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SUPPLEMENTARY INFORMATION
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Arctic winter warming amplified by the thermal inversion and consequent low infrared cooling to space

Methods

There are a number of ways to assess climate feedbacks, each with their own advantages and disadvantages. One of the main difficulties is to separate climate change (and the responsible feedback strengths and characteristics) from climate variability. An often used method to increase the signal-to-noise (i.e. climate change-to-variability) ratio is to study multi-model output, such as those obtained in the CMIP3 initiative for 'realistic' forcing scenarios. The general idea then is to apply statistics on the multitude of independent members (individual models) to reduce the noise[1], and also to use intermodel differences to relate climate processes to feedbacks[2]. Another method, the one employed here, is to use one climate model and apply a sufficiently large forcing (e.g. $2xCO_2$) to obtain a climate change signal that is much larger than the noise. The advantage of this approach is that dedicated experiments can be carried out, including changing certain model processes in order to link these to feedbacks (as is done in this study). Varying the physics in one climate model is a more structured way to create a multi-model ensemble in a perturbed physics framework. These two methods, as well as others, complement each other. There is no indication that one method is superior to others.

We use the global coupled climate model EC-Earth[3], which encompasses the following components: atmosphere, ECMWF's Integrated Forecast System (IFS), resolution T159L62; ocean, NEMO V2, resolution 1 deg.; sea ice, LIM2, resolution 1 deg; coupled through the OASIS3 coupler. The performance of EC-Earth in terms of its simulation of the present-day climate is good[3] in that it significantly outperforms

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the average of the CMIP3 models. EC-Earth is used to perform the various climate change simulations reported here, including simulaitons with changed model physics We did two simulations per set: one for the current climate (1xCO_2) and one for a doubled CO_2 climate (2xCO_2). Three sets of experiments were done to assess the effect of varying stable boundary layer mixing. One set of experiments was done with the standard model, the other two with 'increased' and 'decreased' mixing in the stable boundary layer, respectively (as part of the perturbed physics approach). The increased (decreased) mixing case involves changes in the revised Louis scheme[29] such that the exchange coefficient for heat and moisture increases (decreases) by ~50%. All runs started in 1990 and were 44 year long; results presented in this paper represent means over the final 10 year of these runs, so as to ensure that the difference between the 2xCO_2 and 1xCO_2 run was representative of the climate signal.

The regression method to separate surface and atmospheric contributions of OLR

Because the surface and atmosphere may have opposite contributions in terms of longwave cooling[4], we will quantify the surface and the atmospheric parts separately. This separation of the total TOA outgoing longwave radiation (OLR) into an atmospheric (L_{Tatm}) and a surface contribution (L_{Tsur}) involves a linear regression of OLR against the emitted longwave radiation at the surface (L_S):

$$OLR = (1 - \varepsilon_e)\, L_S + L_{Tatm}$$

where $L_{Tsur} = (1 - \varepsilon_e) L_S$ represents the portion of the longwave flux emitted by the surface that reaches TOA, and ε_e is the effective emissivity of the atmosphere. This regression was made for data grouped seasonally per three months (e.g. December–January–February) and per latitude band; hence resulting values of L_{Tatm} represent seasonal and zonal averages. This procedure assumes that the coefficients of the fit do not vary within the subgroups for which each fit is made. In the extratropics, this regression method yields very good correlations, with R^2 values over 0.9 in most seasons and latitudes, both for clear-sky and all-sky conditions. This method allows us to quantify the atmospheric cooling efficiency $\gamma = L_{Tatm}/(L_{Datm} + L_{Tatm})$, with L_{Datm} being the atmosphere-emitted infrared radiative flux to the surface.

One possible systematic source of error is caused by the atmospheric temperatures (especially close to the surface) being correlated to the surface temperature, which may cause the regression technique to erroneously supplement the surface part by an atmospheric contribution. In other words, because the surface and atmospheric temperatures are correlated, the regression technique might overestimate the surface contribution, and thereby underestimate the atmospheric contribution. We tested this by using a 1-D broadband longwave radiative transfer model[5]. This involves broadband emissivity computations in six spectral intervals, taking into account all major greenhouse gases[6]. This radiative transfer code has been used in the ECHAM-4 climate model; more details can be found in ref (7).

