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Precipitation scaling with temperature in warm and cold climates: An analysis of CMIP5 simulations

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[1] We investigate the scaling between precipitation and temperature changes in warm and cold climates using six models that have simulated the response to both increased CO2 and Last Glacial Maximum (LGM) boundary conditions. Globally, precipitation increases in warm climates and decreases in cold climates by between 1.5%/°C and 3%/°C. Precipitation sensitivity to temperature changes is lower over the land than over the ocean and lower over the tropical land than over the extratropical land, reflecting the constraint of water availability. The wet tropics get wetter in warm climates and drier in cold climates, but the changes in dry areas differ among models. Seasonal changes of tropical precipitation in a warmer world also reflect this “rich get richer” syndrome. Precipitation seasonality is decreased in the cold-climate state. The simulated changes in precipitation per degree temperature change are comparable to the observed changes in both the historical period and the LGM. Citation: Li, G., S. P. Harrison, P. J. Bartlein, K. Izumi, and I. Colin Prentice (2013), Precipitation scaling with temperature in warm and cold climates: An analysis of CMIP5 simulations, Geophys. Res. Lett., 40, 4018–4024, doi:10.1002/grl.50730.

1. Introduction

[2] Changes in the hydrological cycle are expected to scale with temperature changes. Recent observations, as well as model simulations of the 20th century and the response to anthropogenic increases in CO2, have shown that precipitation increases in a warming world [Meehl et al., 2007]. The water vapor holding capacity of the lower troposphere increases by ~7% per degree of warming following the Clausius-Clapeyron relationship, which is well approximated by

\[ e_v = 0.6108e^{\frac{
\frac{T}{a} - 17.27}{b}} \]  

where \( e_v \) is the saturation vapor pressure (kPa), \( T \) is air temperature (°C), \( a = 17.27 \), and \( b = 237.3°C \). The observed and simulated changes in precipitation, however, are consistently smaller than the changes in the saturation vapor pressure [Allan and Soden, 2007, 2008; Adler et al., 2008; DiNezio et al., 2011]. The difference between the two reflects energetic constraints on evaporation [Allan, 2009; Allen and Ingram, 2002; Previdi, 2010; Richter and Xie, 2008]. Equilibrium evaporation, which is the theoretical rate of evaporation (including transpiration) from a large, uniform, wet or well-watered surface, is given by

\[ \dot{E}_q = R_n \frac{d e_v}{dT} = \frac{\dot{R}_n}{\gamma} \]  

where \( \dot{E}_q \) is the latent heat of vaporization of water (~2.45 MJ kg\(^{-1}\)), \( R_n \) is net radiation, and \( \gamma \) is the psychrometer constant (~0.067 kPa K\(^{-1}\) at sea level). The maximum evaporative fraction (the fraction of \( R_n \) that can be used for evaporation under equilibrium conditions) increases less steeply than the saturation vapor pressure, ~1%–4% per degree for temperatures in the range of 0°C–30°C (Figure 1). Equation (2) emphasizes the surface energy-balance constraint on evaporation. In contrast, vapor pressure deficit (a key predictor of evaporation in standard bulk formulae [e.g., Richter and Xie, 2008]) can be regarded as an outcome of rather than a constraint on evaporation [Raupach, 2000]. Indeed, Richter and Xie [2008] showed how key boundary layer properties influencing evaporation can change in response to large-scale changes in the surface energy balance. Water availability can place an additional constraint on evaporation from the land surface and hence further mute the increase of continental precipitation as temperature increases [Trenberth and Shea, 2005].

