

PALEOCLIMATE SIMULATIONS FOR NORTH AMERICA OVER THE PAST 21,000 YEARS: FEATURES OF THE SIMULATED CLIMATE AND COMPARISONS WITH PALEOENVIRONMENTAL DATA

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Abstract — Maps of upper-level and surface winds and of surface temperature and precipitation illustrate the results of a sequence of global paleoclimatic simulations spanning the past 21,000 years for North America. We review a) the large-scale features of circulation, temperature, and precipitation that appear in the simulations from the NCAR Community Climate Model Version 1 (CCM 1), b) the implications of the simulated climate for the past continental-scale distributions of three plant taxa (*Picea* spp., *Pseudotsuga menziesii*, and *Artemisia tridentata*), which are broadly representative of the vegetation across the continent, and c) the potential explanations in terms of atmospheric circulation or surface energy- and water-balance processes for mismatches between the simulations and observations. Most of the broad-scale features of previous paleoclimatic simulations with the NCAR CCM 0 for North America are present in the current simulations. Many of the elements of a conceptual model (based on previous climate simulations) that describes the controls of paleoclimatic variations across North America during the past 21,000 years are found in simulations reviewed here. These include 1) displacement of the jet stream by the Laurentide Ice Sheet to the south of its present position in both winter and summer, 2) generation of a “glacial anticyclone” over the ice sheet at the LGM, and the consequent induction of large-scale sinking motions induced over eastern North America, 3) changes in the strength of surface atmospheric circulation features through time, including weakening of the Aleutian low in winter, and strengthening of the eastern Pacific and Bermuda high-pressure systems in summer as the ice sheet decreased in size, 4) development of a “heat low” at the surface and a strengthened ridge in the upper-atmosphere over the continent at the time of the maximum summer insolation anomaly, 5) increases in summer temperature earlier in regions remote from the ice sheet (these increases appear earlier in the present (CCM 1) simulations than in the previous (CCM 0) ones, however), and 6) continuation of negative winter temperature anomalies into the middle Holocene. In general, simulated surface conditions that are discordant with paleoenvironmental observations can be attributed to the simulation of particular atmospheric circulation patterns (e.g. those that suppress precipitation or advect warm air into a region), with these mismatches amplified in Beringia and the southeastern United States by surface energy- and water-balance processes.

INTRODUCTION

Coordinated analysis of paleoclimatic simulations and observations provides a perspective on how the climate system has responded to past changes in its controls. Such a perspective, however, must be qualified by the uncertainties that exist in both simulations and observations. Paleoenvironmental data document the scope and nature of past climatic variations and sometimes can be interpreted quantitatively, in terms of particular climate variables. Paleoenvironmental data are unable, however, to point unambiguously to the ultimate cause of the climatic variations, because there may be more than one way to generate a specific variation of climate. Climate models, inasmuch as they are physically or mechanistically based, can be used to infer the causes of past climatic changes, but only when the models are known to be correct, a situation as yet not realized. Neither approach can therefore provide a full picture of the controls of past climatic variations or the consequences of these variations, but the *iterative* examination of simulations and observations (“data-model comparisons”) allows progress toward the general objective of understanding how the climate system operates and how it has varied in the past. The purpose of this paper is to examine a set of climate-model simulations and paleoclimatic syntheses in order to understand the controls of continental-scale climatic variations for North America from the last glacial maximum (LGM) to the present, and to identify and explain the cases where the simulations and paleoclimatic data do not agree.

A previous iteration of a coordinated analysis that focused on North America examined a sequence of simulations performed with CCM 0, an early version of the NCAR (National Center for Atmospheric Research) Community Climate Model (Wright *et al.*, 1993; Webb *et al.*, 1993b). The individual analyses that were part of that iteration included: 1) the quantitative comparison of model output with fossil-pollen and lake-level data in eastern North America (Webb *et al.*, 1987; Webb *et al.*, 1993a), and 2) the qualitative comparisons of simulated climate and regional-scale synthesis of fossil-pollen, plant-macrofossil and/or lake-level data for the western

United States and northern Mexico (Barnosky *et al.*, 1987; Thompson *et al.*, 1993), and Beringia (i.e. northwest Canada, Alaska and eastern Siberia; Barnosky *et al.*, 1987; Anderson and Brubaker, 1993, 1994). These comparisons showed how atmospheric circulation, as controlled by changes in the size of the Laurentide Ice Sheet (LIS) and seasonal cycle of insolation, explains many of the changes in regional-scale patterns evident in the paleoenvironmental data. These changes include, for example, those in the broad-scale patterns of effective moisture in western North America (Thompson *et al.*, 1993), and the general reorganization of the vegetation of eastern North America, between the LGM and present (Webb *et al.*, 1993a and b).

Some aspects of the previous simulations were strikingly different from the observations. These aspects included the simulation during glacial times of conditions warmer than those apparent in the paleoenvironmental data for the southeastern United States (Webb *et al.*, 1993a) and Beringia (Barnosky *et al.*, 1987; Anderson and Brubaker, 1994), and the absence of strong gradients in simulated effective moisture anomalies over the western United States and northern Mexico during glacial times (Thompson *et al.*, 1993). Moreover, there was a general tendency in North America for simulated glacial-to-interglacial warming to occur earlier in the simulations than is apparent in the data. General sources for potential discrepancies between simulations and observations include: 1) inadequacy of the climate model, 2) misinterpretation of the paleoenvironmental data, and 3) shortcomings in experimental design, including the specification of boundary conditions (Webb *et al.*, 1987; Thompson *et al.*, 1993). In the set of new simulations described here, the potential sources of mismatch between simulations and observations were addressed, in the hope of diminishing the discrepancies between simulations and observations, and consequently increasing our general understanding of the past climatic variations registered by the data.

In this iteration of a coordinated analysis of paleoclimatic simulations and observations for North America, we use the sequence of paleoclimatic simulations performed with CCM 1, and described in detail by Kutzbach *et al.* (this vol.). New sets of

boundary conditions (e.g. ice-sheet topography, atmospheric carbon dioxide concentrations, etc.) were developed for these simulations, and the motivations for this development are reviewed by Kutzbach and Ruddiman (1993), Kutzbach *et al.* (1993b), and Kutzbach *et al.* (this vol.). Potential model inadequacies in previous simulations were addressed by the application of CCM 1, which contains explicit representations of many processes that operate at the land or ocean surface, but which were parameterized or held constant in the earlier simulations. Sea-surface temperatures, which were prescribed in CCM 0, are computed in CCM 1 using a mixed-layer ocean representation. (CCM 1 still has the same spatial resolution as CCM 0 (4.4 degrees latitude by 7.5 degrees longitude), and crude depiction of topography (the western Cordillera is represented as a broad dome, 1.5 km high at the center).

The paleoclimatic simulations are compared here with paleoecological observations (fossil-pollen abundance and plant macrofossils). We used ecological response surfaces (Bartlein *et al.*, 1986; Prentice *et al.*, 1991) to generate continental-scale maps of estimated probabilities of occurrence of three plant taxa (based on the climate simulations), and compared these with observed distributions. The observed distributions are based on syntheses of the paleoecological data that included pollen relative-abundance data from the less-mountainous eastern and northern parts of North America, as well as the plant-macrofossil and pollen data from the sparser network of sites from the mountainous western North America. This strategy enabled us to make continental-scale comparisons between simulations and observations, which is important in diagnosing the performance of the relatively coarse-scale climate model. The paleoecological data were drawn for the most part from on-line data bases (e.g. the North American Pollen Data base at <http://www.ngdc.noaa.gov/paleo/napd.html>), and research data sets at Brown University and the U.S. Geological Survey.

An additional step in the present analysis is the examination of areal averages of climate-model output for regions where discrepancies between the simulations and observations remained large (the

southeastern United States (see Webb *et al.*, this volume) and Beringia), or where features in the inferred paleoclimate that had not previously been well simulated appear better represented in the current simulations (e.g. the southwestern United States). The specific climate variables examined in these areal averages were chosen to represent the atmospheric circulation controls that directly influence the climate of a region, along with the surface water- and energy-balance processes within the region that may modify the impact of those circulation controls (see Webb *et al.*, this vol.).