We performed two tests: one in which we varied the surface temperature only, and one in which we varied the surface temperature and the atmospheric temperature simultaneously (the same amount up or down), and calculated the changes in

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longwave fluxes. The tests were designed such that both cases should result in the same fluxes. Then, for both experiments, we performed the regression method by fitting OLR against L_s. The result, shown in Supplementary Figure 1, is that the difference between the regression parameters is ~2.5%. In other words, this regression technique overestimates the surface contribution to OLR by about 2.5%. This is a small, but not entirely negligible amount (in the example shown it amounts to ~4 W m^{-2}). However, the main arguments put forward in this study revolve around the climate-induced *changes* in OLR contributions (see Figs. 2 and 4). Since this uncertainty always works in the same direction it is highly likely that its effect on the difference fluxes is very small.

Sensitivity of OLR to changes in the vertical temperature profile

In an attempt to understand the inability of the Arctic wintertime atmosphere to increase its upward emitted longwave radiation, resulting in strong decreases in cooling efficiency, we performed several tests with the 1-D broadband longwave radiative transfer model[5] described above. Specifically, we tried to determine how OLR responds to perturbations in atmospheric temperature (and moisture; we employ fixed relative humidity). We used three 'standard' atmospheric profiles: the standard midlatitude temperature profile (MLP), an Arctic profile (defined here as MLP minus 25 °C), and finally an Arctic profile with an inversion of 10 °C in the layer closest to the surface (Supplementary Figure 2a), representative of Arctic winter conditions, by reducing the temperature of the near-surface layers relative to the Arctic profile. Each of these profiles have been perturbed, per vertical layer, with an 1 °C temperature increase, and the resulting changes in OLR are shown in Supplementary Figure 2b. All calculations here were done for clear-sky conditions.

It is clear that for any profile considered here the maximum sensitivity is situated in the middle atmosphere. The sensitivity decreases towards lower altitudes because the atmosphere aloft increasingly absorbs the additionally emitted radiation. This general picture agrees very well with detailed radiation calculations[4,8]. Evidently, the colder (and dryer) the atmosphere, the less sensitive it is, culminating in an extremely low sensitivity in the inversion layer in Arctic winter. Another contributing factor is that the overlying atmosphere – which in case of an inversion is warmer and more moist than the surface – absorbs part of the upward radiation and thereby reduces the emitted infrared radiation by the cold surface layers that reach TOA. Hence, with Arctic wintertime warming being confined to the near-surface layers, OLR will hardly increase and the majority of the additionally emitted energy will be directed downward (hence a strong decrease in γ, see Figure 3a).

If the warming signal is entrained to higher levels, the OLR will increase more strongly because of the enhanced sensitivity at higher levels. This explains the increased atmospheric radiative cooling in the 'increased mixing' case (Fig. 4a). Finally, Huang et al. (2007) (ref 4) demonstrate that the sensitivity of OLR to surface temperature changes generally increases towards the pole, presumably because polar atmospheres are less opaque to longwave radiation. In any case, the concurring facts that in Arctic winter, near-surface atmospheric OLR sensitivity is very low, whereas the surface OLR sensitivity is relatively high, both contribute to the unexpected result

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depicted in Figure 2, demonstrating that in the Arctic winter most additional OLR originates from the surface.

Effect of inversion strength on the atmospheric cooling efficiency

Our results show that the change in atmospheric cooling efficiency (γ) decreases strongly with inversion strength (see Supplementary Figure 3). This implies that the stronger the inversion is, the less radiation will be radiated out to space by the warming atmosphere. Hence, the longwave feedback is less negative in case of strong inversions, thereby amplifying the warming.

