Lake surface temperatures in a changing climate: a global sensitivity analysis

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Abstract

We estimate the effects of climatic changes, as predicted by six climate models, on lake surface temperatures on a global scale, using the lake surface equilibrium temperature as a proxy. We evaluate interactions between different forcing variables, the sensitivity of lake surface temperatures to these variables, as well as differences between climate zones. Lake surface equilibrium temperatures are predicted to increase by ~70-85% of the increase in air temperatures. On average, air temperature is the main driver for changes in lake surface temperatures, and its effect is reduced by ~10% by changes in other meteorological variables. However, the contribution of these other variables to the variance is ~40% of that of air temperature, and their effects can be important at specific locations. The warming increases the importance of longwave radiation and evaporation for the lake surface heat balance compared to shortwave radiation and convective heat fluxes. We discuss the consequences of our findings for the design and evaluation of different types of studies on climate change effects on lakes.

Keywords

climate change, lakes, heat fluxes, modeling, sensitivity analysis, palaeotemperatures, space-for-time

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1 Introduction

The most direct effect of climate change on lakes is a warming of the lake surface temperature (T_{surf}). Warming trends of T_{surf} mostly between 0.01 and 0.1 °C yr^{-1} have been observed in the last decades both *in-situ* (Shimoda et al. 2011), and from satellites, with a spatial distribution similar to air temperature (T_{air}) trends (Schneider and Hook 2010). Increased T_{surf} affects the timing, duration and intensity of stratification and seasonal deep convective mixing (Livingstone 2003; Fang and Stefan 2009), as well as the timing and duration of ice cover (Weyhenmeyer et al. 2011). Lakes may shift from one seasonal mixing type to another if minimum temperatures exceed a certain threshold. Dimictic lakes may become monomictic (Livingstone 2008), while winter mixing may become incomplete or sporadic in monomictic lakes (Straile et al. 2003).

Changes in temperature and stratification further affect biological and geochemical processes. Oxygen concentrations may decrease in deep stratified lakes (Matzinger et al. 2007) and increase in ice-covered lakes (Fang and Stefan 2009). Nutrient availability for primary producers is expected to increase (Trolle et al. 2011), and harmful algal blooms are predicted to occur more frequently (Paerl and Paul 2012). Changes in the fish assemblage composition have already been observed (Jeppesen et al. 2012), while more complex interactions between physical and biological processes, such as a disruption of the linkage between different trophic levels (Winder and Schindler 2004) need further investigation. These and other effects of climate change on lakes have been discussed in several recent reviews (Adrian et al. 2009; George 2010; Shimoda et al. 2011; Winder and Sommer 2012). Furthermore, as reviewed by Adrian et al. (2009), lakes can be used as sentinels for current and past climate change, and near-surface temperature trends and ice phenology can be valuable indicators for changes in T_{air}. In summary, there is great interest in understanding past and future effects of climate change on physical, chemical and biological processes in lakes.

The Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change (IPCC 2007) provided an overview of the state of climate change predictions. Climate change was simulated for the same greenhouse gas emission scenarios by 23 different atmosphere-ocean general circulation models (AOGCMs). Predictions made for future T_{air} trends are relatively robust, with uncertainties on the order of ~50% (Knutti 2008), while those for other forcing variables such as wind speeds are less reliable, as the variation between models is often of a similar magnitude as the predicted changes.

The aim of the present study is to investigate the sensitivity of T_{surf} to climate variables on a global scale. The sensitivity of T_{surf} to T_{air} has previously been investigated for individual lakes (e.g. Robertson and Ragotzkie 1990; Peeters et al. 2002; Verburg and Hecky 2009) or on a regional to continental scale (e.g. Stefan et al. 1998; Fang and Stefan 2009). All these studies concluded that T_{surf} is expected to increase by 50 to 90% of T_{air}. Livingstone and Imboden (1989) showed that most of the interannual variation in T_{surf} in Lake Aegeri was caused by variations in cloud cover and relative humidity, rather than T_{air}. Austin and Allen (2011) evaluated the sensitivity of T_{surf} in Lake Superior to changes in T_{air}, wind speed, and the extent of ice cover in the previous winter. All three variables were shown to affect T_{surf} to a similar extent.

In the present study, we aim to investigate the following questions:

1. How sensitively does T_{surf} react to changes in meteorological forcing variables?
2. What is the relative importance of meteorological variables other than T_{air} for future trends in T_{surf} on a global scale?
3. What are the differences in the forcing mechanisms between different climate zones?
4. Do the answers to questions 1 to 3 depend on the choice of the AOGCM used to predict T_{surf}?
5. What are the resulting implications for climate change studies on lakes?

We use the lake surface equilibrium temperature (T_{eq}), i.e., the temperature at which the net heat flux at the lake surface is zero for given meteorological conditions, as a proxy for T_{surf}. The implications of this choice are discussed in section 4.1. It is important to note that the focus of the present study is not on trying to precisely predict changes in T_{surf} globally and even less for specific lakes.

2 Materials and methods

2.1 Climate scenarios

Calculations are based on climate predicted for AR4 by the six AOGCMs BCCR-BCM2.0, CGCM3.1, CNRM-CM3, GISS-EH, INM-CM3.0, and Miroc3.2(hires), downloaded from the World Climate Research Programme's (WCRP) Coupled Model Intercomparison Project (CMIP3) database (Meehl et al. 2007). The models were chosen to represent a wide variety of the AOGCMs in AR4, by including models with different grid resolutions, and by selecting models which predict different spatial distributions of present-day surface temperatures (Figure 2 in Knutti 2008). Details on model resolution, the number of grid cells considered and the downloaded variables are given in the supplementary material. Scenario 20C3M was used for the baseline climatology of 1961 to 1990, and SRES A1B for the future climate of 2070 to 2099. SRES A1B has a medium impact, with a globally averaged increase in T_{air} of 2.8 °C (1.7 to 4.4 °C). 30-yr arithmetic means were calculated from the monthly data for each grid cell and each month for both scenarios. The term anomaly, indicated with the symbol Δ, is used in the following for the difference of the arithmetic means of any variable between the future climate and the baseline climatology.