We first review the large-scale features of the simulated climate, as represented by maps of atmospheric circulation and surface temperature and precipitation. Next, response surfaces are used to transform the climate-model simulations into the distributions of three plant taxa which are then compared with those documented by paleoecological data. Finally, the areal averages of climate-model output are examined for several key regions in order to ascertain the role of atmospheric circulation and surface energy- and water-balance processes in contributing to mismatches between the simulations and observations.

GENERALIZED FEATURES OF THE SIMULATED CLIMATE

Previous analyses of climate-model simulations revealed broad-scale features of the simulated paleoclimate that were apparent in the patterns of surface winds, upper-level winds, and of surface temperature and precipitation. Maps of these variables and of sea-level pressure showed the strength and location of atmospheric circulation features such as the Aleutian low and Pacific and Atlantic sub-tropical high-pressure systems (Kutzbach, 1987; Kutzbach *et al.*, 1993a) and the polar jet stream (Kutzbach and Wright, 1985). In preparing new maps to illustrate these features for CCM 1 (Figs. 1-4), we attempted to depict these patterns and features explicitly, by displaying the data at the resolution of the climate model and by choosing a particular representation for each variable that directly describes the features. We did not

contour the simulated climate data or use high-resolution continental outlines, because doing so gives a false sense of high spatial resolution of a model. Moreover, this approach obviates the need to produce schematic or “cartoon” summaries of the simulations as in COHMAP Members (1988) or Thompson *et al.* (1993).

We also “downscaled” the model output to produce an additional set of temperature and precipitation maps for the continent (Figs. 5 and 6); these are used in the second part of our analysis to generate taxon-distribution maps. In preparing these temperature and precipitation maps, we followed the convention of applying anomaly values (paleoclimatic experiment minus control simulations) to a high-resolution (25-km) grid of observed modern climate. Bilinear interpolation was used to interpolate the model anomalies onto the 25-km grid, and we followed the same procedure as Kutzbach *et al.* (this vol.) in adjusting the surface-temperature data to reflect the differing elevations of the model’s grid points and those of the 25-km grid.

This downscaling (and consequent increase in spatial resolution) reduces the sometimes considerable bias exhibited by the GCM (general circulation model) simulations, but carries with it the assumption that this bias is constant over the sequence of experiments. This portrayal of the simulated climate on a 25-km grid also has the desirable property of inducing realistic topographic influence on the simulated fields of temperature and precipitation, but requires the (almost certainly incorrect) assumption that the specific local influence of topography on climate remains constant over time.

The sequence of simulations using CCM 0 and described by Kutzbach *et al.* (1993a) used a heterogeneous set of boundary conditions, with some (e.g. insolation) expressed on a calendar-age scale, and others (e.g. ice-sheet size) on a radiocarbon-age scale. The “18 ka” CCM 0 simulation thus combined insolation appropriate for 18,000 cal yr BP with ice-sheet size and sea-surface temperature conditions appropriate for 21,000 cal yr BP (ca. 18,000 ^{14}C yr BP). (The synthesized fossil-pollen and lake-level data sets used in the previous analyses were also expressed on a radiocarbon-age scale.) The

difference in insolation at the top of the atmosphere between 21,000 and 18,000 (calendar) years ago is not large, and so does not affect the simulation results much. At other times, however, the difference in insolation between, say, 12,000 cal yr BP and 12,000 ^{14}C yr BP (approximately equivalent to 14,000 cal yr BP) is larger, and may affect some interpretations of the difference between the sets of simulations.

For the purpose of comparison of the current sequence of simulations with those done with CCM 0 in Kutzbach *et al.* (1993a), the two sequences of experiments are organized as follows (and are summarized in Table 1): conditions at the LGM are represented by the “18 ka” (radiocarbon yrs) CCM 0 experiment and the 21 ka (calendar yrs) CCM 1 experiment, those for the late full-glacial/early late-glacial by the “15 ka” CCM 0 and 16 ka CCM 1 experiments, those for the late-glacial by the “12 ka” CCM 0 and 14 ka CCM 1 experiments, and those for the early-Holocene conditions by the “9 ka” CCM 0 and 11 ka CCM 1 experiments respectively. The 6 ka experiments that represent mid-Holocene conditions for both models are directly comparable. For the LGM, late-glacial, early Holocene and mid-Holocene, the key features of upper-level winds, sea-level pressure patterns and surface winds, and surface climate (temperature and precipitation), are described for each model in Table 1, along with the principal differences between sequences of simulations. The specific motivation for the design of a sequence of experiments is discussed further in Kutzbach *et al.* (1993b), Webb and Kutzbach (this vol.) and in Kutzbach *et al.* (this vol.). Ages appearing in all figures in this paper are expressed in calendar years before present.

Upper-Level Winds

The CCM 1 simulations show a pronounced control of upper-level winds by the ice sheet and are therefore similar to the simulations done with CCM 0 (Kutzbach *et al.*, 1993a) and other models (e.g. Broccoli and Manabe, 1987; Rind, 1987; Hall *et al.*, 1996), although the circulation patterns are different in detail (Table 1 and Fig. 1). The “split” in the January jet stream, and its displacement to the south of its present position produced by the large ice sheet in CCM 0 is also apparent in the CCM 1 simulations

for 21 ka (Figure 1; see also Kutzbach *et al.*, this vol.). However, the northern branch of the “split” jet is less prominent than in the earlier experiments (there is only a small region of weaker upper-level winds between the two branches). (We note that it was perhaps unfortunate that the expression “split jet” was applied to the ice-sheet induced reorganization of the circulation in previous discussions, because it does not direct attention to the key feature of the circulation reorganization, the southward displacement of the jet in both January and July, which is a fairly robust feature of these and other simulations for the LGM (e.g. Broccoli and Manabe, 1987; COHMAP Members, 1988; Hall *et al.*, 1996).)

Other features of the simulated upper-level winds include: a) simulation of the fastest winds along the southern and southeastern margin of the ice sheet, b) retreat northward across successive simulations of the fastest westerlies over the western United States, with most of the changes evident between 14 and 16 ka, and c) development of a pronounced upper-level ridge in July over the western United States, beginning at 16 ka, and with maximum expression at 11 ka.

Sea-Level Pressure and Surface Winds

The simulated surface-wind and sea-level pressure patterns illustrate the intensification in winter of the Aleutian and Icelandic lows at the LGM, and the persistence of these features into the early Holocene (Fig. 2, January maps, 21 to 11 ka). The “glacial anticyclone” established over the ice sheet at 21 ka (January and July) is more strongly expressed in these simulations by sea-level pressure patterns (e.g. regions with pressures greater than 1025mb) than by the surface winds (in contrast to earlier simulations with CCM 0 in which the anticyclonic circulation was evident in vector-wind maps).

The intensification in summer of the eastern Pacific and Bermuda sub-tropical high-pressure systems during the late-glacial period and early Holocene is also apparent (Fig. 2, July maps). In the southwestern United States, the development of a surface heat low at the time of the summer insolation

maximum (11 ka), and consequently stronger onshore flow in summer is also quite apparent. The surface heat low, onshore flow, and (subtropical) upper-level ridge (see above) together illustrate the development of a stronger-than-present southwestern (or “Mexican”) monsoon in the early Holocene (Fig. 2, July maps, 11 and 6 ka; see also Mock and Bartlein, 1995).

As is the case for the simulated upper-level winds, the surface winds and sea-level pressure patterns in July shift from glacial toward early-Holocene and modern conditions earlier than in January. For example, the region of high pressure over the ice sheet in July at 21 ka is much reduced at 16 ka, whereas in January at 16 ka, the high-pressure system is still evident (Fig. 2).

