Arctic amplification (AA) is a major characteristic of observed global warming, yet the different mechanisms responsible for it and their quantification are still under investigation. In this study, the roles of different factors contributing to local surface warming are quantified using the radiative kernel method applied at the surface after 100 years of global warming under a representative concentration pathway 4.5 (RCP4.5) scenario simulated by 32 climate models from phase 5 of the Coupled Model Intercomparison Project. The warming factors and their seasonality for land and oceanic surfaces were investigated separately and for different domains within each surface type where mechanisms differ. Common factors contribute to both land and oceanic surface warming: tropospheric-mean atmospheric warming and greenhouse gas increases (mostly through water vapor feedback) for both tropical and Arctic regions, nonbarotropic warming and surface warming sensitivity effects (negative in the tropics, positive in the Arctic), and warming cloud feedback in the Arctic in winter. Some mechanisms differ between land and oceanic surfaces: sensible and latent heat flux in the tropics, albedo feedback peaking at different times of the year in the Arctic due to different mean latitudes, a very large summer energy uptake and winter release by the Arctic Ocean, and a large evaporation enhancement in winter over the Arctic Ocean, whereas the peak occurs in summer over the ice-free Arctic land. The oceanic anomalous energy uptake and release is further studied, suggesting the primary role of seasonal variation of oceanic mixed layer temperature changes.

1. Introduction

Arctic amplification is a well-known feature of the effect of increased atmospheric greenhouse gas (GHG) concentrations on climate. It is already observed under current global warming (e.g., Serreze and Barry 2011; Screen and Simmonds 2010), is found in reconstructions of past Earth climate history (e.g., Masson-Delmotte et al. 2006), and is simulated by state-of-the-art climate models for the future (e.g., Manabe and Wetherald 1975; Holland and Bitz 2003; Yoshimori et al. 2009, 2014a; Pithan and Mauritsen 2014).

Several mechanisms are able to explain, at least in part, this Arctic amplification, including the surface albedo feedback (e.g., Crook et al. 2011), the nonlinearity of the Stefan–Boltzmann law to background temperature (e.g., Ohmura 1984; Joshi et al. 2003), atmospheric heat transport (e.g., Solomon 2006; Cai 2005; Graversen...
et al. 2008; Alexeev et al. 2005), ocean heat transport (e.g., Holland and Bitz 2003; Mahlstein and Knutti 2011), water vapor and/or cloud feedbacks (e.g., Graversen and Wang 2009; Vavrus 2004), and lapse rate effect (e.g., Manabe and Wetherald 1975; Yoshimori et al. 2009), among others (e.g., O’ishi and Abe-Ouchi 2009; Shindell and Faluvegi 2009; Crook et al. 2011). Nevertheless, understanding the mechanisms and their relative importance are still investigated (e.g., Yoshimori et al. 2014b; Pithan and Mauritsen 2014). One of the main difficulties is that the evaluation of several mechanisms is not possible from direct model output, especially concerning the roles of imposed GHG concentration, air temperature, water vapor, or cloud changes on radiation.

In this study, we use the radiative kernel method (Soden et al. 2008), applied at the surface, to estimate the relative influence of surface albedo, tropospheric-mean barotropic and nonbarotropic warming, specific humidity, and cloud changes on surface radiations in 32 models from phase 5 of the Coupled Model Intercomparison Project (CMIP5) after 100 years of moderate warming under the representative concentration pathway 4.5 (RCP4.5) scenario (sections 2a and 2c). We also consider other nonradiative terms included in the changes in energy flux at the surface. We convert these contributions to surface energy changes into contributions to surface warming, also allowing us to introduce the role of the local surface warming sensitivity associated with the nonlinearity of the Stefan–Boltzmann law to background temperature (section 2d). Pithan and Mauritsen (2014) have partly used this technique at the surface for mean response of different CMIP5 models but in a general way (annual-mean, land, and oceanic surfaces altogether). Lu and Cai (2009a), Sejas et al. (2014), and Yoshimori et al. (2014a) have analyzed the seasonality of surface Arctic warming amplification but without explicitly separating land and oceanic surfaces. Lu and Cai’s (2009a) study consists of a multimodel analysis from phase3 of CMIP (CMIP3) but uses a less detailed decomposition of feedback mechanisms than allowed by the radiative kernel method. Sejas et al. (2014) and Yoshimori et al. (2014a) use the coupled atmosphere–surface climate feedback–response analysis method (CFRAM; Lu and Cai 2009b), which allows for a detailed decomposition of surface temperature changes (although different from ours), but in individual models. We hereby intend to study the details of the Arctic and tropical warming mechanisms and the associated Arctic amplification (AA), for each surface type (and even for different domains in the Arctic where mechanisms differ; section 2b), as well as their seasonality, by using the radiative kernel method applied to multimodel outputs.

2. Data and method

a. Data

The radiative kernels are calculated following the method given in Soden et al. (2008) (see section 2c for more details). They are derived from the Model for Interdisciplinary Research on Climate, version 4 (medium resolution) (MIROC4m), radiative model in its spectral T42 horizontal resolution (Yoshimori et al. 2014a). The kernels are first calculated on the model’s original 20 vertical sigma levels [details in Yoshimori et al. (2011)], then interpolated onto 17 pressure levels in order to perform the multimodel analysis. Contrary to Soden et al. (2008) or Yoshimori et al. (2011), the radiation fluxes in our case are considered at the surface rather than at the top of the atmosphere (TOA), as in Pithan and Mauritsen (2014). The pressure-interpolated version of the radiative kernels used in this study is available online (http://wwwoa.ees.hokudai.ac.jp/people/myoshimo/data/kernels/laineJC.html).