2.2 Calculation of lake surface equilibrium temperature

T_{eq} is defined as the temperature at which the net heat flux at the lake surface (H_{tot}), given by the sum of the individual heat fluxes (H_i), is zero (Edinger et al. 1968):

$$H_{tot}\left(T_{eq}\right) = H_S + H_A + H_W + H_E + H_C = 0 \qquad (1)$$

Heat fluxes from the atmosphere to the lake are defined to be positive. The lake gains heat from downward shortwave (H_S) and longwave (H_A) radiation, and emits heat by longwave radiation (H_W). The latent heat flux (H_E) due to evaporation or condensation, and the convective heat flux (H_C) can be positive or negative. Numerous equations have been proposed to estimate the individual H_i from the forcing variables T_{air}, vapour pressure (e_a), cloudiness (C), solar radiation at the lake surface (H_{S0}) and wind speed 10 m above the lake surface (u_{10}) (Henderson-Sellers, 1986). The set of equations used in this study is given in the supplementary material, and is mainly based on Livingstone and Imboden (1989). The largest uncertainties due to the choice of equations result from the emissivity of the atmosphere (E_a) for longwave radiation, and the equation for H_E. However, a sensitivity analysis showed that the influence of the choice of equations for these two factors on ΔT_{eq} was small (see supplementary material).

For each grid cell of the six AOGCMs, we calculated the fractions of wetland types defined in the Global Lakes and Wetlands Database (GLWD; Lehner and Döll 2004) and of the climate zones according to the revised Köppen-Geiger classification (Kottek et al. 2006). The climate zones were grouped in A (tropical) B (dry) C (temperate), D (continental), ET (tundra) and EP (ice caps). Each cell was assigned the climate zone of which it contained the largest fraction. The following rules were applied to decide whether a grid cell was included in the global analysis: grid cells with land fraction (*sftlf*) < 10% were discarded except if they contained >10% lakes in GLWD (*sftlf* is defined as the percentage of land surface in BCCR-BCM2.0, GISS-EH, and MIROC3.2(hires) and as either land or water in the other models; large lakes are not considered as land); cells with Köppen-Geiger classification EP were discarded; those with classification B were discarded, except if they contained more than 0.1% lakes, reservoirs or permanent wetlands in GLWD.

Finally, grid cells with $T_{eq} \leq 0$ for either the 20C3M or the A1B scenario were excluded from the analysis. At $T_{eq} \leq 0$ °C shallow lakes are expected to freeze and the used heat budget equations become invalid. In deep lakes, substantial amounts of heat can be stored in the deep water during the warm season, and T_{eq} may need to remain below 0°C for many weeks until enough of this heat has been released to the atmosphere such that an inverse stratification and subsequently an ice-cover can develop. For calculating global and zonal averages, grid cells were weighted by surface area. Data were averaged first spatially and then temporally.

3 Results

3.1 Sensitivity of equilibrium temperature to forcing variables

Before discussing the results of the calculations based on the AOGCMs, we evaluate the sensitivity of T_{eq} to the forcing variables for a range of combinations of T_{air} and relative humidity (h_{rel}) as well as for combinations of T_{air} and u_{10}. The values of the other forcing variables are kept constant as given in the caption of Figure 1.

The difference between T_{eq} and T_{air} (Figure 1a) increases with decreasing $T_{air,}$ and u_{10}, and with increasing h_{rel}. T_{eq} exceeds T_{air} except for hot and dry conditions and for high wind speeds,. Consequently, T_{surf} is expected to be warmer than daily average T_{air} once it has reached equilibrium.

But how long does it take for T_{surf} to approach T_{eq}? A departure of T_{surf} from T_{eq} by 1 °C results in a H_{tot} of -20 to -35 W m^{-2} (Figure 1b). As an example, with H_{tot} = 30 W m^{-2} the temperature of a 5 m thick surface mixed layer would increase by ~0.12 °C d^{-1} and thus exponentially approach T_{eq} with a time scale of 8 days. If the mixed layer is thicker, the time scale increases correspondingly. The net heat flux resulting from a departure of T_{surf} from T_{eq} increases with T_{air} (Figure 1b), because both H_W and H_E increase more than linearly with T_{air}, and therefore also with h_{rel}, because T_{eq} increases significantly with h_{rel} (Figure 1a). Consequently, T_{surf} approaches T_{eq} faster in warm and humid than in cold and dry climates.

Figures 1c to 1e show the sensitivities of T_{eq} to changes in the individual forcing variables. T_{eq} is typically expected to increase by ~70-90% of T_{air} (Figure 1c). The sensitivity function $\partial T_{eq}/\partial T_{air}$ is almost independent of u_{10} and T_{air} but increases under humid conditions, where a smaller fraction of the additional heat input is counterbalanced by increasing evaporation. The sensitivity to changes in h_{rel} is highest under warm and dry conditions (Figure 1d). On average, T_{eq} is predicted to increase by ~0.1 °C if h_{rel} increases by 1%. An increase of u_{10} by 1 m s^{-1} leads to a decrease of T_{eq} by ~1 °C (Figure 1e). This effect is stronger for calm, warm and dry conditions. The sensitivity of T_{eq} to solar radiation (Figure 1f) is directly related to the sensitivity of H_{tot} to T_{surf} (Figure 1b), because any change in H_S needs to be counterbalanced by a corresponding change in T_{surf} to keep $H_{tot} = 0$. The effect of changes in H_S is therefore strongest in cold and dry climates. Typically, an increase in H_S by 1 W m^{-2} leads to an increase in T_{eq} by 0.03 – 0.06 °C.

In summary, we conclude that T_{eq} increases more slowly than T_{air}, that T_{eq} is approached faster in warm climate zones, and that the sensitivity of T_{eq} to changes in h_{rel} and u_{10} increases while that to changes in H_S decreases with T_{air}. The following rules of thumb can be used to derive a first estimate of the relative contributions of different forcing variables to changes in T_{eq}: an increase in T_{eq} by 1 °C is achieved by increasing T_{air} by 1.2 °C, decreasing u_{10} by 1 m s^{-1}, increasing h_{rel} by 10%, or increasing H_S by 25 W m^{-2}.

3.2 Climate scenarios

The anomalies of the meteorological variables predicted by the AOGCMs for the different climate zones are shown in Figure 2. Five of the six models predict similar ΔT_{air}, whereas MIROC3.2(hires) forecasts significantly higher warming, mainly because of a stronger shortwave cloud feedback (Yokohata et al. 2008). The average ΔT_{air} is similar for all climate zones, since the rule T_{eq} > 0 °C excludes winter months at high latitudes, for which generally a strong warming is predicted. Relative humidity and cloudiness are predicted to decrease in all climate zones. Average wind speed anomalies are almost consistently positive except for the continental climate zones. Solar radiation is expected to decrease in the tundra and continental zones with no consistent trends elsewhere.

