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A study of the large-scale climatic effects of a possible disappearance of high-latitude inland water surfaces during the 21st century

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This study evaluates the climatic impact of possible future changes in high-latitude inland water surface (IWS) area. We carried out a set of climate-change experiments with an atmospheric general circulation model in which different scenarios of future changes of IWS extent were prescribed. The simulations are based on the SRES-A1B greenhouse gas emission scenario and represent the transient climatic state at the end of the 21st century. Our results indicate that the impact of a reduction in IWS extent depends on the season considered: the total disappearance of IWS would lead to cooling during cold seasons and to warming in summer. In the annual mean, the cooling effect would be dominant. In an experiment in which the future change of prescribed IWS extent is prescribed as a function of the simulated changes of permafrost extent, we find that these changes are self-consistent in the sense that their effects on the simulated temperature and precipitation patterns would not be contradictory to the underlying scenario of changes in IWS extent. In this "best guess" simulation, the projected changes in IWS extent would reduce future near-surface warming over large parts of northern Eurasia by about 20% during the cold season, while the impact in North America and during summer is less clear. As a whole, the direct climatic impact of future IWS changes is likely to be moderate.

Introduction

Wetlands, lakes and ponds are a typical feature of northern ecosystems of Canada, Alaska and Siberia. These water bodies are under discussion as important carbon sinks or sources (e.g. Zimov *et al*. 1997, Oechel *et al*. 2000, Chapin *et al*. 2000, Walter *et al*. 2006, Walter *et al*. 2007), ecological niches (Gilg *et al*. 2000) and also play an important role in the local and regional climate and hydrology by governing the heat and water fluxes (e.g. Bonan 1995, Lofgren 1997, Krinner 2003, Gutowski *et al*. 2007). For example, Boike *et al*. (2008) found a 50% decrease of latent heat fluxes during dry summer at a Siberian wetland site due to a negative water balance and thus disappearing polygonal ponds. Despite this significance, there are few studies available on their formation and degradation processes associated with climate change. For example, using satellite images, Smith *et al*. (2005) stated a 6% declination of lake surface area (1973–1998) in Siberia despite a concomitant precipitation increase. The deficit averages a greater loss of lakes in the discontinuous permafrost zone (13%) as compared with the gain in the continuous permafrost zone (12%). Riordan *et al*. (2006) observed an 11% decrease of closed ponds in the Alaskan subarctic for the past 50 years concomitant with an increasing air temperature and thus hypothesized enhanced evapotranspiration. Yoshikawa and Hinzman (2003) reported shrinking lakes on the Seward Peninsula in Alaska due to the loss of water through the thawing of permafrost. Jorgenson *et al*. (2006) reported a large increase of permafrost degradation (thermokarst) and pit density from 1989 to 1998 especially in ice wedge areas of northern Alaska. A study from the Canadian High Arctic on Ellesmere Island (Smol and Douglas 2007) demonstrated that ponds that existed for several millennia are now drying. The authors attributed this to changes in the evapotranspiration/precipitation ratio. Thus, an understanding of potential processes causing these changes in water surface extent is still required for local and regional scales.

Transitions between drier steppe-like surfaces and peatland-lake dominated surfaces are not uncommon throughout the Earth's history. During the last maximum (21 ky BP) the arctic landscape was characterized by arctic desert and steppe vegetation with very few lakes (Overduin *et al*. 2007). With the beginning of warming and deglaciation (starting about 15 kyr ago), the number of thermokarst lakes and the areal extent of peatlands increased exponentially to the wet conditions of the Holocene optimum (Walter *et al*. 2007). Thermokarst lake formation and expansion waned in the later Holocene. The formation of thermokarst lakes and their methane emission is currently discussed as the main factor for high methane levels at about 12 kyr ago in ice cores from Greenland and Antarctica (Walter *et al*. 2007, Petrenko *et al*. 2008).

The climatic impact of inland water surfaces (henceforth IWS) depends on the season and on the regional climate itself. In a fairly general way, it can be stated that, due to their high effective thermal inertia, IWS will tend to respond more slowly than normal continental soil to the seasonal forcing, and thus induce a lag in the surface seasonal temperature cycle and reduce its amplitude (e.g. Bonan 1995, Krinner 2003). When IWS replace normal continental soil, there will almost always be a local tendency towards a modified partitioning between sensible and latent turbulent surface fluxes (Pitman 1991). However, the impact of IWS on the hydrological cycle is much more difficult to predict and depends very much on the climate itself and on regional circulation patterns. On one hand, the presence of IWS can cause an intensification of the hydrologic cycle in dry conditions (e.g., Hostetler *et al*. 1994, Krinner *et al*. 2003). On the other hand, the cooling effect of IWS during summer can induce a stabilization of the atmosphere downstream, and thus lead to reduced precipitation rates (e.g. Lofgren 1997). In paleoclimate studies, it has been shown that ice-dammed lakes can either lead to a more positive surface mass balance of the adjacent ice sheet via reduced marginal melt rates in summer due to their cooling effect (Krinner *et al*. 2004), or to more negative ice sheet surface mass balances because of precipitation reductions as a consequence of the more stable structure of the lower atmosphere (Hostetler *et al*. 2000). It is therefore not obvious that the climatic impact of IWS, evaluated for the present in several climate model studies (e.g. Pitman 1991, Bonan 1995, Lofgren 1997, Krinner 2003), will be similar to today's in a possibly warmer future climate, particularly if one keeps in mind that the future climate change is expected to be particularly strong in the high latitudes (IPCC 2007).

Here we carry out atmospheric general circulation model simulations for the end of the 20th century and for the end of the 21st century to assess the impact of a potential future reduction of IWS extent in the boreal continental areas (in this paper, the word "boreal" refers to all ice-free continental regions polewards of the southern limit of the taiga). The simulations are based on the SRES (IPCC Special Report on Emission Scenarios: Nakicenovic *et al*. 2000) A1B greenhouse gas emission scenario. We are particularly interested in impacts on the influence of IWS on the energy and water cycle. The main question is to what degree this potential reduction of IWS could affect the expected climate change during

