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RESPONSE OF NORTH AMERICAN FRESHWATER LAKES TO SIMULATED FUTURE CLIMATES[1]

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ABSTRACT: We apply a physically based lake model to assess the response of North American lakes to future climate conditions as portrayed by the transient trace-gas simulations conducted with the Max Planck Institute (ECHAM4) and the Canadian Climate Center (CGCM1) atmosphere-ocean general circulation models (A/OGCMs). To quantify spatial patterns of lake responses (temperature, mixing, ice cover, evaporation) we ran the lake model for theoretical lakes of specified area, depth, and transparency over a uniformly spaced (50 km) grid. The simulations were conducted for two 10-year periods that represent present climatic conditions and those around the time of CO_2 doubling. Although the climate model output produces simulated lake responses that differ in specific regional details, there is broad agreement with regard to the direction and area of change. In particular, lake temperatures are generally warmer in the future as a result of warmer climatic conditions and a substantial loss (> 100 days/yr) of winter ice cover. Simulated summer lake temperatures are higher than 30°C over the Midwest and south, suggesting the potential for future disturbance of existing aquatic ecosystems. Overall increases in lake evaporation combine with disparate changes in A/OGCM precipitation to produce future changes in net moisture (precipitation minus evaporation) that are of less fidelity than those of lake temperature.
(KEY TERMS: climate change; freshwater lakes; aquatic ecosystem; lake modeling.)

INTRODUCTION

Because lakes and wetlands respond directly to climate, quantifying their sensitivity to possible climate change and variability will provide information crucial to the assessment of water resources in the future (IPCC, 1996a; Covich *et al.*, 1997; Grimm *et al.*, 1997; Hauer *et al.*, 1997; Magnuson *et al.*, 1997; Melak *et al.*, 1997; Mulholland *et al.*, 1997; Rouse *et al.*, 1997; Schindler, 1997; Clair, 1998). Most aquatic processes and the evaporative component of the water balance of lakes and wetlands are controlled by the annual cycle of water temperature. The annual temperature cycle, which is characterized by the evolution of internal temperature structure, mixing, and seasonal ice cover, is governed climatically by the exchange of moisture, heat, and momentum within the planetary boundary layer, and physically by properties of lake water (e.g., transparency or trophic state, chemical density) and the hypsometry (e.g., depth-area relation, surface area-to-depth ratio) of the lake basin.

The sensitivity of lakes to climate change can be assessed using physically based models that simulate seasonal thermal evolution in response to atmospheric forcing (e.g., Croley, 1990; McCormick, 1990; Chang *et al.*, 1992; Stefan *et al.*, 1993, 1998; Hondozo and Stefan, 1993; Vavrus *et al.*, 1996; Hostetler, 1995; Hostetler and Giorgi, 1995; Thompson *et al.*, 1998; Fang *et al.*, 1999). Many modeling studies have focused either on a specific lake or on a number of lakes in a particular region of interest. Here, we apply a lake model to simulate continental scale patterns of lake response to climate change as characterized by two general circulation models. Our modeling is accomplished by applying a physically based lake model at each grid point of an equally spaced rectangular grid that covers most of North America. We simulate hypothetical lakes for which we derive geographically consistent climate forcing from climate models and specify realistic physical characteristics (e.g., depth and turbidity). Our approach yields spatially uniform simulations of lake temperature, the depth and duration of mixing, the duration and thickness of ice cover, and net moisture (precipitation minus evaporation), thereby allowing us to quantify regional to continental-scale lake responses to climatic change. The approach is of particular advantage

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near coastlines, over mountainous terrain, and in areas of nearly balanced net moisture (precipitation equals evaporation) where steep climatic gradients cause accompanying gradients steep of lake responses.

MODEL DESCRIPTION AND METHODS

The lake model is a one-dimensional, energy-balance model that simulates the vertical structure of temperature and stratification, seasonal lake ice, and evaporation (Hostetler and Bartlein, 1990; Hostetler, 1991; Small *et al.*, 1999). Turbulent or wind-driven mixing of the water is parameterized by eddy diffusion (Henderson-Sellers, 1985), and internal convective mixing is simulated as a density-driven process. The one-dimensional heat transport equation for the model is written as:

$$\frac{\partial T}{\partial t}=\frac{1}{A(z)}\frac{\partial}{\partial z}\left\{A(z)\left[k_m+K(z,t)\right]\frac{\partial T}{\partial z}\right\}+\frac{1}{A(z)}\frac{\left(\partial A(z)\Phi\right)/\partial z}{\rho c_p} \tag{1}$$

where T is temperature, t is time, $A(z)$ is the area of the lake at depth z, k_m is molecular diffusion, $K(z,t)$ is eddy diffusion, ρ is the density of the water, and c_p is the specific heat of water. The heat source term Φ represents downwelling solar radiation as a function of water depth and the transparency of the water in accordance with Beer's law:

$$\Phi(z) = (1 - \beta)\,(1 - \alpha_{sw})\,\phi_{sw}e^{-\eta z} \tag{2}$$

where β is the fraction of incoming global (solar) radiation ϕ_{sw} absorbed at the surface, α_{sw} is the shortwave albedo of the lake surface, and η is the light extinction coefficient, which is a function of turbidity caused by suspended particles such as clay and phytoplankton. Highly turbid (trophic) lakes ($\eta > 1.0$) generally exhibit greater surface heating and less heating at depth than transparent or oligotrophic lakes ($\eta < 0.3$) in which radiation is transmitted deeper in the water column.

The boundary condition for Equation (1) is the surface energy balance which is written as:

$$-\left[k_m+K(z,t)\right]\frac{\partial T}{\partial z}=(1-\alpha_{sw})\phi_{sw}+(1-\alpha_{lw})\phi_{lw}-\phi_{lu}-q_{le}-q_h \tag{3}$$

where α_{lw} is the longwave albedo of the water surface, ϕ_{lw} is downwelling atmospheric radiation, ϕ_{lu} is back radiation, and q_{le} and q_h are the latent and sensible heat fluxes, respectively.

A compatible energy balance submodel, modified after Patterson and Hamblin (1988), is used to simulate the formation and ablation of lake ice. The lake model has recently been updated to improve the ice model and other components (Small *et al.*, 1999). The model has been applied in a variety of climate-lake studies (e.g., Hostetler and Bartlein, 1990; Hostetler and Benson, 1990; Vassiljev *et al.*, 1994; Hostetler and Giorgi, 1993, 1995; Boqiang and Qun, 1998; Thompson *et al.*, 1998).

For this study, we used a 92-by-135 rectangular grid with even grid spacing of 50 km that covers most of North America (hereafter the lake model grid). A lake model was run at each grid point for a hypothetical lake for the duration of the simulations (nominally ten years), thus providing geographic coverage of most lakes and wetlands in North America at relatively high spatial and topographic resolution. Because the model was run over all non-ocean points of the lake grid, lake responses were simulated at gridpoints in areas such as the desert southwest and the Great Basin where evaporation greatly exceeds precipitation and few or no natural lakes actually exist. However, some lakes that are supplied runoff from moisture derived at higher elevations (e.g., Pyramid Lake, Nevada) and reservoirs (e.g., those on the Colorado River system) do exist in arid regions, and our simulations are representative of such water bodies.