3.3 Lake surface equilibrium temperatures

The anomalies in lake surface equilibrium temperature (ΔT_{eq}) and their global distribution vary significantly among the different models (Figure 3). As expected from ΔT_{air}, ΔT_{eq} calculated based on the model MIROC3.2(hires) was highest, especially in the continental climate zone (Table 1) and between 40°N and 70°N (see supplementary material). High ΔT_{eq} were consistently predicted in the region of the Laurentian Great Lakes and in the Amazon Basin. Similar to ΔT_{air}, ΔT_{eq} is comparably low at high latitudes for several models because only months with $T_{eq} > 0$ were considered. The average predicted ΔT_{eq} is ~70% of ΔT_{air} in dry and tundra climates, ~80% in tropical and temperate climates, and ~85% in continental climates (Table 1).

Table 1 Annual mean anomaly of the lake surface equilibrium temperature, ΔT_{eq} (°C), in 2070 to 2099 with respect to 1961 to 1990 based on six AOGCMs for five different climate zones and globally. In parentheses: ΔT_{eq} as a percentage of ΔT_{air}, and average number of months per year with $T_{eq} > 0$.

	ΔT_{eq} (°C) *(% of ΔT_{air}, average number of months)*					
	Tropical	Dry	Temperate	Continental	Tundra	**Global**
BCCR-BCM2.0	2.11 *(84%, 12)*	2.07 *(71%, 10.7)*	2.23 *(84%, 11.1)*	2.41 *(86%, 5.9)*	1.79 *(79%, 4.4)*	**2.14** ***(81%, 8.4)***
CGCM3.1	2.67 *(83%, 12)*	2.57 *(74%, 10.6)*	2.53 *(86%, 10.7)*	2.55 *(92%, 5.7)*	2.28 *(73%, 4.1)*	**2.57** ***(82%, 8.3)***
CNRM-CM3	2.77 *(82%, 12)*	2.71 *(73%, 10.8)*	2.57 *(81%, 11.3)*	2.46 *(83%, 6.3)*	2.42 *(78%, 5.0)*	**2.62** ***(80%, 8.7)***
GISS-EH	2.30 *(82%, 12)*	2.17 *(65%, 10.9)*	2.05 *(77%, 11.2)*	2.01 *(76%, 6.2)*	1.75 *(64%, 4.9)*	**2.12** ***(75%, 8.7)***
INM-CM3.0	2.02 *(75%, 12)*	2.33 *(71%, 11.0)*	2.25 *(78%, 11.6)*	2.69 *(87%, 6.4)*	2.37 *(72%, 4.8)*	**2.27** ***(76%, 8.8)***
MIROC3.2(hires)	3.46 *(77%, 12)*	3.53 *(71%, 10.8)*	3.57 *(81%, 11.7)*	4.82 *(85%, 6.6)*	3.66 *(70%, 5.0)*	**3.76** ***(78%, 8.8)***
Multi-model mean	2.56 *(80%, 12)*	2.56 *(71%, 10.8)*	2.53 *(81%, 11.3)*	2.82 *(85%, 6.2)*	2.38 *(72%, 4.7)*	**2.58** ***(79%, 8.6)***

3.4 Heat fluxes

The net heat flux H_{tot} at T_{eq} and therefore also its anomaly ΔH_{tot} are per definition both equal to zero. However, the components H_i of H_{tot} do change (Figure 4). H_A on average increases by ~20 W m^{-2} (~30 W m^{-2} for MIROC3.2(hires)), which is partially offset by a corresponding ΔH_W of ~-15 W m^{-2} (~-20 W m^{-2} for MIROC3.2(hires)). Both H_A and H_W change more at low than at high latitudes. ΔH_E is negative, i.e., evaporation is predicted to increase, for all models in all climate zones. Conversely, positive ΔH_C of a few W m^{-2} are predicted because T_{surf} increases less than T_{air}. For most lakes, H_C is negative for the present climate. On average, the absolute value of H_C decreases by ~20%. The average ΔH_S is small, except for a decrease by ~-5 W m^{-2} in the tundra zone. These changes result in a shift in the importance of the contributions of the individual heat fluxes to the heat budget: longwave radiation gains importance as a heat source compared to shortwave radiation, and evaporation gains importance as a heat loss compared to both longwave radiation and the convective heat flux.

3.5 Contributions of the different driving variables

In order to determine the contributions of the forcing variables to ΔT_{eq}, we recalculated ΔT_{eq} using the future climate for each variable while keeping all other variables at the baseline climatology. For ΔT_{air}, we calculated two scenarios, keeping either e_a or h_{rel} at the baseline. The results are displayed in Figure 5. For all models and climate zones ΔT_{air} is the most important driving variable, and the only one that on average increases T_{eq}. The warming by ΔT_{air} would be reduced by 40-50% due to increased evaporation and a lower increase of H_A if e_a remained constant. However, the AOGCMs consistently predict that e_a follows T_{air} with only a slight average reduction in h_{rel}.

In total, the anomalies of the variables besides ΔT_{air} reduce the warming caused by ΔT_{air} on average by only ~9%, half of which is due to the decrease in h_{rel}. Nevertheless, these variables can be important at specific locations. This is highlighted by the root mean square (rms) of the contributions of the different variables (Figure 5) which is a measure of the variability they induce to ΔT_{eq}. On average, the variability caused by variables other than ΔT_{air} is ~40% of that caused by ΔT_{air}. The variability caused by ΔH_S increases with latitude, while that induced by Δh_{rel} is highest in warm dry regions.

The correlations between anomalies in the driving variables and ΔT_{eq} are, as expected, strongest for ΔT_{air}, positive for ΔT_{air} and ΔH_{S0}, and negative for Δh_{rel}, Δu_{10} and ΔC. Interestingly, the correlation between ΔT_{air} and ΔT_{eq} significantly increases with latitude, reaching values of r = 0.8 for continental and tundra climates, but only 0.5 for tropical climates, even though most of ΔT_{eq} is caused by ΔT_{air} in all climate zones. Conversely, the correlation of ΔT_{eq} with Δu_{10} becomes weaker from -0.5 for tropical climates to -0.2 for tundra and continental climates. The correlations with Δu_{10} are surprisingly strong considering that Δu_{10} contributes little to the average and rms of ΔT_{eq}. The reason for these trends in correlations is the growing importance of H_E and its anomaly compared to H_A and H_W towards lower latitudes (Figure 4). ΔH_{S0} and Δh_{rel} are generally strongly negatively correlated (r ~ -0.6), because of the positive correlation between ΔH_{S0} and ΔC. Since H_{tot} increases both with H_{S0} and h_{rel}, their effects on ΔT_{eq} partially cancel each other.