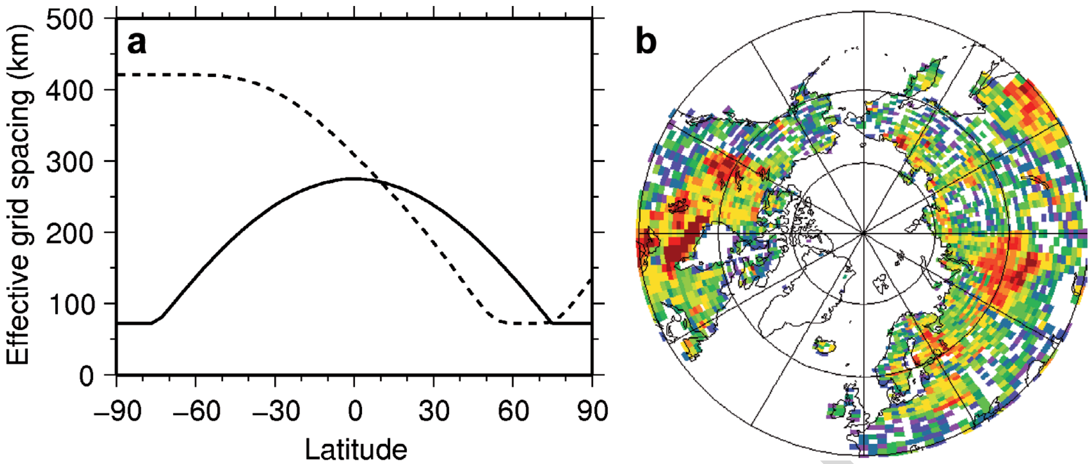

Fig. 1. (**a**) Effective meridional (dashed line) and zonal (full line) grid-point spacing (in km), taking into account the effect of zonal low-pass filtering near the poles which is necessary to ensure numerical stability. (**b**) Present-day inland water surface fraction (permanent lakes plus swamps, dimensionless) after Cogley (2003).White grid points indicate IWS below 0.003, ocean or ice cover.

the 21st century. In a first approach, we carry out sensitivity simulations in which we test the atmospheric response to a total disappearance of all IWS. This allows to compare our results for the 21st century with those of previous sensitivity tests carried out for the 20th century, and to obtain a strong signal which is easy to interpret. A second sensitivity test consists of a 50% reduction of future IWS extent. Finally, we use informed but rough estimates of the future extent of perennially frozen deep soil to produce a more realistic estimate of the future climatic impacts of a possible IWS reduction. These different sensitivity tests allow us to evaluate the dependency of the climate response on the severity of the future IWS reduction.

Methods

We used version 4 of the LMDZ (Laboratoire de Météorologie Dynamique-Zoom) atmospheric general circulation model with 144 × 108 × 19 (longitude × latitude × vertical) grid points. The model is described in detail by Hourdin *et al*. (2006). The irregular meridional grid-point spacing varies from 420 km at the South Pole to less than 75 km between 56°N and 75°N (Fig. 1a), while the convergence of the meridians near the poles automatically increases the zonal resolution. North of 75°N, zonal low-pass filtering is activated to ensure numerical stability, impeding the effective zonal grid spacing to drop below 73 km. Applications with similar grid distortions, allowing comparably high spatial resolution over the region of interest, were successfully carried out for high-latitude regions such as Antarctica and northern Eurasia (e.g. Krinner *et al*. 1997, 2004, 2006, 2007).

This version of LMDZ4 contains a lake column model described by Krinner (2003). The model takes into account molecular and wind-induced turbulent heat conduction below the lake surface, gradual absorption of sunlight, convection and water phase changes. The surface mass balance of IWS is calculated as precipitation minus evaporation minus runoff which occurs when the water column exceeds a prescribed maximum height. This maximum height is 20 m for lakes and 3 m for wetlands. We used the GGHYDRO (ver. 2.3) global hydrographic dataset (Cogley 2003) to prescribe global lake and wetland fractions in LMDZ4. GGHYDRO is a dataset covering the entire surface of the Earth and containing hydrographic data on the areal extent of different kinds of terrain, and on the distribution of mean terrestrial surface runoff. The IWS fraction on the model grid (Fig. 1b) exhibits the well-known pattern of areas of high inland water fraction such as the West Siberian Plain.

The model was run for the periods 1980–2000 and 2080–2100. For the period 1980 to 2000, the prescribed time-dependent boundary conditions are essentially observed sea-ice concentrations and sea-surface temperatures (Rayner *et al*. 2003, http://badc.nerc.ac.uk/data/hadisst/) and observed greenhouse gas (CO_2, CH_4, N_2O, CFCs), aerosol and ozone concentrations. For the period 2080 to 2100, the prescribed greenhouse gas and aerosol concentrations follow the IPCC SRES-A1B scenario (Nakicenovic *et al*. 2000). Among the different SRES scenarios of future greenhouse gas and aerosol emissions supposed to be of equal probability, SRES-A1B is a "middle of the road" estimate. The oceanic boundary conditions are obtained by using the anomalies of SST and sea-ice fraction simulated by the IPSL-CM4 (Institut Pierre-Simon Laplace coupled model ver. 4) atmosphere–ocean general circulation model under the same greenhouse gas forcing scenario (for details, *see* Krinner *et al*. 2008). Atmospheric ozone concentrations are prescribed following Meehl *et al*. (2006). The first year of each simulation is discarded as spinup. According to Simmonds (1985), this is an appropriate spin-up time for an atmosphere-only model. This leaves 20 years (periods 1981–2000 and 2081–2100) for analysis, coherent with previous applications of this model for polar climate studies (Krinner *et al*. 2007, 2008).

The setup of our simulations is summarized in Table 1. For each period (end of the 20th and end of the 21st century), we carried out one reference simulation with the present-day observed IWS extent (referred to as C20IWS and C21IWS, respectively), and one without IWS (referred to as C20 and C21, respectively). Moreover, for the end of the 21st century, we produced a simulation in which lake and wetland fractions were divided by two (C21IWS/2).

A uniform 50% or 100% reduction of IWS extent between now and the end of this century is, of course, a highly idealized scenario. Moreover, a complete IWS disappearance is rather improbable. An assessment of the influence of glacial history, permafrost, topography and peatland on the prevalence of lakes in the Arctic by Smith *et al*. (2007) has shown that spatial lake fraction statistics "are surprisingly similar across continuous, discontinuous and sporadic permafrost zones", and decrease sharply in permafrost-free areas. These authors suggest that a "permafrost-free" Arctic would see its IWS area reduced by about 40%. It is not clear how soon lake drainage occurs after a transition from a permafrost to a non-permafrost environment, although some observations (Smith *et al*. 2005) suggest that this may happen rapidly. It is not clear either whether permafrost-induced lakes require a critical minimum depth of the permafrost base, a maximum depth of the permafrost table or minimum ground ice content. The retreat of permafrost will of course occur slowly and, at least under a SRES-A1B-like scenario, the permafrost extent at the end of the 21st century will clearly be out of equilibrium with the evolving surface climate. But the annual mean temperature at a reasonable depth in the

Table 1. Overview of the simulations described in this paper.

