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Monsoon changes for 6000 years ago: Results of 18 simulations from the Paleoclimate Modeling Intercomparison Project (PMIP)

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Abstract. Amplification of the northern hemisphere seasonal cycle of insolation during the mid-Holocene causes a northward shift of the main regions of monsoon precipitation over Africa and India in all 18 simulations conducted for the Paleoclimate Modeling Intercomparison Project (PMIP). Differences among simulations are related to differences in model formulation. Despite qualitative agreement with paleoecological estimates of biome shifts, the magnitude of the monsoon increases over northern Africa are underestimated by all the models.

Introduction

The reliability of climate models in predicting climate change on the regional scales where ecological and economic impacts of global climate change occur needs to be assessed. Current models exhibit a range of skill in simulating today's climate [*Gates et al.*, 1998] and can produce quite different regional responses to a given forcing [*Kattenberg et al.*, 1995]. Even if models could simulate today's climate perfectly, this would not guarantee accurate simulation of climate change. Attempting to simulate substantially different past climatic conditions allows one to evaluate model responses to climate changes. The Paleoclimate Modeling Intercomparison Project (PMIP) [*Joussaume and Taylor*, 1995] has coordinated carefully designed simulations of past climates. The aim is to compare the climate response of different models, to determine, where possible, the reasons for differences in response, and to use paleodata to discriminate, where appropriate, between realistic and unrealistic responses. The PMIP project was strongly influenced by the pioneering work of the Cooperative Holocene Mapping Project [*COHMAP-Members*, 1988].

The mid-Holocene period (6000 years ago, 6 ka BP) was chosen as one focus for PMIP because there is compelling evidence that regional climates were significantly different from today [*Wright et al.*, 1993], extensive quality-controlled data syntheses are available [e.g., *Yu and Harrison*, 1996; *Prentice and Webb III*, in press] and the external forcing can be precisely defined (Tab. 1). Each of the 18 PMIP models (Tab. 2) simulated equilibrium climates under present-day forcing and with 6 ka BP orbital parameters [*Berger*, 1978] and atmospheric CO_2 concentration [*Raynaud et al.*, 1993]. Sea surface temperatures (SSTs), vegetation and land-surface characteristics were prescribed as for present day in both simulations. Note that fixing the SSTs limits the impact of CO_2 changes. The estimated differences in SST between 6 ka BP and today are generally small [e.g., *Ruddiman and Mix*, 1993]. This simple design aids intercomparison but can not be expected to yield complete agreement with paleodata. The results for each model were averaged over at least a 10-year period, following a one-year

Table 1. Forcing Conditions for the 6 ka BP Simulations.

	Present-day	6 ka BP
orbital parameters		
Eccentricity	0.016724	0.018682
Axial tilt (°)	23.446	24.105
perihelion-180° (°)	102.04	0.87
CO_2 concentration	345 ppm or C_{ctrl}	280 ppm or C_{ctrl}*(280/345)

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Table 2. Model Designation, Horizontal (Spectral Modes or Longitude x Latitude Number of Grid Points) and Vertical (L) Resolutions. Ancillary Information Indicates Whether the Model Versions have been used for Future Climate Simulations (C), or in : (A) Atmospheric Model Intercomparison Project, (F) Feedback Analysis and Intercomparison for GCMs and Observations for Cloud and Snow Feedbacks, (P) Project for Intercomparison of Land-surface Parametrization Schemes. Model Descriptions are Available at the PMIP web Site (http://www-pcmdi.llnl.gov/pmip).