Output from two transient climate simulations conducted with coupled ocean-atmosphere models (A/OGCMs) by the Max Planck Institute (URL http://www.drkz.de/forschung/reports.html) (the ECHAM4/OPYC3, hereafter ECHAM4) and the Canadian Climate Center (URL http://www.cccma.bc.ec.gc.ca/mdels/cgcml.html) (hereafter the CGCM1) was used to derive climate inputs for the lake model. The ECHAM4 and CGCM1 transient climate simulations, which have been used widely for climate impacts assessments, were conducted for the Intergovernmental Panel on Climate Change (IPCC) climate assessment following the IS92A protocol for increasing trace gases (CO_2, CH_4, N_2O, and hydrofluorocarbons) in the models by 1 percent/year beginning in (simulated) 1990 (IPCC, 1992, 1994, 1996b). The direct radiative effects of sulfate aerosol loading are included in CGCM1, but not in the ECHAM4.

Lake responses and their sensitivity to warmer climate conditions are evaluated by two continuous ten-year simulations conducted with the A/OGCM output: one for the present climate defined here as the climate model years of 1970 through 1979 (hereafter the control simulation) and one for the climate model years of 2070 through 2079 spanning the approximate year (2072) of CO_2 doubling (hereafter the 2xCO_2 simulation). Daily averages of screen height (2-m) values for air temperature, wind speed, and atmospheric specific humidity and net surface short-wave and atmospheric (longwave) radiation that are used to drive the lake model were derived from the monthly values of A/OGCM output obtained from the Data Distribution Center that is maintained by the IPCC (URL http://ipcc-ddc.cru.uea.ac.uk/dkrz/dkrz_index.html) (IPCC-DDC). We used bi-linear interpolation to distribute the monthly values of climate model output on the lake model grid, and daily values at each lake grid point were computed by linear interpolation between bracketing months while the model was running.

Air temperatures from the coarser grid and thus lower topographic resolution A/OGCMs were adjusted to reflect the higher resolution topography of the lake model grid by applying a linear lapse-rate correction:

$$T_a = T_{gcm} - 0.0065\delta z \tag{4}$$

where T_a is the 2-m air temperature at the lake grid points, T_{gcm} is the air temperature from the A/OGCM, δz (m) is the difference in elevation between the target grid point on the lake grid and the mean elevation of the topography of the A/OGCM grid boxes used in the interpolation, and 0.0065°C m^{-1} is a nominal free atmospheric lapse rate for the mid-latitudes.

The 10-meter winds from the ECHAM4 output were scaled to a height of 2 meters using the relation:

$$u_2 = 0.794u_{10}, \tag{5}$$

where u_2 and u_{10} are the 2- and 10-meter winds, respectively. The factor 0.794 is derived from a standard (logarithmic) equation that is used to scale wind speeds under neutral atmospheric conditions as a function of height and surface the roughness of the water (e.g., Brutsaert, 1982).

Specific humidity q for the ECHAM4 is unavailable at the IPPC-DDC site. We computed it from available dew-point temperature values (Sutton, 1953):

$$q = \frac{0.622e}{psurf - 0.378e} \tag{6}$$

where e = 101325 $exp(13.3185\Delta - 1.9760\Delta^2 - 0.6445\Delta^3 - 0.12990\Delta^4)$, $\Delta = 1.0-(373.15/T_{dew})$ (e.g., Brutsaert, 1982), and psurf is surface pressure (~101325 Pa).

Longwave (atmospheric) radiation is not reported for either model at the IPCC-DDC site, so we derived it from air vapor pressure e and air temperature T_a using the empirical method of Idso and Jackson (1969):

$$\phi_{lw} = 0.97\varepsilon\sigma T_a^4 \tag{7}$$

where atmospheric emissivity $\varepsilon = 1.24(e/T_a)^{1/7}$ and σ is the Stephan Boltzmann constant (5.67x10^{-8} W m^{-2} °K^{-4}).

For the control period (1970-1979), the amplitude of the seasonal cycle of the solar radiation simulated by the ECHAM4 over North America is similar to the historical estimate reported in the Vegetation/Ecosystem Modeling and Analysis Project (VEMAP) data set (Kittel *et al.*, 1997), but the magnitude of the values is ~25 percent lower than those of the VEMAP data, particularly over high latitudes during summer. The bias is not surprising because solar radiation in AGCMs is adjusted to produce realistic zonal averages, so that simulated values may over- or underestimate observations over any particular region. For input to the lake model, the ECHAM4 solar radiation values were scaled by a factor of 1.25 to obtain values in better agreement with the VEMAP values.

In the CGCM1 control simulation, monthly averaged 2-m wind speeds are unrealistically low (<1 m sec^{-1}) over much of western North America. As a result, very low simulated lake evaporation rates (which are dependent on wind speed) caused unrealistically high lake temperatures in the lake model. Based on values of wind speed in the VEMAP data set, monthly wind speeds from the CGCM1 were restricted to be ≥ 1.25 m sec^{-1}. As we discuss below, the modified radiation and wind-speed values produce simulated lake responses that are in good agreement with observations. To produce self-consistent lake responses to climate change, we applied the adjustments to both the control and 2xCO_2 GCM values.

We present three lake simulations: one each conducted with the ECHAM4 and CGCM1 output for a relatively small (surface area of 0.5 km^2), shallow (4-m) lake of intermediate turbidity (η = 0.6 in Equation 2), and an additional simulation of a deeper (30-m) lake of relatively low turbidity (η = 0.3 in Equation 2) conducted using only the CGCM1 output. A similar simulation of the 30-m lake with the ECHAM4 output is not reported because intramodel comparison sufficiently illustrates depth-dependent responses. The specified lakes provide a broad range of conditions over which to evaluate the sensitivity to climate change. At the 1-m vertical resolution of the lake model, the 4-m lake remains essentially free of stratification, and, owing to its limited thermal inertia, the simulated annual temperature cycle is in phase with the climatic forcing, so the lake simulations are nominally representative of wetlands, albeit without the effect of emergent vegetation.

CLIMATOLOGY OF THE LAKE MODEL INPUTS

Although the lake model was run only over non-ocean grid points, for perspective on large-scale climate patterns and their changes, we present the climate inputs over the entire lake grid that includes both land and ocean gridpoints (Figure 1). For the ECHAM4, the mean annual air temperature over the land points is 8°C for control and 13°C for $2xCO_2$, compared with temperatures of 7°C for control and 11°C for $2xCO_2$ for the CGCM1. The 2°C difference between the $2xCO_2$ experiments reflects model differences and the cooling effect of sulfate aerosols in the CGCM1 simulation. Although the magnitude of the air temperature anomalies ($2xCO_2$ minus control) are similar for the ECHAM4 and CGCM1, there are substantial differences in their spatial patterns (Figure 1).

The warmer atmosphere of the $2xCO_2$ simulations increases the moisture-holding capacity of the air and leads to accompanying increases of 25 percent or more in atmospheric mixing ratios (Equation 6, Figure 1). The largest changes in mixing ratios correspond to the largest temperature increases. Changes in future wind speeds, attributed to atmospheric warming, and reductions or amplifications of regional temperature gradients are relatively small and range from a maximum decrease of ~10 percent to increases of > 3 percent (Figure 1).