Simulation	Period	Prescribed inland water surfaces	Note
C20IWS	1981–2000	Present-day observed IWS extent (after Cogley 2003)	Reference simulation for this period
C20	1981–2000	No IWS	
C21IWS	2081–2100	Present-day observed IWS extent (after Cogley 2003)	Reference simulation for this period
C21	2081–2100	No IWS	
C21IWS21	2081–2100	IWS reduced by 40% south of "equilibrium permafrost limit", increased by 10% in remaining "equilibrium permafrost" regions	"Best guess" simulation
C21IWS/2	2081–2100	IWS reduced by 50% as compared with present-day observed extent	Not shown

soil (such that it is approximately in equilibrium with the surface climate) can give an idea which permafrost extent would be consistent with the simulated surface climate, and it may serve as a basis for a rough estimate of large-scale changes of high latitude IWS extent during this century. Furthermore, as results by Smith *et al*. (2005, 2007) suggest, a temporary increase of IWS is possible where continuous permafrost remains in spite of the warming. These considerations led us to carry out a "best estimate" simulation (called C21IWS21) for 2081–2100 in which IWS fractions are modified in accordance to the simulated future climate change and the analysis of the dependence of IWS extent on climate as suggested by Smith *et al*. (2007). The region where the present-day annual mean temperature at the 50-cm depth ($T_{soil,50cm}$) is below 0 °C corresponds reasonably well to observed permafrost limits as given by Zhang *et al*. (2008). Therefore, this $T_{soil,50cm}$ annual mean 0 °C isotherm is referred to as the "equilibrium permafrost limit" in the following. IWS extent is reduced by 40% in regions which are north of the equilibrium permafrost limit during the period 1981–2000 (simulation C20IWS) and south of that limit at the end of the 21st century (simulation C21IWS) (green and yellow in Fig. 2). The regions remaining north of the equilibrium permafrost limit in 2081–2100 (red and brown in Fig. 2) can similarly be identified as those where an increase of IWS by about 10% might be expected according to the "thought exercise" by Smith *et al*. (2007). Consequently, the prescribed IWS fraction was increased by 10% in these areas. Except for the IWS extent, this simulation C21IWS21 is identical to the simulations C21IWS, C21IWS/2 and C21.

As the principal aim of this paper is to assess the climatic effect of a potential future decrease of the high-latitude IWS extent, and as the simulations including IWS are those in which the present surface type distribution is represented more realistically, we will present our results systematically with the "observed present inland water surfaces" case as reference simulations. In other words, our reference simulations for 1981–2000 and 2081–2100 are C20IWS and C21IWS, respectively.

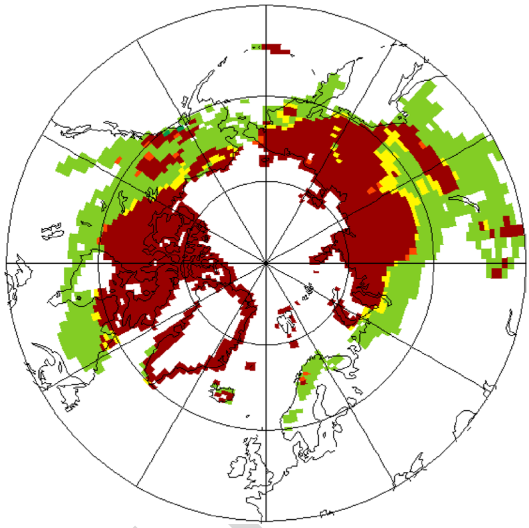

Fig. 2. Continental, ice-free areas with simulated annual mean soil temperature at 50 cm depth below 0 °C for different simulations. Green: C20IWS only; yellow: C20IWS and C21IWS21; red: C20IWS and C21IWS; brown: C20IWS, C21IWS and C21IWS21. All coloured pixels approximately correspond to the present-day simulated equilibrium permafrost extent, while pixels in red, yellow and brown indicate equilibrium permafrost in at least one of the 21st century simulations.

Results

Climate sensitivity to total IWS disappearance

If all inland water surfaces (IWS) disappeared today, warming in summer would occur (Fig. 3c). The reason is the missing large heat capacity of the IWS, which take more time to respond to seasonal forcing. For the same reason, cooling occurs in autumn (Fig. 3d). It partly extends into winter (Fig. 3a). Annual mean precipitation changes induced by IWS disappearance at the end of the 20th century (not shown) are mostly weaker than ±10%. The impact of a total disappearance of IWS on the simulated surface air temperature in the 21st century is shown in Fig. 4. Similar to the situation for the present (Fig. 3), the disappearance of IWS induces cooling during autumn and winter (Fig. 4a and d), and warming during summer (Fig. 4c). Despite justified criticism (e.g. Nicholls 2001), Student's