The models used in this analysis consist of CMIP5 models (Taylor et al. 2012), interpolated on the same T42 horizontal grid and the same 17 atmospheric pressure levels as the radiative kernels. We compare the climatology of the last 20 years of the twenty-first century following the default RCP4.5 scenario to the climatology of the last 20 years of the twentieth century of historical runs (Taylor et al. 2012). The input data for the kernel method consists of the monthly climatology of the multimrun ensemble mean of each model for each period (Table 1). The kernel method is then applied to each model, and output fields are then averaged among the 32 different models with equal weights to give the ensemble mean results. Some data needed for extra analyses were not available for all 32 models. This concerns the snow cover plotted in section 3c (21 models, Table 1) and data needed in the analysis performed in section 3e (11 models, Table 1).

b. Definitions of domains

In this analysis, spatial averages over land and oceanic surfaces are considered over grid cells with 100% of each surface type under the T42 resolution and weighted according to their areas. AA is defined as the ratio of temperature changes in the Arctic (latitudes greater than 60°N) relative to the tropics (between 30°S and 30°N). The overall oceanic domain north of 60°N is called the Arctic oceans in this study. It is further separated into the Greenland–Icelandic–Norwegian–Labrador (GINL) Seas (vertically striped region in Fig. 1) and the Arctic Ocean (horizontally striped region in Fig. 1). We also separate the land surfaces north of 60°N (domain referred to as Arctic land) into Greenland and the rest of the domain (ice-free Arctic land). These separations of
the Arctic oceanic and land areas were performed because mechanisms were found a posteriori to be very different in these separate regions. The way we define the GINL Seas region is very pragmatic, as we noticed that the annual-mean 4°C surface temperature change line over the oceans north of 60°N (solid white line in Fig. 1) separated the regions with clear separate warming mechanisms very well for all months.

c. Decomposition of local surface energy budget change

The radiative kernel method allows for a decomposition of the radiation budget into terms related to temperature, albedo, water vapor, and cloud changes for both shortwave (SW) and longwave (LW) radiation (Soden et al. 2008). Imposed GHG concentration changes are evaluated indirectly according to the proportion of CO2-equivalent changes relative to a doubling of CO2 in MIROC4m, for which a precise quantification of this effect has been performed [adjusted from stratospheric effects as in Yoshimori et al. (2009)]. In our case, after one century of integration following the RCP4.5 scenario of CMIP5, we estimate the radiative effect of imposed GHG changes relative to a doubling of CO2 to be about 0.74:

\[
\frac{\sum_{i=1}^{20} \log(CO_2^{eq_i}) - \sum_{j=1}^{20} \log(CO_2^{eq_j})}{\log(2)} R_{2\times CO_2} \sim 0.74 R_{2\times CO_2},
\]

(1)

where CO2^{eq_i} and CO2^{eq_j} correspond to the mean CO2-equivalent effect of GHG concentration for year i, ranging from 2080 to 2099 for the RCP4.5 scenario, and for year j, ranging from 1980 to 1999 for the historical scenario. The radiative effect of a doubling of CO2 (from preindustrial levels) in the MIROC4m radiative model is represented by \(R_{2\times CO_2}\). The nearly logarithmic radiative effect of GHG concentration is considered in Eq. (1). The

<table>
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direct effect of imposed GHG is small compared to other terms in these simulations (see section 3) so that the precise way to evaluate its effect is not decisive.

Given large biases for the term associated with the atmospheric warming when using interpolated pressure levels instead of model original vertical levels, a correction is applied to this term (see appendix A for details). The local corrected effect of atmospheric warming on surface downward LW radiation is then decomposed into a part associated with a vertically homogeneous warming of the troposphere equivalent to the local tropospheric-mean temperature change (referred to here as barotropic warming), and a part associated with the departure from the local tropospheric-mean value (referred to as nonbarotropic warming). It is different from the usual decomposition considering a vertically uniform temperature change and its departure relative to the surface temperature change (e.g., Soden et al. 2008). We think a decomposition relative to the tropospheric-mean temperature change is more meaningful for radiative considerations at the surface. Indeed, we expect the lapse rate effect to be positive in polar regions, at least during the coldest months, where and when atmosphere is on average very stable and the warming is trapped at the surface (e.g., Persson et al. 2002), hence leading to a very efficient surface warming (for a given value of vertical-mean tropospheric warming) through atmospheric LW emissions (e.g., Ohmura 2012). On the contrary, in the equatorial regions, the atmosphere is often unstable (although the mean temperature profile is kept close to a moist adiabatic lapse rate by deep convection mixing), and the warming spreads and amplifies with height through latent heat release. The warming effect of the atmosphere through LW emission toward the surface is therefore less efficient (for a given value of vertical-mean tropospheric warming), as the atmospheric warming is relatively weak at low levels. Considering deviations relative to surface temperature changes would lead to opposite signs of lapse rate effects in these regions for surface considerations and would therefore be misleading. To make clear that our decomposition of the atmospheric warming is different from previously established methodologies (e.g., Soden et al. 2008), we refer to nonbarotropic warming instead of “lapse rate effect.”

Albedo, water vapor, and cloud feedback derivations follow the methodology of Soden et al. (2008), especially concerning the masking correction of the cloud effects. The remaining difference between the actual radiation changes in a given model and the sum of the estimates for the role of imposed GHGs, temperature, albedo, water vapor, and cloud changes (referred to here as kernel residual) may include the following:

- the different sensitivities to the different factors of the given model compared to MIROC4m used for kernels calculation, which can partly result from differences in baseline climates;
- the nonlinearity effects of actual factor changes compared to the linearized kernel method;
- the interactions between the different factors; and
- the effects of aerosols and ozone concentration changes.

Apart from the radiative budget, the surface energy budget also includes sensible and latent heat fluxes and “other surface effects” that may include the following:

- seasonal and long-term (disequilibrium) variability of surface heat storage;
- horizontal and/or vertical heat advection (in the case of oceans) and diffusion;
- phase change between water and ice or snow; and
- artificial energy sink over ice sheets, when subject to melting, as these imposed ice sheets are fixed and cannot melt in the models, imposing the surface temperature not to exceed the freezing value: This artificial energy sink is expected to be equivalent to a melting of the ice sheet, although indirect effects like the lowering of the ice sheet, the presence of liquid water easier to evaporate than ice, and its eventual lubricating effect on the glaciers sliding (in any case not simulated by these models) are not taken into account. In case the ice sheet would have totally melted away locally if it had been correctly simulated, the surface energy budget would be completely
different from the actual simulated one. Special caution is therefore needed about the model results over Greenland.