Mean-annual values of future incident solar radiation increase by ~5 percent over a large area of the mid-latitudes in the ECHAM4 (Figure 1), probably as a result of change in cloud cover and surface albedo associated with a loss of snow cover. Future solar radiation is largely unchanged in the CGCM1 simulation, although there are limited areas of reduced radiation that suggest some effect of increased cloudiness or the scattering by sulfate aerosol loading, or both. Future longwave radiation values are higher in both models than their respective control values. Increases ranging up to 20 percent are coincident with areas of warmer air temperatures and increased atmospheric mixing ratios.

In contrast to the CGCM1 simulation, future annual precipitation rates in the ECHAM4 simulation exhibit a fairly large range of change (Figure 1). Drier conditions that prevail over the deserts of southern California and the Southwest diminish to the north over the Great Basin and into the Pacific Northwest. Wetter conditions generally prevail east of the Rocky

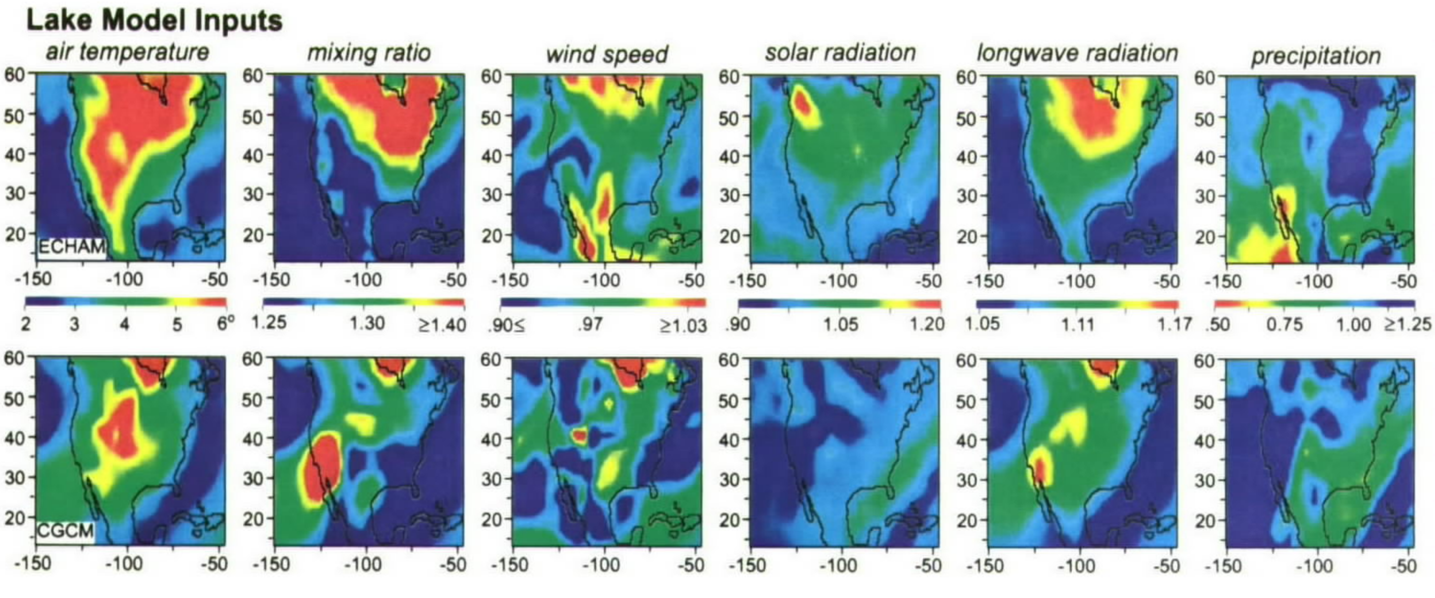

Figure 1. Changes in Lake Model Inputs for the ECHAM4 Model (top row) and for the CGCM1 Model (bottom row) Expressed as Differences [$2xCO_2$ (2070-2079) Minus Control (1970-1979)] for Air Temperature, and as Ratios [$2xCO_2$ (2070-2079):Control (1970-1979)] for Other Variables. Vertical and horizontal axis labels indicate latitude and longitude, respectively.

Mountains, with maximum increases located over the Midwest and Southeast. In agreement with the ECHAM4 simulation, the future climate of the CGCM1 displays increased precipitation east of the Rocky Mountains and over the Midwest. In contrast, however, large increases in future precipitation are simulated by the CGCM1 along the west coast, while decreases are simulated over the Southeast.

While there is broad agreement regarding the direction and the general location of the simulated future changes in some of the lake inputs, disagreements between the model results are apparent at the regional scale. Some of the differences in the spatial distribution of the climate anomalies are attributable to differences in the topographic resolution of the A/OGCMs (2.8° latitude by 2.8° longitude in the ECHAM4 versus 3.75° latitude by 3.75° longitude in the CGCM1). In addition, substantial interannual to interdecadal variability exists in precipitation patterns and rates simulated by A/OGCMs, so a given ten-year period may differ from other ten-year periods.

CONTROL LAKE MODEL RESULTS

For the control period (1970-1979), the geographic patterns of lake surface temperature (including both open and ice-covered periods) simulated with the ECHAM4 and CGCM1 model output are similar (Figure 2) and reflect the topographic and latitudinal effects on climate patterns. For the shallow lake, however, the ECHAM4 output yields mean-annual water temperatures of about 0°C as far south as about 40°N latitude, and temperatures well below 0°C over parts of the Cordilleran and higher latitudes. These surface temperatures are colder than both the CGCM1 values