[3] Analyses of recent changes in tropical rainfall have shown that precipitation has increased markedly in wet regions and has decreased in subtropical dry regions [Adler et al., 2008; Allan and Soden, 2007; Wentz et al., 2007; Zhang et al., 2007; Allan et al., 2010]. This is also a feature of seasonal climates, with summer (monsoon) precipitation increasing more than winter (dry season) precipitation [Giorgi and Bi, 2005; Chou et al., 2013]. This phenomenon has been referred to as “the rich get richer” syndrome [Trenberth, 2011] and can be explained either as a result of increasing the amount of atmospheric water vapor [Held and Soden, 2006] or from diversion of moisture into regions of atmospheric convergence associated with changes in atmospheric circulation [DiNezio et al., 2011]. Trenberth and Shea [2005] suggested that a similar syndrome is also characteristic of extratropical regions, which is expected since these are regions of net moisture import from lower latitudes. Model simulations of the 20th and 21st centuries from the last round of the Coupled Model Intercomparison Project (CMIP3) show similar tendencies, with wetting in convergence regions and drying in the subtropics associated with weakening of the Hadley circulation. Recent simulations suggest that the “rich get richer” tendency is also expected in the 21st century, independent of the strengthening of the Walker circulation under warming climate conditions [Giorgi and Bi, 2005; Chou et al., 2013].
with a strengthening of the Walker circulation [DiNezio et al., 2011]. However, the response of the tropical circulation is influenced by multiple processes operating on different time scales (e.g., water vapor [Bony et al., 2013]), and the response of precipitation is weaker than that shown by the observational record and differs among different models [Allan and Soden, 2008].

[3] The observational record is short, and the strength of the precipitation response to temperature has been controversial [Wenz et al., 2007; Adler et al., 2008; Gu et al., 2007; Huffman et al., 2009; Trenberth, 2011]. Thus, it is still unclear whether the discrepancy between CMIP3 model results and observations is significant. Recent analyses [Izumi et al., 2013] have shown that the simulated large-scale patterns of temperature changes at the Last Glacial Maximum (LGM) are remarkably similar (though of opposite sign) to those shown in raised CO2 experiments. These signals include changes of comparable magnitude in the land-sea temperature contrast, in the magnitude of high-latitude amplification of temperature changes, and changes in seasonality in response to year-round forcing, and the simulated patterns are consistent with those in paleoclimatic or instrumental observations. Thus, the LGM experiments provide an opportunity to examine precipitation scaling with temperature and the regional patterns of precipitation changes and to determine whether these patterns are consistent with paleo-observations.

[5] Here we analyze outputs from six models that have run both LGM and raised CO2 experiments in CMIP5. We evaluate whether the raised CO2 experiments show similar changes in precipitation to the earlier CMIP3 experiments and then examine whether consistent changes are also present in the cold-climate state of the LGM. The LGM is an equilibrium experiment comparable to the CMIP5 4xCO2 experiment, but we also use the CMIP5 1% CO2 per year transient experiment in our analyses. Finally, we examine the consistency between simulated and observed changes in precipitation scaling at the LGM and during the historic period to determine whether simulated changes in precipitation are realistic.

2. Methods

[6] Six CMIP5 models (IPSL-CM5A-LR, MPI-ESM-P, MIROC-ESM, CCSM4, MRI-CGCM3, and GISS-E2-R) have performed both the LGM and raised CO2 experiments. In our analyses, we use five simulations. Following the CMIP5 nomenclature, these are the Last Glacial Maximum (lgm), a preindustrial control simulation (piControl), a 20th century simulation (historical), a transient 1% per year increase in CO2 over the simulation (1pctCO2), and an abrupt change to 4xCO2 (abrupt4xCO2). lgm, piControl, and abrupt4xCO2 are equilibrium experiments, and historical and 1pctCO2 are transient experiments. The boundary conditions for each experiment are described in Taylor et al. [2012]. The lgm experiment represents a cold-climate state, in response to low greenhouse gas concentrations and expanded Northern Hemisphere ice sheets. The 1pctCO2 and abrupt4xCO2 experiments represent warm-climate states, in response to increased greenhouse gas concentrations. The CO2 concentration at the end of 1pctCO2 is similar to the CO2 concentration used in the abrupt4xCO2 experiment. To provide an alternative realization of a warm-climate state, we therefore sampled the middle part of the 1pctCO2 experiment (model years 86–115) when the CO2 level was approximately 750 ppm. The total forcing in the lgm and 4xCO2 experiments is similar but, although greenhouse gases are the dominant contributor to the tropical forcing at the lgm experiment, they contribute only about half (2.85 Wm−2) of the total global forcing [Braconnot et al., 2012].