Relative to the earlier sequence of experiments with CCM 0, the transitions from the LGM circulation configuration to a Holocene one appear to occur earlier in the CCM 1 simulations. Explanation for this pattern may lie in 1) a combination of more detailed specification of ice-sheet topography (i.e. Peltier, 1994), which shows a greater change (decrease in elevation) between 21 and 14 ka than was assumed in the previous simulations (see Kutzbach *et al.*, this vol.), and 2) a tendency for the current simulations to have responded significantly to the increase in summer insolation between 21 and 16 ka (see next section). Overall, however, the conceptual model of changes in circulation and surface climates over North America that these simulations support is remarkably similar to the one described earlier (Table 1; COHMAP Members, 1988; Thompson *et al.*, 1993).

Temperature and Precipitation

The most striking feature of the simulated temperature patterns is the early warming in summer, beginning with the 16 ka simulation in regions distant from the ice sheet (Fig. 3, July maps). This trend culminates at 11 ka, when warmer-than-present conditions in July were simulated nearly everywhere in North America, despite the continued presence of a substantial ice sheet in northeastern North America. In Beringia at 16 ka, simulated July temperatures were near present, and at 14 and 11 ka were 5-10°C

greater than present. Similarly, simulated July temperatures in the southeastern United States were close to present values by 16 ka, and reached greater than present values at 14 ka (see Webb *et al.*, this vol.). In contrast, January temperatures remained lower than present in many regions until 6 ka or present, and were quite low over the ice sheet from 21 to 11 ka. As was the case in the previous simulations, a first-order consequence of these trends in January and July temperatures was the simulation of greater seasonality of temperature than present, with the greatest expression at 14 and 11 ka.

Simulated precipitation patterns illustrate conditions that in most places were drier than present, in both January and July, almost continuously throughout the sequence of simulations (Fig. 4). The specific circulation mechanisms that underlie these precipitation variations seem to be the ice-sheet induced reconfiguration of the winter circulation at the LGM, and the insolation-related amplification of the subtropical high-pressure systems in summer during the late-glacial and early-Holocene intervals. The major exceptions to this persistent pattern of dryness are: 1) the simulation of greater-than-present January precipitation along the west coast accompanying the shift in the jet stream at the LGM (Fig. 4, January maps, 21 to 14 ka), 2) the simulation of greater-than-present July precipitation in Beringia during the time of much warmer-than-present conditions there (16 to 11 ka), and 3) the simulation of greater-than-present July precipitation in the Southwest at 11 and 6 ka, accompanying the simulated enhancement of the southwestern monsoon in summer then.

Discussion

The primary large-scale features of the past climates simulated by CCM 1 for North America are quite similar to those produced in the previous (CCM 0) simulations. The Laurentide Ice Sheet affected circulation and temperature, particularly at the LGM, with the simulation of the displacement of the winter jet stream to the south of its present location, the presence of a glacial anticyclone, and very cold temperatures over and adjacent to the ice sheet. The sequence of simulations after the LGM also reveals how the decrease in the height and

albedo of the ice sheet, and concomitant changes in sea ice and sea-surface temperatures, influences circulation, surface temperature and precipitation. The sequence of simulations also shows the direct effect of the amplification of the seasonal cycle of insolation on both January and July temperatures, and its indirect effect on the strength of the sub-tropical high pressure systems.

CCM 1 also simulated several secondary features (e.g. of temperature, precipitation, winds, etc.) of the sequence of simulations noted previously (Webb *et al.*, 1987; Barnosky *et al.*, 1987; Thompson *et al.*, 1993) including: 1) strong northeast-to-southwest temperature gradients between the ice sheet and ice-free areas; 2) a shift through time in the location of positive temperature anomalies in July toward the center of the continent as ice-sheet size decreased; 3) a later increase of January temperatures to present values relative to that in July temperatures; 4) a decrease in precipitation in the continental interior in the early Holocene in response to simulation of the northwesterly flow around the heat low (and the existence of the upper-level ridge); 5) the simulation of easterly winds and colder and drier conditions at the LGM in the Pacific Northwest; and 6) the simulation of an effective moisture gradient across Beringia (wetter in the west, drier in the east) from 21 to 14 ka.

In summary, the principal differences between the current and previous simulations are: 1) reduced influence of the ice sheet on January upper-level circulation patterns in CCM 1 than in CCM 0; 2) more distinct variations in surface wind and sea-level pressure patterns in CCM 1, including the clear simulation of an enhanced Mexican monsoon; and 3) warming in July earlier in the sequence in CCM 1 than in CCM 0. Persistent “problems” or mismatches between simulations and observations (see below and Webb *et al.*, this vol.) include the apparently overly warm conditions simulated in summer in the southeastern United States and in Beringia. A major apparent improvement in the present sequence of simulations relative to the previous ones is the clearer depiction of the enhanced southwestern monsoon at 11 and 6 ka.

This examination of the generalized features of the simulated climate across sequences of simulations leaves open the question of whether the simulated features are indeed correct. In the next section of the paper we address this question by comparing the simulations with paleoenvironmental observations.

SIMULATIONS OF PLANT TAXA DISTRIBUTIONS AND COMPARISONS WITH OBSERVATIONS

Two general strategies exist for the comparison of paleoclimatic simulations and observations. In the *inverse-model* approach, paleoclimatic data, such as fossil-pollen data, are interpreted in climatic terms, and the resulting inferred-climate values are compared with simulated values. Webb *et al.* (1993a) and Huntley and Prentice (1993) used this approach to examine the CCM 0 simulations performed by Kutzbach *et al.* (1993a), and Webb *et al.* (this vol.) and Prentice *et al.* (this vol.) used it for the CCM 1 simulations. In the *forward-model* approach, climate-model output is transformed into simulations or estimations of paleoclimatic data, and these are directly compared with the observations. Webb *et al.* (1987; this vol.) used this approach to compare the observed abundances of fossil-pollen in eastern North America with those estimated using model output and ecological response surfaces. This forward-model approach is also implicit in the application of the Biome model by Kutzbach *et al.* (this vol.) to summarize the multivariate model output in terms of a single variable (i.e. biome type).

Each approach has its own set of merits and disadvantages. The inverse approach has the potential to provide documentation of the actual paleoclimate in terms that can be directly compared with model output. However, the application of this approach requires extensive modern observations to calibrate the relationships used to transform paleoecological data into paleoclimatic estimates. Furthermore, more than one combination of paleoclimatic variations potentially can produce the same responses in vegetation. Consequently, it may not be possible to obtain unique or unambiguous reconstructions of past climatic variations using the

inverse-model approach, although this situation can be mitigated to some extent by using “multiproxy” approaches. The forward-model approach has the advantage of being more generally applicable than the inverse approach, at a minimum requiring only a simple listing of the climatic requirements of a particular plant taxa. The forward method also lends itself to applications where the network of paleoclimatic data is sparse. In such instances it may be possible to use the sparse network of data to subjectively “fill-in” a map that can be compared with forward-model simulations. However, the forward-model approach involves the comparison of simulations and observations in terms of a climatic indicator, such as the distribution of a climatically sensitive plant species, as opposed to a climatic variable, such as July temperature. At first glance, this approach may therefore seem less appealing than the inverse approach, if the goal is to critique the simulation of a particular climate mechanism that might be described by one or more climate variables. A comparison of simulated with observed *Picea* pollen, for example, may seem less meaningful than a comparison of maps of simulated and inferred July temperature. However, because the correct simulation of a particular plant taxon requires the proper simulation of several climate variables simultaneously, the forward approach may actually be the more powerful one. (This point also holds for biome simulations, as in Kutzbach *et al.*, this vol.)

In the present comparison of paleoclimatic simulations and observations, we used the forward-model approach for the reasons outlined above and because it is the only one currently applicable to the whole of North America. Although the inverse approach based on fossil-pollen data can be applied in eastern North America (Webb *et al.*, this volume) it is not yet feasible to use this approach in the more topographically complex western and northwestern North America. We applied the forward-model approach by examining how the sequence of climate-model simulations altered the estimated distributions of plant taxa. Three taxa were chosen for these continental-scale comparisons: *Picea* spp. (spruce) which has a transcontinental distribution, is often an important taxon in the late-Quaternary vegetation history of North America, and has been used previously to argue for the validity of some aspects of

the CCM 0 simulations (COHMAP Members, 1988); and *Pseudotsuga menziesii* (Douglas-fir) and *Artemisia tridentata* (sagebrush) that together illustrate the interplay between forest and steppe in western North America.