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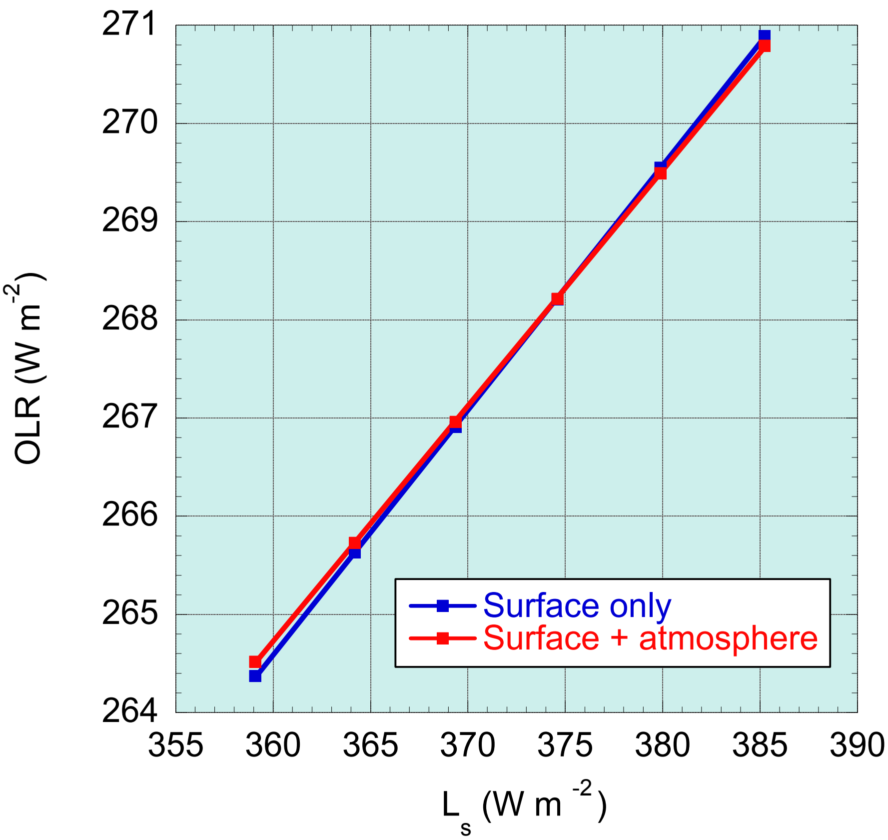
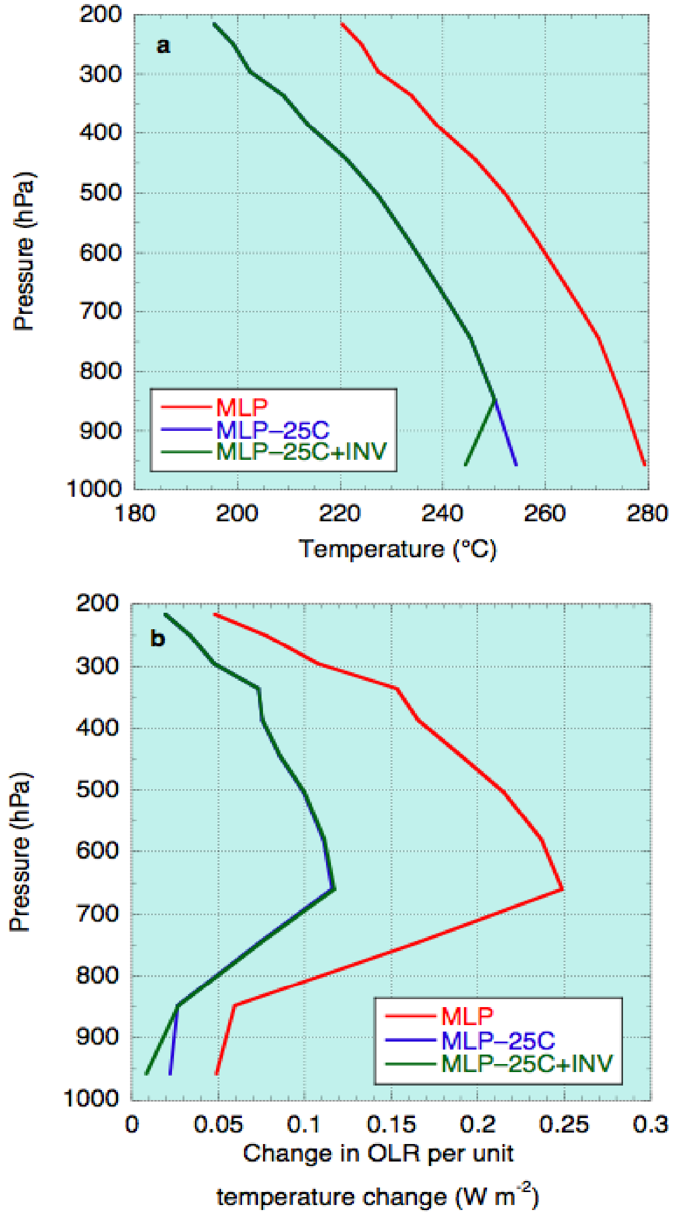