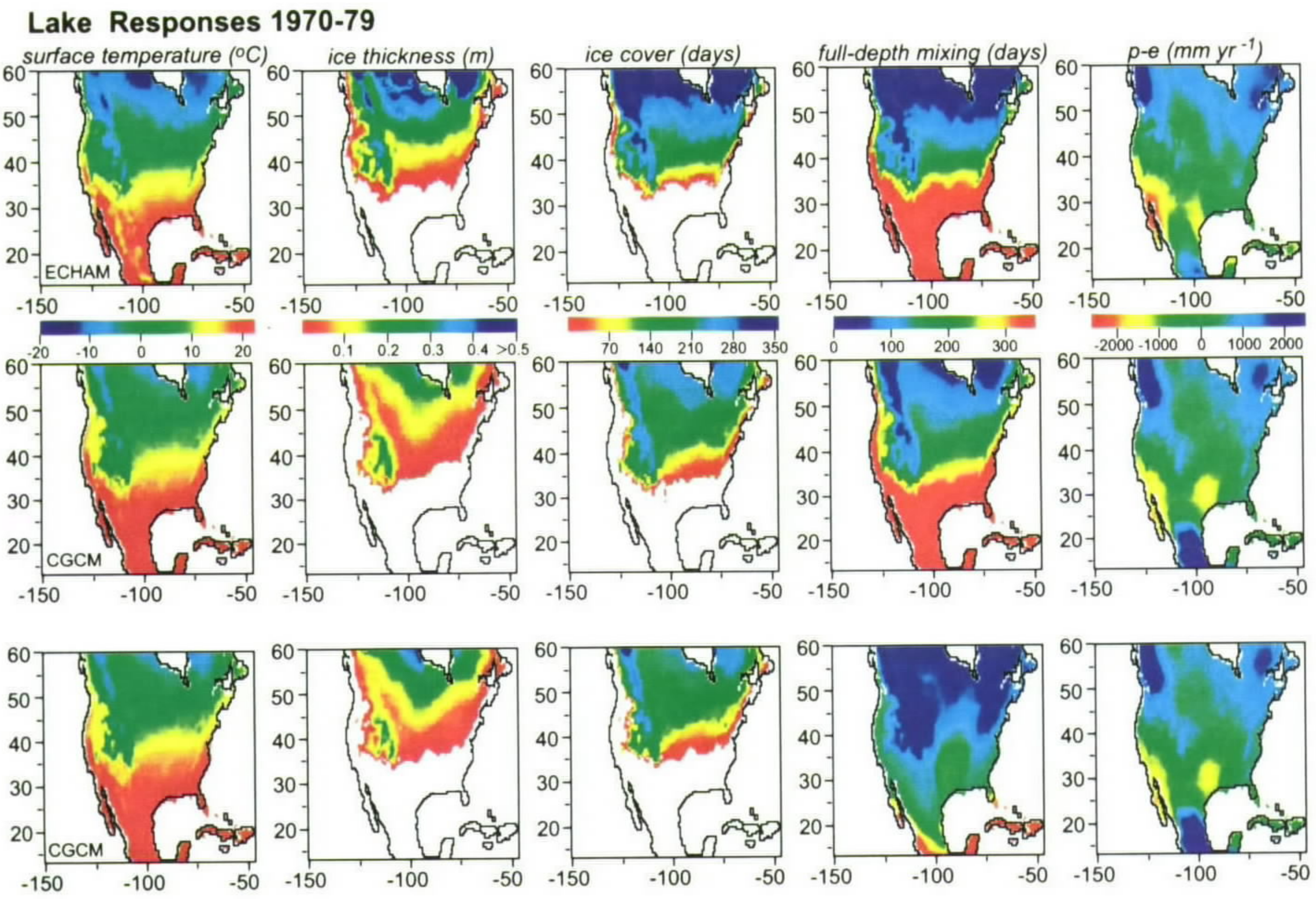

Figure 2. Ten-Year Averages of Selected Lake Responses as Simulated with the Control (1970-1979) Input Data Sets from the ECHAM4 and CGCM1. Surface temperature and ice thickness are expressed as annual averages. Ice cover and full-depth mixing (to the bottom) are expressed in the total number of days annually. The top and middle rows are responses for a 4-m deep lake and the bottom row are responses for a 30-m deep lake. Net moisture (P-E) is precipitation minus evaporation.

and observed values. The cold bias in the ECHAM4 is also evident in the simulated thickness and duration of winter ice cover. At 45°N over the Great Plains and Midwest (the approximate latitude of the Great Lakes), the annual duration of simulated ice cover exceeds 200 days, substantially longer than observed values that range between 120 to 150 days (Robertson *et al.*, 1992; Vavrus *et al.*, 1996; Stefan *et al.*, 1998). Essentially permanent lake ice is simulated over high latitudes and portions of the northern Rocky Mountains. The CGCM1 output, on the other hand, produces a more realistic geographic pattern of ice properties on the shallow lake, with an average duration of about 140 days at 45°N in the Midwest, and maximum durations of around 275 days over the higher elevations and latitudes.

In general, the responses of the deep lake to the CGCM1 forcing are similar to those for the shallow lake. There are small differences in simulated surface temperature, ice thickness, and ice duration that reflect the effects of thermal inertia and mixing in the deeper lake. Relative to the shallow lake, the greater thermal inertia of the deep lake tends to reduce the amplitude of the seasonal cycle of surface temperature and delay freezing in late fall and early winter.

In both the ECHAM4 and CGCM1 simulations, wind-driven turbulence mixes the 4-m lake to the bottom more-or-less continuously during ice-free periods in temperate regions and throughout the year along the coasts and in the south. Differences in the number of simulated days of full-depth mixing between the models are explained primarily by differences in the duration of simulated ice cover.

In general, fewer days of full-depth mixing are simulated for the 30-m deep lake with the CGCM1 output. The latitudinal gradient in mixing apparent in the 4-m lake simulation is much less well defined for the deeper lake. In ice-free areas along the coasts and in the south, the 30-m deep lakes mix for a total of around four months of the year. In more continental areas where winter ice cover occurs, the duration of mixing ranges from a month or less up to two to three months. The general reduction in the duration of full-depth mixing in the deep lake illustrates how the shallow turbulent mixing that was sufficient to mix the shallow lake effectively maintains a stable thermocline in the deeper lake. In the stratified lake, convection plays a more dominant role in producing seasonal overturn, particularly in areas of winter ice cover. Modeled convection is caused by subsurface density (temperature) instabilities that are set up by the penetration of solar radiation (η = 0.3 in Equation 2) and the loss of heat in the fall and gain of heat in the spring while the temperature of the water column is around 4°C (neutral stability). In contrast, over the warmer areas of the southern Great Plains and Texas, the deep lake mixes for over six months of the year, indicating a lack of stable stratification in the generally warmer lake water.

The patterns of lake net moisture (precipitation minus evaporation, P-E) produced by the ECHAM4 and CGCM1 output are similar in the control simulations (Figure 2), indicating agreement in simulated lake evaporation rates (Figure 3). The most negative values of P-E are simulated in both models over the western deserts and southwest regions where reservoirs, ephemeral lakes, and lakes that derive water from higher elevation catchments exist. Negative values also occur over the the southern Great Plains and parts of the Caribbean islands and Puerto Rico. Values that are less negative, or nearly balanced (P = E), extend northward through the Midwest and eastward to the Southeast. The north-to-south line running through the Midwest and Great Plains that approximately divides positive P-E values to the east from negative P-E values to the west (prairie-forest border) is well captured by both models. Positive P-E values occur in the high-precipitation and low evaporation areas of the Pacific Northwest, along the east coast, and at higher latitudes.

2xCO_2 LAKE MODEL RESPONSES

For both the ECHAM4 and CGCM1 output, mean-annual temperatures for the lake surface increase from about 2°C to more than 7°C in the 2xCO_2 simulations (Figure 4). The largest temperature changes are located over the Rocky Mountains and the Midwest where there is a loss of seasonal lake ice in the 2xCO_2 simulations. Warming of greater than 3°C occurs in areas where the lakes were ice free in the control simulation. No increases in the duration of ice cover were simulated. The warmer temperatures in areas that were ice free in the control simulation are attributed to generally warmer air temperatures and higher levels of longwave radiation throughout the year. In areas where simulated ice cover occurs in the control, warming is attributed in part to higher summer air temperatures and higher levels of longwave radiation, and in part to a reduction (or complete loss) of ice cover, which allows to warming to commence earlier in spring and persist later in autumn.

