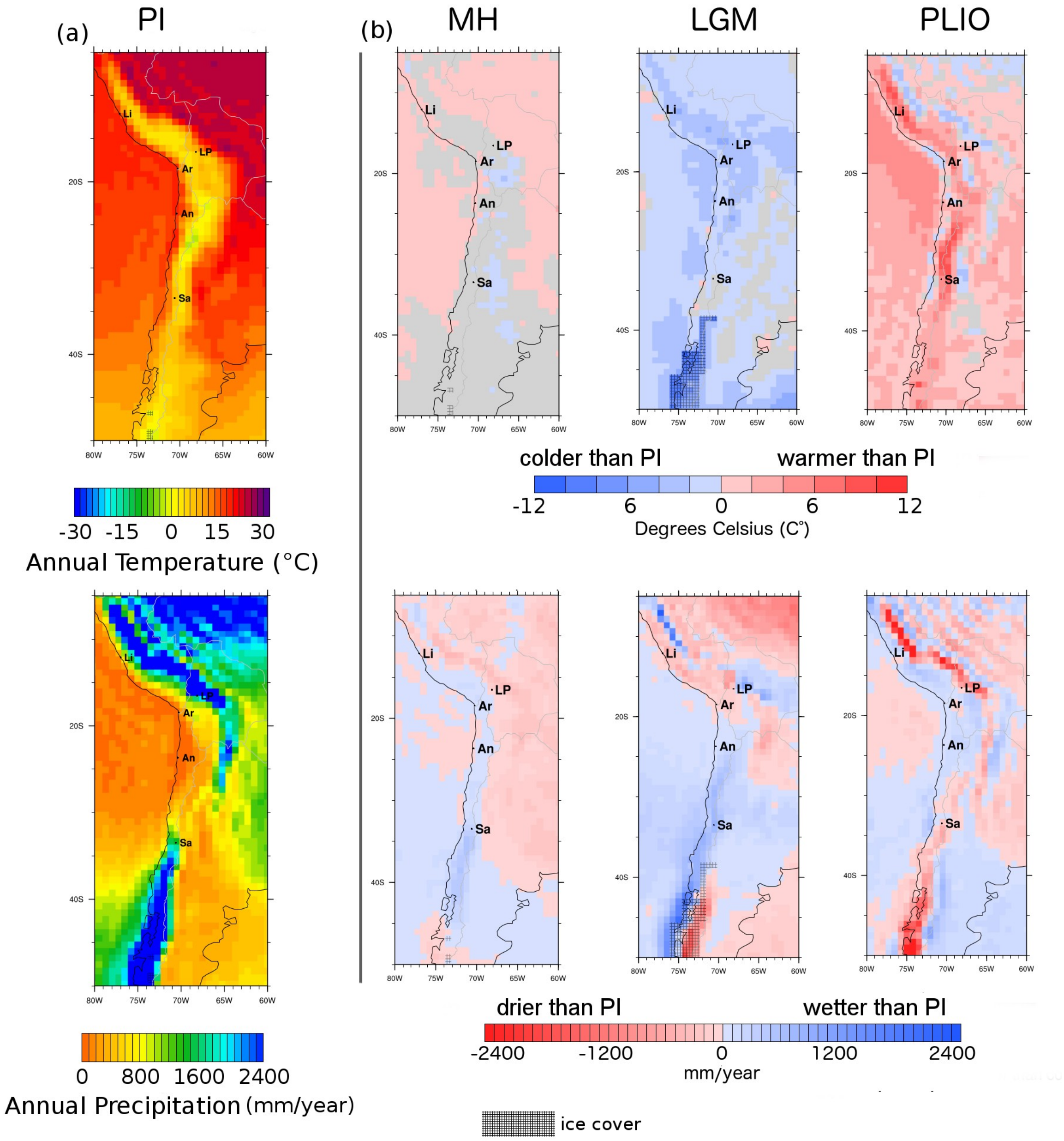

Mean Annual Precipitation
Mean Annual Temperature
South Asia (Tropical)
South Asia (Temperate)
South Asia (High-Altitude)
mm/year
°C
a
b
c
d
e
f
800
600
400
200
0
30
20
10
0
-10
-20
-30
PI
MH
LGM
PLIO

Figure 5



Himalaya-Tibet Climate Clusters
Climate Characterization
PI
MH
LGM
PLIO
C1 C2 C3 C4 C5 C6
Prec P-amp T2m T-amp U-Jan U-Jul V-Jan V-Jul
High Moderate Low
PI variable ranking
MH variable ranking
LGM variable ranking
PLIO variable ranking
Du Nn Ku Ka Lh Na Ch
60E 90E 120E
0 20N 40N 60N


Figure 6



(a)
PI
(b)
MH
LGM
PLIO
Li
LP
Ar
An
Sa
20S
40S
80W
75W
70W
65W
60W
-30 -15 0 15 30
Annual Temperature (°C)
colder than PI
warmer than PI
-12
6
0
6
12
Degrees Celsius (C°)
0 800 1600 2400
Annual Precipitation (mm/year)
drier than PI
wetter than PI
-2400
-1200
0
1200
2400
mm/year
ice cover