[7] To compare the results from different models, with different spatial resolutions, the outputs of each model were regridded onto a common 2° × 2° grid. Land grid-cells were defined as those 2° × 2° cells with a land fraction of ≥40%. The near-surface air temperature (tas) was used over the land and sea ice–covered sea surface (sic ≥40%), and the sea surface temperature (tos) was used over the ocean (sic < 40%). (We use tas for ocean temperatures to facilitate comparisons with historical and paleoreconstructions of sea-surface temperatures; differences in tas and tos in ice-free areas are negligible.) The changes in precipitation and temperature for each experiment and model are expressed as anomalies from that model’s piControl (experiment minus control), except in the case of the historical simulation, where the anomaly is calculated as the difference between the first and last 27 years of the simulation. We adopted this approach because the temperature at the beginning of the historical run is different from the corresponding PI simulation for most of the models (CCSM4, GISS-E2-R, IPSL-CM5A-LR, and MIROC-ESM); the 27 year interval is the length of the baseline period used for the calculation of anomalies in the HadCRUT4 data set. Area-averaged values are calculated for the globe, the tropics (here defined as 30°N–30°S) and the extratropics (≥30°N and >30°S). We analyzed the seasonal climate changes in terms of changes in the wettest and driest month ([mean precipitation of the wettest month (MPWE) and mean precipitation of the driest month (MPDR)]. The delimitation of wet and dry regions was made using precipitation deciles of the
Figure 2. The change in precipitation (%) as a function of the change in global temperature (°C) as simulated by each of the six CMIP5 models (IPSL-CM5A-LR, MPI-ESM-P, MIROC-ESM, CCSM4, MRI-CGCM3, and GISS-E2-R) at the Last Glacial Maximum (LGM), from the historical run (average for period 1979–2005 CE), the 1% CO₂ run (1pctCO₂, average for model years 86–115), and the 4xCO₂ run. The left-hand plot shows the global relationship, while the right-hand plots shows the change in global precipitation (%) over (red) land and (blue) ocean as a function of the change in global land and ocean temperature (°C).

piControl for each model in order to capture the most extreme states (although similar results are obtained using, e.g., the upper and lower quartiles of precipitation). This results in different regions being defined as wet or dry in each model.

[8] We evaluate the realism of the lgm and historical simulations using paleoclimate reconstructions and historical observations over the land. (There are no reconstructions of precipitation over the ocean.) We use a data set of quantitative climate reconstructions for the LGM from S. P. Harrison et al. (Climate model benchmarking with glacial and mid-Holocene climates, submitted to Climate Dynamics, 2013). This data set provides reconstructions (including uncertainties) of several climate variables; here we use mean annual temperature over the land and ocean and mean annual precipitation over the land. The historical data are derived from two data sets: temperature data are from the HadCRUT4 combined land and ocean temperature data set [Morice et al., 2012], which covers the period from 1850 to 2009; precipitation data are from the GHCN (Global Historical Climatology Network, Version 2) product, which provides land precipitation data and covers the interval from 1900 to 2010 [Peterson and Vose, 1997]. The earliest part of the record is based on very few actual observations; the observed historical change is therefore taken as the difference between the mean for 1979–2005 and the mean for 1941–1970, and the simulated climate is the difference between the same years in the simulations.