Summer lake temperatures increase substantially over much of the domain in both 2xCO_2 simulations. August temperatures simulated with the CGCM1 output generally range from 30°C to 35°C and higher over most of the domain south of ~45°N (Figure 5). Surface temperatures of ~25°C in the 2xCO_2 simulation occur as far north as Hudson's Bay and over a large area of the Rocky Mountains. Average summer

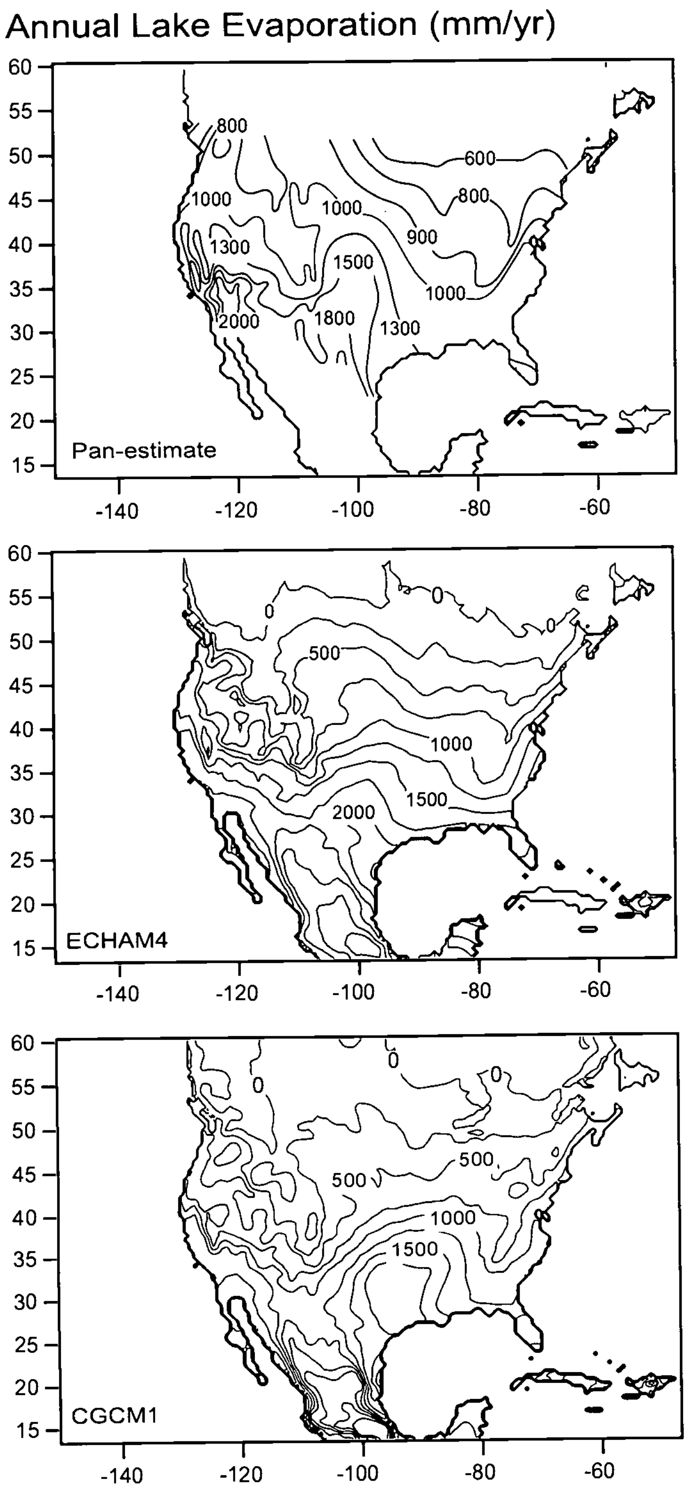

Figure 3. Annual Evaporation Rates for Present Derived from Pan Estimates and as Simulated by the Lake Model Using the ECHAM4 and CGCM1 Control (1970-1979) Input Data Sets. The contour intervals for the pan estimates are not evenly spaced; a contour interval of 250 mm/yr is used for the maps of simulated values. Pan-based estimates from the U.S. Weather Bureau (1968).

temperatures (June through August) increase from 2°C to greater than 4°C in the $2xCO_2$ simulation. Coincident with relatively low increases in air temperature (Figure 1), less summer warming is simulated over the Pacific Northwest and Southeast. Cooler temperatures over the southern Great Plains are attributed to higher wind speeds in the $2xCO_2$ simulation (Figure 1), which enhance evaporation and thus cool lake surface temperatures.

The magnitude of lake warming simulated by both the ECHAM4 and CGCM1 over the upper Midwest agrees with results obtained by Stefan *et al.* (1998), who used CGCM1 output to model temperature responses of Minnesota lakes. The warming simulated here is also in general agreement over the western U.S. with results of Hostetler and Giorgi (1995) and Thompson *et al.* (1998) in which the lake model was driven by output from a regional climate model.

Widespread loss of winter ice cover in the Rocky Mountain region, the western mountains, throughout the Great Plains, and in the East is evident in both the ECHAM4 and CGCM1 $2xCO_2$ simulations (Figure 4). Although there is a general decrease in the duration and thickness of ice in both simulations, there are areas where the ECHAM4 and CGCM1 produce different results. The differences are primarily attributed to changes in the simulated snow cover over ice. In winter, snow cover insulates the ice from extreme air temperatures and, in general, reduces the thickness of simulated ice relative to what it would be in the absence of snow. The area of maximum reduction in the duration of ice (>100 days) is centered over the Rocky Mountains and extends eastward over the high plains and into the Midwest. Relative to the control, the boundary of ice-free conditions is shifted northward by 10° latitude or more over much of the Midwest and East.

Reductions in the duration of ice cover are accompanied by substantial increases in the number of days of mixing in the 4-m lake (Figure 4). Smaller increases in mixing are simulated for the 30-m lake because summer stratification that was present in the control simulation continues to inhibit mixing in the $2xCO_2$ simulations. There is a tendency for no increase or a decrease in the number of days of mixing simulated for the deep lake over the South. The largest reductions, corresponding to lower wind speeds in the $2xCO_2$ simulation, occur over the Southeast and Florida where reduced evaporation during summer and wind driven mixing during winter combine to increase the stability of a warmer water column.

Changes in mixing and internal heating associated with climate are illustrated in the simulated interannual evolution of temperature with depth in an ice-free lake (located at 30°N latitude and 105°W longitude; Figure 6) and a dimictic lake (seasonal lake