Methods and Data Sets

We used ecological response surfaces (Bartlein *et al.*, 1986; Prentice *et al.*, 1991) to construct relationships between the probability of occurrence (incidence) of a particular taxon, and climate, as represented by January and July mean temperature and January and July total precipitation. (These four climate variables were used to represent, albeit in a crude fashion, the seasonal energy- and moisture-related controls of plant growth and abundance.) The specific method of determining the “fitted” or estimated (probability) value for a particular combination of climate values is illustrated for pollen relative abundance data in Fig. 17.7 of Webb *et al.* (1993a), and Huntley *et al.* (1995) discuss the extension of this approach to presence/absence data.

The response surfaces were calibrated using observed values of taxon presence/absence and of climate on a 25-km grid (Fig. 7). Modern climate values from over 8000 climate stations were interpolated onto the 25-km grid using a locally weighted trend surface regression procedure that allowed the effects of elevation on precipitation and temperature to be expressed in the interpolated data. These interpolated data appear in Figs. 5 and 6 (January and July maps for present). Taxon presence-absence data were obtained by digitizing range maps (Little, 1971, 1976), and placing these data on the 25-km grid. The gridded data for *Picea* spp. (spruce) were developed from the superimposition of all North American species.

The observed distributions of the three taxa appear in Fig. 7, as do the estimated probabilities of occurrence for each taxon obtained by evaluating the response surfaces using the gridded modern climate values. The response surfaces themselves are depicted by 2-dimensional “slices” (July temperature by January temperature) through the 4-dimensional climate space, with the slices arranged according to July and January precipitation. On both the maps and

response surfaces, “probability” should be taken to mean the proportion of sites near a particular location in climate space at which the taxon is observed.

Comparisons of the observed and simulated distributions of these taxa shows that the distributions are reasonably well estimated using the response surfaces and observed modern climate (Fig. 7). The observed continuous range of each taxon is matched well by the distribution of grid points with relatively high estimated probabilities of occurrence, and low, but nonzero probabilities outline regions where the taxa are discontinuously present. The proportions of grid cells for which a taxon was correctly predicted as present (i.e. a probability of occurrence > 0.40) if present, and correctly predicted as absent if absent, are 0.928 for *Picea* spp., 0.978 for *Pseudotsuga menziesii*, and 0.972 for *Artemisia tridentata*. The threshold probability value for declaring presence at a particular point in climate space, 0.40, was determined by evaluating different values using the modern data set (see also Huntley *et al.*, 1995).

The response surfaces were used to estimate the past distributions of the three taxa by “plugging-in” the simulated temperature and precipitation values that appear in Figs. 5 and 6, to produce the resulting estimated distributions that appear in Figs. 9 and 10. For combinations of values of the climate variables that are represented in the calibration (modern) data set, the response surfaces may be evaluated in a straightforward manner. For combinations of climate values that fall outside of the range of modern values, an alternative evaluation approach must be used. In previous analyses (e.g. Prentice *et al.*, 1991; Webb *et al.*, 1993a), the trends evident in the response surfaces were simply extended into regions of climate space without modern data. In the present analysis we used a different method.

This alternative method involves reducing the dimension of the climate space until enough data points are present in the neighborhood (in climate space) of the target point to estimate the probability of occurrence of a particular taxon in that region of climate space. In other words, if the climate being evaluated falls outside the 4-dimensional (January and July temperature and precipitation) climate space of the present, a 3-dimensional space (omitting

January precipitation) is next considered, and if there are still no points nearby, a 2-dimensional space (omitting July precipitation), and finally a 1-dimensional space (only July temperature) is considered. If no points are available in the 1-dimensional space, then the taxon presence/absence of the closest point in climate space is assigned. This alternative approach is faster than the previous extrapolation approach, and obviates the need to extrapolate the response surfaces.

The dimension of the climate space used to estimate the probability of occurrence of the taxa is shown in Fig. 8. Most of the time, little reduction in the dimensionality of the climate space is required, but in some regions (i.e. southwest and northwest of the ice sheet at 21 ka, and along the northern edge of the ice sheet from 21-11 ka) particularly unusual (relative to present) paleoclimates prevail. For those climates, the estimated probabilities of occurrence of the different taxa were assigned from the closest point in climate space for which a fitted value existed.

The observed paleoecological data consist of gridded (spatially interpolated) pollen data in eastern North America (see Webb *et al.*, 1993a for a description), and pollen and plant-macrofossil data plotted on a site-by-site basis in western North America (Figs. 9 and 10). The number of sites and the lesser topographic complexity of eastern North America allows the pollen data to be contoured or interpolated onto a grid, whereas in western North America, gridding the pollen and plant macrofossil data is not as straightforward. Simulated distributions in these figures are illustrated by probabilities, and a value of 0.40, selected by examination of the simulations of the present distributions, is used to indicate the simulated presence of a taxon.

The attribution of *Picea* pollen percentages in the fossil-pollen data to the taxon *Picea* spp. is straightforward, because we are not distinguishing among the different species of *Picea*. In the case of *Pseudotsuga*, however, its pollen is not distinguishable from that of *Larix* spp. (larch), and so some of the pollen percentages that describe the past distribution of *Pseudotsuga* may actually represent

Larix. *Larix* spp. produces very low abundances of pollen, which largely mitigates this source of uncertainty in the fossil-pollen data. Similarly, *Artemisia* pollen percentage data from Alaska and northwest Canada include pollen from *Artemisia* species other than *A. tridentata* (i.e. *A. frigida* and *A. alaskana*). For these reasons, we will confine quantitative comparisons of observed and predicted incidences of *Pseudotsuga menziesii* and *Artemisia tridentata* to regions south of the area occupied by the Cordilleran Ice Sheet. The plant-macrofossil data generally includes identifications of the taxa present at the species level.

For eastern North America and Beringia, we rely on visual comparisons of maps of observed pollen abundance and simulated incidence of *Picea*. In western North America south of the Cordilleran Ice Sheet, the paleoecological data consists of both fossil-pollen abundance and the presence (or absence) of *Artemisia* and *Pseudotsuga*, and the topographic complexity of the region leads to more involved map patterns of both simulated and observed incidence. Therefore, we expressed both simulated and observed incidences as presence/absence values, and examined some simple statistics to summarize the maps. In particular, for each taxon and time, we examined the proportion of sites in the paleoecological data sets (including the modern) for which the taxon was correctly predicted as present or absent by the response surfaces (Fig. 11). We also examined trends over time in the proportions of sites with the individual taxa observed or simulated to be present.

***Picea* spp. Observations and Simulations: Eastern United States and Beringia**

Picea spp. (spruce) is a widely distributed taxon at present, and regions of high incidence of spruce define well the boreal forest and montane forests north of Mexico. For eastern North America at the LGM, the simulated distribution of spruce resembles the observed one fairly well (Fig. 9). From 16 to 11 ka, the simulated distribution occupies a narrower band south of the ice sheet than does the observed, however. Inspection of the simulated climate (Figs. 5 and 6), and the response surface for *Picea* spp. (Fig. 7), along with the simulated distributions and response surfaces for taxa not shown, indicates that

this pattern reflects the simulation of a steep temperature gradient south of the Laurentide Ice Sheet during this interval, with cold conditions near the ice sheet, and near-present conditions in summer in the southeastern United States (see Webb *et al.*, this vol.). In the Pacific Northwest, a somewhat higher incidence of *Picea* spp. than observed occurs at 21 and 16 ka, and this difference will be discussed further below.

The most striking discrepancy between the simulated and observed distributions of *Picea* spp. occurs in Beringia. There, a high incidence of *Picea* spp. is simulated as early as 16 ka, with a suggestion of a subsequent west-to-east spread of boreal forest. The pattern contrasts sharply with the paleoecological data, which shows an east-to-west migration of *Picea* spp., but not until the early Holocene (i.e. after 11 ka) (Anderson and Brubaker, 1994). This early simulation of a high incidence of *Picea* spp. in Beringia can be easily attributed to the simulation of warmer- and wetter-than-present conditions in summer there at 16 ka (Figs. 5 and 6). There is also an evident west-to-east gradient in the simulated distribution of *Picea* spp. in this region from 16 to 11 ka, that reflects similar gradients in simulated July temperature and precipitation during that interval.