Because we are using monthly mean data, the notion of surface is expected to represent something quite diffuse and/or mixed, like the mixed layer in the case of oceanic surfaces and a few centimeters to a few meters of upper surface layer in the case of sea ice, snow, or ground surfaces.

d. Conversion to local surface temperature changes

Results given in this study consist of a decomposition of temperature changes directly rather than energy flux changes themselves, as in Pithan and Mauritsen (2014). To do so, we consider that the LW emission changes by the surface are the adjusted result of the changes in other terms and that they depend linearly on surface temperature changes $\Delta T_{\text{surf}}$, with $\Delta$ representing changes associated with the climate change, $\varepsilon$ the emissivity of the surface, $\alpha$ the Stefan–Boltzmann constant, and $T_{\text{surf}}$ the surface temperature). The surface warming sensitivity to surface LW emission change ($\Delta T_{\text{surf}}/\Delta L_{\text{Wup}}$) is calculated locally for each month and separately for each model. The local effect of a factor influencing the energy budget at the surface in terms of surface temperature change is then obtained by the following:

$$\Delta T_{\text{surf}} = \frac{\Delta T_{\text{surf}}}{\Delta L_{\text{Wup}}} \sum_{\text{fact}} \Delta R_{\text{surf}}^{\text{fact}} \varepsilon (\Delta T_{\text{surf}}/\Delta L_{\text{Wup}})^{\prime} \sum_{\text{fact}} \Delta R_{\text{surf}}^{\prime \prime} \tag{2}$$

where the sign of $\Delta L_{\text{Wup}}$ is defined positively upward, contrarily to other terms ($\Delta L_{\text{Wup}} - \sum_{\text{fact}} \Delta R_{\text{surf}}^{\text{fact}} = 0$), and $\Delta R_{\text{surf}}^{\text{fact}}$ represents the surface energy flux effect of each of the factors cited in section 2c (denoted by fact): surface albedo changes; vertically homogeneous atmospheric warming equal to the tropospheric vertical-mean value (barotropic warming); atmospheric temperature change departing from the barotropic warming (nonbarotropic warming); water vapor, cloud, imposed GHGs concentration, and latent and sensible heat flux changes; other surface effects, including the different factors listed in section 2c; and the residual of the radiation decomposition from the kernel method (kernel residual), including SW only (since the residual for the downward LW radiation is included in atmospheric warming terms as explained in appendix A).

Another factor that has been proposed to contribute to AA concerns the different surface warming sensitivity to a given energy flux imbalance depending on the background temperature (e.g., Ohmura 1984): that is, the local dependence of $\Delta T_{\text{surf}}/\Delta L_{\text{Wup}}$ to $T_{\text{surf}}$. The colder the surface, the more sensitive it is to a given energy flux change (i.e., Stefan–Boltzmann law). This effect, which acts on the warming sensitivity rather than on the energy budget, cannot be directly compared to other terms cited above. To quantify its role compared to other factors, we need to consider the local warming sensitivity w.r.t. some reference. As in Pithan and Mauritsen (2014), we consider the local departure (primes) of the surface warming sensitivity from the global mean (overbars) to separate this effect:

$$\Delta T_{\text{surf}} = \frac{\Delta T_{\text{surf}}}{\Delta L_{\text{Wup}}} \sum_{\text{fact}} \Delta R_{\text{surf}}^{\text{fact}} \varepsilon (\Delta T_{\text{surf}}/\Delta L_{\text{Wup}})^{\prime} \sum_{\text{fact}} \Delta R_{\text{surf}}^{\prime \prime} \tag{3}$$

or

$$\Delta T_{\text{surf}} = \frac{\Delta T_{\text{surf}}}{\Delta L_{\text{Wup}}} \sum_{\text{fact}} \Delta R_{\text{surf}}^{\text{fact}} + \left(\frac{\Delta T_{\text{surf}}}{\Delta L_{\text{Wup}}} \sum_{\text{fact}} \Delta R_{\text{surf}}^{\text{fact}}\right)^{\prime} \Delta L_{\text{Wup}} + \left(\frac{\Delta T_{\text{surf}}}{\Delta L_{\text{Wup}}} \sum_{\text{fact}} \Delta R_{\text{surf}}^{\text{fact}}\right)^{\prime \prime} \Delta L_{\text{Wup}} \tag{4}$$

The first term on the right-hand side (RHS) represents the sum of the different factors cited on the list above (assuming a global-mean surface warming sensitivity). The second term of the RHS represents the effect of the local departure of the surface warming sensitivity from the global mean (assuming a global-mean surface energy flux imbalance), referred to as surface warming sensitivity. The third term represents the interaction or synergy between the local surface energy imbalance and the local surface warming sensitivity, both departing from global mean (synergy effect). Note that, in Pithan and Mauritsen (2014), the second and third terms are considered altogether as an effect of the surface warming sensitivity alone, although the third term is also associated with the local anomalies of surface LW emission changes. It leads to an overestimation of the surface warming sensitivity effect that we have evaluated using a similar analysis as theirs, although using different data (appendix B). We found that the proper contribution of the surface warming sensitivity in Fig. 2c of Pithan and Mauritsen (2014) would probably be unchanged in terms of tropical warming but could be halved in terms of Arctic warming, leading to a possible reduction in annual-mean AA contribution of about 40% (appendix B).

e. Contributions to Arctic amplification

AA can be decomposed relative to the different factors following:
Arctic oceans, the mean warming (solid blue lines in Figs. 2a,b) is mostly influenced by the Arctic Ocean (orange line in Fig. 2b), compared to the GINL Seas (green line in Fig. 2b), as the former domain is more than 3 times larger in area than the latter (Table 2). Over land (Figs. 2c,d), the total Arctic mean (blue line) is mostly determined by the ice-free Arctic land (orange line in Fig. 2d) compared to Greenland (green line in Fig. 2d), as the former area is about 6 times larger than the latter (Table 2).

