Atmospheric and Oceanic Contributions to Irreducible Forecast Uncertainty of Arctic Surface Climate

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(Manuscript received 17 June 2015, in final form 9 October 2015)

ABSTRACT

Uncertainty of Arctic seasonal to interannual predictions arising from model errors and initial state uncertainty has been widely discussed in the literature, whereas the irreducible forecast uncertainty (IFU) arising from the chaoticity of the climate system has received less attention. However, IFU provides important insights into the mechanisms through which predictability is lost and hence can inform prioritization of model development and observations deployment. Here, the authors characterize how internal oceanic and surface atmospheric heat fluxes contribute to the IFU of Arctic sea ice and upper-ocean heat content in an Earth system model by analyzing a set of idealized ensemble prediction experiments. It is found that atmospheric and oceanic heat flux are often equally important for driving unpredictable Arctic-wide changes in sea ice and surface water temperatures and hence contribute equally to IFU. Atmospheric surface heat flux tends to dominate Arctic-wide changes for lead times of up to a year, whereas oceanic heat flux tends to dominate regionally and on interannual time scales. There is in general a strong negative covariance between surface heat flux and ocean vertical heat flux at depth, and anomalies of lateral ocean heat transport are wind driven, which suggests that the unpredictable oceanic heat flux variability is mainly forced by the atmosphere. These results are qualitatively robust across different initial states, but substantial variations in the amplitude of IFU exist. It is concluded that both atmospheric variability and the initial state of the upper ocean are key ingredients for predictions of Arctic surface climate on seasonal to interannual time scales.

1. Introduction

Arctic climate dynamics, with its strong natural variability on time scales from seasons to decades (e.g., Bengtsson et al. 2004; Döschel and Koenigk 2013; Beitsch et al. 2014) and its amplified response to global climate change (e.g., Serreze and Francis 2006; Pithan and Mauritsen 2014), is still poorly understood. Because the Arctic is characterized by sustained large-scale heat flux convergence in the ocean and the atmosphere and, at the same time, exhibits strong surface heat flux variations due to extremely seasonal incoming solar radiation, useful insights into coupled atmosphere–ocean processes can be gained from analyzing anomalies in the Arctic heat budget (Tietsche et al. 2011). Here, we explore which heat fluxes contribute to anomalies of the Arctic heat budget in idealized ensemble prediction experiments with a state-of-the-art Earth system model.

For initial-value predictions of the Arctic climate, on time scales from seasons to years, both atmospheric and oceanic heat fluxes are potentially important. A strong contribution from oceanic heat flux would raise the chances of pushing back the predictability horizon, owing to larger thermal inertia and slower time scales of the ocean. The mean lateral oceanic heat transport into the Arctic is much smaller than the mean atmospheric heat transport (Serreze et al. 2007), but anomalies on interannual to decadal time scales might be of comparable magnitude (Shaffrey and Sutton 2006). Model results indicate that there is indeed correlation between ocean heat transport anomalies and sea ice anomalies in the Arctic (Holland et al. 2006; Zhang 2015). It has been suggested that the drastic sea ice decline observed in the late 2000s was partly triggered by enhanced ocean heat...
transport into the Arctic some years before (Schauer et al. 2004; Polyakov et al. 2010). However, given the strong stratification of the upper 500 m of the Arctic Ocean with colder and fresher water at the surface and warmer and more saline water at depth, the release of heat transported into the Arctic at depth to the surface is expected to be slow. There is an ongoing debate about the magnitude and variability of this vertical heat exchange, with tidal mixing over rough topography (Rippeth et al. 2015; Lique 2015), double-diffusive mixing (Sirevaag and Fer 2012), and Ekman pumping (Yang 2009) as possible contributing processes.

This uncertainty in the relative role of surface heat fluxes, ocean lateral heat transport, and vertical heat exchange at depth in shaping variability of the Arctic surface climate is relevant to the practical problem of predicting the state of the Arctic sea ice cover months to years ahead. Recent seasonal prediction experiments (Sigmond et al. 2013; Chevallier et al. 2013; Msadek et al. 2014; Peterson et al. 2015) have shown partly contradictory levels of skill. The association between lack of skill and potential sources of uncertainty remains elusive. However, it is highly likely that all three fundamental sources of uncertainty are important: model errors, initial-condition uncertainty, and the irreducible forecast uncertainty (IFU). The IFU is caused solely by the chaotic nature of the climate system and manifests itself in trajectory divergence of prediction ensembles with almost identical initial conditions (often referred to as the ensemble spread). In practice, the three sources of uncertainty are hard to separate, and estimates of IFU depend on the model and initial conditions used.