4 Discussion

4.1 Equilibrium temperatures versus real surface temperatures

For shallow lakes, i.e., for the majority of lakes worldwide, T_{eq} is a good proxy for T_{surf} on the seasonal time scale (Wilhelm et al. 2006). It has also been shown to be linearly related to observed stream temperatures, although these are often modified by groundwater, shading, or meltwater (Bogan et al. 2003). Furthermore, ΔT_{eq} is an even better proxy for ΔT_{surf}, as system-specific differences between T_{eq} and T_{surf} are largely cancelled by taking the difference between the future and the baseline scenario.

In deep lakes, however, T_{surf} is strongly modified by mixing processes. When they are stratified and surface mixed layers are thin, e.g. during spring and summer in temperate lakes, T_{eq} is a good measure of T_{surf} (see supplementary material). But during periods of deep mixing larger water volumes are involved in the surface heat balance and T_{surf} responds slowly to changes in T_{eq} (Livingstone and Lotter 1998). For most temperate and boreal deep lakes, however, periods of deep mixing occur mainly during months when $T_{eq} \leq 0$ °C and are thus not included in our evaluation.

As wind strongly affects mixing in lakes, the relative effect of wind speed on T_{surf} should be greater than that on T_{eq}. The effects of wind speed could be additionally underestimated for two reasons: Local winds can exceed the average large scale winds predicted by AOGCMs; and the predicted changes in the difference between air and lake temperatures can modify winds created by large lakes. For example, Desai et al. (2009) showed that winds created by Lake Superior have intensified due to the decreasing difference between T_{air} and T_{surf}, and that the effects of wind speed and T_{air} on trends in T_{surf} were of similar importance.

Water transparency further modifies T_{surf}. In clear, stratified lakes, a significant fraction of H_S penetrates the surface mixed layer and does not contribute directly to its heat budget. Consequently, T_{surf} can be significantly below T_{eq} (Rinke et al. 2010). Furthermore, an increase in H_A, which is absorbed at the lake surface, has a different effect on the thermal stratification of a clear lake than an increase in H_S, although they both equally affect T_{eq}. The predicted ΔT_{eq} is mainly due to changes in heat fluxes that are absorbed or emitted directly at the lake surface (Figure 4). In combination these effects contribute to the trend toward shallower thermocline depths and stronger stratification that has been observed in clear, stratified lakes (Coats et al. 2006; Livingstone 2003). Conversely, in turbid lakes T_{surf} decreases with increasing turbidity, as sunlight can only warm up a fraction of the layer that is mixed every night (Houser 2006). Finally, T_{surf} can be modified by inflows, especially in lakes with short residence times, or by extensive lake level variations (Rimmer et al. 2011).

These individual response mechanisms highlight that the results of our study should not be used to predict changes in T_{surf} for a specific lake. For this purpose, it is more advisable to use a specifically calibrated lake model in combination with climate forcing derived from a regional climate model, as it has been done for example by Komatsu et al. (2007) for Lake Biwa. Nevertheless, lake surface temperatures, especially during summer, have been shown to be regionally highly coherent, and largely driven by the regional climate forcing (Livingstone et al. 2010).

4.2 Sensitivity of lake surface temperature to air temperature

The predicted ΔT_{eq} of 70 to 85% of ΔT_{air} agrees well with previous studies. Hondzo and Stefan (1993) simulated temperature and stratification in Minnesota lakes of different size and depth for a climate scenario with doubled atmospheric CO_2 concentrations. Average epilimnion temperatures increased by ~3.0 °C or ~70% of the increase in T_{air} of 4.4 °C. Fang and Stefan (2009) made analogous simulations for the contiguous United States. Mean annual T_{air} increased by 6.7 °C, mean annual T_{surf} by 5.2 °C, or 78% of the increase in T_{air}. River temperatures have been predicted to increase by approximately 65% of the increase in T_{air}, provided that discharge remains constant (van Vliet et al. 2011).

4.3 Implications for studies of climate change effects on lakes

In the following we discuss the implications of our results for different approaches to estimate climate change effects on lakes. Following the space-for-time approach, climate change effects are inferred from existing climate gradients, either altitudinal (Karlsson et al. 2005) or latitudinal (Weyhenmeyer 2008). However, lakes over a latitudinal gradient are not only subject to different T_{air}, but also to different day lengths that are not affected by climate change, as well as to different contributions of individual H_i (Figure 4). Livingstone et al. (2005) showed a clear relationship between T_{surf} and smoothed T_{air} in summer for 29 Swiss alpine lakes spanning an altitude range of 2000 m. In this study the lapse rate of T_{surf} exceeded that of T_{air}, in contrast to the expectations from our calculations. This suggests that the decrease in T_{surf} with altitude results not only from decreasing T_{air}, but also from gradients in other variables such as cloud cover or snowmelt. Lakes at different altitudes or latitudes are therefore not necessarily a good surrogate for predicting climate change effects.

Modeling studies are another common approach to predicting climate change effects on physical and biogeochemical processes in lakes. Often, the output from a global or a regionally downscaled climate model is used to drive such models (Fang and Stefan 2009). This has the advantage of covering the influences of all major driving variables. However, the different processes can be difficult to disentangle, and uncertainties in the prediction of some climate variables may blur the effects of variables such as T_{air} that can be predicted with higher confidence. For example, large and highly uncertain wind speed anomalies in autumn significantly affected predictions by Stefan et al. (1996). These issues can be avoided by modifying individual variables, such as T_{air} or u_{10} (Trolle et al. 2011), but then negative or positive feedbacks caused by correlations between climate variables may be overlooked. Furthermore, large lakes influence the local climate, and coupled lake-climate models may be required to understand how global change modifies such climate-lake systems (Martynov et al. 2010).