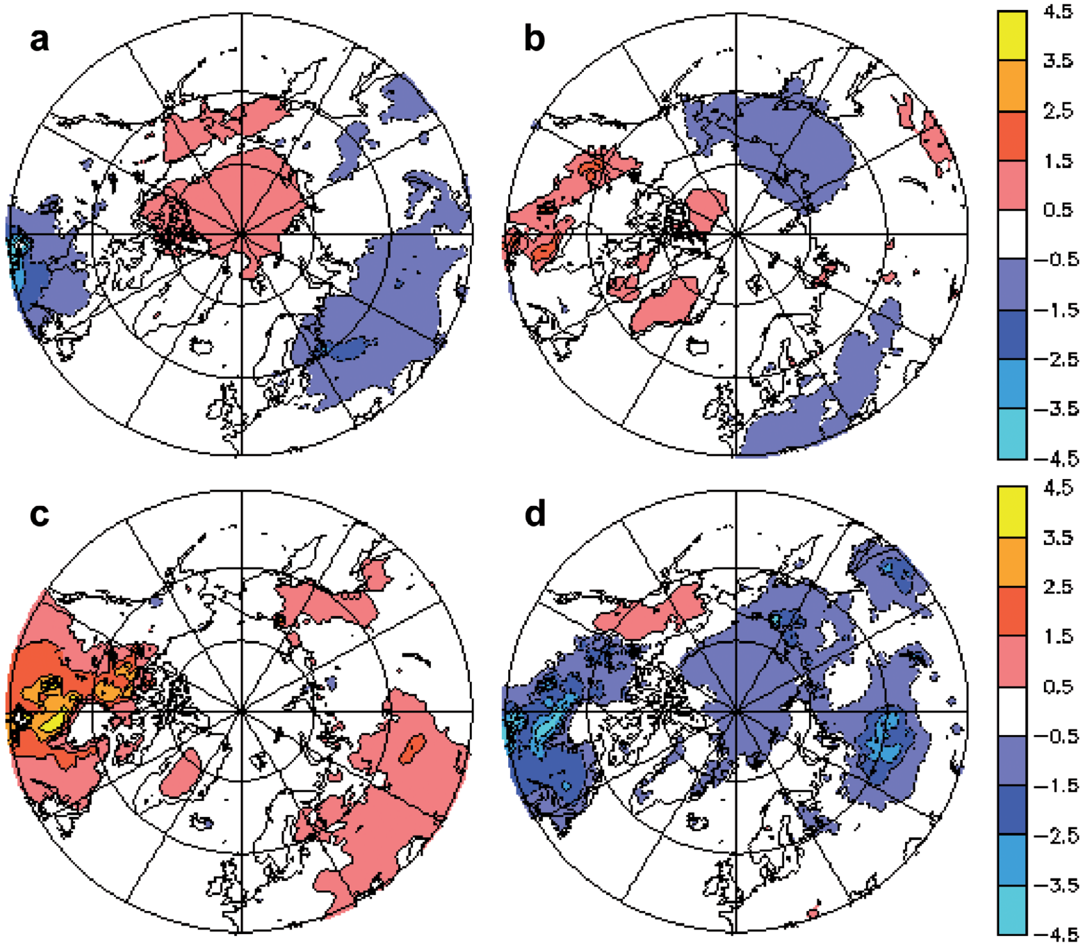

Fig. 3. Sensitivity of the simulated 1981–2000 surface air temperature (°C) to the total disappearance of inland water surfaces (IWS): $T_{2m,C20} - T_{2m,C20IWS}$. (**a**) winter (December–January–February), (**b**) spring (March–April–May), (**c**) summer (June–July–August), (**d**) autumn (September–October–November).

t-test is frequently used in atmospheric sciences to assess the likelihood of the difference between two simulations to be merely due to chance (e.g. von Storch and Zwiers 1999). A two-sided *t*-test of the temperature change signal indicates statistical significance (less than 10% probability of the signal to be due to chance) over large parts of Siberia and North America in summer and autumn, and more locally in winter. The weak spring temperature change is not significant at this 10% level.

The climatic effect of a potential future decrease of IWS has to be put into perspective with the amplitude of the expected climate change itself. On an annual mean basis (Fig. 5), the total disappearance of IWS would result in a weakening of the future warming over the largest parts of the boreal continents. The effect would be in its maximum (future warming reduced by more than 20%) over the West Siberian Plain, Karelia, Lake Baikal and western China. The largest effect in North America occurs in Québec, where the annual mean temperature increase over the 21st century is reduced by more than 15%. The opposite effect (reinforcing of the future warming) occurs over western Europe. However the absolute amplitude of this signal is weak, because the simulated annual mean climate change in that region is relatively small as a consequence of a strong projected reduction of the meridional overturning in the North Atlantic.

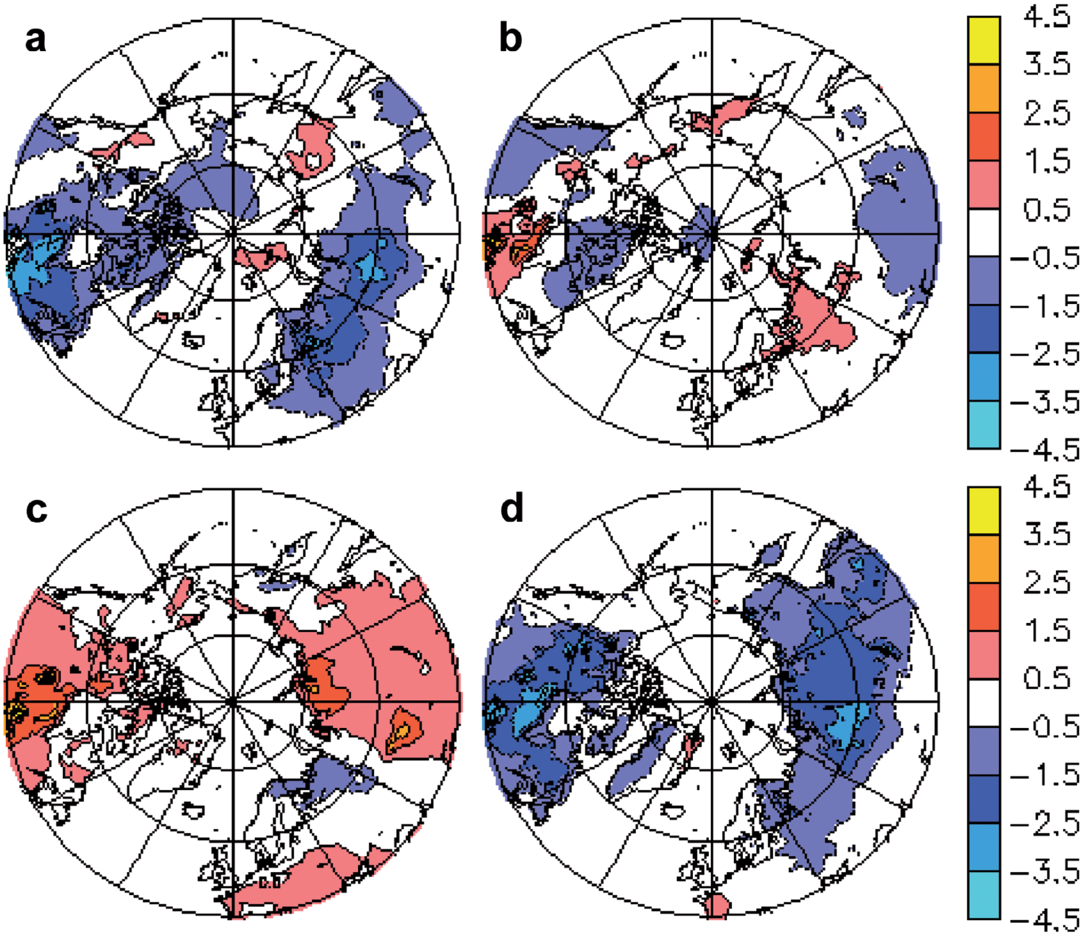

Fig. 4. Sensitivity of the simulated 2081–2100 surface air temperature (°C) to the total disappearance of inland water surfaces (IWS): $T_{2m,C21} - T_{2m,C21IWS}$. (**a**) winter (December–January–February), (**b**) spring (March–April–May), (**c**) summer (June–July–August), (**d**) autumn (September–October–November).

On a seasonal basis, the total disappearance of IWS would lead to a strong reduction of the autumn and winter warming (Figs. 6a, d and 7a, d) reaching more than 50% over the West Siberian Plain and south of the Hudson Bay in autumn. In turn, the absence of the cooling effect of the IWS in summer would lead to an increase of up to 50% of the future surface air temperature increase (Figs. 6c and 7c). Because future warming will be stronger in autumn and winter than in summer (Fig. 6), the cold season reduction of future warming dominates the annual mean signal (Fig. 5).

The simulated annual-mean boreal precipitation at the end of the 21st century is almost nowhere affected by the total disappearance of IWS and therefore not shown. Similarly, the circulation changes as illustrated by the 500-hPa height changes are weak. In terms of surface pressure, they correspond to changes of the order of 2 hPa and therefore not shown either.