Figure 7

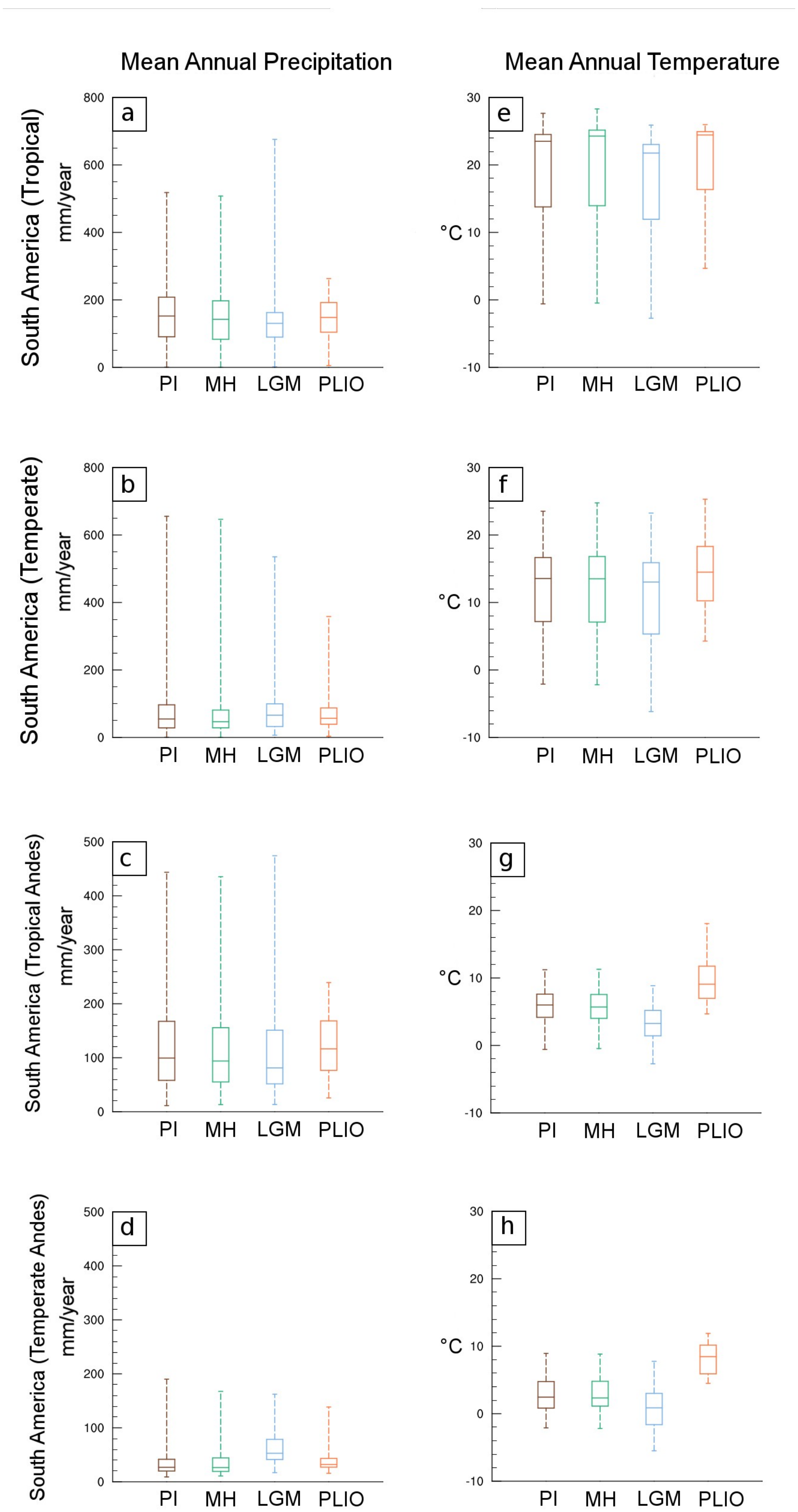

Mean Annual Precipitation
Mean Annual Temperature
South America (Tropical)
South America (Temperate)
South America (Tropical Andes)
South America (Temperate Andes)
mm/year
°C
a
b
c
d
e
f
g
h
PI
MH
LGM
PLIO