Our results confirm that, on average for a large number of sites, changing only T_{air} should result in good approximations for predicting climate change effects on T_{surf}. However, there are three caveats: first, this is not necessarily true for specific locations, where changes in other driving variables might be as important as changes in T_{air}. Second, it is conceivable that climate models fail to correctly predict changes in variables such as u_{10}, e_a, C, or H_{S0}. Third, if a model is driven by specific humidity, it is important to increase specific humidity to ensure that h_{rel} remains constant, even though this might lead to ΔT_{eq} being overestimated in cases where there is not sufficient water available for humidity to follow T_{air}. Keeping specific humidity constant on average results in an underestimation of ΔT_{eq} by ~40%.

The comparably weak correlation between ΔT_{eq} and ΔT_{air} in tropical and dry climate zones might be regarded as a caveat for the use of biotic and abiotic proxies in sediment records for reconstructing past T_{air}. Such records are especially valuable in the tropics, where the potential of other climate archives such as ice cores and tree rings is limited (Verschuren 2003). However, reconstructing past climate from lake sediments requires conceptual models for the dependence of sediment proxies on climatic forcing. Often complex interactions are involved, and for some proxies, e.g. branched glycerol dialkyl glycerol tetraethers (Loomis et al. 2012), or chironomid assemblages (Eggermont et al. 2010), it is not yet clear to what extent they record air or water temperatures. If they record water temperatures, our results indicate that these are not necessarily well correlated to T_{air}, especially in tropical regions. This highlights the importance of understanding all processes connecting the driving variable (usually T_{air}) with the formation of a proxy variable in lacustrine sediments.

Finally, even though predictions derived from different AOGCMs generally agree, significant deviations for specific models and parameters exist. As discussed above, MIROC3.2(hires) results in much higher ΔT_{eq}, and the spatial distribution of ΔT_{eq} differs between models. All models predict negative correlations between ΔC and ΔT_{eq} except INM-CM3.0. The reason is that ΔC is strongly related to Δe_a (r = 0.50) in INM-CM3.0, whereas this correlation is weak or even negative in the other models (r between -0.27 and 0.18) where ΔC is much more related to Δh_{rel}. As a consequence, also the correlation between ΔH_{S0} and ΔT_{eq} is much weaker in INM-CM3.0 than in the other models. Furthermore, the models significantly differ in the predicted effects of forcing variables other than T_{air}. For example, the rms of the contribution of Δu_{10} to ΔT_{eq} is almost twice as high for CGCM3.1 as for CNRM-CM3, and that of ΔH_{S0} is ~50% higher than that of the other models. It is therefore advisable to use the output of several climate models for predicting climate change effects on lake surface temperatures and for evaluating the importance of different drivers.

5 Conclusions

On a global scale, the warming of lakes during the 21st century is expected to be mainly driven by increasing T_{air}. In agreement with previous studies, T_{eq} are predicted to increase by ~70 to 85% of the increase in T_{air}. The warming caused by T_{air} is on average reduced by ~10% by changes in other meteorological variables. However, on a local scale and for specific lakes, the effects of other variables can be of similar importance as those of T_{air}. Especially in tropical regions, the correlations between ΔT_{eq} and ΔT_{air} are relatively low, and lake temperature proxies are not necessarily good proxies for past changes in T_{air}. The relative contributions of the individual heat fluxes to the heat budget of lakes are expected to change, as incoming longwave radiation and evaporation gain importance compared to the other heat fluxes. The predicted changes are generally similar for all AOGCMs used in this study, but some relevant differences exist. It is therefore advisable to use the output of a few different AOGCMs for robust predictions of climate change effects on lakes.

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Figure Captions

Fig. 1 (a) Difference between lake surface equilibrium temperature T_{eq} and air temperature T_{air}; (b) sensitivity of the total heat flux, H_{tot}, to changes in lake surface temperature, T_{surf}; (c-f) sensitivity of T_{eq} to changes in (c) T_{air}, (d) relative humidity, h_{rel}, (e) wind speed, u_{10}, and (f) solar radiation, H_{S0}; all as functions of T_{air} and h_{rel} (left panels) and T_{air} and u_{10} (right panels); The other forcing variables were kept constant at C = 0.5, H_S = 200 W m^{-2}, p_{air} = 1000 hPa, u_{10} = 2 m s^{-1} (left panels), and h_{rel} = 0.5 (right panels). Note the different color scales

Fig. 2 Mean anomalies of the forcing variables in six AOGCMs in the SRES A1B scenario in 2070 to 2099 with respect to 1961 to 1990, for different climate zones and averaged globally. Only land grid cells with $T_{eq} > 0$ °C were considered for the averaging

Fig. 3 Annual mean predicted ΔT_{eq} in 2070 to 2099 with respect to 1961 to 1990 based on six AOGCMs. Only grid cells and months with $T_{eq} > 0$ °C were considered. Cells containing less than 0.1% wetlands in the GLWD with Köppen-Geiger classification polar (marked in blue) or dry (marked with black dots) were excluded from the calculations of global and zonal averages

Fig. 4 Average lake surface heat fluxes H_i (individual components i of H_{tot}) and their anomalies (absolute and relative) in 2070 to 2099 with respect to 1961 to 1990, calculated from the output of six AOGCMs for five different climate zones and globally. The number of grid cells and average number of months considered for each model and climate zone are specified in Table S1 and Table 1, respectively. Bars indicate multi-model means, symbols values for individual models

Fig. 5 Effects of the anomalies (2070 to 2099 with respect to 1961 to 1990) of single meteorological forcing variables on ΔT_{eq}, calculated from the output of six AOGCMs for five different climate zones and globally. From top to bottom: average effect on ΔT_{eq}; root-mean squared (rms) effect on ΔT_{eq}, and correlation coefficients (r) between ΔT_{eq} and the anomalies of the forcing variables. The number of grid cells and average number of months considered for each model and climate zone are specified in Table S1 and Table 1, respectively. Bars indicate multi-model means, symbols values for individual models. Note that the contributions of variables other than T_{air} were stretched by factors of 10 (red panel; right axis) and 5 (blue panel; right axis) compared to those of T_{air}