The changing degree of resemblance between the simulated and observed *Picea* spp. distributions through time provides a strong motivation for examining sequences of simulations in evaluating the performance of climate models, as opposed to those generated just for the extreme states. For the extreme climate states in the sequence of simulations, 21 ka (the LGM) and 6 ka (middle Holocene, with the ice sheet gone and summer insolation still greater than present), the observed and simulated patterns do not look particularly discordant. However, at the intermediate times (16 to 11 ka) the mismatch is quite striking, particularly in Beringia. Had only 21 and 6 ka been examined, a much more optimistic view of the accuracy of the past climates simulated by CCM 1 would have been gained.

***Pseudotsuga menziesii*, *Artemisia tridentata*, and *Picea* spp.: Simulations and Observations: Western United States**

The distributions of *Pseudotsuga menziesii* and *Artemisia tridentata*, two widespread taxa in western North America (Fig. 10), along with those of *Picea* spp., provide a generalized description of the spatial patterns of forest and steppe in this region. Several aspects of the paleoecological data for these taxa must be kept in mind while comparing the observed distributions maps with the simulated ones. *Pseudotsuga* pollen is indistinguishable from *Larix* pollen, and other species of *Artemisia* appear in the pollen data (particularly in Beringia), but are not simulated by the response surfaces (which are calibrated using the range of *Artemisia tridentata* alone). Similarly, because the observed modern *Picea* spp. presence/absence data was developed by superimposing the range maps of all North American species, and the western United States and northern Mexico pollen data represent the pollen contributed by several taxa, we have no way of knowing whether the appropriate species of *Picea* is being simulated correctly at each site. None of these situations, however, is likely to influence the observed maps in a major way in the present context. Also, as is apparent from the maps, the observations are sparse at times in western North America, and large areas are not represented.

The proportion of all sites in the western United States and northern Mexico (south of latitude 49°N and west of longitude 100°W), where a taxon is correctly simulated as present if present, or absent if absent, is shown for each taxon and climate-model experiment in Fig. 11. In general, each of the taxa is reasonably well simulated, with the exception of *Artemisia tridentata* for the 21 and 16 ka experiments, and *Pseudotsuga menziesii* and *Picea* spp. for 21 ka. The differences between the proportions correctly predicted for each taxon here, and those reported above in discussing the goodness-of-fit of the response surfaces arises from the fact that here we are considering a much smaller number of sites (than the ca. 32,000 25km grid points), and a distribution of sites that varies from experiment to experiment.

At 21 ka, the observed distribution of *Pseudotsuga menziesii* is much reduced from its present range and is confined mainly to the southern part of the western U.S., and is virtually absent on the Pacific Slope of Washington and Oregon. From 16 to 11 ka, the observed distribution is further reduced relative to present, and it increases from then until present (Figs. 10 and 11). In contrast, the simulated distribution of Douglas-fir at 21 ka is almost as extensive in the Pacific Northwest as at present, and low probabilities of *Pseudotsuga menziesii* are even simulated in Beringia from 16 to 6 ka. The least extensive simulated distribution of *Pseudotsuga menziesii* occurs at 16 ka, and the general observed increase in the abundance of this taxa following 11 ka is also apparent in the simulations.

The observed distribution of *Artemisia tridentata* is high south of the Cordilleran Ice Sheet at 21 ka and, much like *Pseudotsuga menziesii*, it subsequently decreases to 11 ka and then increases toward present. From 11 ka to present, the observed abundance of *Artemisia tridentata* progressively decreases in the Pacific Northwest, and the center of the range of *Artemisia tridentata* shifts toward the south. As was true for *Pseudotsuga menziesii*, the simulated distribution of *Artemisia tridentata* differs from the observed. Simulated *Artemisia tridentata* incidences are low in the Pacific Northwest from 21 to 14 ka (when observed abundances there were high, but these could have been due to species of *Artemisia* other than *A. tridentata*). From 11 ka to present, both the simulated and observed incidences increase. The observed distribution of *Picea* spp. is low between 21 and 11 ka, and it increases subsequently. As was the case for *Pseudotsuga menziesii*, the simulated incidence of *Picea* spp. in the western United States at the LGM was greater than observed.

In general, the observed increase in abundances of all three taxa from 11 ka to present seems to be matched by the simulations, both in terms of map patterns (Figs. 9 and 10), and summary statistics (Fig. 11). However, at 21 ka, the simulated distributions of the two tree taxa are more extensive than observed, while that for the steppe shrub *Artemisia tridentata* seems less extensive than observed. At 16 ka this pattern is essentially reversed.

Examination of the response surfaces for *Picea* spp., *Pseudotsuga menziesii*, and *Artemisia tridentata* indicates that the preferred locations in climate space of these taxa differ more in terms of precipitation than in temperature (Fig. 7). Examination of the patterns of simulated temperature and precipitation indicates that the simulated changes in abundance of these taxa (decreasing incidence of *Pseudotsuga menziesii* and increasing incidence of *Artemisia tridentata* from 21 to 16 ka) are the result of the simulation of moist (and to some extent, warm) conditions south of the Cordilleran Ice Sheet at 21 ka relative to those that might be inferred from the data, while at 16 ka, these conditions were replaced by the colder and drier conditions that favor sagebrush.

Comparisons among the simulated January surface winds (Fig. 2), precipitation fields at the resolution of the model (Fig. 4) and on the 25-km grid (Fig. 6), and the simulated distributions of *Pseudotsuga menziesii* and *Artemisia tridentata* (Fig. 10) suggest that the mismatch between the simulations and observations may be attributable to the coarse spatial resolution of the model. Although the westerly component of January surface winds is weaker than present at 21 ka, precipitation in the Pacific Northwest is not significantly lower than present, and the band of high precipitation along the west coast extends two or three grid cells inland, owing to the great smoothing of the complex topography of western North America in the model. As a consequence, precipitation anomalies in this region are automatically of large spatial scale (relative to the scale of the real topography). When these precipitation anomalies are superimposed on the more realistic topography of the 25-km grid, the result is the simulation of a larger region of high “coastal” precipitation than was probably actually experienced. In other words, in the real climate system of western North America, the influence of the Pacific Ocean does not extend as far inland as it does in the model. The coarse resolution of the model therefore favors the simulation of the more moisture-dependent *Pseudotsuga menziesii* and *Picea* spp. at the expense of the more drought-tolerant *Artemisia tridentata* in the western interior at 21 ka.

This same insufficient-resolution explanation may also apply to the apparent failure of the

simulations to generate sufficient precipitation in the southwestern United States to support the pluvial lakes that were maintained at relatively high levels during the interval from 21 through 14 ka (Thompson *et al.*, 1993). In the previous simulations with CCM 0 as well as in the present ones with CCM 1, simulated precipitation in the southwestern United States and northern Mexico does not seem to be appreciably augmented in either winter or summer (Fig. 4) by the southward shift of the jet stream or other circulation changes (Figs. 1 and 2). Because of the very smooth topography in the model, the orographic precipitation enhancement that played a role in maintaining the lakes is absent, as is also the lake-effect precipitation that must have occurred in the real climate system (Hostetler *et al.*, 1994).

Simulated precipitation in the Pacific Northwest decreased at 16 ka (relative to 21 ka) in both January and July (Fig. 4), in response to the circulation changes accompanying the beginning of the post-LGM northward shift of the jet stream (Fig. 1), reduction in prominence of the Aleutian low in winter, and expansion of the subtropical high in summer (Fig. 2). The consequences of these circulation changes for the pattern of climate anomalies across the western United States and northern Mexico will be explored further below.