Figure 8

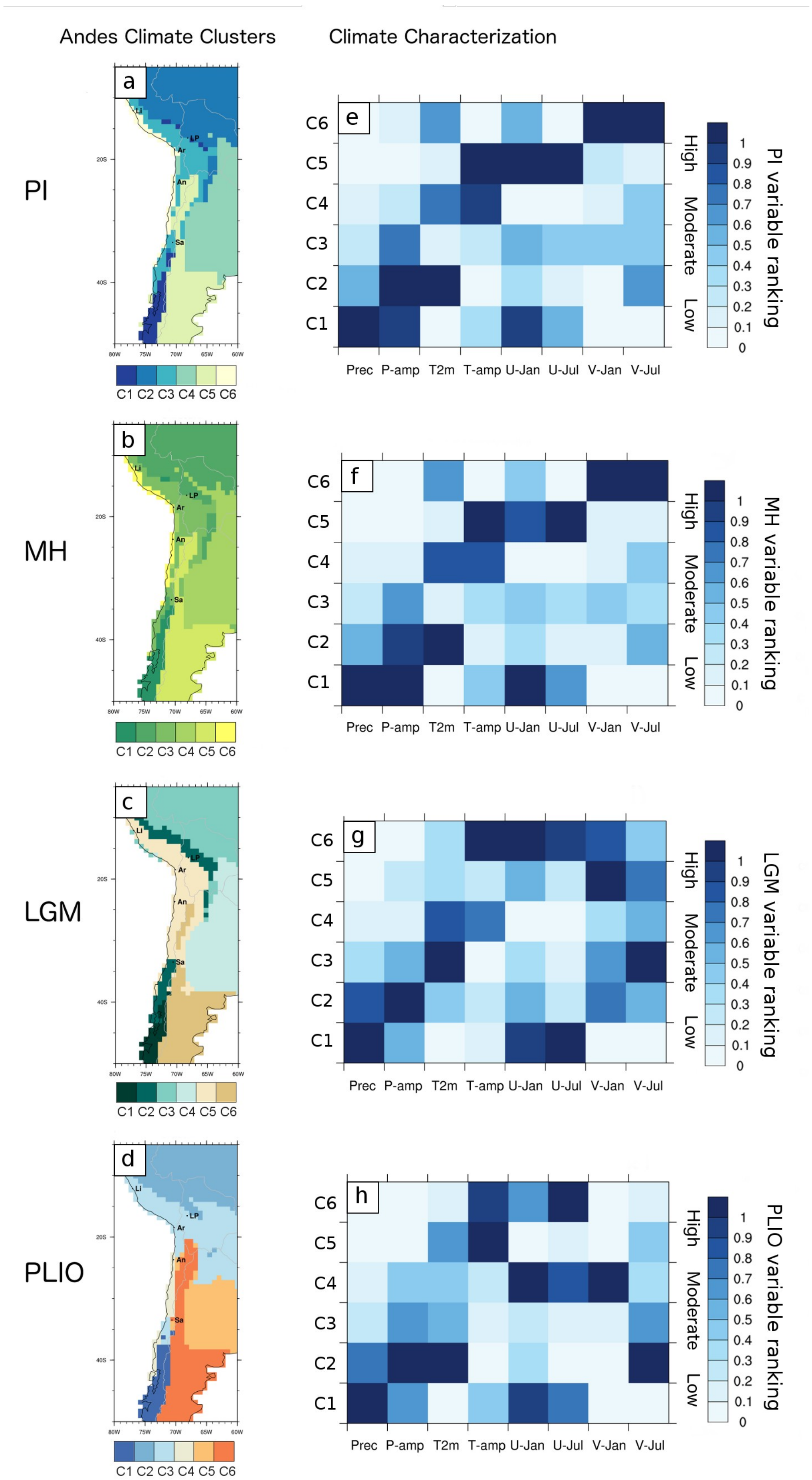

Andes Climate Clusters
Climate Characterization
PI
MH
LGM
PLIO
a
b
c
d
e
f
g
h
C1 C2 C3 C4 C5 C6
C6
C5
C4
C3
C2
C1
Prec P-amp T2m T-amp U-Jan U-Jul V-Jan V-Jul
High
Moderate
Low
1
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0
PI variable ranking
MH variable ranking
LGM variable ranking
PLIO variable ranking