Linearity of the climatic effect of IWS disappearance

How does the climatic impact of IWS vary with their extent? The surface air temperature change induced by the disappearance of 50% of all inland water surfaces (experiment C21IWS/2, not shown) is roughly half the signal we obtain with a total disappearance (experiment C21,

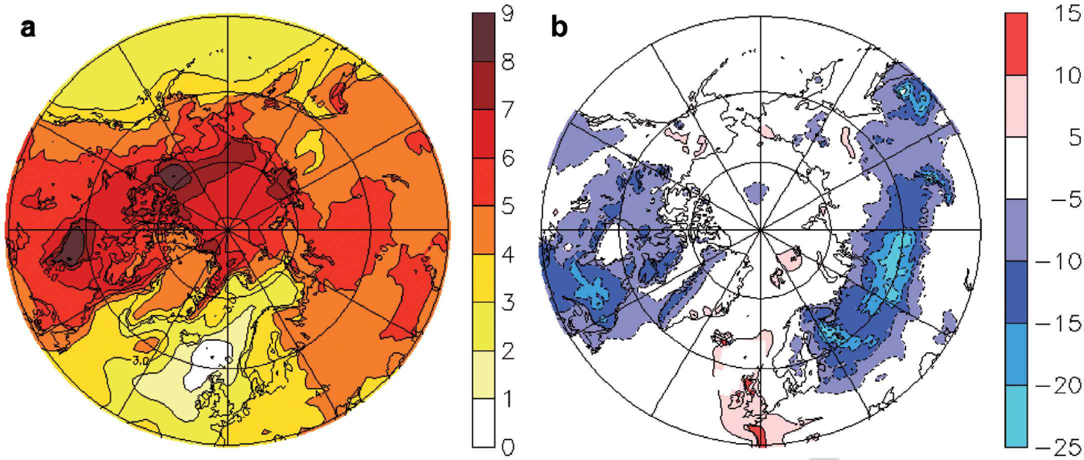

Fig. 5. (**a**) Annual mean surface air temperature change over the 21st century (from 1981–2000 to 2081–2100): $T_{2m,C21IWS} - T_{2m,C21}$ (°C). (**b**) Relative sensitivity (%) of the simulated annual mean surface air temperature change to the disappearance of inland water surfaces: $(T_{2m,C21} - T_{2m,C21IWS})/(T_{2m,C21IWS} - T_{2m,C20IWS})$.

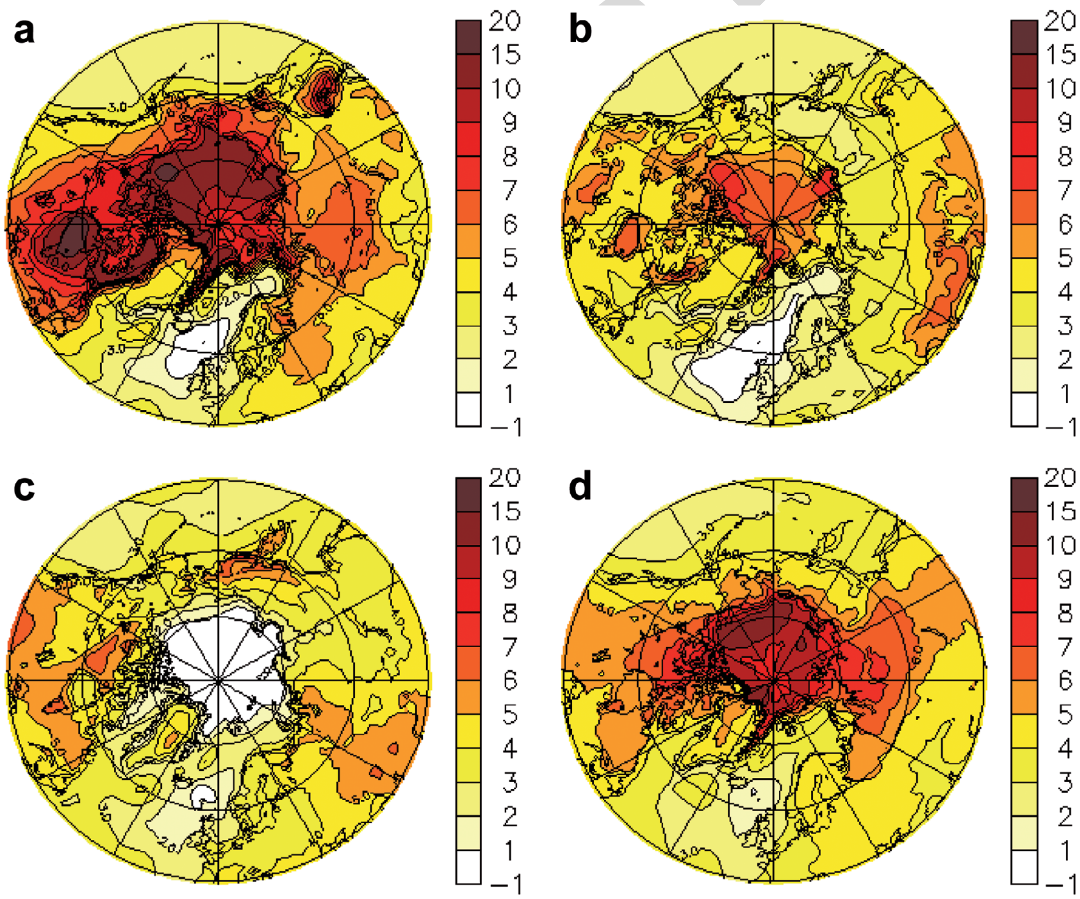

Fig. 6. Simulated seasonal surface air temperature change over the 21st century (from 1981–2000 to 2081–2100): $T_{2m,C21IWS} - T_{2m,C20IWS}$ (°C). (**a**) winter (December–January–February), (**b**) spring (March–April–May), (**c**) summer (June–July–August), (**d**) autumn (September–October–November).