Model designation		Version	Resolution	CAFP Yes/No
1- BMRC	Bureau of Meteorology Research Centre (Australia)	V.3.3	R21, L9	YYYN
2- CCC2	Canadian Centre for Climate Modelling and Analysis (Canada)	V.2	T32, L10	YYYN
3- CCM3	NCAR Climate Community Model (USA) run at CCR	V3	T42, L18	YNNY
4- CCSR1	Center for Climate System Research (Japan)	V.5.4.02	T21, L20	YYNN
5- CNRM2	Centre National de Recherches Météorologiques (France)	V.2	T31, L19	YNNY
6- CSIRO	Common wealth Scientific and Industrial Research Organisation (Australia)	V.4-7	R21, L9	YYNN
7- ECHAM3	Max-Planck Institut fuer Meteorologie (Germany) run at Bremen Univ.	V.3.6	T42, L19	YYYY
8- GEN2	National Center for Atmospheric Research (USA) GENESIS model	V. 2	T31, L18	YYNN
9- GFDL	Geophysical Fluid Dynamics Laboratory (USA)		R30, L20	YNNY
10- GISS	Goddard Institute for Space Studies (USA)	V.II prime	72*46, L9	YYYY
11- LMD4	Laboratoire de Météorologie Dynamique (France) run at LSCE	V. 4ter	48*36, L11	YNYY
12- LMD5	Laboratoire de Météorologie Dynamique (France) run at LSCE	V. 5.3	64*50, L11	YYYY
13- MRI2	Meteorological Research Institute (Japan)	V. II.b	72*46, L15	YYNN
14- MSU	Moscow State University (Russia)		10x15, L3	YNNN
15- UGAMP	UK Universities Global Atmospheric Modelling Programme	V.2	T42, L19	YYNN
16- UIUC11	University of Illinois Urbana-Champaign (USA)		72*46, L11	YNNN
17- UKMO	UK Meteorological Office Unified Model	V. 3.2	96*73, L19	YNYY
18- YONU	Yonsei University (Korea)	Tr7.1.1	72*46, L8	NNNN

spin up. The seasons were defined using a modern calendar with vernal equinox fixed at 21 March [*Joussaume and Braconnot*, 1997].

Model Results

We focus on the response of the northern hemisphere monsoons, which are a consistent feature of the model simulations and paleodata. At 6 ka BP, insolation increases (decreases) in the northern hemisphere summer (winter) by about 5% over the northern hemisphere. The seasonal cycle of temperature over land is enhanced in the northern hemisphere and reduced in the southern hemisphere. The summer warming reaches a maximum of 2-2.5 °C over eastern Asia (not shown). The resulting increased land-sea temperature contrast enhances low-level convergence into the monsoon low over northern Africa and Eurasia. This increases precipitation along the southern side of the monsoon flow over subtropical Africa and northern India

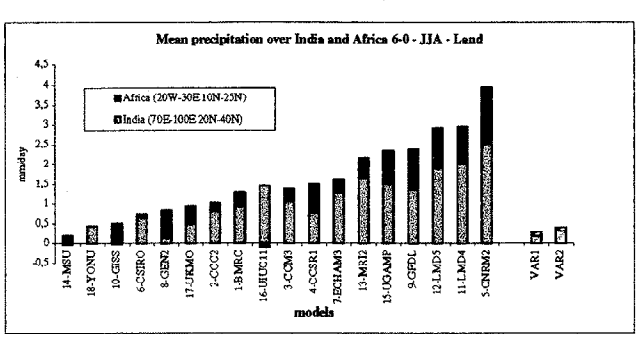

Figure 1. Simulated precipitation changes over northern India (70-100E, 20-40N; grey color) and northern Africa (20W-30E and 10-20N; black), compared to interdecadal variability estimated from standard deviations based on long coupled simulations of the GFDL (VAR-1) [*Manabe and Stouffer*, 1996] and UKMO (VAR-2) [*Tett et al.*, 1997] models.

(Figure 1). These changes are consistent with earlier simulations [e.g. *Kutzbach and Otto-Bliesner*, 1982] and qualitatively robust across all PMIP models. Figure 1 shows that almost all models produce an increase in precipitation over northern India (70-100E, 20-40N) and over northern Africa (20W-30E, 10-25N). Precipitation increases vary from model to model, from 0. to 2.5 mm/day over northern India and 0. to 1.5 mm/day over northern Africa, changes that are generally larger than interdecadal variability simulated by coupled atmosphere-ocean models (Figure 1). The simulated increase of monsoon precipitation also differ spatially which explains why a few models exhibit no change in these limited-area averages.