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Figure1

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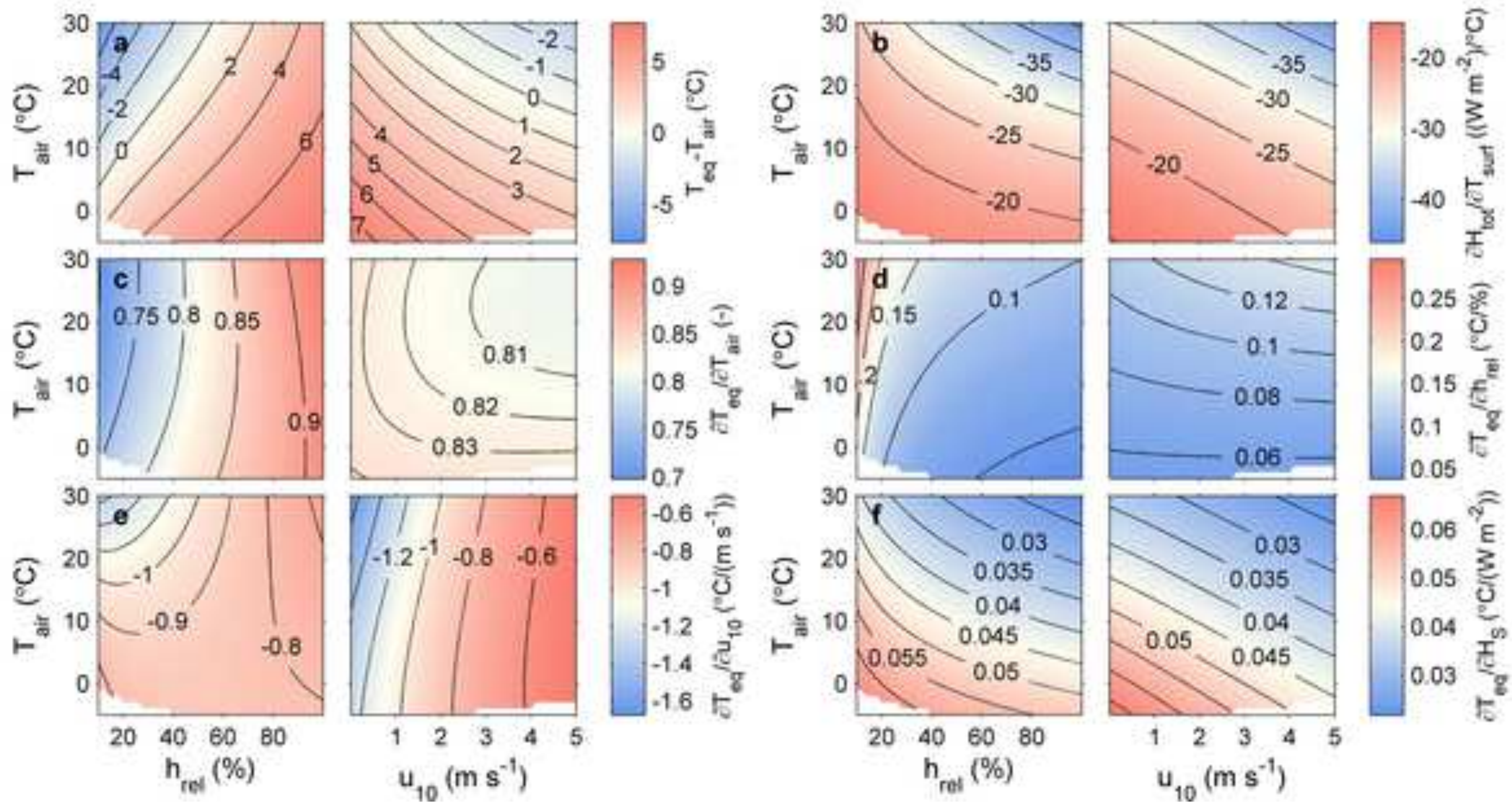

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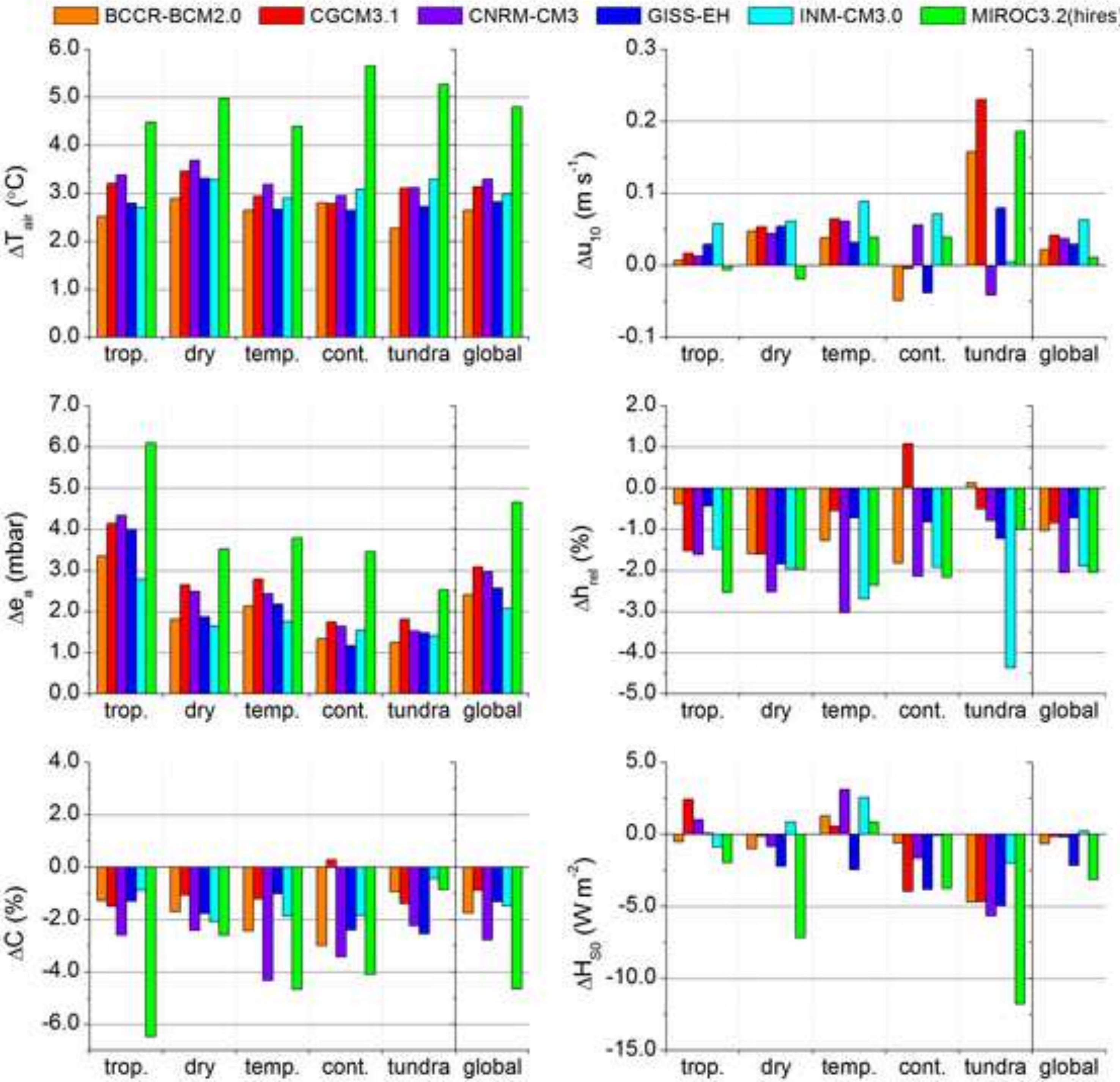