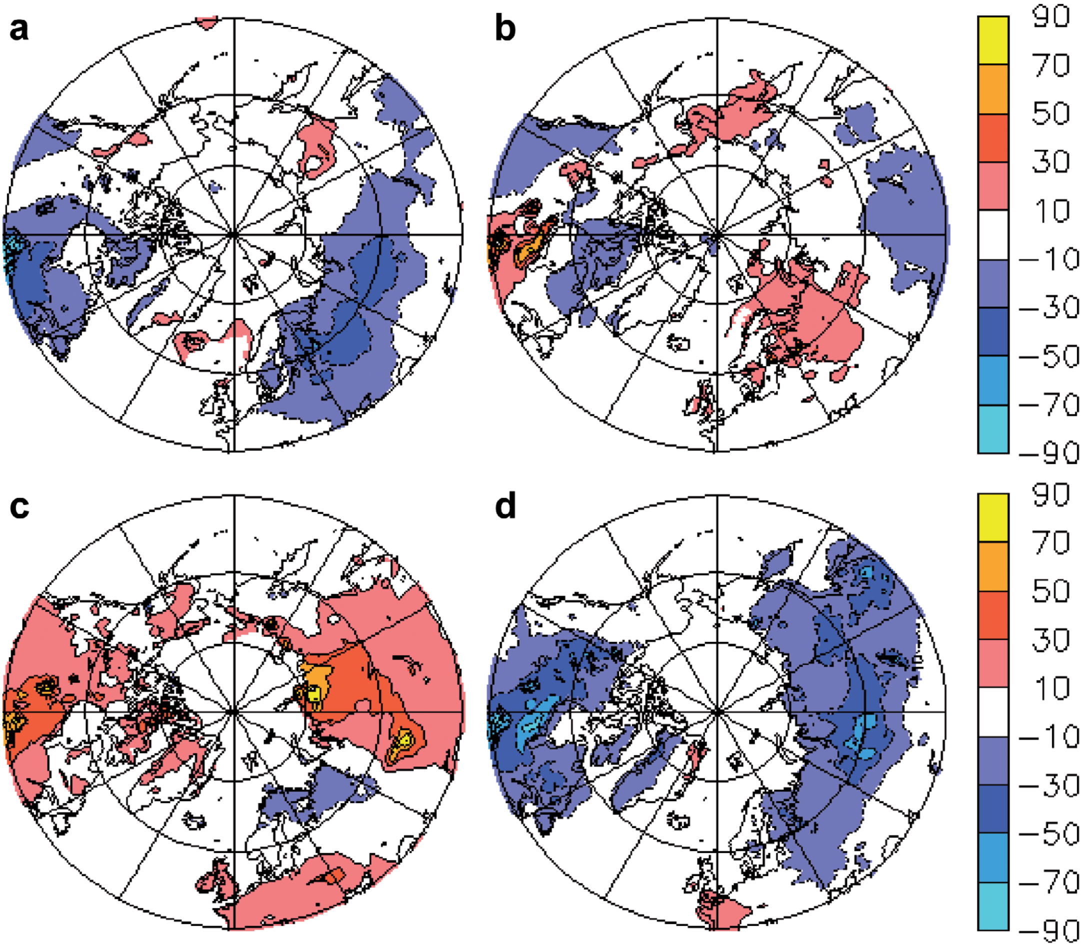

Fig. 7. Relative sensitivity (%) of the simulated seasonal surface air temperature change (from 1981–2000 to 2081–2100) to the disappearance of inland water surfaces: $(T_{2m,C21} - T_{2m,C21IWS})/(T_{2m,C21IWS} - T_{2m,C20IWS})$. (**a**) winter (December–January–February), (**b**) spring (March–April–May), (**c**) summer (June–July–August), (**d**) autumn (September–October–November).

Fig. 4). The climatic effect of IWS at the end of the 21st century therefore does appear to depend in a roughly linear manner on their extent.

With this in mind, and reminding the geographical patterns of the prescribed IWS extent changes (Fig. 2), it is fairly straightforward to interpret the temperature change signal obtained in the "best estimate" IWS extent experiment (C21IWS21: 40% IWS reduction prescribed in regions where the "equilibrium permafrost" disappears at the end of the 21st century, together with a 10% increase in the remaining "equilibrium permafrost" areas). The climatic impact of these moderate changes of IWS extent is rather weak (Fig. 8). The strongest signal appears in the northern Ural/West Siberian Plain region, where the prescribed reduced IWS leads to a clear and statistically significant cooling. Similarly, there is slight autumn cooling over central Siberia, linked to the reduced IWS and consistent with the results of experiment C21 (total disappearance of all IWS). The almost total absence of any signal south of the Hudson Bay is expected because no change of IWS is prescribed in this region in C21IWS21 (Fig. 2). In general, the statistical significance of the simulated temperature change signal is low, simply because of the weakness of the signal. However, because the signal is clearly consistent with that obtained when all IWS are removed (simulation C21),

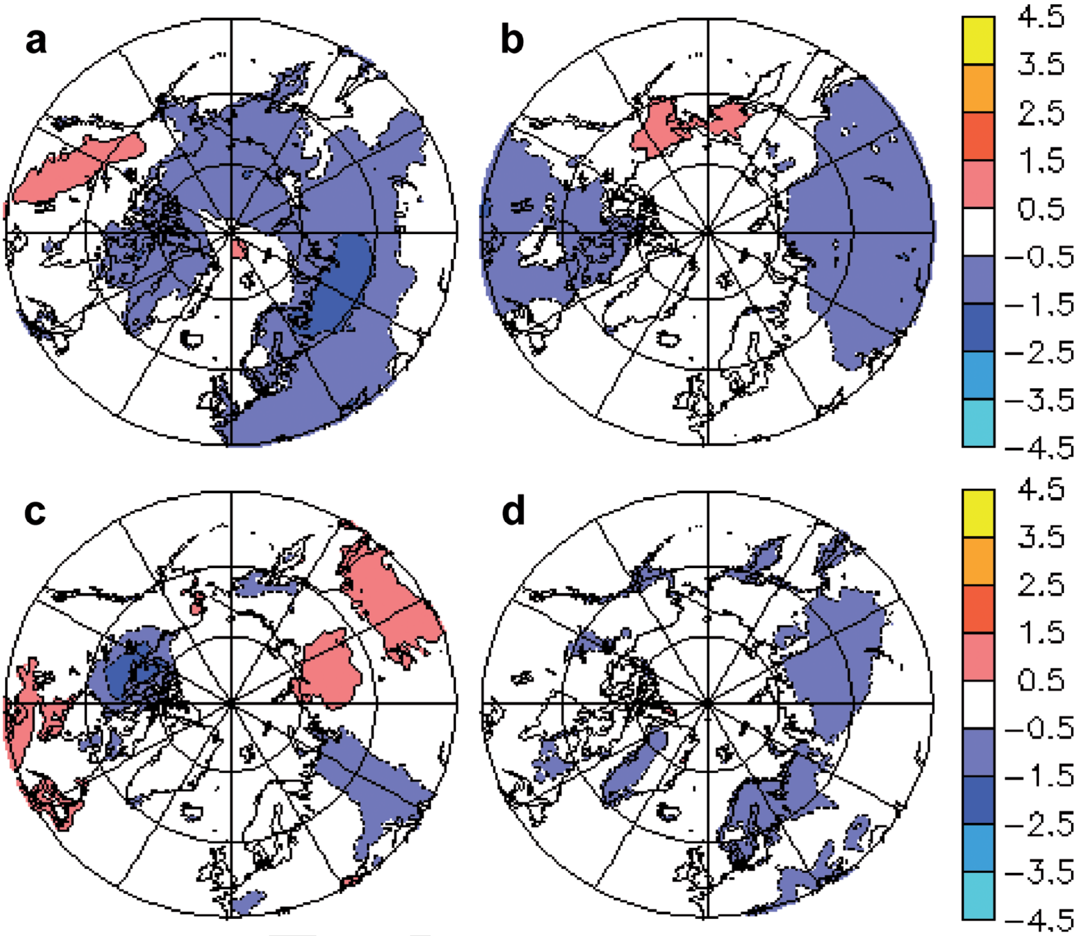

Fig. 8. Sensitivity of the simulated 2081–2100 surface air temperature (°C) to "best estimate" changes of inland water surfaces: $T_{2m,C21IWS21} - T_{2m,C21IWS}$. (**a**) winter (December–January–February), (**b**) spring (March–April–May), (**c**) summer (June–July–August), (**d**) autumn (September–October–November).

its physical significance (Nicholls 2001) is less questionable.