The range of precipitation increases over northern India is correlated (correlation coefficient r= 0.75, significant at the 95% level) with the range of warming over Central Asia (40-150E, 30-60N). The latter varies across models from 0.6 to 2.1°C, which is greater than the standard deviation of decadal variability estimated by coupled climate models (0.16°C for GFDL and 0.40 °C for UKMO) (Figure 2a). This correlation is consistent with land/sea thermal contrast driving the monsoon [*Prell and Kutzbach*, 1987]. Although all the models are forced by the same changes in incoming radiation (ca 22 Wm^{-2} over central Asia in JJA), model differences in present-day planetary albedo lead to differences in the shortwave forcing (defined here as the insolation-induced change in net shortwave flux at the top of the atmosphere when present day planetary albedo is assumed). This explains part of the temperature range (Figure 2b). Differences in cloud feedbacks [e.g. *Cess et al.*, 1990] also modulate the net top of the atmosphere radiation changes and account for some of temperature differences (Figure 2b). For example, model 2, with the smallest warming, has a negative cloud feedback arising from increased reflection of solar radiation, whereas model 12, with the second largest warming, has the largest positive cloud feedback. Differences in cloud feedbacks can arise for many reasons, including differences in land-surface parameterizations [*Masson and Joussaume*, 1997] or cloud schemes [e.g. *Liao et al.*, 1994].

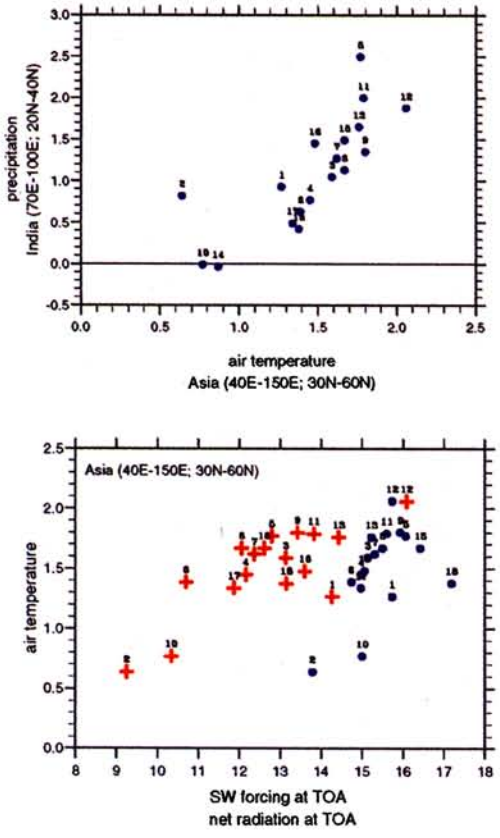

Figure 2. Monsoon changes in June-July-August. a) Relationship between precipitation changes over northern India (70-100E, 20-40N) (mm/day) and changes of surface air temperatures over central Asia (40-150E, 30-60N) (°C) for all the simulations. b) Relationship over central Asia between changes in surface air temperature (°C) and net short wave forcing (dots) (Wm^{-2}) and changes in the net top of the atmosphere radiation (crosses) (Wm^{-2}) for all the models except model 14 (for which these diagnostics were unavailable).

Comparison with Paleoclimate Data from Africa

Geomorphic and biostratigraphic evidence indicates that northern Africa was considerably wetter than today and extensively vegetated in the mid-Holocene and that lakes were present in now-dry basins as far north as 27N, and in the northwestern part of north Africa [*Jolly et al.*, 1998a]. Pollen- and plant macrofossil-based reconstructions of 6 ka BP vegetation in northern Africa show that grasslands (steppe) were found at least as far north as 23N, and that no plants characteristic of desert were present south of 20N at that time [*Jolly et al.*, 1998b; Figure 3b].

In order to compare model results with paleovegetation data, we consider the annual mean changes in precipitation averaged over northern Africa from 20W to 30E. All the models produce increased precipitation in the northern part of their present-day rainbelt (Figures 3a and 3c), in qualitative agreement with reconstructed vegetation shifts (Figure 3b). Maxima of precipitation changes vary from model to model (Figure 3a). The models in which the maximum change at 6 ka BP is furthest north (models 1, 8, 11, 13 and 18) are also those in which the northern limit of the rainbelt is furthest north in the control. Using the known moisture-balance requirements of grassland vegetation (Figure 3a), we estimate that at least an additional 100 mm/year would be required for most models to sustain grassland at 23N, i.e. more than twice as much as simulated in this area. Thus none of the model precipitation changes would sustain steppe as far north as reconstructed from the data [see e.g. *Harrison et al.*, 1998]