3. Results: Simulated Changes

[9] The ensemble averages of the six models (see Figure S1 in the supporting information) illustrate the large changes in temperature and precipitation characteristic of the cold- and warm-climate states. The lgm simulations show changes of comparable magnitude (though opposite sign) to the 4xCO₂ simulations, consistent with the fact that the overall forcing is of comparable magnitude [Braconnot et al., 2012], and historical and 1pctCO₂ show changes intermediate in magnitude. There are consistent patterns in the large-scale temperature response in warm- and cold-climate states [Izumi et al., 2013]: the land warms/cools more than the oceans, and the high latitudes warm/cool more than the tropics. Izumi et al. [2013] also showed that there is a different seasonal response to year-round climate forcing in both warm and cold climates. These large-scale temperature patterns are broadly reflected in the changes in precipitation (Figure S1). In general, there are bigger changes in precipitation over the land than over the ocean in both warm- and cold-climate states. Changes in precipitation in the high latitudes (north of approximately 50°N) are larger than those in the midlatitudes (30°N–50°N), although the response of precipitation in the tropics does not scale straightforwardly with temperature.

[10] There is a strong relationship between changes in global temperature and precipitation, with increased precipitation in a warm climate and decreased precipitation in a cold-climate state (Figure 2). The estimate of the scaling across all the climate states and all models indicates a 2.06% ± 0.09% change per degree (Figure 2); estimates based on individual models across the climate states vary between 1.63% and 2.51% per degree. The range of values (Table S1) obtained for the lgm experiment (1.80%–2.89%) is similar to that obtained for the 4xCO₂ experiment (1.37%–2.43%). The values for an individual model are always larger in the lgm experiment than in the 4xCO₂ experiment, however, consistent with the fact that the energetic limitation on evaporation is smaller in the colder state (Figure 1). The values from the 1pctCO₂ experiment are not consistently larger than those from the 4xCO₂ experiment, but the differences in scaling between the two experiments are small.

[11] The historical simulation is the only experiment to include volcanic and solar forcing and changes in aerosols and land use. The simulated changes in temperature over the historic period are small (<1°C), as is the magnitude of the forcing (relative to the lgm or 4xCO₂ simulations), though consistent with the magnitude of changes shown by the HadCRUT4 data (Figure S2). The results obtained for the historical simulations are anomalous: while some models show an increase in precipitation over the course of the simulation, three models (GISS-E2-R, MIROC-ESM, and
MRI-CGCM3) show a negative relationship between temperature and precipitation.

[12] To quantify the impact of water availability as a constraint in the *lgm* and raised CO₂ experiments, we estimated the precipitation scaling over the land and ocean separately. The estimate of the scaling across all climate states and all models (Figure 2) indicates a 2.42 ± 0.09% change per degree over the ocean and a 1.75 ± 0.16% change over the land. This finding suggests that the change in global precipitation with temperature is slightly reduced because of the additional constraint of water supply on evaporation over land areas. Estimates of the relationship between temperature and precipitation obtained from individual models and experiments generally show that the scaling over the ocean is greater than that over the land (Table S2). Thus, the values obtained for the model ensemble mean for ocean and land, respectively, are 2.64% and 2.28% for the *lgm* experiment, 2.04% and 1.39% for the *lpctCO₂* experiment, and 2.32% and 1.33% for the *4xCO₂* experiment. Model responses over the ocean are more consistent than those over the land (Figure 2). The variability over the land probably reflects larger differences in treatment of the land among the different models (e.g., number of vegetation types, treatment of soil moisture, effective rooting depth, and inclusion of carbon cycle).

[13] The role of water limitation can also be examined by comparing the scaling over the tropical and extratropical land areas, with the expectation that water supply constraints might be less prominent in extratropical regions. In warm-climate states, the scaling of precipitation with temperature over the extratropical land (mean value, *lpctCO₂*: 2.61%°C and *4xCO₂*: 2.77%°C) is indeed greater than over tropical land areas (mean value, *lpctCO₂*: 1.00%°C and *4xCO₂*: 0.72%°C). This is also generally the case for individual

![Figure 3. Simulated changes in precipitation (%) in the wettest and driest areas of the tropical (30°N–30°S) and extratropical (>30°N and >30°S) land and ocean. The wettest and driest areas are defined separately for each individual model as those grid cells that fall in the top and bottom deciles of precipitation in the control simulation (piControl). Simulated changes in tropical precipitation (%) during the wettest month (MPWE) and the driest month (MPDR) are shown in the bottom panels for comparison.](image-url)
However, the scaling between temperature and precipitation in the 