All three sources of uncertainty have been discussed in the literature: model errors mainly in the context of developing new and better representations of physical processes in the Arctic (e.g., Tsamados et al. 2014; Vihma et al. 2014), initial-condition uncertainty in data-assimilation and perturbation studies (Day et al. 2014; Massonnet et al. 2015), and IFU in idealized ensemble prediction experiments (e.g., Blanchard-Wrigglesworth et al. 2011; Tietsche et al. 2013b). However, to our knowledge the nature of the irreducible forecast uncertainty of Arctic surface climate has not been adequately characterized. What are the principal mechanisms responsible? Is predictability lost purely through atmospheric “weather noise,” or is there a role for internal ocean dynamics that is potentially less well observed and understood?

Here, we characterize IFU of Arctic surface climate by analyzing the heat budget of the upper Arctic Ocean and sea ice in a series of idealized ensemble prediction experiments. This allows a direct attribution of IFU to either surface heat fluxes, ocean lateral heat transport, or vertical heat exchange at depth. This knowledge of the relative contributions of these heat fluxes to IFU can inform prioritization of model development and expansion of the observational network.

Two examples might serve to illustrate that. First, if it was found in idealized ensemble prediction experiments that the vertical ocean heat exchange at depth contributed significantly to IFU, the aforementioned interdependency of the three types of forecast uncertainty would imply that both initial conditions and the dynamical evolution of the subsurface ocean are important for predictions of Arctic surface climate, and a strong case could be made for focusing efforts on improving observation capacity and model fidelity in the subsurface ocean. On the other hand, if IFU in a certain region, season, or lead time was found to be caused by the atmospheric surface heat flux alone, it would suggest that improved atmospheric observations and modeling fidelity should to be prioritized.

The remainder of this article is structured as follows: section 2 describes the Earth system model used, the ensemble prediction experiments, and the simulated Arctic heat budget; section 3 presents the attribution of irreducible forecast uncertainty to heat fluxes; and section 4 summarizes the results.

2. Model and methods
a. MPI-ESM

We use the low-resolution version of the Max Planck Institute Earth System Model (MPI-ESM) as used for phase 5 of the Coupled Model Intercomparison Project (CMIP5). The atmosphere component of MPI-ESM is ECHAM6 with a horizontal resolution of approximately 1.5° and 47 vertical levels (Stevens et al. 2013). This atmospheric model is the successor of ECHAM5 (Roeckner et al. 2003), with the most notable improvements being related to the radiative transfer code. In comparison to its predecessor, ECHAM6 has more vertical levels and extends higher into the upper atmosphere so that it is capable of representing internal variability in the stratosphere, which potentially has an impact on Arctic predictability via stratosphere–troposphere coupling.

The ocean component of MPI-ESM is MPI-OM as described by Marsland et al. (2003) with a few minor modifications. The horizontal resolution of the ocean model is between 20 and 100 km in the Arctic. There are 40 vertical levels; the upper 100 m of the ocean are represented by 8 levels with thicknesses ranging from 10 to 15 m. For a description of the ocean circulation in MPI-ESM, see Jungclaus et al. (2013). A single-category dynamic–thermodynamic sea ice model based on Hibler (1979) is embedded in the ocean model.
details on the sea ice model, see Notz et al. (2013) or Tietsche et al. (2013a).

b. Setup of idealized ensemble prediction experiments

We carry out a set of idealized ensemble prediction experiments to estimate the oceanic and atmospheric contributions to the heat budget of the upper Arctic Ocean. Verifying Arctic predictions against observations is still in its infancy, given sparse observations and strong model biases. These problems will cause any heat budgets calculated from real predictions to be polluted by initialization shocks and drifts. For this study, we entirely rely on simulations with MPI-ESM, which allows a physically consistent interpretation of the results. However, caution is advisable for aspects of the simulation where the model can be demonstrated to deviate from the observed climate. Model output from the simulations is openly available from the British Atmospheric Data Centre (Day et al. 2015).