Discussion

The mismatches for *Picea* spp., *Pseudotsuga menziesii*, and *Artemisia tridentata* between their simulated distributions and those described by the paleoecological record raise questions concerning why the mismatches occurred. There are three potential explanations: 1) the paleoclimatic simulations were correct, but the observed paleoecological data does not adequately reflect climate, and therefore cannot be used as a standard of comparison, 2) the paleoclimatic simulations were correct, but the method of estimating the past distributions of the taxa is faulty, and 3) the paleoecological data adequately reflect past climates, but the simulations are not correct. This latter case in turn would result either from model inadequacies or from an incomplete experimental design.

The first explanation could hold if migrational lags or vegetation inertia were important. For example, the simulation of the high probability of *Picea* spp. in Beringia at 11 ka, when only low incidences were observed, might have arisen because of migrational lag (i.e. both the real climate, and that simulated by CCM 1, may well have been suitable for spruce, but spruce simply had not arrived yet (Anderson and Brubaker, 1994)). However, this mismatch (more *Picea* simulated than observed) occurred at 16 and 14 ka as well, and while a 5000-yr migrational lag (from 11 ka to 6 ka) could be marginally possible, a 10,000 year-long lag (from 16 to 6 ka) would not. Moreover, when *Picea* does migrate across Beringia (Anderson and Brubaker, 1994) it does so at a rate great enough to spread rapidly across the region (i.e. in a span of time less than that between experiments here). Moreover, the observation that large-scale patterns of vegetation change in eastern North America seem parsimoniously explainable by climate (Prentice *et al.*, 1991) further discounts this potential explanation.

As to the second explanation (the use of an inappropriate method for estimating taxon distributions), the results obtained using a more mechanistically based biome model (Kutzbach *et al.*, this volume) were similar to those produced here using response surfaces: extensive areas of taiga in Beringia were simulated by Biome 1 at 16, 14 and 11 ka, and trends in the relative areas of forest and steppe in the western United States were also similar in both approaches. This similarity in result between approaches suggests that the specific mismatches noted here are not attributable to the method of transforming climate model output into simulated plant-taxon distributions. However, the use of only January and July temperature and precipitation to describe the climatic controls of plant taxon distributions opens the possibility that climate anomalies that involve combinations of other variables (such as soil moisture or net radiation) would not be adequately reflected by the simulated plant distributions.

The third explanation, incorrect paleoclimatic simulations, therefore remains as a plausible explanation for the mismatches between simulations and observations. As indicated above, incorrect

aspects of the simulations might be attributable to fundamental model inadequacy or to shortcomings in experimental design. For the simulations of *Pseudotsuga menziesii*, *Artemisia tridentata*, and *Picea* spp. in the western United States, the latter may be likely, to the extent that the spatial resolution of the model can be considered an experimental design choice. The simulations of warmer conditions in summer than those that can be inferred from the paleoecological data in other regions (the southeastern and western United States and Beringia) do not seem so easily attributable to the spatial resolution of the model. In the next section, we attempt to illustrate the climate mechanisms that underlie these specific simulations.

PROCESS ANALYSIS OF THE SIMULATIONS IN KEY REGIONS

The sequence of comparisons between the simulations and observations in this study and in Webb *et al.* (this vol.), shows, in both the southeastern United States and Beringia, simulated conditions that were warmer in summer than those implied by the data. Simulated conditions in the western United States at the LGM were milder and moister than implied by the data, but through time, trends in the simulated distributions of forest (as represented by *Pseudotsuga menziesii* and *Picea* spp.) and steppe (as represented by *Artemisia tridentata*), were similar to those observed in the data. What are the causes of these simulated regional climate anomalies that sometimes appear consistent with the paleoclimatic evidence, but at other times sharply disagree?

Two sets of related processes and components in the simulations govern the simulated climate of a particular region—atmospheric circulation and the surface energy and water balances. Inadequate representation of either or both processes in the model, could lead to the kinds of mismatches observed above. Therefore, an important element of the comparison of simulations and observations is the examination of the climatic processes potentially responsible for mismatches (as well as agreements) between simulations and observations. Webb *et al.*

(this vol.) completed a diagnostic analysis (see also Zhang *et al.*, 1996) of the simulations in the southeastern United States (reviewed briefly below) that implicated amplification by surface energy and water balances of moisture anomalies, generated by particular atmospheric circulation anomalies, as the explanation for the simulations of conditions as warm or warmer than present in summer for the LGM and later. In this section we review the simulations for Beringia and the western United States in a similar fashion. Webb *et al.* (this vol.) used a “flow-chart” representation of the surface energy and water-balance components and their dynamic controls to illustrate their analysis. Here we use a more compact representation of these data that emphasizes their trends over the sequence of simulations (Figs. 12-17), and briefly review the results for the southeastern United States in this manner for comparison. All points in the individual regions were chosen to remain ice-free throughout the sequence of simulations. “Significant” or large anomalies (judged relative to the model’s natural variability) are indicated by dots on Figs. 12-17.

Southeastern United States

In the southeastern United States, the glacial anticyclone at the LGM (Fig. 2), and its subsequent diminution, along with the contemporaneous expansion of the Bermuda high, jointly maintained large-scale sinking motions in summer throughout the sequence of simulations (Fig. 12; see also Webb *et al.*, this vol.). This large-scale subsidence resulted in the simulation of conditions that were less cloudy and drier than present in summer, and these in turn produced positive net radiation anomalies in spring and early summer (despite greater-than-present net longwave radiation loss). The greater-than-present net radiation anomalies, coupled with the negative precipitation anomalies, led to pronounced negative surface-wetness anomalies, and to negative latent heating anomalies, and thus to positive sensible heating anomalies in summer. This heating resulted in the simulation of near-surface air temperatures higher than they would have been otherwise, and also higher than had they been controlled solely by air mass temperature (as represented by the temperature at the 850mb level).

In this region, surface energy and water-balance processes converted a circulation-induced anomaly in cloudiness, precipitation and effective moisture into a positive summer temperature anomaly from 16 to 11 ka. The extent to which this same pattern may be repeated in other regions may provide clues as to whether it is the circulation, the surface energy and water-balance processes, or both, that lead to inaccurate simulations.

Beringia

The regional climates of eastern Beringia (Alaska and northwestern Canada) and western Beringia (easternmost Siberia) are governed today by the interactions among the upper-level east Asian trough and western North American ridge, and the surface Aleutian low- and Pacific subtropical high-pressure systems (Mock *et al.*, submitted). Changes in the location and strength of these circulation features, along with the size and prominence of the glacial anticyclone, produced different sequences in the simulated climatic responses between eastern and western Beringia in a fashion similar to the way surface climates are controlled by these features today. In eastern Beringia, the glacial anticyclone at the LGM, and the strengthening subtropical high pressure system in summer thereafter (Fig. 2), produced stronger-than-present southeasterly flow into the region, along with large-scale rising motions, and consequently increased cloudiness and precipitation, and decreased longwave radiation loss from the surface (Fig. 13). In fall, winter, and early spring at the LGM (21 ka), simulation of a strong Aleutian low (Fig. 2), and westward shift of the western North American ridge (Fig. 1) produced stronger-than-present easterlies and regional subsidence, and consequently clearer and drier conditions than present (Fig 13). From 16 to 11 ka, the greater simulated cloudiness in summer, along with greater-than-present snow depth in spring and fall (and consequently higher surface albedo) reversed the sign of the net shortwave radiation anomaly (from that of the insolation anomaly). The simulation of cloudy and moist conditions during the interval of large positive anomalies of summer insolation (14 to 6 ka) resulted in positive P-E and soil-moisture anomalies in eastern Beringia. Consequently, no amplification of the warming trend

by surface energy and water-balance processes seems evident in the simulations.

Comparison of near-surface air temperature (T_{air}) and airmass temperature (T_{850}) (Fig. 13) indicates that the profiles of the seasonal cycles of the anomalies of these variables are quite similar to one another over time. These observations jointly suggest that there was little amplification (or damping) of circulation-controlled temperature anomalies by the processes that operate at the surface in the simulations for eastern Beringia. Instead, the relatively high near-surface air temperatures seem relatable to the advection of warm air into the region and the consequent reduction in the loss of longwave radiation from the surface.