Discussion and Conclusion

The qualitative consistency in the monsoon enhancement at 6 ka BP is remarkable given the wide range of resolution and the variety of subgrid-scale parameterizations in the models. It contrasts with the wide range of response shown in simulations with doubled CO_2 [*Kattenberg et al.*, 1995]. At 6 ka BP, the patterns of response are anchored by the land/sea distribution because the increase in land-sea temperature contrast drives the hydrologic response. This is true even in simulations with computed SSTs [e.g. *Liao et al.*, 1994]. Nevertheless, there are large differences in model sensitivity. We have shown that differences in precipitation intensity can be related to differences in both the present day albedo and cloud feedbacks. To fully understand the origin of differences in these feedbacks on climate, specific sensitivity experiments performed with individual models will be required.

None of the PMIP models produces a large enough northward shift of the monsoon rainfall, consistent with the reconstructed changes in vegetation over northern Africa. We have

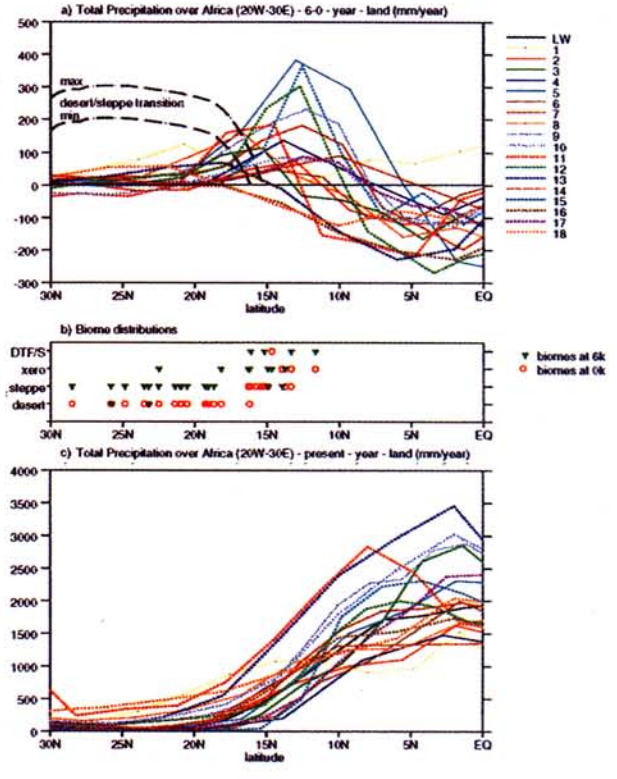

Figure 3. Zonal average climate and distribution of biomes in northern Africa (20W-30E). (a) 6 ka BP minus present annual precipitation (mm/year). The hatched lines are estimated upper and lower bounds for the excess precipitation required to support grassland, based on present climatic limits of desert and grassland taxa in paleoecological records (D. Jolly and I.C. Prentice, unpublished results). b) Biome distributions (desert, steppe, xerophytic and dry tropical forest/savannah (DTF/S), 6 ka BP and present c) Simulated present day annual precipitation (mm/year) compared to climatology [*Legates and Willmott* (LW), 1990].

deliberately omitted the vegetation, lake and ocean temperature feedbacks on climate. Experiments in which mid-Holocene vegetation changes in the Sahara were (a) represented by decreases in surface albedo [*Street Perrott et al.*, 1990], (b) prescribed from observations [*Kutzbach et al.*, 1996], or (c) calculated interactively [*Texier et al.*, 1997; *Claussen and Gayler*, 1997] produced larger increases and a greater northward extension of the rainfall over northern Africa. Including the observed occurrence of large lakes [*Coe and Bonan*, 1997] or coupling with a full ocean model [*Kutzbach and Liu*, 1997; *Hewitt and Mitchell*, 1998] has the same effect though no experiment produces the northward extent derived from paleodata. Future work will determine if this failure is model-dependent or results from the omission of synergistic interactions between the ocean, vegetation and surface-water storage. Such investigations are particularly important, given the state-of-the-art simulations of future climates also do not include vegetation and surface-water storage feedbacks.

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