Figure 9

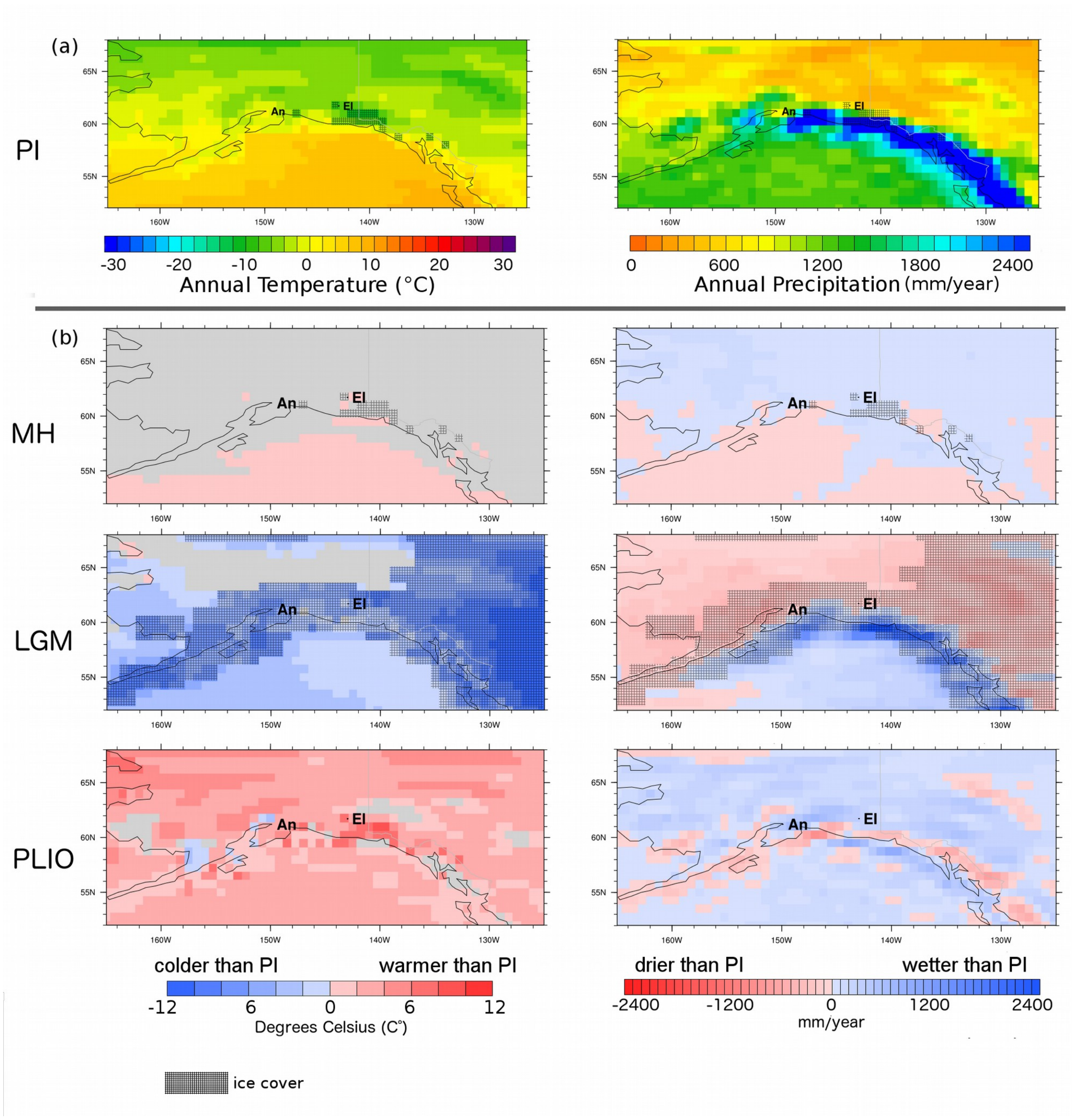



Figure 10

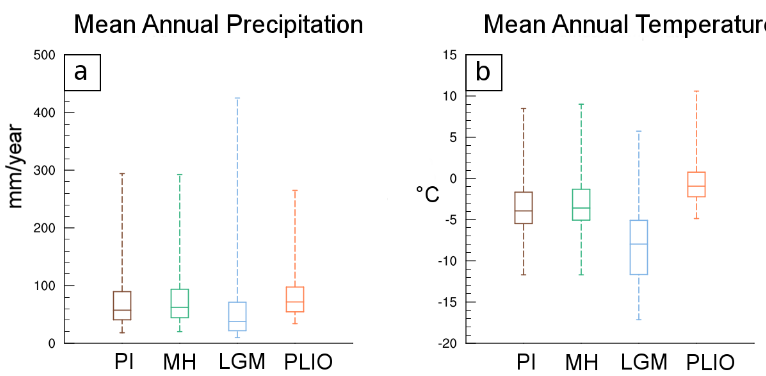
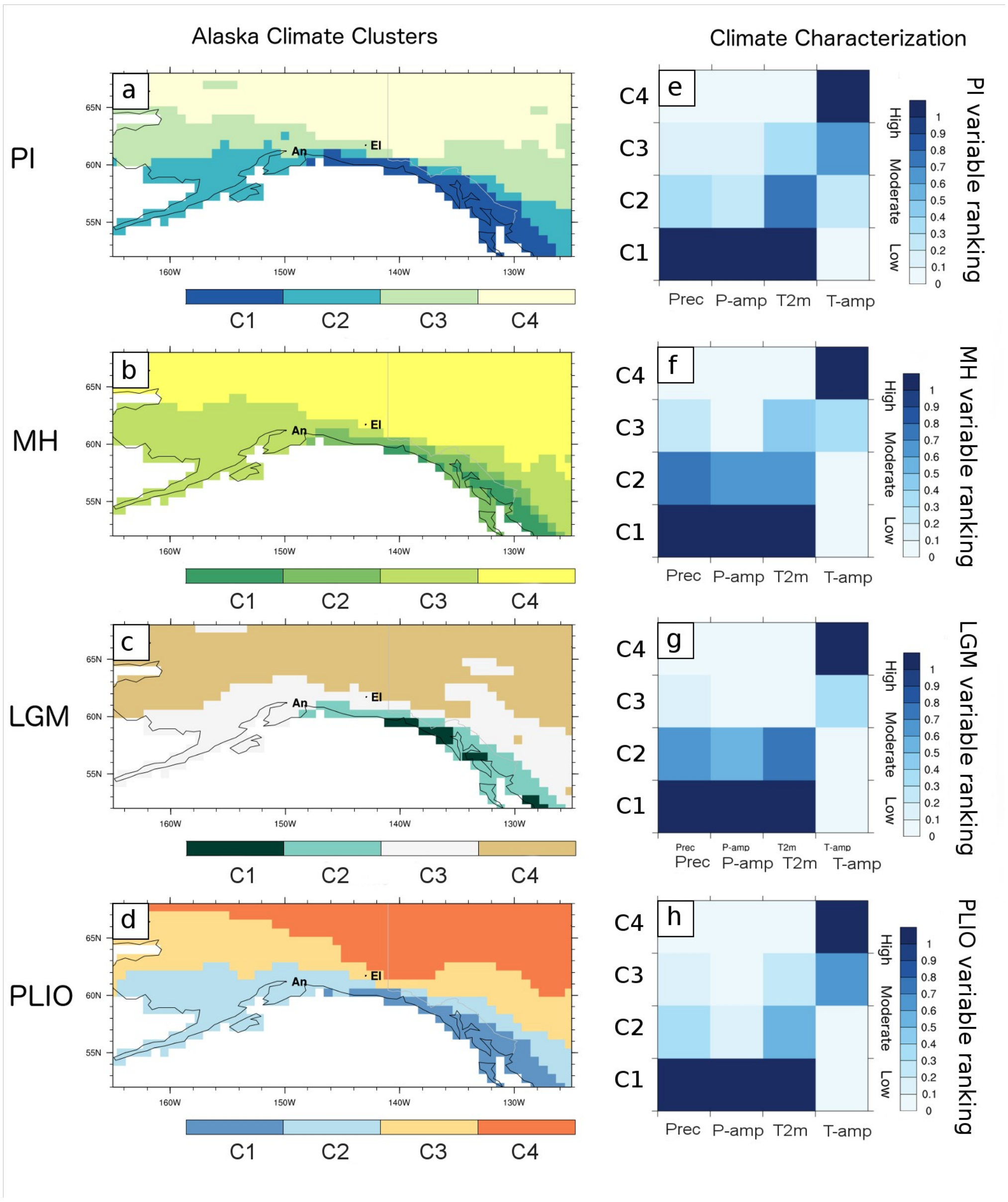



Figure 11



Alaska Climate Clusters
Climate Characterization
PI
MH
LGM
PLIO
a
b
c
d
e
f
g
h
An
El
65N
60N
55N
160W
150W
140W
130W
C1
C2
C3
C4
Prec
P-amp
T2m
T-amp
High
Moderate
Low
1
0.9
0.8
0.7
0.6
0.5
0.4
0.3
0.2
0.1
0
PI variable ranking
MH variable ranking
LGM variable ranking
PLIO variable ranking