The experimental protocol is the same as in Tietsche et al. (2014); we perform a long control simulation with radiative forcing fixed at conditions representative of the year 2005. From this control simulation, we select initial states so that the dates of these initial states are at least several years apart and can be considered independent of each other for the purpose of seasonal-to-interannual predictions. We also deliberately sample the range of natural variability in the Arctic in the model in terms of sea ice extent and volume and ocean heat transport into the Arctic. There are 16 ensembles of 16 members starting on 1 November and 12 ensembles of 9 members starting on 1 July. This means we can also compare the difference between predictions started in the melting and freezing seasons. The length of each simulation is 36 months.

Perturbed initial conditions for the ensembles are created by adding a very small amount of spatially uncorrelated noise to the sea surface temperature field (Gaussian distribution with $10^{-4}$ K standard deviation). The perturbation is so small that it is equivalent to assuming perfect knowledge of the initial conditions; the subsequent evolution of the ensemble is solely determined by the chaotic dynamics of the climate system.

c. Some aspects of the simulated Arctic climate

The setup of the control simulation means that the state of the Arctic in our experiments does not correspond to the observed recent decades but rather to a future state under a “commitment” radiative forcing scenario (IPCC 2013, section 12.5), where some future warming is expected (Li et al. 2013). Nevertheless, the main features of the Arctic climate simulated here are close to observation over the recent decades.

The mean sea ice extent in the control simulation is $4.5 \times 10^6$ km$^2$ in September and $13 \times 10^6$ km$^2$ in March, which is close to the average ice extent observed over the last 10 years (2006–15): $5 \times 10^6$ km$^2$ in September and $15 \times 10^6$ km$^2$ in March (Fetterer et al. 2002). The sea ice volume, however, is substantially smaller than current best estimates for the last 10 years: $3 \times 10^3$ km$^3$ and $17 \times 10^3$ km$^3$ in September and March in our simulations versus $6 \times 10^3$ km$^3$ and $23 \times 10^3$ km$^3$ in September and March in PIOMAS, respectively (Schweiger et al. 2011). For further details on the simulated sea ice state, see supplementary figures in Tietsche et al. (2014).

Transport through the Barents Sea Opening and Fram Strait is an essential component of the heat budget of the Arctic Ocean. Both sections are resolved by the model with approximately 20 grid points. Inspection of the full-depth temperature and velocity sections (not shown) reveals that the model is able to simulate the main features of the observed flow through these sections, but velocities tend to be too low and temperatures too high in general. The first bias is to be expected given the relatively low resolution of the model, and the second bias is related to the commitment-scenario warming.

The simulation of the Arctic sea level pressure (SLP) and 2-m air temperature in MPI-ESM is shown in Fig. 10 in Notz et al. (2013), who conclude that the simulation of both fields is much more consistent with reanalysis products than in previous model versions. Notable remaining biases are a too weak expression of the Aleutian low and the Beaufort High in winter, and a slight cold bias in surface air temperature over the Arctic Ocean both in summer and winter.

d. Definition and simulation of the upper–Arctic Ocean heat budget

We are interested in forecasting anomalies in Arctic sea ice and the upper-ocean waters. To trace where these anomalies come from, we define an Arctic Ocean domain and attribute the changes in heat content in the domain to heat fluxes across its boundaries. The domain is the same as the one used by Serreze et al. (2007), who synthesized a variety of atmospheric and oceanic data available at the time to estimate the large-scale energy budget of the Arctic. As shown in Fig. 1a, the lateral boundaries of the domain are the Nares Strait and the Parry Channel in the Canadian Archipelago, the Bering Strait, the Fram Strait, and the Barents Sea Opening. Its boundaries in the vertical are the ocean surface and a depth of 90 m. The domain includes the upper 8 levels of the ocean model, as well as sea ice and snow on sea ice. The lower boundary at 90-m depth is chosen so that it is deep enough to encompass the Arctic Ocean surface mixed layer throughout the year (with exception of the
Barents Sea) and allows for some lateral ocean heat transport. At the same time, it is shallow enough to be relevant for surface climate and exclude the Atlantic layer, which has its own very strong variability and hardly communicates with the surface layers.