Dispersion among models is relatively small for tropical regions and large over Arctic domains and tends to be proportional to the mean value, with the exception of the dispersion over the GINL Seas, which is large, especially during late winter, despite a moderate warming.

f. Significance test

The robustness of a multimodel ensemble mean result is tested by considering if the sign of its value is consistent among models. It consists of a chi-square test with one degree of freedom. We consider that individual model results are independent from each other for simplicity and because dependency is difficult to assess. The confidence level chosen is 99%. In our case of 32 models being used, it is equivalent to considering if at least 24 model results have the same sign as the ensemble mean.

3. Results and interpretations

a. Surface temperature changes

Figure 1 shows the ensemble annual-mean surface temperature change after one century of moderate warming following the RCP4.5 scenario. As commonly simulated and already observed under current global climate change (e.g., Serreze and Barry 2011; Screen and Simmonds 2010), surface temperature warming is usually enhanced over the continents compared to the oceans at similar latitudes and enhanced at high northern latitudes compared to other latitudes (Arctic amplification). In the northern North Atlantic, however, a relatively weak warming is simulated compared to other Arctic regions and even globally (Drijfhout et al. 2012; Rahmstorf et al. 2015). We have decided to isolate this region, corresponding roughly to the GINL Seas (vertically striped region in Fig. 1), from the rest of the Arctic oceanic regions north of 60°N (horizontally striped region). Indeed, the feedback mechanisms are substantially different in these two domains (see section 3b below).

In terms of seasonality, both over Arctic land and Arctic oceans, the mean warming (solid blue lines in Fig. 2) is the largest during boreal winter, with a peak in November, and minimum during boreal summer, the period during which AA is not simulated. In terms of amplitude, the seasonal cycle of surface warming is larger over oceanic surfaces than over land surfaces in the Arctic region. Over the oceans, the total Arctic mean warming (blue line in Figs. 2a,b) is mostly influenced by the Arctic Ocean (orange line in Fig. 2b), compared to the GINL Seas (green line in Fig. 2b), as the former domain is more than 3 times larger in area than the latter (Table 2). Over land (Figs. 2c,d), the total Arctic mean (blue line) is mostly determined by the ice-free Arctic land (orange line in Fig. 2d) compared to Greenland (green line in Fig. 2d), as the former area is about 6 times larger than the latter (Table 2).

Dispersion among models is relatively small for tropical regions and large over Arctic domains and tends to be proportional to the mean value, with the exception of the dispersion over the GINL Seas, which is large, especially during late winter, despite a moderate warming.

b. Quantifying factors over oceans

Figure 2 shows the decomposition of surface temperature changes over oceans for the different months of the year. Over the Arctic Ocean (Fig. 2a), the sum of positive contributions is the largest during winter months with a peak in November, whereas the sum of negative contributions is the largest in summer, explaining most of the seasonality of surface temperature changes (black line). Most factors exhibit a clear seasonal cycle, the largest seasonal amplitudes being related to the albedo and other surface effects.

The positive albedo feedback effect is the largest in summer, with a peak of +5°C in July, linked with a large retreat of the mean sea ice cover (Fig. 4a, blue solid line) and a large solar irradiance, whereas it has no effect in winter when solar irradiance is weak or nonexistent. Other surface effects have a strong negative contribution in summer, partly balancing the albedo feedback, and a strong positive effect in winter, averaging to a small annual-mean effect of +0.34°C. A detailed analysis for better understanding this term is performed in section 3e. The barotropic warming contribution is relatively large (annual mean around +2°C) and positive all year-round, indicating the warming effect for the surface of an overall warmer troposphere through LW emission. The nonbarotropic warming contribution exhibits positive values most of the year, especially in late fall and early winter months (up to +2.5°C) but almost null contributions during summer months. It indicates that the warming over the Arctic Ocean in winter is the largest in the lower troposphere (which impacts the
Surface radiation budget more than the changes in the higher troposphere, associated with the usually strong vertical stability of the atmosphere (Fig. 5a). In summer, the atmosphere is less stratified, and the warming is not concentrated at the surface (Fig. 5a), leading to weak contributions of the nonbarotropic warming (Fig. 3a). Cloud feedback has a warming effect of up to $1.25^\circ$C in winter and a cooling effect of up to $1.5^\circ$C in summer (Fig. 3a). It is consistent with an increased mean cloud cover over the Arctic Ocean in the models (solid blue line in Fig. 4c), leading to both increased greenhouse and parasol effects, favoring backward LW radiation in winter but decreasing incoming SW radiation in summer. Other changes in cloud properties (e.g., the amount or phase of condensates or the mean altitude or depth of clouds) may also play a role in the total cloud feedback. The latent and sensible heat flux changes mostly tend to cool the surface, especially in winter (up to $2^\circ$C for latent cooling), associated with more evaporation and a greater warming of the surface than the atmosphere above during cold months. In summer, the contribution of surface heat fluxes is near zero. The large evaporation enhancement in winter is consistent with the sea ice cover reduction associated with a very large difference

**Table 2.** Percentages of total Arctic or tropical areas (all grid cells; first row) or total surface type (including fractions from grid cells with mixed surface types; second row) covered by grid cells with 100% of the surface type considered, for different domains, for T42 resolution.