As compared with the anthropogenic climate change signal in absence of any changes of IWS (that is, in our case, the difference between C21IWS and C20IWS), the impact of the "best estimate" reductions of the inland water extent indeed appears fairly substantial (Fig. 9). As a consequence of the reduced thermal inertia of the surface, the future winter and spring warming is reduced by more than 15% (up to 40%) over a large part of Eurasia. A similar but weaker signal appears in autumn. Over North America the signal is not so clear. Similarly, in summer, the signal is spatially rather noisy, casting doubt about its physical significance.

Discussion

Comparison with previous work

The sensitivity of the present-day (boreal) climate to the presence of IWS has been evaluated in several previous studies (e.g. Pitman 1991, Bonan 1995, Lofgren 1997, Krinner 2003). The work by Krinner (2003) was carried out with an older version of the climate model used here and the results suggested a very strong cooling effect of IWS in summer, which was not reported by the earlier studies. It is not reproduced here, either. The sensitivity of the boreal climate obtained in the present study with the present version of LMDZ is indeed more similar to the response

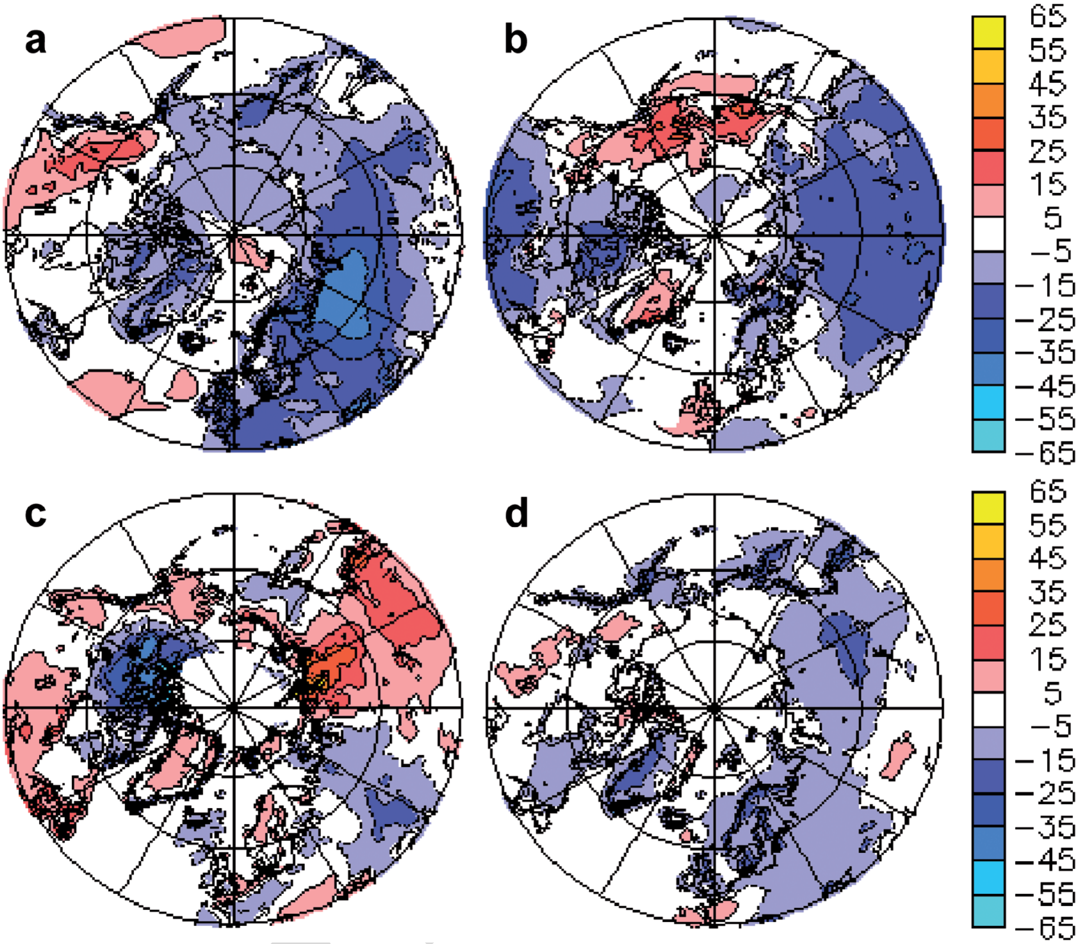

Fig. 9. Relative sensitivity of the simulated 2081–2100 minus 1981–2000 surface air temperature change (°C) to "best estimate" changes of inland water surfaces: $(T_{2m,C21IWS21} - T_{2m,C21IWS})/((T_{2m,C21IWS} - T_{2m,C20IWS}))$. (**a**) winter (December–January–February), (**b**) spring (March–April–May), (**c**) summer (June–July–August), (**d**) autumn (September–October–November).

obtained by Pitman (1991) and Bonan (1995) than to the response obtained with the older version of the same model. The strong summer cooling reported by Krinner (2003) can in fact be traced back to a dry bias of the summer soil in the older version of LMDZ. This dry summer bias led to a considerable overestimate of boreal summer temperatures in the simulations without IWS, which was corrected in the simulations with IWS. The more recent version of LMDZ used here does not exhibit such a bias. The simulated boreal summer temperatures are in reasonable agreement with observations even in the version without IWS, leading to a damped thermal response of the climate model to the inclusion of IWS.