The total heat flux $F$ into the domain is the sum of the surface heat flux $F_{\text{sfc}}$ and the ocean heat flux $F_{\text{ocn}}$, the latter consisting of the ocean lateral heat flux $F_{\text{ol}}$ and the ocean vertical heat flux $F_{\text{ov}}$:

$$F = F_{\text{sfc}} + F_{\text{ocn}} = F_{\text{sfc}} + F_{\text{ol}} + F_{\text{ov}}.$$  (1)

The direction of all heat fluxes is defined so that a positive heat flux increases the heat content of the domain with respect to liquid water at a reference temperature of $0^\circ\text{C}$. See Fig. 1b for a schematic of the heat budget.

The atmospheric surface heat flux comprises the net shortwave radiation, the net longwave radiation, and turbulent sensible and latent heat fluxes. The lateral component $F_{\text{ol}}$ of the ocean heat flux is the sum of the latent heat transport associated with the export of sea ice and snow on sea ice through the Fram Strait and the sensible heat transport associated with the transport of ocean water with a temperature differing from the reference temperature through all the lateral ocean boundaries. The vertical component $F_{\text{ov}}$ is due to vertical motion of ocean water across the lower boundary of the Arctic Ocean domain at 90-m depth. We do not specify which process is responsible for the vertical ocean heat flux but expect that the largest contribution will be entrainment of warm water at depth during winter mixed layer deepening and possibly Ekman pumping in some locations (Yang 2009).

The climatological average of the simulated heat fluxes into the Arctic Ocean domain over the course of a year is shown in Fig. 2a. The average has been calculated from the last September–August period in ensembles started in November, which gives a sample of 256 simulated model years. The net surface heat flux is comparable to the one diagnosed by Serreze et al. (2007) from reanalysis data. However, it is difficult to meaningfully assess the realism of simulated surface heat fluxes in more detail for two reasons: first, the simulations here should resemble a future slightly warmer climate because the initial conditions were taken from a 300-yr-long control simulation started from the end of a CMIP5 historical twentieth-century simulation, with radiative forcing fixed to 2005 conditions. Second, the reanalysis data themselves have large uncertainties in surface heat fluxes. For instance, Lindsay et al. (2014) demonstrate that reanalyses simulate radiative heat fluxes at the surface that can be biased by several tens of W m$^{-2}$.

The realism of the vertical ocean heat flux $F_{\text{ov}}$ at 90-m depth shown in Fig. 2a is, to our knowledge, almost impossible to evaluate, since no large-scale observations exist. However, the $F_{\text{ov}}$ simulated by MPI-ESM seems to be compatible with the expected magnitude and seasonal behavior; the climatological mean of $F_{\text{ov}}$ is practically zero in summer, when the surface mixed layer in the Arctic is very shallow. In winter, the modeled $F_{\text{ov}}$ reaches up to 16 W m$^{-2}$. This high value is due to the Barents Sea being included, where convection deeper
than 90 m is the norm. If the Barents Sea is excluded, the maximum winter value for the upward heat flux is 4 W m$^{-2}$, which is compatible with earlier large-scale estimates (see Polyakov et al. 2010).

The contributions to the lateral ocean heat flux (Fig. 2b) are calculated for a reference temperature of 0°C. This temperature makes interpretation of heat fluxes easy because (i) it is close to the average temperature of the domain and (ii) it is close to the phase transition temperature of water so that heat fluxes can be interpreted as causing extra melting or freezing. The same reference temperature, or only slightly deviating values, are often used in the literature (Aksenov et al. 2010; Tsubouchi et al. 2012; Lique and Steele 2013; Koenigk and Brodeau 2014). Note, however, that the values shown in Fig. 2b are only for the upper 90 m and are not directly comparable to the full-depth heat transport given in these studies.

The largest lateral ocean heat flux component in our simulations is the inflow of warm Atlantic water through the Barents Sea Opening, which provides around 6–7 W m$^{-2}$ to the Arctic Ocean domain. Ocean heat transport through the Fram Strait is 2 W m$^{-2}$. Sea ice and snow export through the Fram Strait constitutes a positive heat flux of up to 2 W m$^{-2}$ in late winter. The ocean heat transport through the Canadian Archipelago is small (less than 2 W m$^{-2}$). The Bering Strait contributes a positive heat flux of up to 2 W m$^{-2}$ in summer but a negative heat flux of about −0.5 W m$^{-2}$ in winter, when the water entering the Arctic through the Bering Strait is at the seawater freezing temperature and hence colder than the reference temperature 0°C.