Figure 12

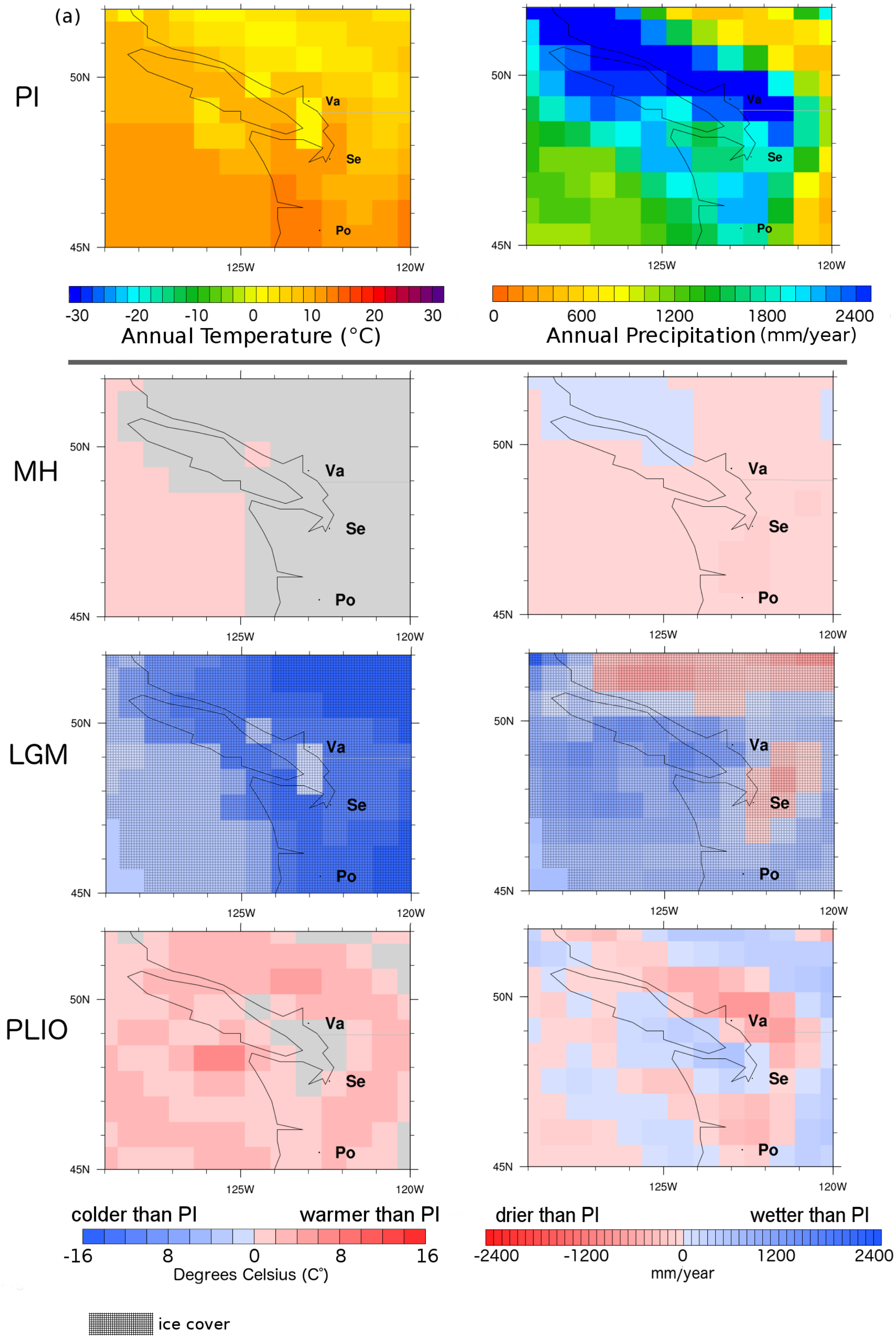

(a)
PI
MH
LGM
PLIO
Va
Se
Po
50N
45N
125W
120W
-30 -20 -10 0 10 20 30
Annual Temperature (°C)
0 600 1200 1800 2400
Annual Precipitation (mm/year)
colder than PI
warmer than PI
-16 8 0 8 16
Degrees Celsius (C°)
drier than PI
wetter than PI
-2400 -1200 0 1200 2400
mm/year
ice cover