<table>
<thead>
<tr>
<th>Domain for T42 grid cells with 100% of the surface type considered</th>
<th>Oceans</th>
<th>Land</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Arctic</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Arctic Ocean</td>
<td>GINL Seas</td>
</tr>
<tr>
<td>Arctic or tropics total domain (land and oceans)</td>
<td>25.4</td>
<td>7.6</td>
</tr>
<tr>
<td>Arctic or tropics domain type (total ocean or total land)</td>
<td>53.5</td>
<td>16.0</td>
</tr>
</tbody>
</table>

![Fig. 2. Surface temperature changes averaged (a) over Arctic and tropical oceans, (b) over the GINL Seas and the Arctic Ocean, (c) over Arctic and tropical land areas, and (d) over Greenland and the ice-free Arctic land. Solid lines represent the ensemble-mean values; dashed lines indicate the mean dispersion among models (plus or minus one std dev around ensemble mean).](image-url)
in specific humidity between the open-ocean and sea ice surfaces (i.e., large temperature difference and the nonlinearity of the Clausius–Clapeyron relationship, leading to a saturation level of specific humidity lower over sea ice than open water for a given temperature). The surface warming sensitivity has a moderate positive contribution to surface warming during cold months (up to more than $1$°C), associated with cold background temperatures (i.e., Stefan–Boltzmann law and section 2d). The water vapor and imposed GHGs effects are found to be relatively small in winter (up to about $0.5$°C for water vapor) and almost null in summer. The synergy effect, which represents the interaction between the local surface warming sensitivity and the local energy flux imbalance (see section 2d) tends to be relatively large and positive during cold months (around $+2.5$°C in winter, when surface energy flux changes and the surface warming sensitivity effects are the largest) and small to null during summer months.

Over the GINL Seas (Fig. 3b), the contributions of the different factors to the surface warming are very different from those over the Arctic Ocean. The amplitude of the seasonal variability of the surface temperature changes and of the different factors is much smaller. The albedo effect is moderate (up to $1.5$°C) and concentrated in the late spring months when sea ice retreat is the largest (Fig. 4a, dashed blue line), with sea ice melting entirely by midsummer even in historical runs (Fig. 4a, dashed green line). The atmospheric warming effect is positive all year-round and occurs mostly through barotropic warming (typically around $+2$° to $+2.5$°C), with the nonbarotropic warming effect being...
very small. The mean cloud cover increases slightly in summer (Fig. 4c, dashed blue line), leading to less SW radiation reaching the surface and contributing at least in part to the small cloud cooling effect (−1°C in June). Other surface effects are relatively large and negative all year-round (between −1°C and −4°C), although a relatively large dispersion between models is found in February and July when nine models simulate a positive contribution. The ensemble-mean negative values are consistent with a general decrease in the Atlantic meridional overturning circulation (AMOC) in most models (Drijfhout et al. 2012; Rahmstorf et al. 2015) and, hence, a reduction of the oceanic heat advection in the GINL Seas, where deep water convection occurs. This large cooling effect, partly compensating the atmospheric warming effect, leads to a weak warming of the ocean in this region (Fig. 1). The surface warming is therefore smaller than the surface air warming and leads to large positive sensible heat flux anomalies, contributing to more than +3°C surface warming in winter. The sign of the mean surface latent heat flux change is not robust among models all year-round.

As stated in section 3a, the averages over the total Arctic oceans (Fig. 3c) mostly bear the characteristics of the Arctic Ocean, as it represents a large fraction of the overall domain. One exception concerns the sensible heat flux effects, which are particularly large over the GINL Seas. Over tropical oceans (Fig. 3d), seasonality is very weak. The main contribution to surface warming occurs through barotropic warming (between +2°C and +2.5°C) and, to lesser extent, through the water vapor feedback (around +0.7°C). The nonbarotropic warming effect is negative all year-round (around −0.75°C), indicating that the atmospheric warming is weaker in the lower troposphere than at upper levels, resulting in negative temperature anomalies relative to the tropospheric-mean value close to the surface. It is consistent with the equatorial atmosphere being regularly unstable, leading to deep convection bringing surface heat excess upward, while adding extra heating through condensation. The net radiative warming effect is partly balanced by increased evaporation (around −1°C).

Contributions to oceanic AA (Fig. 6a) mostly follow the contributions to Arctic warming, as the amplification
effects are usually the greatest in this region. Exceptions include barotropic warming (almost the same warming effect over tropical and Arctic domains, resulting in a limited effect on AA), latent heating (almost the same cooling effect during boreal winter months in both regions but greater in the tropics during boreal summer, resulting in a moderate positive contribution to AA during this period), and nonbarotropic warming effect (usually the opposite effect in the two regions, resulting in a moderate to large positive contribution to AA, especially in winter).

c. Quantifying factors over land

Contributions to surface temperature changes over land are shown in Fig. 7. Contribution factor amplitudes are usually smaller than over oceans in the Arctic region (vertical axis reduced by half in Fig. 7 compared to Fig. 3). Over the ice-free Arctic land (Fig. 7a), the seasonality of factors is more complex than over the Arctic Ocean, with different contributions peaking at different times of the year. Some factors have similar effects as over Arctic oceanic regions, like the barotropic and nonbarotropic warming contributions, the surface warming sensitivity, or some other secondary factors (imposed GHGs, synergy, water vapor, and, to some extent, sensible heating), whereas major differences are also simulated. The albedo feedback peaks in spring over the ice-free Arctic land, similar to over the GINL Seas, but differently from over the Arctic Ocean where it peaks in summer. As in the GINL Seas, the feedback is due to a larger snow cover retreat in spring than in summer (blue solid line in Fig. 4b), since snow is fully melted even in historical runs at midsummer (Fig. 4b, solid green line). We can therefore consider that this difference in the timing of the albedo feedback between the Arctic Ocean and the ice-free Arctic land (and the GINL Seas) can be indirectly attributed to a difference in their mean latitude. A large snow cover difference is also found in fall as snow starts to accumulate later in late twenty-first-century runs (Fig. 4b), but solar irradiance is weaker than in spring and only results in a small secondary peak in albedo feedback. The timing of the cloud feedback is also partly different from over the Arctic Ocean, with the weak or negative contributions occurring in spring rather than in summer. The positive cloud contribution during the cold season, consistent with a larger cloud cover, is similar for both surface types (Figs. 4c,d). Cloud property changes other than the cloud cover itself may also play a role in the total cloud feedback. The latent heat flux effect is very different between the ice-free Arctic land and the Arctic Ocean, with increased evaporation occurring mostly in summer over the former domain and in winter for the latter. Finally, other surface effects have much less impact on the surface warming over land than over oceans.