The particular variations in the circulation controls of western Beringia are such that no systematic pattern of anomalies of cloudiness or precipitation occur across the sequence of simulations (Figs. 14a and 14b). At the LGM, a strong positive snow-depth anomaly in summer (and consequently greater albedo) contributes to the simulation of negative net shortwave and net radiation anomalies. This effect in turn produces negative near-surface temperature anomalies in summer that are larger than those of airmass temperatures. From 14 to 11 ka, a weakly negative soil-moisture anomaly prevails throughout the year, but there does not seem to be as strong an effect of soil-moisture anomalies on the partitioning of net radiation into latent and sensible heating as in the southeastern United States (compare Figs. 12 and 14).

The simulation of warmer-than-present summer temperatures in both eastern and western Beringia from 14 to 6 ka (and in turn the simulation of a more extensive distribution of spruce than observed) seems best explained as a consequence of dynamic (i.e. atmospheric circulation) effects alone, with amplification by surface energy and water-balance processes apparently of little importance. In general, the variations during the year in near-surface air temperatures resemble those of the airmass temperatures across the sequence of simulations. This is in sharp contrast to the southeastern United States (Fig. 12) (Webb *et al.*, this vol.) where

simulated near-surface July temperatures are greater than present from 16 to 11 ka (while airmass temperatures are only near present) (Fig. 12), and also greater than those implied by paleoecological evidence. These differences between regions seem attributable to the specific consequences of the surface energy- and water-balance processes.

Western United States

Relative to the southeastern United States and Beringia, mismatches between simulations and observations are less striking in the western United States, particularly after 14 ka. The main mismatch is the simulation in the Pacific Northwest of conditions milder and moister than those that would be inferred from the observations at the LGM (Thompson *et al.*, 1993), followed by the simulation of colder and drier conditions at 16 ka, in contrast to what can be inferred from paleoecological data. This mismatch seems most simply explained by the simulations of January and July precipitation that were not significantly lower than present at the LGM (Figs. 4 and 15) owing to the absence of realistic topography in the model. In the 21 ka simulation, net shortwave radiation and net radiation are both lower than present in spring, in response to the effects of a greater-than-present snow depth on albedo (Fig. 15). This in turn, leads to greater-than-present surface wetness in summer. However, because the response surfaces use July precipitation as a predictor, this pattern of surface energy- and water-balance components cannot serve to explain the simulations of more extensive distributions of tree taxa than are apparent in the paleoecological data for 21 ka.

In general, the trends through time in the simulations of near-surface air temperature resemble those at the 850mb level in the western United States more so than in other regions, implying greater control of near-surface air temperatures by circulation (i.e. through the advection of differing airmasses into the region) than by surface energy- or water-balance interactions (although the sensible heating rates in the Pacific Northwest between 14 and 6 ka suggest some modification of airmass temperatures by surface energy- and water-balance processes). The importance of this advective control is especially evident for the southwestern United

States (Fig. 16) and for Mexico (Fig 17). Comparison of the trends over time of the energy- and water-balance components in the Pacific Northwest (Fig. 15), southwestern United States (Fig. 16), and Mexico (Fig. 17) with those for the southeastern United States and Beringia indicates little modification of the dynamically influenced aspects of regional climate in the western United States as compared with the situation in those latter regions.

The simulations for Mexico at 11 and 6 ka reveal the structure of an enhanced monsoonal circulation, with increased rising motions and increased cloudiness and precipitation in summer relative to present (Fig. 17). Despite the simulation of greater cloudiness (than present) and the reduced net shortwave radiation that results, net radiation remains positive in summer, in likely response to the simulation of increased longwave flux toward the surface that the additional clouds induce. In the previous simulations with CCM 0, some elements of a strengthened monsoon were simulated (e.g. Kutzbach, 1987; see also Mock and Bartlein, 1995), but the current CCM 1 simulations show the individual features more clearly. Overall, apart from those aspects of the simulations that are directly attributable to the low spatial resolution of the model, the simulated paleoclimates in the western United States do not appear to include the simulation of particularly unusual circulation features or their amplification by surface water- and energy-balance processes that characterized the other regions.

Discussion

These examinations of the seasonal variations in surface energy- and water-balance components, their controls, and temperature in the regions with large differences between simulated and observed conditions offer some insights into the causes of these mismatches. In the southeastern U.S. and to a much lesser extent in western Beringia, dynamically induced anomalies (i.e. ones attributed to atmospheric circulation anomalies) seem to have been amplified by surface energy- and water-balance processes. In eastern Beringia and to a lesser extent in the Pacific Northwest at the LGM, dynamically induced anomalies were apparently the source of

mismatches, with little apparent amplification or modification by surface energy- and water-balance components. This pattern suggests that all cases of mismatches here are ultimately circulation related, with surface physics amplifying the circulation-induced anomalies only in some cases. In both the southeastern United States and Beringian cases, the simulated circulation changes involve the glacial anticyclone at 21 ka and the subsequent expansion of the subtropical high pressure systems, which becomes evident in both regions at 16 ka. However, there are other cases, such as western North America from 14 ka to present, where the simulated circulation-induced climatic anomalies seem appropriate, so atmospheric circulation does not universally generate mismatches. In none of the cases did surface energy- or water-balance processes apparently reverse or dampen circulation-induced anomalies (although the sign of the insolation forcing was frequently reversed in the simulated net radiation). Consequently, the simulation of positive temperature anomalies that were too early, or too large, seems systematically related to atmospheric circulation mechanisms, sometimes amplified by the surface-physics simulation in the model.

SUMMARY AND CONCLUSIONS

This paper has described a second iteration of “data-model” comparison for North America, following those described by Kutzbach and Wright (1985), Webb *et al.* (1987), Barnosky *et al.* (1987), Webb *et al.* (1993a), and Thompson *et al.* (1993). Two particular objectives of this analysis were to review the previously described conceptual model of the controls of regional paleoclimatic variations for North America from the LGM to present and to diagnose the causes of mismatches of the regional climates simulated by CCM 1 and those recorded in paleoenvironmental observations.

In general, many of the broadscale features of the simulated paleoclimates recognized in the previous analyses emerged in the present one. These include:

- displacement of the jet stream by the Laurentide Ice Sheet to the south of its present position in

both winter and summer; (The expression of the northern branch of a “split” jet stream was not great in the present (CCM 1) simulations, however.)

- generation of a “glacial anticyclone” over the ice sheet at the LGM, and the consequent large-scale sinking motions induced over eastern North America; (Although, in the present analysis, the anticyclone was less evident in the surface wind fields than in the previous analyses, and began to disappear earlier.)
- changes in the strength of surface atmospheric circulation features through time, including weakening of the Aleutian low in winter, and strengthening of the eastern Pacific and Bermuda high-pressure systems in summer as the ice sheet decreased in size;
- development of a “heat low” at the surface, and a strengthened ridge in the upper-atmosphere over the continent at the time of the maximum summer insolation anomaly;
- increases in summer temperature earlier in regions remote from the ice sheet; (These increases appear earlier in the present (CCM 1) simulations than in the previous (CCM 0) ones, however.)
- continuation of negative winter temperature anomalies into the middle Holocene.

A number of other secondary features, such as changes in the orientation of surface temperature gradients across the continent evident in the earlier simulations, were also visible in the current set of simulations.

The sequence of simulated climates was compared with paleoenvironmental observations employing a forward modeling approach in which the sequence of climate-model output was used to simulate the distributions of three plant taxa. These simulated distributions were in turn compared with paleoecological data sets. This comparison revealed:

- broad agreement in the simulated and observed ranges of spruce over much of eastern North America, with locally less-good agreement in regions adjacent to the ice sheets, where the simulation of very steep temperature gradients produced a band of simulated *Picea* spp. that was too narrow relative to that displayed in the paleoecological data;
- general agreement in the trends of observed and simulated distributions of forest and steppe in western North America, with some regional anomalies;
- severe mismatches between the simulations and observations in the southeastern United States (discussed in Webb *et al.*, this vol.) and in Beringia, attributable to the simulation of (summer) conditions warmer than those implied by the paleoecological data.