These ocean lateral heat transports appear reasonable but slightly too high when compared to observational estimates (see Table 1 in Beszczynska-Möller et al. 2011). For the Bering Strait, observational estimates of heat transport are 1–2 W m$^{-2}$ (albeit referenced to the seawater freezing temperature). For the Fram Strait and Barents Sea Opening, observations suggest full-depth heat transports of 3 and 5–7 W m$^{-2}$, respectively.

The monthly heat fluxes at the boundaries of the Arctic Ocean domain cause a corresponding monthly change in its heat content $\Delta Q$. It comprises changes in sensible heat content of the upper 90 m of ocean water $\Delta Q_{\text{sens}}$ and changes in latent heat content from changing amounts of sea ice and snow $\Delta Q_{\text{lat}}$:

$$\Delta Q = \Delta Q_{\text{sens}} + \Delta Q_{\text{lat}}.$$  \hspace{1cm} (2)

The heat content change over a seasonal cycle is roughly equipartitioned into sensible heat change of seawater and latent heat change of sea ice (Fig. 2c). Contributions of snow on sea ice are small by comparison. Each heat content change contribution on the right-hand side of Eq. (2) is strongly correlated with the combined heat content change on the left-hand side throughout the whole lead time of the ensemble predictions. On average, there is a correlation of 0.75 between $\Delta Q_{\text{lat}}$ and $\Delta Q$ and a correlation of 0.87 between $\Delta Q_{\text{sens}}$ and $\Delta Q$. This implies that the conclusions on the combined upper-ocean heat budget that we are going to derive are highly relevant both for upper-ocean water temperatures and for sea ice in the Arctic.
The central point of this study is to establish the contribution of ocean lateral heat transport and vertical heat fluxes at the surface and at depth to IFU of Arctic sea ice and upper-ocean heat content. In the same manner as for the monthly values considered in the previous section, heat flux into the domain accumulated since the start of the prediction must equal the heat content change within the domain:

$$\Delta Q + R = F = F_{\text{ocn}} + F_{\text{sfc}}.$$  \hspace{1cm} (3)

This is valid for every single member of a prediction ensemble at all times during the prediction. The term $R$ is the residual of the heat budget (i.e., the mismatch between the diagnosed heat content change and the diagnosed heat flux). In our simulations, there are non-negligible heat budget residuals, which remain even after careful consideration of all potentially contributing components. These residuals are discretization errors arising from approximating the monthly change of heat content by the monthly mean of heat content. Test simulations with a few ensemble members (not shown) clearly demonstrate that saving the instantaneous states at the beginning of each month and using them to diagnose monthly heat content change leads to negligible residuals. We note that this problem can be easily avoided in future studies by including the instantaneous beginning-of-month state in the model output. However, even the nonnegligible residuals we are working with in this study are small enough to allow robust attribution of heat content change to heat fluxes, as we will see in the following paragraph and the next section.

To illustrate the richness of possible trajectories and the relative magnitude of the residuals, we show in Fig. 3 the anomalies of surface heat flux and ocean heat flux accumulated since initialization together with the resulting heat content change for three example members of a single initial-condition ensemble. Ensemble member 1 has a consistent negative anomaly in ocean heat flux, while the atmospheric heat flux is smaller but nevertheless leaves seasonal imprints on the heat content change. Ensemble member 2 has a relatively weak ocean heat flux anomaly, and its heat content anomaly is almost exclusively determined by the atmospheric heat flux. Finally, heat content change in ensemble member 3 is very close to the ensemble mean, while atmosphere and ocean have large and opposing contributions. In all cases, residuals are small compared to the anomalies in heat content and heat flux that we want to analyze.

To quantify the IFU of the accumulated heat fluxes and heat content changes over many realizations as shown in Fig. 3, we calculate their average ensemble variance at each lead time. Because of the idealized setup of the experiments (see Tietsche et al. 2014), the ensemble variance of any simulated quantity $X$ directly measures the mean-square error (MSE) of the ensemble mean forecast and hence the IFU:

$$\text{var}(X) = \text{MSE}(X) = \text{IFU}(X).$$  \hspace{1cm} (4)

Calculating the variance of the heat budget [Eq. (3)], we see that any variance in accumulated heat flux within an ensemble can be attributed to one or more of the contributing heat fluxes together with their covariances:

$$\text{var}(\Delta Q) + \text{u.v.} = \text{var}(F) = \sum_{ij} \text{cov}(F_i, F_j).$$  \hspace{1cm} (5)