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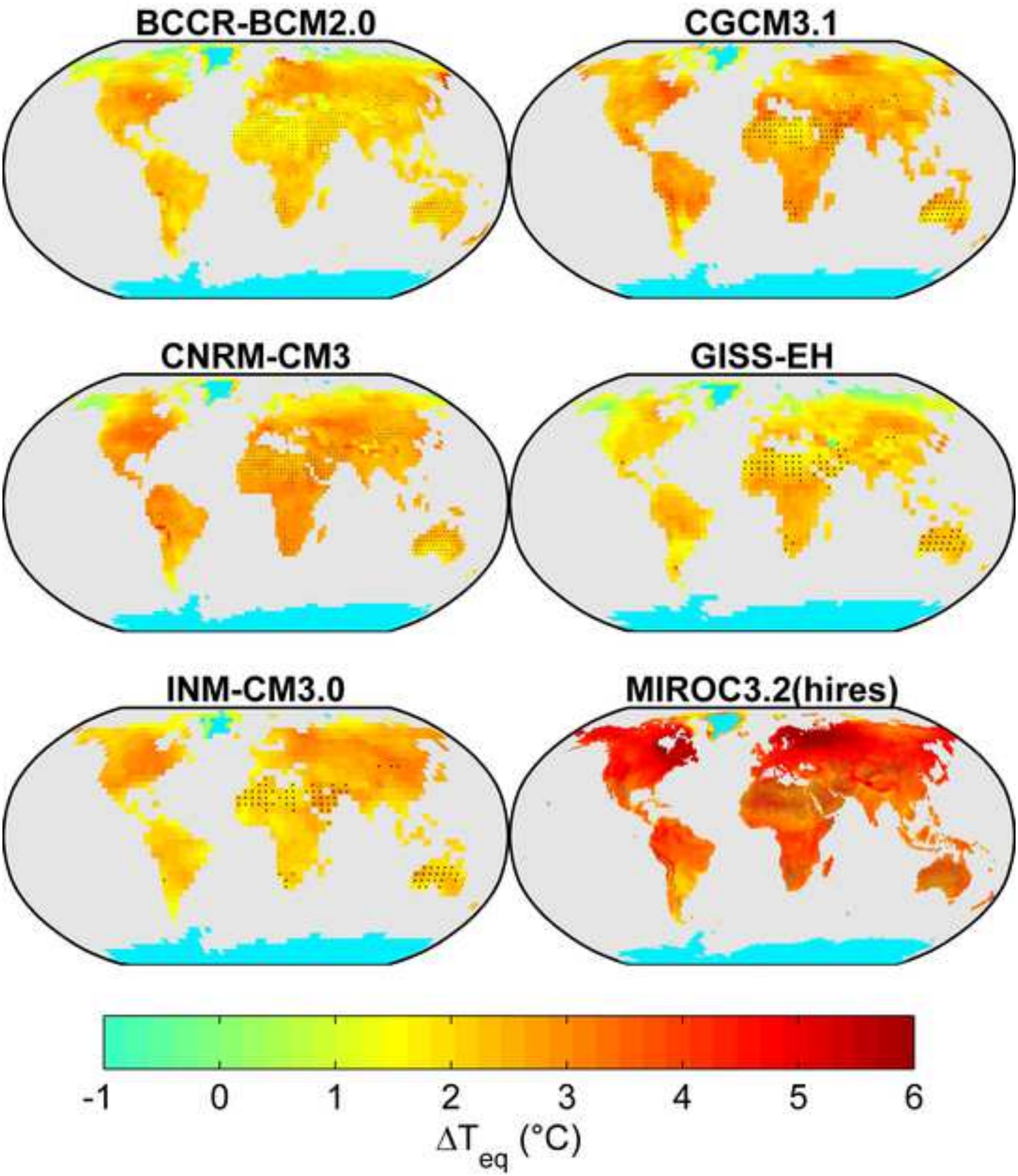