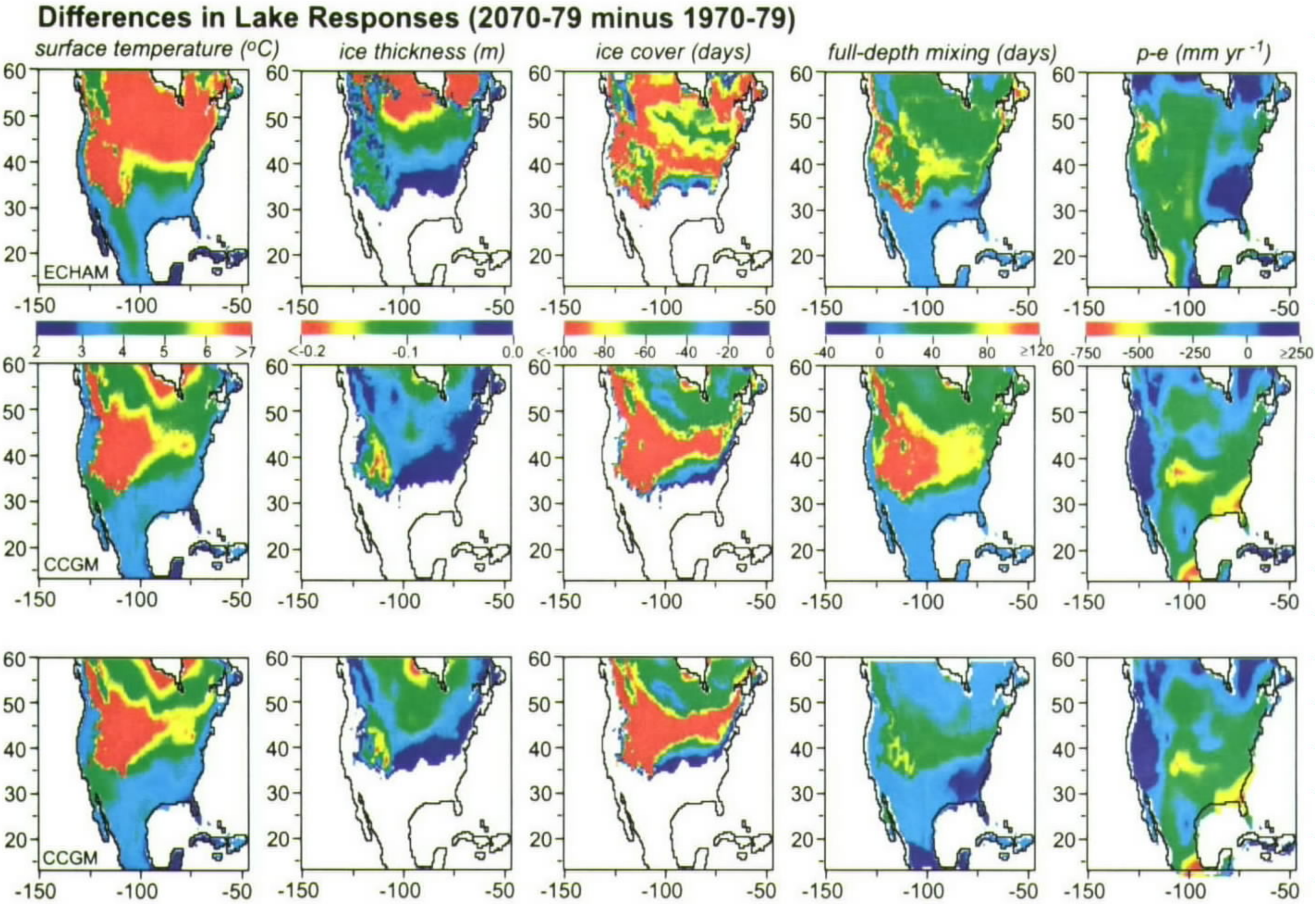

Figure 4. Changes in Responses as Simulated with the ECHAM4 and CGCM1 Input Data Sets Expressed as Differences Between the $2xCO_2$ (2070-2079) and Control (1970-1979) Annual Average Values. The top and middle rows are for the 4-m deep lake and the bottom row is for the 30-m deep lake.

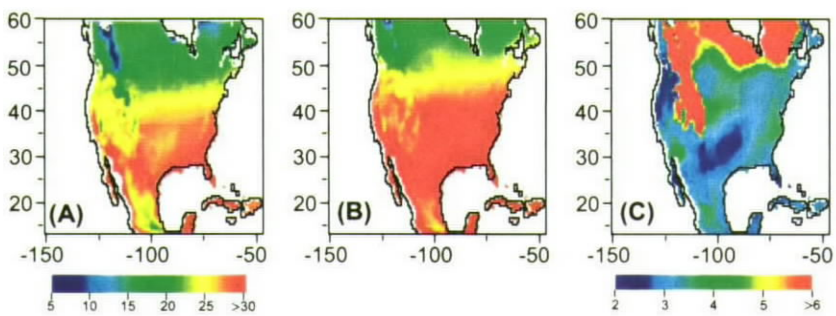

Figure 5. Ten-Year Average Lake Surface Temperatures Simulated with the CGCM1 Input Data Sets. August control (A); August $2xCO_2$ (B); and summer (June through August) differences ($2xCO_2$ minus control) (C).

ice, located at 41°N latitude and 88°W longitude; Figure 7). Over the last nine years of the control simulation, the ice-free lake displays a regular seasonal cycle of heating. Limited stratification occurs only during the fourth summer. In the $2xCO_2$ simulation, the lake is 2-4°C warmer than the control during all seasons (Figure 6). During some periods, changes in the seasonal onset or cessation of mixing cause larger changes in temperature (≥ 6°C) at various depths. Warming near the bottom during the spring-through-

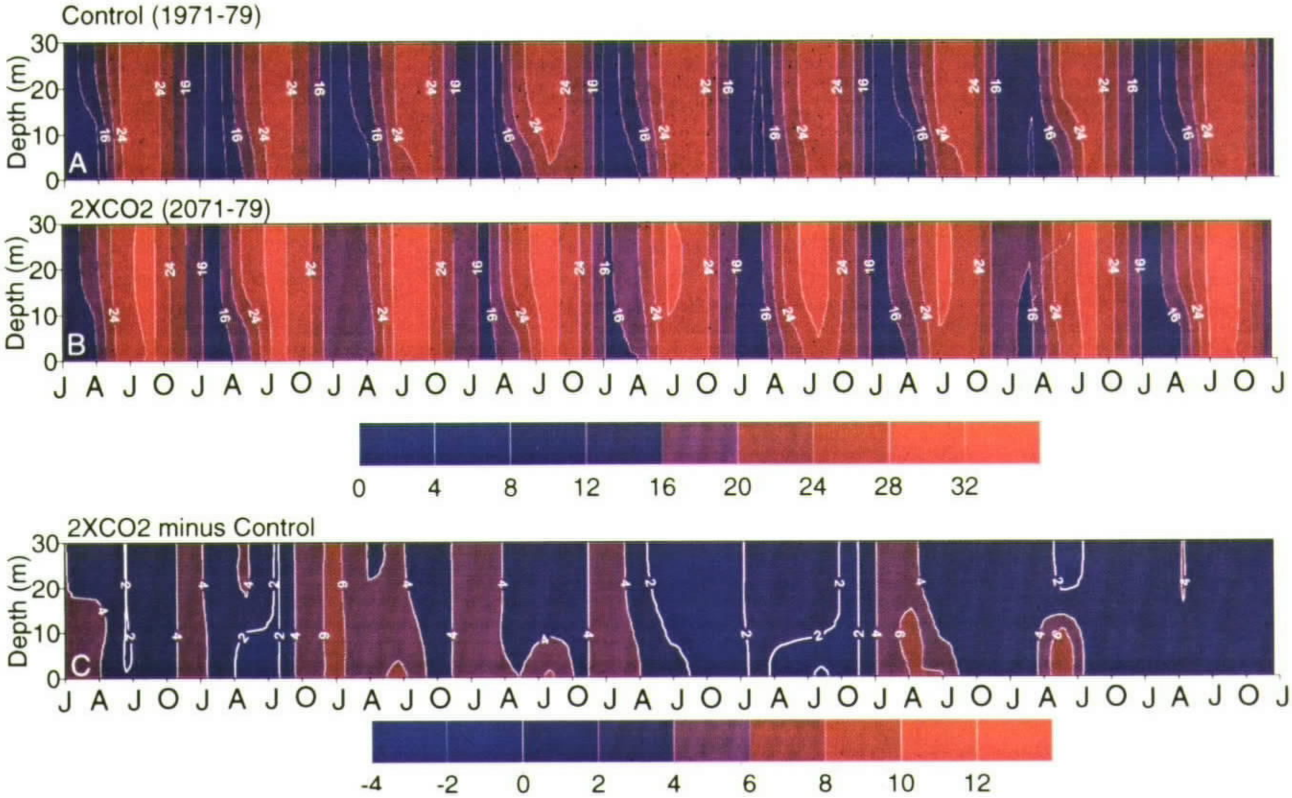

Figure 6. Simulated Evolution of Depth-Temperature Structure for a 30-m Deep Lake Over the Last Nine Years of Simulation Conducted with the CGCM1 Output. The lake is located at 30°N latitude and 105°W longitude on the model grid. Control (A), $2xCO_2$ (B) and anomalies (control minus $2xCO_2$) (C). Time axes are labeled for January, April, July, and October (J, A, J. O). The upper color scale corresponds to the control and $2xCO_2$ plots, and the lower scale corresponds to the anomalies. Units of °C.