The largest warming over the ice-free Arctic land during the coldest months is attributed to many positive contributions peaking at this time of the year and to weak cooling factors, whereas the minimum found in July–August is associated with weaker warming factors, especially albedo (compared to spring) and surface warming sensitivity, nonbarotropic warming, cloud or synergy effects (compared to winter months), and a relatively large latent cooling (compared to colder months).

Over Greenland (Fig. 7b), the seasonality of surface warming is weaker than over the ice-free Arctic land. The barotropic warming, surface warming sensitivity, and, to a lesser extent, cloud, water vapor, and imposed GHGs effects contribute to warmer surface temperature all year-round. The albedo feedback peaks in summer, consistent with the largest snow cover retreat over Greenland at this time of the year (dashed blue lines in Fig. 4b). Cooling factors are weak and mostly consist of a small sensible cooling during cold months and relatively large other
surface effects in summer, consistent at least in part with extra snow melting (Fig. 4b). It should be remembered that the simulated surface energy budget over ice sheets is subject to potentially large errors, as the melting of land ice is not taken into account in the models (see section 2c).

The total Arctic land surface temperature budget (Fig. 7c) is dominated by the ice-free land, as it represents a much larger proportion of the domain (Table 2). The factors contributing to the warming of land surfaces in the tropics (Fig. 7d) mostly consist of the mean tropospheric warming and water vapor feedbacks, partly balanced by the nonbarotropic warming effect, as for tropical oceans. Evaporation increase is much weaker than over tropical oceans, although compensated by an increase in heat transfer to the atmosphere through sensible heat flux.

The seasonality and the contributors to the land AA (Fig. 6b) are similar to the ones for ice-free Arctic land (Fig. 7a), with the exceptions of the mean tropospheric warming and water vapor feedbacks (both larger over tropical land than Arctic land, implying negative contributions to AA), sensible heat flux (larger cooling effect over tropical land, leading to positive contributions to AA), and nonbarotropic warming effects, usually of opposite sign over the two regions, leading to a relatively large contribution to positive AA over land. The amplitude of the AA over land is much smaller than for oceanic surfaces.

d. Total surface types

In the tropics, the proportion of oceans is much larger than land surfaces (Table 2), but the surface warming contributions are not much different, with the exception of latent and sensible heat fluxes (Figs. 3d and 7d). The total contributions over the whole tropics (Fig. 8b) are close to the ones over the oceans but also not very different from the ones over land. In the Arctic, the proportion of land and oceans is relatively similar (Table 2). Contributions to total surface Arctic warming (Fig. 8a) and to total AA (Fig. 6c) are therefore close to an average of respective contributions for oceanic and land surfaces with a relatively equal weight (Figs. 3c and 7c and Figs. 6a, b, respectively).

e. Interpreting other surface effects over the Arctic Ocean

To better understand the changes associated with the term other surface effects over the Arctic Ocean (dark blue bars in Fig. 3a), we try to reconstruct it [$R_{surf}^\text{oth}$ in Eq. (6) below] by adding estimates of the warming of the oceanic mixed layer [first term of the RHS of Eq. (6)] and of the sea ice surface (second term), the temperature
change associated with the deepening of the mixed layer (hence the mixing with deeper water of different temperature, third term), and the melting or formation of sea ice (fourth term):

\[
R_{\text{surf}}^\text{th} = -C_w \rho_w \left( (1 - \text{sic}) ml + \text{sic} \times ml^\text{si} \right) \frac{\partial T_{ml}}{\partial t} - C_w \rho_w h_{\text{si}} \text{sic} \frac{\partial T_{\text{si}}}{\partial t} \\
- C_p \rho_w \left[ (1 - \text{sic}) \frac{\partial ml^{\text{si}}}{\partial t} + \text{sic} \frac{\partial ml^\text{si}}{\partial t} \right] \left( T_{ml}^{\text{si}} - T_{dl}^{\text{si}} \right) + L_m \frac{\partial \text{sim}}{\partial t} + \text{res},
\]

with \( C_w = 4200 \text{ J kg}^{-1} \text{ K}^{-1} \) and \( C_{\text{si}}^{\text{si}} = 2060 \text{ J kg}^{-1} \text{ K}^{-1} \) the specific heat of water and sea ice at constant pressure; \( \rho_w = 10^3 \text{ kg m}^{-3} \) and \( \rho_{\text{si}} = 917 \text{ kg m}^{-3} \) the density of water and sea ice, respectively; \( L_m = 334 \text{ kJ kg}^{-1} \) the latent heat of fusion of ice; sic and sim the sea ice cover and mass, respectively; ml the mixed layer depth defined by sigma level \( T \) (standard CMIP5 data output variable mlotst); ml^{\text{si}} = \max(ml - \text{sid}, 0), \) the part of the mixed layer depth under sea ice (sid is sea ice depth); \( h_{\text{si}} = \min(1, \text{sid}) \), an arbitrary sea ice depth of 1 m (if sea ice is thick enough) over which energy fluxes are expected to spread [typical depth of seasonal temperature variations]
within sea ice in Perovich and Elder (2001); $T_{si}$ the surface sea ice temperature; $T_{ml}$ and $T_{ml}^{+1}$ the vertical-mean ocean potential temperature within the mixed layer for the current month considered and for the following month, respectively; $T_{dl}^{-1}$ the vertical-mean ocean potential temperature of the previous month in the portion of the ocean into which the mixed layer has grown during the month (under the mixed layer the month before but within the mixed layer the month after); $(\partial ml/\partial t)^{\bullet} = \max(\partial ml/\partial t, 0)$, the deepening of the mixed layer, if applicable; and $(\partial ml_{si}^{\bullet}/\partial t)^{\bullet}$, the deepening of the portion of the mixed layer under sea ice, if applicable. The residual term includes the horizontal heat advection, errors from approximations of other terms and other unaccounted mechanisms like a small-scale turbulent sensible heat flux between the mixed layer and the deeper oceanic layer.