The sum on the right-hand side expands into the flux variances and covariances. It can be taken over each set

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**Fig. 3.** Example of accumulated heat flux and heat content change anomalies. Shown are three ensemble members that start from almost identical initial conditions. Anomalies are calculated with respect to the ensemble mean. Accumulated atmospheric surface heat flux $F_{\text{sfc}}$ is in blue, accumulated oceanic heat flux $F_{\text{ocn}}$ is in red, and heat content change $\Delta Q$ is in black. The gray dashed line is the total heat flux $F$. The difference between $\Delta Q$ and $F$ is the residual $R$. 

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of flux components introduced in Eq. (1) that adds up to give the total flux: either the set \( \{ F_{\text{sfc}}, F_{\text{ocn}} \} \) with two variance terms and one covariance term or the set \( \{ F_{\text{sfc}}, F_{\text{al}}, F_{\text{ov}} \} \) with three variance terms and three covariance terms. The term \( \text{u.v.} \) on the left-hand side is the unexplained variance ascribed to the residuals of the heat budget and is defined by \( \text{u.v.} = \text{var}(R) + 2 \text{cov}(\Delta Q, R) \).

Whereas the variances are by definition always positive, the covariances can be negative if the corresponding heat fluxes are anticorrelated. If all heat fluxes were uncorrelated, their variances would simply add to give the total forecast uncertainty of heat flux and hence forecast uncertainty of heat content change. Negative covariances will tend to reduce the forecast uncertainty; positive ones will increase it.

This decomposition provides us with the desired tool to estimate how important the different atmospheric and oceanic heat flux contributions in Eq. (3) are for irreducible forecast uncertainty of heat content in the Arctic Ocean domain. If, for example, at a particular lead time the term \( \text{var}(F_{\text{ocn}}) \) was small, the IFU of heat content would originate entirely from the atmospheric surface heat flux. If, on the other hand, there was a sizable contribution from \( \text{var}(F_{\text{ocn}}) \), it would indicate that ocean dynamics plays an important role in the prediction problem. Any strong covariance between \( F_{\text{sfc}} \) and \( F_{\text{ocn}} \) would indicate inherently coupled atmosphere–ocean–sea ice dynamics, with inextricably linked heat fluxes.

### 3. Heat flux contributions to irreducible forecast uncertainty

#### a. Mean across ensembles

The result of the above-described procedure of decomposing IFU of heat content into atmospheric and oceanic heat flux is shown in Fig. 4. For reference, we can convert heat variances into equivalent changes of ocean water temperatures or ice volume by performing a simple back-of-the-envelope calculation with the material constants for specific heat and density of ice and water. A typical heat variance of \( 5.5 \times 10^{15} \text{J}^2 \text{m}^{-4} \) is equivalent to a temperature change of 0.2 K in the upper 90 m or an ice volume change of 2500 km$^3$ in the Arctic Ocean domain. It is evident that IFU for Arctic Ocean domain heat content grows throughout the lead time of 3 years but does not do so at a constant rate; it tends to be higher in summer than in winter. This applies regardless of whether predictions are started in July (Fig. 4a) or in November (Fig. 4b). High IFU during summer is clearly dominated by the atmospheric contribution \( F_{\text{sfc}} \), but in winter the oceanic contribution \( F_{\text{ocn}} \) becomes important. The large seasonality of IFU is consistent with the notion that in the melt season, positive feedbacks acting in the Arctic amplify perturbations of the initial conditions, while in the freezing season, prevailing negative feedbacks lead to slower growth of initial condition perturbations and hence IFU (Tietsche et al. 2011).

The atmospheric contribution to IFU dominates in the first lead year (Figs. 4a,b). The oceanic contribution is almost as large as the atmospheric contribution in the second lead year and becomes larger in the third lead year. The contribution from the negative covariance between \( F_{\text{sfc}} \) and \( F_{\text{ocn}} \) is initially weak, but in the third lead year it is as large as either of the two alone and reduces IFU substantially. Hence, one must conclude that, in these simulations, atmospheric and oceanic heat fluxes are inextricably linked: whenever there is an unpredictable variation at the surface, a large part of it is compensated by a variation of ocean heat flux of the opposite sign. If this strong negative covariance is indeed a real feature of the Arctic surface climate, it implies that dynamical forecast systems cannot be skillful unless they simulate the heat exchange with deeper ocean layers and subarctic seas realistically. In a forecast system that underestimates oceanic heat exchange, even a perfect simulation of Arctic surface atmospheric heat fluxes would lead to erroneous forecasts because the resulting change in heat content would be too large. Conversely, if the system was somehow tuned to forecast the correct amount of heat content change, the simulated atmospheric surface heat fluxes would necessarily be too small.