Figure 13

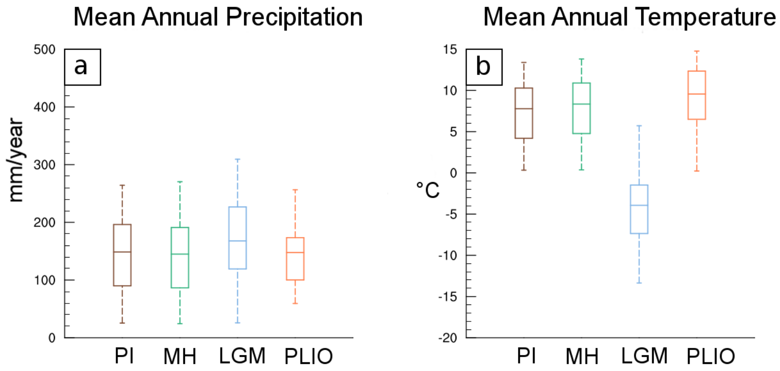



Figure 14

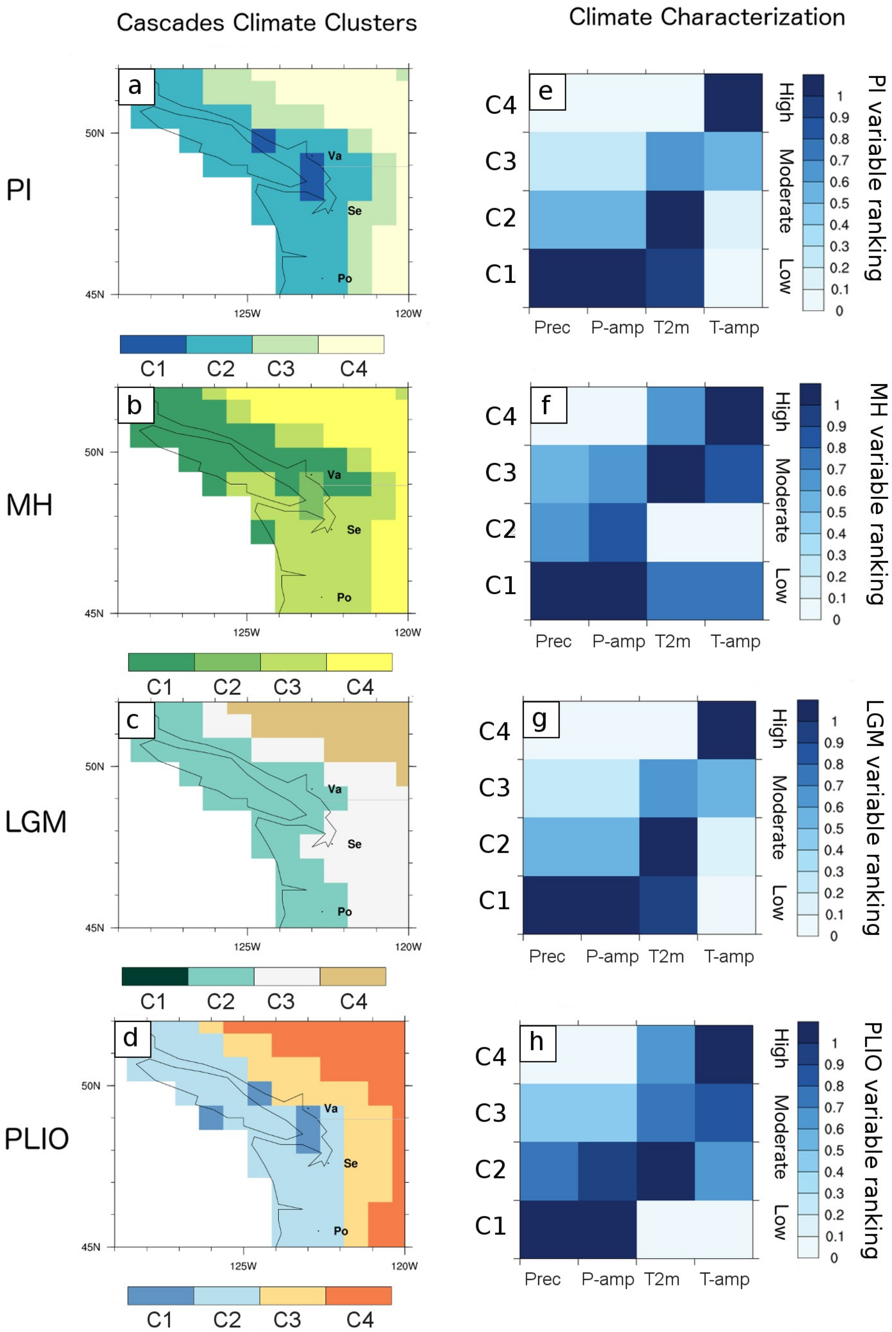



Figure 15

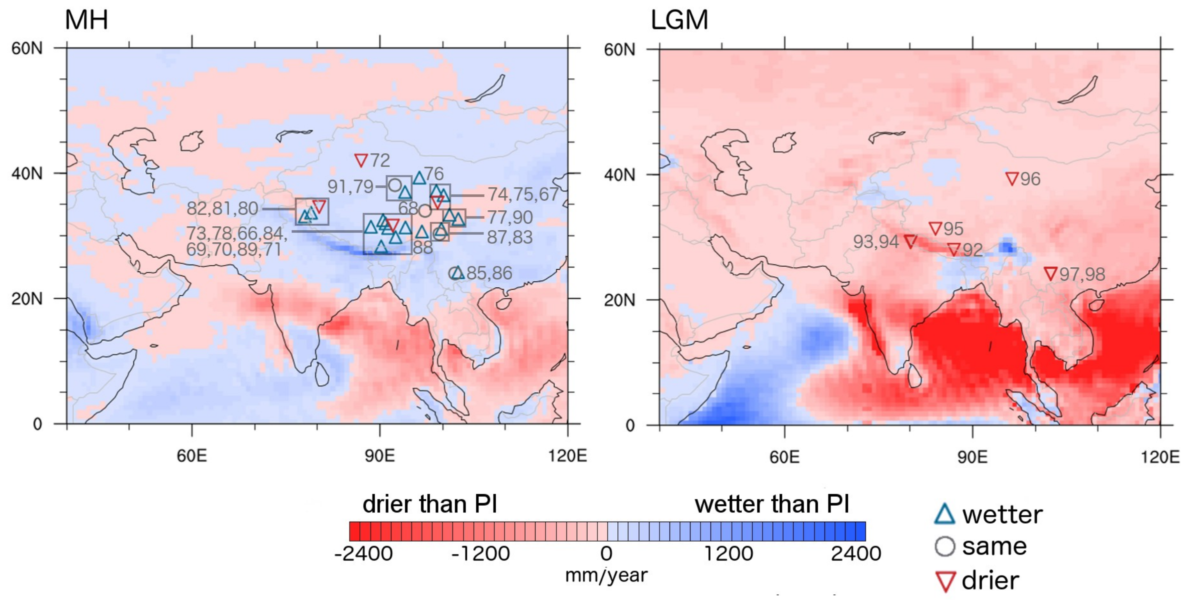