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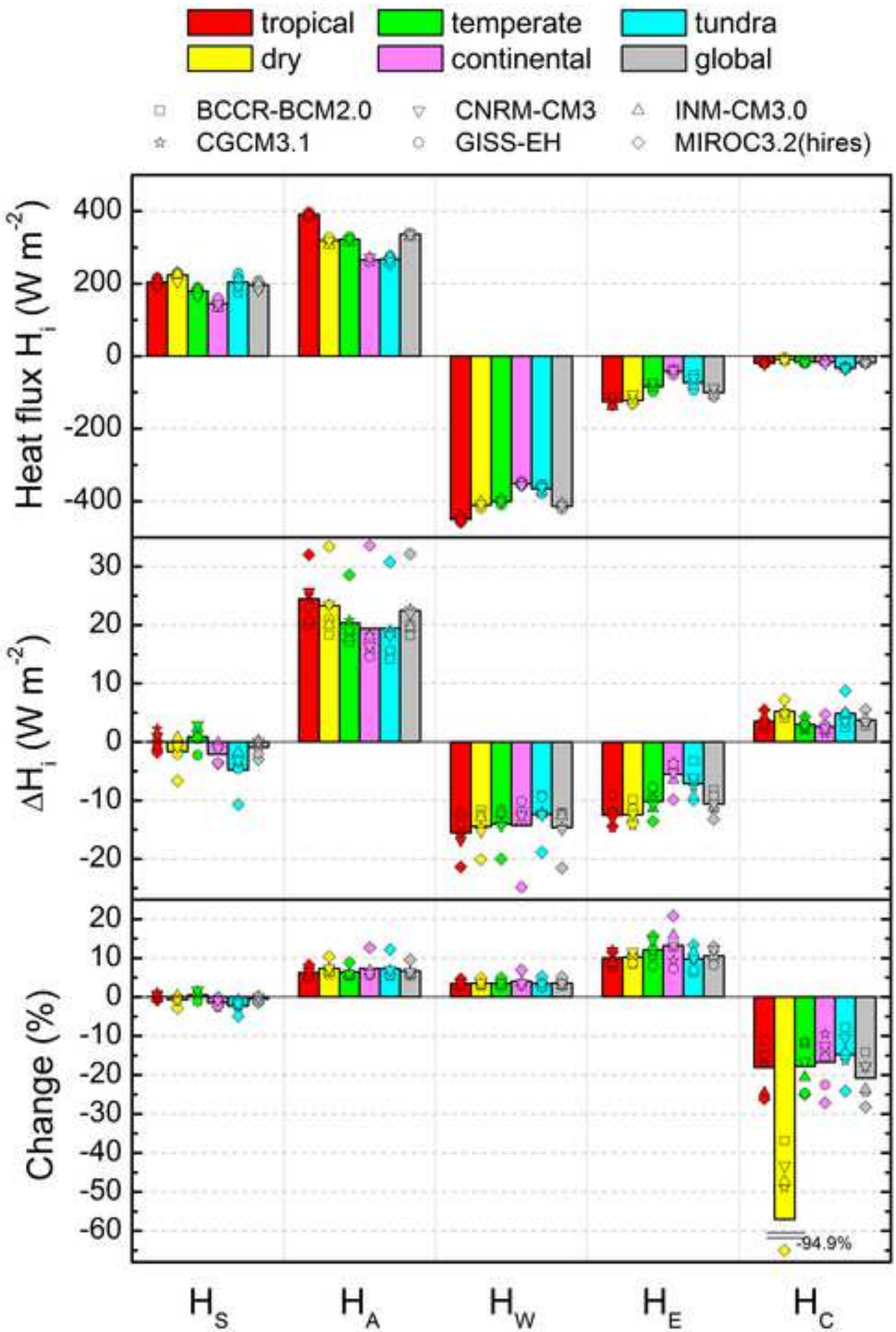

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Figure5

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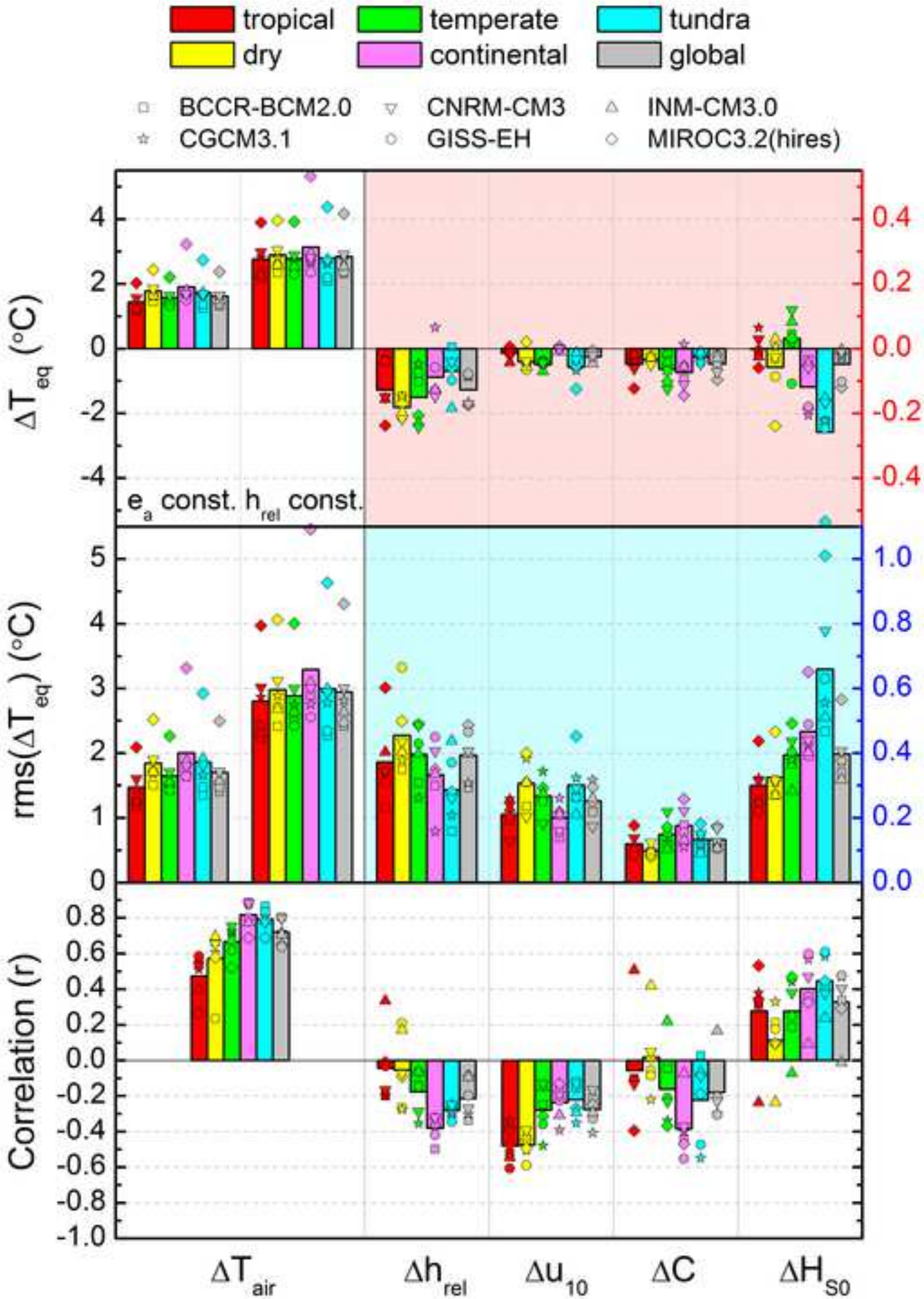

Supplementary Material

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