summer period of the fourth, seventh, and eighth years suggests inhibited deep mixing, but the pattern of anomalies is not persistent. Inhibited deep mixing would likely be more important in deeper lakes (~50-100 m) that mix less frequently, and may lead to a complete loss or intermittent occurrence of seasonal overturn (Hostetler and Giorgi, 1995; Thompson *et al.*, 1998).

Substantially more change and variability in both temperature and mixing is simulated for the dimictic lake (Figure 7). Winter ice cover that is present in the control simulation is absent in the $2xCO_2$ simulation, and the lake warms throughout the water column by ≥ 4°C in all seasons. As in the control simulation, under $2xCO_2$ climatic conditions the lake stratifies during most summers, but at higher temperatures. The occurrence of smaller warm or cold anomalies (≤ 2°C), suggests a tendency for stratification at warmer temperatures to limit the duration of mixing in some years.

Combined with changes in precipitation, broad increases in lake evaporation in the $2xCO_2$ simulations (Figures 3 and 8) cause widespread changes in simulated net moisture. In the ECHAM4 simulation, drier conditions occur over much of the West and Great Plains, whereas wetter conditions occur over the Southeast, Midwest, and Caribbean. The CGCM1 produces similar drying over the Great Plains, but conditions wetter than those of the ECHAM4 simulation are produced over California and the Pacific Northwest. In opposition to the ECHAM4 results, somewhat drier conditions are simulated by the CGCM1 output over the East and the Southeast. Without a complete water-balance model that accounts for stream- and groundwater flow, exact details of lake-level changes cannot be quantified; however, the general pattern of the P-E anomalies suggests the overall direction of change that could be expected.

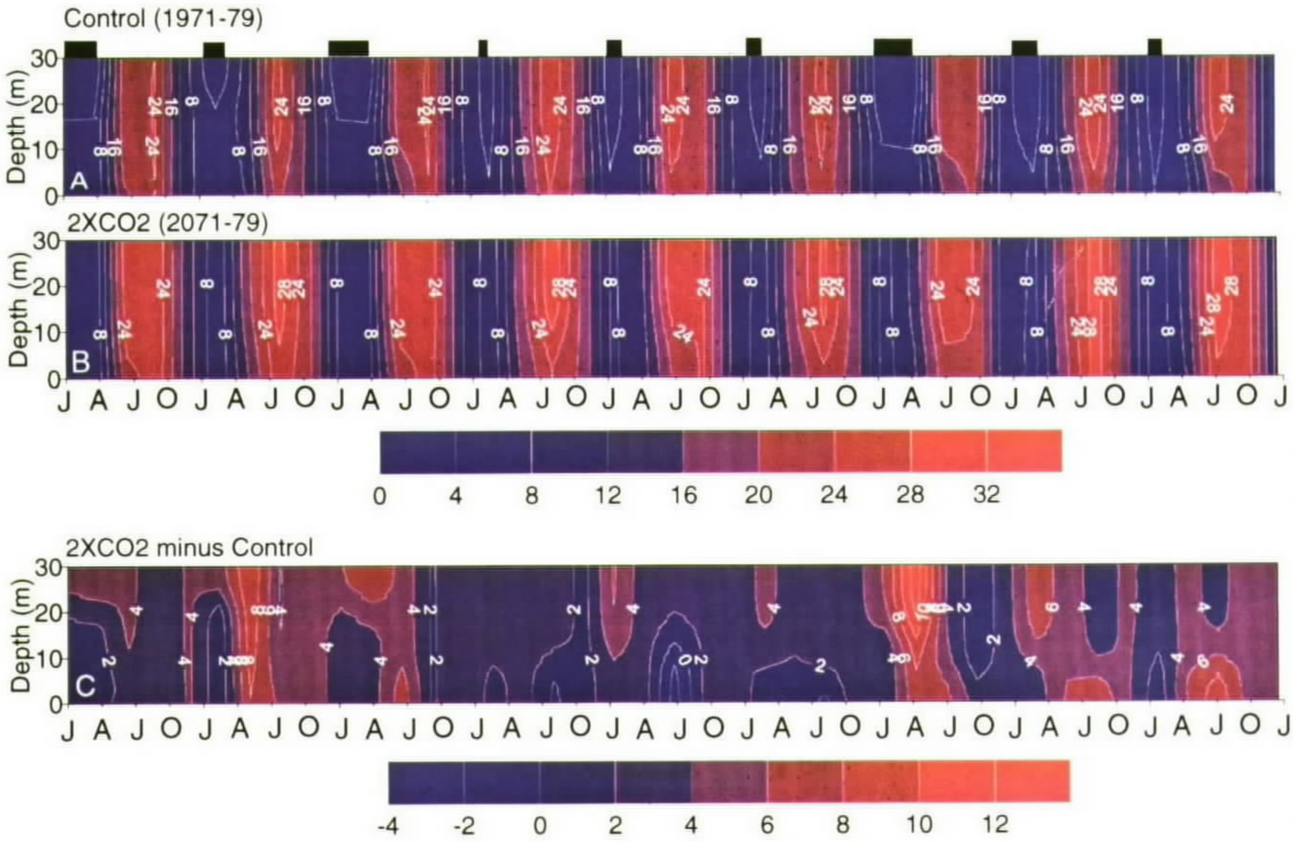

Figure 7. Simulated Evolution of Depth-Temperature Structure for a 30-m Deep Lake Over the Last Nine Years of Simulation Conducted with the CGCM1 Output. The lake is located at 41°N latitude and 88°W longitude on the model grid. Control (A), $2xCO_2$ (B) and anomalies (control minus $2xCO_2$) (C). Black bars on the control figure indicate the duration of winter ice cover, which displays substantial interannual variability. Time axes are labeled for January, April, July, and October (J, A, J, O). The upper color scale corresponds to the control and 2xCO2 plots, and the lower scale corresponds to the anomalies. Units of °C.

DISCUSSION

Although in detail the lake responses to climate warming differ in the ECHAM4 and CGCM1 simulations, several broad areas of agreement emerge. Both simulations exhibit a general warming of lakes of ~3°C in ice-free areas and 5°C or more over the Rocky Mountains, the Great Plains, and higher latitudes. Greater warming of the latter regions is attributed to the loss or reduction of winter ice cover. Substantial summer warming is simulated in response to longer heating seasons, warmer air temperatures, and increased longwave radiation in the $2xCO_2$ climates. Changes in the duration of mixing in the shallow lake are broadly similar in both simulations. There is less agreement between simulated changes in P-E, primarily due to differences in precipitation rates produced by the different A/OGCMs (Figure 1).