Figure 16

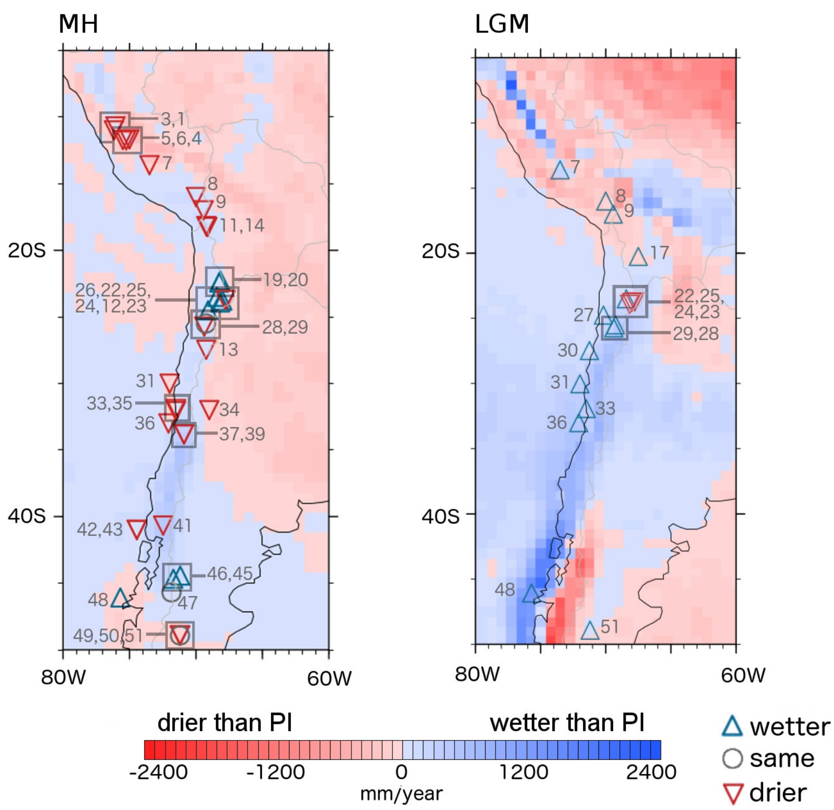
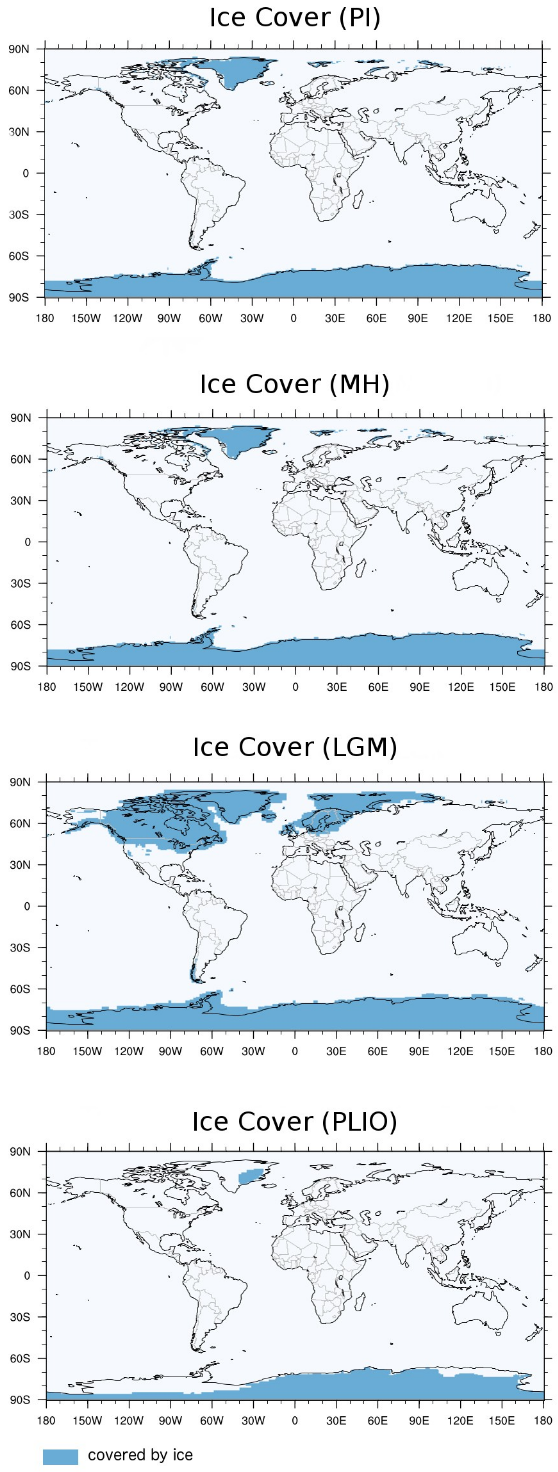



Figure 17




Supplemental Figure S1