In the case of the southeastern United States these mismatches were also present in the earlier comparisons between CCM0 simulations and observations at about the same magnitude, while those in Beringia seem larger than observed previously.

In order to identify the sources of these mismatches, we completed a “process analysis” of surface energy- and water-balance components in several key regions. These analyses revealed that:

- the large mismatches in the southeastern United States and Beringia persisted across several times in the sequence of simulations;
- in both the southeastern United States and Beringia the mismatches seem attributable to particular atmospheric circulation anomalies, amplified in some cases by surface energy- and water-balance processes (e.g. in the southeastern United States and to a lesser extent in western Beringia);
- in the western United States the simulated circulation patterns do not seem inappropriate for the most part;
- the common features in the circulation-induced mismatches are the glacial anticyclone at the LGM, and the strengthening subtropical high-pressure systems thereafter; these produce surface winds and vertical motion fields that create conditions that are apparently too warm or too dry.

Several other observations can be made based on our analysis of simulations and observations. The analysis of a suite of experiments ranging from the LGM to present illustrates the importance of not focusing strictly on the climatic “extremes” (e.g. the LGM and 6 ka (a time when the ice sheets had disappeared but the seasonal cycle of insolation was still larger than at present)), despite the simpler experimental designs these times afford. The comparisons of simulations and observations at 16, 14 and 11 ka, particularly in Beringia and the southeastern United States, revealed patterns of disagreement between of simulations and observations that would not have been noted had only the 21 and 6 ka experiments been analyzed.

The analysis of the simulated surface energy- and water-balance components, near-surface and upper-air temperatures, and their various controls illustrate the rich variety of paleoclimatic anomalies that can occur and present a major challenge to inverse-model approaches for “data-model” comparison. There may be any number of ways to generate the same responses in surface temperature and moisture variables, and it may not always be possible to constrain the reconstructions sufficiently (e.g. Cheddadi *et al.*, 1997) in order to yield unambiguous or physically consistent results.

This same complexity of surface-climate responses lends support for the forward-modeling approach as described here, because the physical consistency that is implicit in the surface energy- and water-balance components probably reduces the number of combinations of anomalies that would likely occur. This in turn reduces the possibility of apparent agreement in data-model comparisons arising from chance combinations of climate variables.

Several extensions to the present combination of visualizations, data-model comparisons and process analyses should be explored. Although the specific mapping techniques used here were successful in portraying such features as the glacial anticyclone or surface heat low over the southwestern United States in summer without resorting to “cartoon” representations, such visualizations are still qualitative in nature. There are emerging techniques for representing and comparing features or objects on series of maps that can be adapted for the specific problems that arise in data-model comparisons (e.g. MacEachren and Taylor, 1994). A second area of improvement is in the simulation of the distribution of plant taxa, done here using an empirically based statistical procedure. A recently developed simple mechanistically based “bioclimatic” model (i.e. Sykes *et al.*, 1996) can now be used and its results compared to those generated by the ecological response surfaces employed here.

The process analysis of the climate model output would seem to be a powerful tool for understanding *why* the specific climates simulated in key regions arise (Zhang *et al.*, 1996). However, the interpretation of these results here was largely qualitative in nature, resting on generally accepted concepts of how surface energy- and water-balance components interact, as well as the major implicit assumption that the model has captured these interactions correctly. Consequently, an important adjunct to this analysis would be the parallel analysis of the modern climate in the key regions, along with the exploration of modern analogues of paleoclimatic anomalies (e.g. Mock and Bartlein, 1995). In general, the coordinated analysis of model output with paleoenvironmental observations, coupled with diagnostic studies of both the present observed and simulated past climates can offer an important perspective on the current capabilities of climate models and on the pathways that should be followed for their improvement.

ACKNOWLEDGMENTS

Pat Behling and Rich Selin were instrumental in furnishing the climate-model output in a

comprehensible and useable form. We also wish to thank Bev Lipsitz, Lynn Songer, J.J. Shinker and Victor Barnett for assistance with the maps, J.A. Mohr for assistance with the paleoecological data, and S.L. Shafer for editorial assistance. This research was supported by grants to the University of Oregon from the National Science Foundation and U.S. Geological Survey.

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FIGURE LEGENDS

FIG. 1. Simulated January and July upper-level (500mb) winds. Wind directions are shown by arrows, and velocity by shading. The outline of the region covered by ice in the model is also shown. Polar stereographic projection, principal meridian 120 W.

FIG. 2. Simulated January and July sea-level pressure and surface winds. Surface wind directions are shown by arrows, sea-level pressure by shading.

FIG. 3. Simulated January and July near-surface air temperature (C).

FIG. 4. Simulated January and July monthly total precipitation (mm).

FIG. 5. Simulated January and July near-surface air temperature (C), generated by interpolating anomalies (paleoclimatic experiment minus present control) onto a 25km gridded data set of the present values of these variables. The "Present" map panels are those observed present values. Albers projection, standard parallels at 66.66 N and 33.33 N, principal meridian 120 W. Paleogeography from Dyke and Prest (1987) and Peltier (1994).

FIG. 6. As in Fig. 5, only for January and July monthly total precipitation.

FIG. 7. Observed distributions of *Picea* spp. (all North American species of spruce), *Pseudotsuga menziesii* (Douglas-fir), and *Artemisia tridentata* (sagebrush), simulated probabilities of occurrence obtained using the response surfaces for these taxa and the observed present climate, and illustrations of the response surfaces for each.

FIG. 8. Dimension of the climate space (i.e. number of climate variables) used in the evaluation of the response surfaces at each grid point and time.

FIG. 9. Observed (top) and simulated (bottom) incidence and abundance of *Picea* spp.

FIG. 10. As in Fig. 9, only for *Pseudotsuga menziesii* and *Artemisia tridentata*.

FIG. 11. Proportions of sites with paleoecological data from the western United States correctly predicted present or absent for each taxon at each time (top), observed proportion of sites with each taxon present (middle), and predicted proportion of sites with each taxon present (bottom).

FIG. 12. Anomalies (paleoclimatic experiment minus modern control simulations) of energy- and moisture-balance components (Table 2) for grid points representing the southeastern United States (Table 3). The similarity across experiments in the seasonal profiles of the anomalies of net SW radiation, latent and sensible heating rates, soil moisture and P-E reveal the influence of the surface energy- and water-balance processes in near-surface air temperatures that in summer were not significantly cooler than present.

FIG. 13. As in Fig. 12, for grid points representing eastern Beringia. In contrast to the southeastern U.S. where surface energy- and water-balance components acted to increase near-surface temperatures relative to airmass temperatures (Fig. 12), the relatively high near-surface air temperatures after 16 ka in eastern Beringia seem related to increased LW radiation gain (or decreased loss) from warm airmasses advected into the region.

FIG. 14. As in Fig. 12, for grid points representing western Beringia. The similarity across the 16 ka to 6 ka experiments in the seasonal profiles of the anomalies of near-surface and 850mb temperatures together with the small anomalies of sensible heating rate suggest that as in eastern Beringia, the relatively warm conditions in summer in western Beringia are related to the advection of warm airmasses into the region.

FIG. 15. As in Fig. 12, for grid points representing the Pacific Northwest. In this region the similarity across simulations between the seasonal profiles of near-surface and 850mb temperatures suggest that the near-surface temperature anomalies are determined mainly by the advection of airmasses into the region, although the positive sensible heating anomalies in summer (apparently related to negative surface wetness anomalies) also suggest some modification of the advected airmass temperatures by surface energy- and water-balance processes.

FIG. 16. As in Fig. 12, for grid points representing the southwestern United States. This region displays the greatest similarity across simulations between the seasonal profiles of near-surface and 850mb temperatures, which suggests little modification of the advected airmass temperatures by surface energy- and water-balance processes.

FIG. 17. As in Fig. 12, for grid points representing Mexico. The greater-than-present anomalies in summer at 11 ka and 6 ka in rising motions, cloudiness, precipitation, and latent heating rate reflect the development of a stronger summer monsoon at those times.