In our modeling study, we used the present and future climates produced by two A/OGCMs. Many other A/OGCMs have been used to simulate future climate change (IPCC, 1996a). Given the relatively wide range of simulated present and future climate produced by the various climate models (IPPC, 1996a), we expect that in detail we would obtain a similar range of lake responses that in aggregate would yield the same level of broad agreement we obtained with the ECHAM4 and CGCM1.

The lake responses presented here were obtained with a lake model that was applied in what is termed a "stand-alone mode." In such an application, output from a climate model is used to drive the lake model, but there is no feedback between the lake and the atmosphere. Both positive and negative feedbacks between a lake and the atmosphere govern the overall response and sensitivity of a lake to climate. Because the lake cannot modify the atmospheric inputs, and because the climate inputs were obtained from atmospheric models in which lakes are unresolved, the simulated responses obtained in our stand-alone application maximize the sensitivity of lakes to climate change. Nonetheless, our results provide physically consistent estimates of the sensitivity of lakes to future climate states. We are addressing the issue of interactive feedback by coupling the lake model with a high-resolution atmospheric model (e.g., Hostetler *et al.*, 1993, 1994; Bates *et al.*, 1993, 1995; Small *et al.*, 1999).

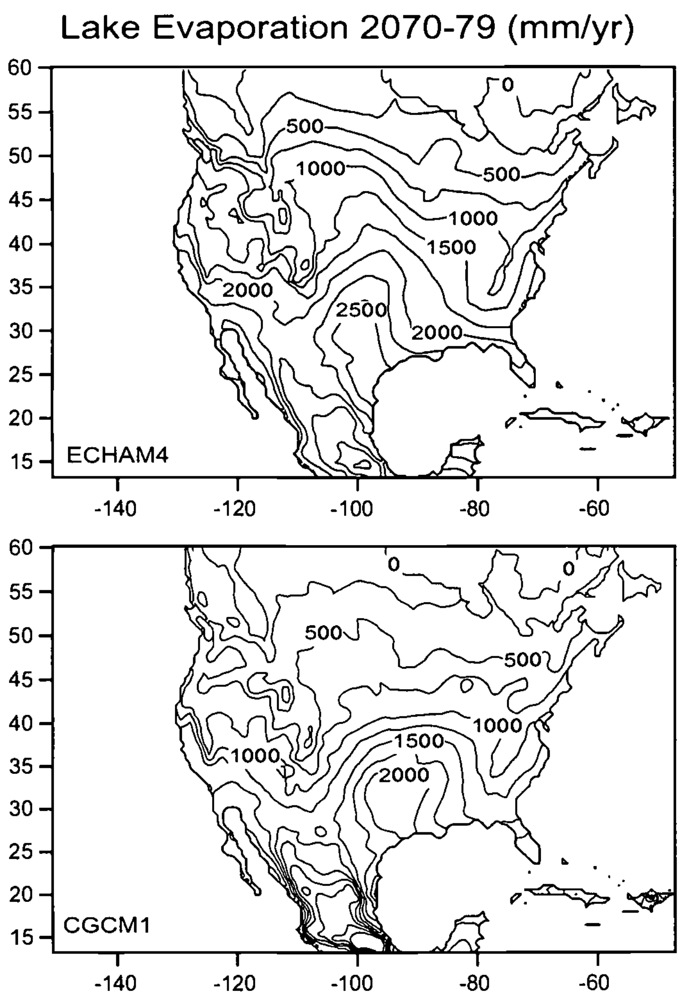

Figure 8. Annual Evaporation Rates as Simulated for $2xCO_2$ Conditions (2070-2079) Using the ECHAM4 and CGCM1 Input Data Sets. Contour interval of 250 mm/yr.

Our focus here has been on assessing changes in mean lake thermal properties in response to changes in the mean climate state as portrayed in the $2xCO_2$ simulations. It is likely that climatic variability will play an equal or greater role in determining future conditions in lakes. Extended and more frequent periods of climatic extremes, for example, could amplify or offset changes in the mean state. Although it is an inherent property of some lakes and wetlands to attenuate the effects of short-term climatic variability, if the amplitude and duration of such variability is large, the effect on the ecology of lakes and wetlands may override responses to change in average conditions (e.g., Katz and Brown, 1992). Assessing the effects of climate variability on lakes and wetlands warrants further study.

The effect of average warming on the aquatic ecology of a given lake would be determined by many factors including geographic location, physiographic setting, original trophic state, and water chemistry. Some basic qualitative responses to overall warmer lake conditions can be inferred by comparing our model results with more site-specific studies, but a complete analysis of the effects of climate-driven changes in lake thermal structure on the aquatic habitats of specific lakes or wetlands requires additional modeling of parameters such as dissolved oxygen, biologic productivity, water chemistry, and the complete water balance.

Over the southern and south-central regions of the domain, August water temperatures increase from ~28°C to 32°C or higher. Combined with sustained increases of 3-4°C throughout summer, such warming would likely present a substantial stress on fish, particularly if thermal refugia at depth were reduced or eliminated (Figures 6 and 7). Where productivity is not limited by high water temperatures, warmer conditions may lead to more eutrophic states. Warmer water could affect dissolved oxygen levels, especially if warmer conditions were accompanied by increased productivity, leading to increased oxygen demands for aerobic decomposition and possibly anoxic conditions at depth (Lam *et al.*, 1987; McCormick, 1990; Schertzer and Sawchuck, 1990; Jewell, 1992; Fang *et al.*, 1999). Warmer conditions may also lead to degraded water quality (McCormick, 1990), increased summer anoxia (Stefan *et al.*, 1993; Stefan and Fang, 1994), and loss of productivity in boreal lakes (Schindler *et al.*, 1996).

Loss of winter ice cover and associated warming may be beneficial to fish populations where productivity and growth presently are limited by the duration of open water periods and cold summer temperatures (e.g., Magnuson *et al.*, 1990; Hostetler and Giorgi; 1995; Porter *et al.*, 1996; Stefan *et al.*, 1998; Fang *et al.*, 1999). Productivity may be negatively affected, however, if longer growing periods for fish are not accompanied by increased food availability (McDonald *et al.*, 1996). A reduction in the duration of ice may also be beneficial to northern lakes in which winter anoxia under ice causes fish kills (Fang *et al.*, 1999).

Ecological communities that rely on physical processes that currently occur under ice before spring breakup could be disrupted by a substantial loss of lake ice (Kilham *et al.*, 1996).

Reduced P-E values in the $2xCO_2$ simulation indicate the potential for lower water levels in lake basins and wetlands, whereas increased P-E would probably raise the level of most lakes and wetlands. Decline in lake depth, coupled with associated lower flows from surface and groundwater sources, may exacerbate increases in temperature and related problems with loss of thermal refugia and decreased levels of dissolved oxygen during summer and winter. In addition, increased evaporative losses coupled with decreased freshwater inputs may alter lake chemistry, causing freshwater lakes to become more saline, and perhaps lead to meromixis (Romero and Melak, 1996). A trend to drier conditions in one region may accompany a trend to wetter conditions in a neighboring region; thus migratory waterfowl that rely on wetlands for nesting grounds may be able to adapt to drier conditions by finding adequate habitat elsewhere.

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