4. Results: Comparison With Observations

Paleoclimate reconstructions show generally colder and drier conditions over the land at the LGM. The simulated changes in temperature (at grid cells where there are paleoclimate reconstructions) are colder than observed (Figure 4). Nevertheless, the simulated change in precipitation is systematically less than observed, both in the tropics and
extratropics. Thus, the scaling between temperature and precipitation changes in the models appears to be somewhat weaker than observed. However, the comparison for the historical period (Figure 4) does not show a marked discrepancy between the observed and simulated changes in precipitation scaling with temperature. In the tropics, the models showing greater warming over the land than the observations are also wetter, and those that show cooler conditions are drier. In the extratropics, most models show increased temperatures and little or no change in precipitation, whereas the observations show modest increases in both temperature and precipitation. Given the failure to identify a systematic bias in the historical simulations, it seems likely that differences between simulated and observed changes at the LGM are within the range of observational uncertainty.

5. Discussion and Conclusions

[17] Global precipitation increases with warming and decreases with cooling. The relationship, as estimated here from the individual models and simulations, varies between 1.5 and 3% per degree. This range is consistent with previous model-based estimates of precipitation changes during the 20th century [Held and Soden, 2006], using future scenarios [Allen and Ingram, 2002; Held and Soden, 2006], and based on PMIP2 LGM experiments [Boos, 2012].

[18] There are consistent patterns in the nature of the scaling of large-scale precipitation changes with temperature in warm- and cold-climate states. Thus, the change in precipitation with temperature is greater over the ocean than over the land in both warm and cold climates. Similarly, over land areas, the change in precipitation per degree temperature change is larger in the extratropics than the tropics. Changes in tropical precipitation are greatest in areas that are currently wet, resulting in increased precipitation in warm-climate states and decreased precipitation in cold-climate states. The seasonality of precipitation in the tropics also changes in a consistent way, with increased seasonality in warm-climate states and decreased seasonality in cold-climate states.

[19] At global and regional scales, the scaling of precipitation change with temperature is consistently much less than the 7% per degree change in atmospheric water vapor predicted by the Clausius-Clapeyron relationship, but consistent with the values expected taking into account energetic constraints on evaporation (~1%–4% per degree for temperatures in the range of 0°C–30°C). The steeper scaling over the ocean compared to the land, and over the extratropical land compared to the generally more arid tropical land, suggests that water limitations reduce modeled precipitation/temperature scaling by about a quarter.

[20] Both the spatial patterns and the scaling relationships are broadly consistent with earlier analyses [see, e.g., Held and Soden, 2006; Liu et al., 2009; Previdi, 2010; Boos, 2012]. The response of precipitation in 20th century simulations is generally weaker than that shown by the observational record [Allan and Soden, 2008]. The results presented here show that the same applies to the response of precipitation differences between LGM and present. However, evaluation using historical observations suggests that differences between the observed and simulated precipitation scaling do not exceed the reconstruction uncertainty. Broadly speaking, the evaluations suggest that models are able to capture the large-scale constraints on precipitation scaling in a realistic way.

[21] The inclusion of paleosimulations in the CMIP5 suite of model experiments makes it possible to demonstrate the robustness of simulated behavior across a wider range of climates. More importantly, it offers additional possibilities for model evaluation. Our analyses show that the energetic constraints on evaporation (and water limitation over the land) constrain the simulated changes in precipitation scaling with temperature in a realistic way. While improvements in the availability of paleoclimate reconstructions, and analysis of precipitation scaling over a wider range of paleoclimates, would be useful, these analyses demonstrate the utility of inclusion of paleoclimate simulations as CMIP5 experiments.

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References


Giorgi, F., and X. Bi (2005), Updated regional precipitation and temperature changes for the 21st century from ensembles of recent