Figure 9a shows the other surface effects over the Arctic Ocean as calculated in Fig. 3a but for 11 models (see Table 1; dark blue curve in Fig. 9a) and the result of its reconstruction following Eq. (6), but for changes expressed in terms of temperature [energy flux change multiplied by $\Delta T_{surf}/\Delta LW_{up}$ in Eq. (6)]. The reconstruction is very close to the actual values, bringing some confidence in the accuracy of the reconstruction. The different terms of the decomposition suggest that the main contributor to the overall energy absorption in summer is associated with the larger warming of the mixed layer [yellow curve in Fig. 9a; first term of Eq. (6)], and to a lesser extent with a larger melting of sea ice in early summer (green curve) and a larger vertical mixing in late summer (gray curve). The excess release of heat in winter seems mostly associated with an enhanced cooling of the mixed layer (yellow curve) and, to a lesser extent, with a larger sea ice formation (green curve), balanced by increased vertical heat mixing (gray curve) as winter progresses.

Among the different terms influencing the total change in the mixed layer temperature changes (Fig. 9b), the main contributor is associated with the amplitude of the temperature changes $\{-C_{w}^{\rho}[(1 - sic)ml + sic \times ml_{si}^{\bullet}]\Delta(\partial T_{ml}/\partial t), \text{red curve in Fig. 9b}\}$, rather than changes in the mixed layer depth (green and gray curves for open-ocean and sea ice-covered portions of the grid cells, respectively) or changes in the sea ice cover (blue curve).

The changes associated with vertical mixing due to a seasonal deepening of the mixed layer (gray curve in Figs. 9a,c) are mostly associated with changes in the temperature difference between the water that is going to be mixed in and the final mixed layer temperature $\{C_{w}^{\rho}[(1 - sic)(\partial ml/\partial t)^{\bullet} + sic(\partial ml_{si}^{\bullet}/\partial t)^{\bullet}]\Delta(T_{ml}^{+1} - T_{dl}^{-1})$, red curve in Fig. 9c\}$. It is due to a greater warming of the surface mixed layer in winter than the underlying water mass (not shown). The winter mixing therefore contributes to a larger relative cooling of surface waters.

4. Conclusions

AA occurs for both oceanic and land surfaces, except during boreal summer (Figs. 2 and 6). The seasonality of the Arctic surface warming is similar in terms of timing for both land and oceanic surfaces, with a minimum in summer and a maximum in winter (peak in November). Nevertheless, the amplitude of the warming seasonal cycle is much larger over oceanic regions (0°–6°C warming range) compared to land surfaces (0°–2°C). Accordingly, during summer, when the Arctic warming...
is minimum, AA is very weak or nonexistent, whereas it is the largest in winter.

The main reason for the warming minimum in summer over the Arctic Ocean (excluding the GINL Seas) is because of a large energy uptake by the ocean, compensating in large parts the excess energy absorbed by the surface due to the lower albedo (Figs. 3 and 8; sections 3b and 3e). Over land, the mechanisms explaining the summer minimum are different. Over the ice-free Arctic land (excluding Greenland), the warming minimum is mostly associated with a relatively small albedo effect (because all snow has already melted away in early summer) and a relatively strong evaporation increase (Fig. 7; section 3c).

In winter, the oceanic warming maximum is associated with the large extra oceanic heat storage and, to a lesser extent, with high surface warming sensitivity (Stefan–Boltzmann law effect), warming trapped close to the surface because of usually strong atmospheric vertical stability, and larger cloud cover contributing at least in part to the positive cloud feedback by trapping more longwave radiation (Figs. 3 and 8; sections 3b and...
The reasons for the land winter warming maximum are also associated with high surface warming sensitivity, strong atmospheric stability, and positive cloud feedback, but also with cooling factors being weak (small evaporation increase), whereas the heat storage effect is almost null (Fig. 7; section 3c).

The main difference between the GINL Seas compared to the rest of the Arctic Ocean concerns the role of the oceanic heat advection, which tends to decrease in the simulations (slowing down of the AMOC). It implies a weaker surface warming in this region, which is partly compensated by sensible heating from the warmer atmosphere (Fig. 3b; section 3b).

Over Greenland, the albedo change peaks in summer compared to spring over the ice-free Arctic land but is partly compensated by extra snow melting. Evaporation changes are relatively small over Greenland compared to other Arctic land surfaces (Fig. 7b; section 3c). It should be remembered that the physics of ice sheets is not properly taken into account in the models as they cannot grow, melt, or move and that the simulated surface energy budget over land ice is therefore subject to potentially large discrepancies (see section 2c).

A next step toward a deeper understanding of processes responsible for surface temperature changes in polar regions would be to separate the analysis over sea ice and open-ocean (snow and snow free) portions of oceans (land) within each grid cell, as the changes in surface energy budget should be substantially different for these different types of surface.

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**APPENDIX A**

**Correction Technique for Surface Radiative Kernel Analysis Using Pressure Levels and Comparisons between Radiative Decompositions**

*a. Problem using the kernel method on interpolated pressure levels*

We compare the results of the surface radiation decomposition for MIROC4m only for a doubling of CO₂ [further details in Yoshimori et al. (2009)], using different methods. The partial radiative perturbation method (PRP; Wetherald and Manabe 1988), corrected for a so-called decorrelation bias following Colman and McAvaney (1997), is expected to be the most accurate method. The kernel method on the model's 20 original sigma levels (kernel L20) and for the interpolated version on the 17 pressure levels as used in this study (kernel p17) are also applied for comparison. The results are in relatively good agreement w.r.t. the PRP method for kernel L20 and kernel p17 methods for most components of the decomposition (see details in section c below), except for the one associated with the
atmospheric temperature change for the kernel p17 method, for which a very large discrepancy is found for all regions and for all months (not shown). We think that it results from a large sensitivity of the surface LW radiative budget to the temperature change in the lowest part of the atmosphere, where the vertical interpolation induces substantial loss of accuracy. Nevertheless, since there is only one component of the decomposition that suffers such a large discrepancy, the difference (residual) between the actual radiative budget change and the sum of the terms of the decomposition almost entirely opposes this bias. Moreover, this discrepancy resulting from the atmospheric warming effect only impacts the downward LW component of the surface radiative budget. Therefore, adding the residual from the downward LW radiative budget $D_{LW_{res \_dn}}$ to the component associated with the atmospheric temperature change $D_{LTatm}$ provides a good correction effect for this term, which then lays in good agreement w.r.t. the PRP and kernel L20 methods.