Supplementary Figure 1 | Comparison of the linear regression to separate TOA longwave fluxes (see Methods), between surface only temperature variations and surface and atmospheric temperature variations combined. Dots represent model simulated longwave radiative fluxes, while the lines represent the linear fits. A standard Arctic profile was used to calculate the fluxes. In the 'Surface only' case, atmospheric temperatures were kept fixed while surface temperature were varied. In the 'Surface + atmosphere' case, atmospheric temperatures in the lowest layers (below 725 hPa) of the troposphere were changed in concert with the surface temperature changes. The y-axis intercept equals the atmospheric contribution to OLR (which in both cases has quite normal values), which in the 'Surface + atmosphere' case increases by 2.5% with respect to the 'Surface only' case.



Supplementary Figure 2 | Sensitivity of OLR as a function of unit temperature anomalies as calculated by a 1-D radiative transfer model. **a**, The three representative atmospheric temperature profiles, with a midlatitude profile (MLP), an Arctic profile (MLP–25C) and an Arctic winter profile with a 10 °C surface inversion (MLP–25C+INV), **b**, change in OLR per unit temperature increase per height level for the respective temperature profiles.

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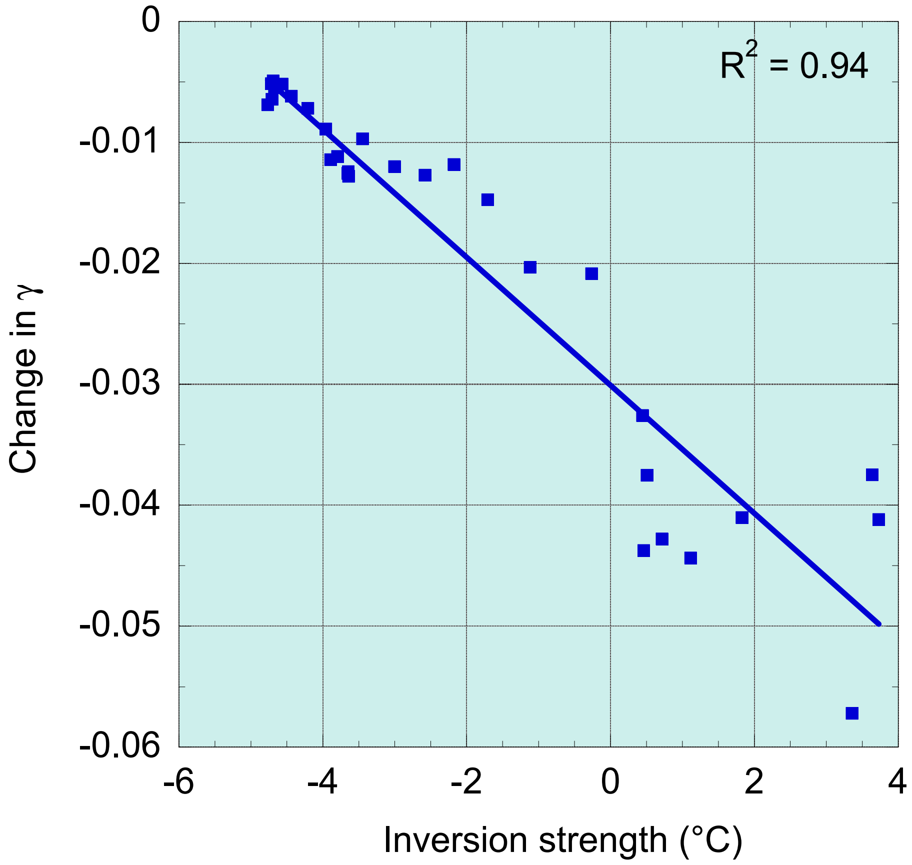

Supplementary Figure 3 | Change in cooling efficiency (γ) for the CONTROL experiment ($2xCO_2$ – $1xCO_2$) as a function of inversion strength in the current ($1xCO_2$) climate in DJF. Each square denotes a zonal mean value over a 2° latitude zone, from 30° – 90° N. The line represents the linear fit to the data. Inversion strength is defined here as the temperature difference between the 925 hPa and 1000 hPa level).