Importance of the underlying IWS map

There are several global-scale compilations of wetland and lake extents (e.g. Cogley 2003, Lehner and Döll 2004 and references therein, Prigent *et al*. 2007). For practical reasons, we used the Cogley (2003) GGHYDRO compilation in this work. A recent study by Grosse *et al*. (2008) shows a major discrepancy between these larger scale datasets and distribution and coverage of lakes obtained from high resolution remote sensing. At three study sites in eastern Siberia, they found that the extent of water bodies was underestimated by a factor of two to seven. While these differences between the vari-

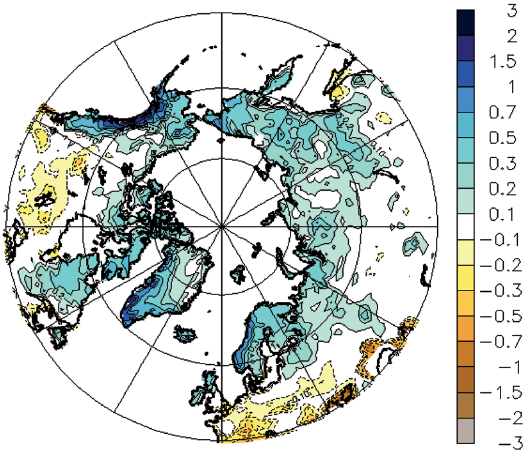

Fig. 10. Difference of annual mean precipitation minus evaporation ($P – E$, mm day^{-1}) between simulations C21IWS21 and C20IWS.

ous databases exist on small scales, the regional-scale characteristics of the distribution of IWS are similar. That is, on spatial scales represented by the climate model used here, the simulated patterns of the atmospheric response to a change in IWS extent will not be critically dependent on the underlying present-day map of IWS.

Coherence of the prescribed IWS changes with the simulated climate changes

Following Smith *et al*. (2007), we based our scenario of future changes in the IWS extent only on soil temperature changes. In reality, precipitation and evapo-transpiration contribute to determine the extent of IWS (Prigent *et al*. 2007), and will certainly continue to do so in the future. It is therefore of interest to assess the consistency of the simulated future precipitation and evaporation changes with our scenario of changes of IWS. For this to be done properly, we would have to use a detailed hydrological model, and this would be clearly beyond the scope of this work. Therefore, we base our assessment on precipitation and evaporation changes only. The simulated annual mean precipitation minus evaporation ($P – E$) changes over the 21st century (Fig. 10) indicate an increase of $P – E$ in the high northern latitudes, and a tendency to a more negative surface water balance further south. The $P – E$ changes are dominated by the precipitation changes. The evaporation changes exhibit the same spatial structure, but they are somewhat weaker. In other words, in regions where we prescribed a temperature-induced 10% increase of IWS extent in C21IWS21 because of the continuing prevalence of permafrost conditions (Fig. 2), the model tends to simulate a $P – E$ increase. In regions where the simulated annual mean soil temperature changes led us to prescribe a 40% reduction of IWS extent in C21IWS21, the annual mean $P – E$ changes tend to be more neutral. This means that our simulated $P – E$ changes do not appear in gross contradiction to the imposed scenario of changes of IWS extent.

Our results have shown that the disappearance of IWS would lead to significant, seasonally varying surface temperature changes. The scenario of IWS extent changes used in simulation C21IWS21 is based on future soil temperature changes which were simulated supposing no change of IWS extent (simulation C21IWS). Therefore it is interesting to ask whether the thermal impact of a possible future reduction of IWS would alter the future soil temperature distribution sufficiently to induce major discrepancies between the future soil temperatures simulated in C21IWS21 and those simulated in C21IWS. In our experimental setup we used the annual mean soil temperature at 50 cm depth ($T_{\mathrm{soil,50cm}}$) as a proxy for the frost-induced changes of IWS extent. Concerning the $T_{\mathrm{soil,50cm}}$ 0 °C isotherm, the main differences between C21IWS and C21IWS21 are a consequence of the cooling induced by the reduction of IWS extent, most pronounced in parts of central Siberia (yellow in Fig. 2). In these fairly small regions, $T_{\mathrm{soil,50cm}}$ is below 0 °C in C21IWS21, but not in C21IWS. An iterative approach would require to carry out an additional simulation in which the prescribed IWS extent be increased by 10% (instead of being reduced by 40%) in regions which are yellow in Fig. 2 (and vice versa for the isolated regions which are red). However, such an iterative approach does not appear warranted for several reasons. First, the approach we used to prescribe future changes of IWS extent is a very rough one. Second, the absolute impact of changes of IWS extent is not extremely strong. Third, the regions concerned (yellow and red in Fig. 2) are not very large.

Indirect effects

While most studies hypothesize an increase in northern methane emission as a result of the air temperature and precipitation increase, the expected reduction of IWS will probably, at least partially, compensate for this effect. However, our study shows that the expected IWS reduction would reinforce the future summer warming. Because wetland methane emission usually peaks in summer, this means that methane emission intensity from the remaining wetlands (that is, emission per unit area of water surface) would probably be somewhat stronger than in the case without IWS reduction.

Conclusion

The aim of this work was to contribute to the ongoing discussion about possible future changes in the IWS extent in the high northern latitude by evaluating the climatic impact of such changes at the end of the 21st century (supposing greenhouse gas concentrations according to the SRES-A1B emission scenario). In an extreme (and improbable) scenario supposing the total disappearance of all IWS, our GCM simulates surface cooling locally in excess of 3 °C during winter and spring both in Eurasia and North America. Our results suggest that the climatic impact of a future reduction of the IWS extent is roughly proportional to the severity of the prescribed scenario of IWS extent changes. The impact of a potential reduction of the IWS extent on precipitation appears negligible, while the induced near-surface temperature changes are significant in large regions, and vary with season. In the annual mean, the total disappearance of IWS would reduce the expected future warming by about 10% (and locally more than 20%) over large parts of northern Eurasia and North America.

We carried out a "best guess" simulation in which we prescribed a fractional reduction of high latitude IWS extent as a function of the simulated reduction of the area of annual-mean subsurface temperatures. We insist that for several reasons detailed in the discussion, this "best guess" simulation should only be regarded as a very crude estimate of the potential climatic impact of changes of IWS extent. Taking the simulated impact of our "best guess" inland water reduction scenario at face value, we conclude that a reasonable future change of IWS would reduce future near-surface warming by up to 30% regionally and seasonally. However, in summer, future warming might be intensified due to reduced cooling induced by the decrease of IWS extent. The impact of these changes of IWS on the simulated surface water balance (precipitation minus evaporation) would not be in contradiction to the expected changes of IWS: in the most northerly regions, where our scenario would suggest an increase of IWS, the simulated annual mean $P - E$ increases. Further south, there are no systematic changes of $P - E$.

As a whole, this study suggests that the direct climatic impact of expected future IWS changes will be relatively moderate on large spatial scales. In particular, the effect on precipitation will be very weak, and circulation changes will be fairly weak. Near-surface temperature appears to be more sensitive. Future annual mean warming in the continental regions of the high northern latitudes might be reduced by about 10% by the effect of warming on the future extent of IWS. Seasonal characteristics of the future temperature changes can be more strongly affected, for example the summer surface temperature increase. This study makes no attempt to make the potential effects of these changes on the regional carbon fluxes, which could also influence the amplitude of the future climate change, not only locally, but globally (Walter *et al*. 2006). This study demonstrates, however, that the complex interactions can both amplify and counteract the predicted warming trend.

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