**FIG. A1.** Decomposition of surface temperature changes (K; black dashed lines) for a doubling of CO$_2$ in MIROC4m, using (left) the PRP, (center) the kernel L20, and (right) modified kernel p17 methods over different oceanic domains. Empty bars and dashed lines are identical in the three methods.
b. Correction technique

Since $\Delta LW_{\text{atm}}$ is further split into a barotropic and a nonbarotropic warming component (see section 2c), we need to make a choice regarding how to split the correction term between these two components. Since we think that the original discrepancy arises from a loss of vertical accuracy for levels close to the surface, it is logical to consider that the weakest bias will be for the effect of the nonbarotropic atmospheric warming departing from the surface air temperature change, as its value will be close to zero for the lowest levels of the atmosphere. We therefore add the entire corrective term $\Delta LW_{\text{res, dn}}$ to the term associated with a uniform barotropic warming (Barotr) equal to the surface air temperature change $\Delta Tas$ 

$$\text{Barotr}_{\text{corrected}}(\Delta Tas) = \text{Barotr}_{\text{uncorrected}}(\Delta Tas) + \Delta LW_{\text{res, dn}}$$

Nevertheless, the barotropic warming component defined in this study corresponds to atmospheric temperature changes equal to the mean vertical tropospheric value $\Delta T_r$. It can be obtained linearly from

$$\text{Barotr}_{\text{corrected}}(\Delta T_r) = \frac{\Delta T_r}{\Delta Tas} [\text{Barotr}_{\text{uncorrected}}(\Delta Tas) + \Delta LW_{\text{res, dn}}]$$

and

---

Fig. A2. As in Fig. A1, but for different land domains.
NonBarot corrected($\Delta T_r$) = NonBarot uncorrected($\Delta T_a$) + $\left(1 - \frac{\Delta T_r}{\Delta T_a}\right)$ [Barot uncorrected($\Delta T_a$) + $\Delta L W_r e s_{dn}$]

as the sum of the two terms should be equal to Barot uncorrected($\Delta T_a$) + NonBarot uncorrected($\Delta T_a$) + $\Delta L W_r e s_{dn}$.

c. Comparisons between PRP, kernel L20, and modified kernel p17 methods

Figures A1–A3 show the surface temperature change decompositions for the PRP (left panels), the kernel L20 (center panels), and the modified kernel p17 (right panels) methods, averaged over the same subregions as in the present study. Contributions that are the same for the three methods (surface warming sensitivity, other surface effects, surface latent, sensible, imposed GHGs, and synergy terms) are plotted as empty bars. Note that the barotropic and nonbarotropic warming contributions for the PRP and kernel L20 methods are not obtained directly for temperature changes equal to $\Delta T_r$ but approximated linearly from vertically uniform temperature changes equal to $\Delta T_r$ and $\Delta T_a$ and NonBarot corrected($\Delta T_r$) = NonBarot corrected($\Delta T_a$) + $\Delta L W_r e s_{dn}$.

The surface albedo effect is relatively well approximated through the kernel methods (terms identical for kernel L20 and modified kernel p17) but can lead to substantial overestimations, indicating that nonlinear effects are relatively important. Barotropic and nonbarotropic warming
terms usually lead to the largest differences between the kernel L20 and modified p17 methods. The modified p17 kernel method usually results in values being closer to the PRP ones compared to kernel L20, although some substantial discrepancies regarding the former method are sometimes found, for example in June over the Arctic Ocean for the nonbarotropic contribution [red bars in Fig. A1 (top)]. Water vapor and cloud contributions are well approximated in both kernel methods. The residual term is usually substantially larger for both methods compared to PRP results, although remaining small.

In conclusion, the precise quantification of the different contributions presented in this article may suffer substantial biases (especially over oceans for barotropic and nonbarotropic terms, as in June), but the main features and orders of magnitude should be consistent. Note that the loss of accuracy using the modified p17 kernel method is balanced by the use of a multimodel analysis, which is expected to lead to more robust results than single model results.

**APPENDIX B**

**Comparison of Current Methodology to Pithan and Mauritsen (2014)**

Pithan and Mauritsen (2014) present some energy flux analyses to explain Arctic amplification in simulations for which CO₂ concentration level in the atmosphere is quadrupled. The main conclusion is that temperature feedbacks dominate the annual-mean AA and that surface albedo feedback is the second largest contributor, both from surface and top-of-atmosphere perspectives. We want to point out some misleading quantification of the term called “surface warming” in their paper (and which we think should better be referred to as “surface warming sensitivity”). Indeed, they include the second and third terms of the RHS of Eqs. (3) and (4) as a sole effect of the surface warming sensitivity, although the third term is also associated with the local anomalies of surface LW emission changes. We argue that the surface warming sensitivity effect should only correspond to the second term, whereas the third one could be referred to as “interaction” or “synergy” between the local surface energy imbalance and the local surface warming sensitivity, both departing from global mean.

To consider how their conclusions might be changed when making the appropriate separation, we performed a similar plot as their Fig. 2c (but with the data and modified p17 kernel method presented in the present article), using their decomposition, and by separating the synergy effect from the surface warming sensitivity effect (Fig. B1). Although the models, scenarios, and radiative kernels are different between their analysis and ours, results are very similar, showing that global warming effects on AA are robust and that we can make an appropriate comparison between our results and theirs. The dashed orange diamond represents the surface warming sensitivity effect according to their decomposition, whereas the solid orange and pink diamonds represent the separate contributions of the surface warming sensitivity only and of the synergy, according to our interpretation. Even if the corrected surface warming sensitivity contribution to the annual-mean AA remains important (second largest), it is substantially reduced, by about 40%, as its contribution to Arctic warming itself is halved.

**REFERENCES**