The attribution of IFU during the first few months of the predictions started in July (Fig. 4a) deserves some further attention. Only for this particular case is it possible to attribute IFU predominantly to the failure to predict atmospheric surface heat fluxes because oceanic heat flux variance is negligible. This result suggests that domainwide predictions of the surface Arctic Ocean (including sea ice) from summer to autumn are—in principle—only a problem of getting the atmospheric surface heat fluxes right, while internal ocean heat fluxes are of little importance. Note that this is only true under the idealizing assumptions made here. In practice, the initial-state uncertainty of the poorly observed Arctic Ocean and sea ice as well as model errors always causes erroneous ocean heat fluxes that degrade forecast quality.

By considering the two components of the ocean heat flux \( F_{\text{al}} \) and \( F_{\text{ov}} \) separately, it becomes evident that the strong negative covariance between \( F_{\text{sfc}} \) and \( F_{\text{ocn}} \) is mainly due to the strong negative covariance between \( F_{\text{sfc}} \) and \( F_{\text{ov}} \) (Figs. 4c,d). Their negative covariance develops in winter and stays roughly constant over summer. Although our methodology does not allow
specifying the process that is responsible for the vertical heat exchange, we suggest that it is mainly the deepening of the surface mixed layer and the corresponding entrainment of warm water from below during the freezing season, as discussed by Tietsche et al. (2013a). This will predominantly happen in regions where the mixed layer of the model is seasonally deep enough to cross the chosen lower boundary of the domain at 90-m depth (Barents and Kara Seas and directly north of Svalbard and Franz Josef Land). It might also be that Ekman pumping plays a significant role regionally, as suggested by Yang (2009).

From Figs. 4c,d we also see that the lateral ocean heat transport $F_{ol}$ has small but nonnegligible covariance with the vertical ocean heat flux $F_{ov}$ and the surface heat flux $F_{sfc}$. In the third lead year, the covariance between $F_{sfc}$ and $F_{ol}$ tends to be negative, which is reminiscent of the Bjerknes compensation known to be active at longer time scales (e.g., Shaffrey and Sutton 2006). Note that the variances of $F_{ov}$ and $F_{ol}$ are very similar. This, of course, is due to the particular choice of 90 m as the lower boundary—a shallower choice would result in the vertical heat flux being more important than the lateral heat transport, and a deeper choice would make
the vertical heat flux less important than the lateral transport.

How would the statements made in this section change if an ocean depth different from 90 m was chosen as the lower boundary of the domain? Conceptually, it is straightforward to repeat the same calculations for different lower boundaries. To illustrate the key outcomes of these calculations, we show in Fig. 5 the heat flux decomposition at lead month 34 (September) for ensembles started in November as a function of the chosen bottom boundary ocean depth.

The depth dependence of the heat flux decomposition is nontrivial, but we discern three broad regimes. In the surface regime, \( F_{sfc} \) dominates, and the contribution of \( F_{ol} \) is larger than that of \( F_{ov} \). However, with increasing depth, \( F_{ol} \) and its covariances become gradually more important. As argued in section 2d, the chosen depth of 90 m limits this surface regime, and for this depth var(\( F_{ov} \)) becomes equal to var(\( F_{ol} \)). Below 90 m, the variance of \( F_{ov} \) rises sharply, becoming larger than the variance of \( F_{sfc} \) at roughly 200-m depth. At the same time, the coupling with the surface becomes weaker, as evident from a continued decline of the covariance between \( F_{sfc} \) and \( F_{ov} \). At around 400-m depth, this covariance is zero, whereas the variance of \( F_{ol} \) reaches a maximum. This is the Atlantic layer regime, where internal ocean processes dominate. If the lower boundary of the domain chosen is even deeper than that, we reach the full-depth regime: \( F_{ol} \) stabilizes at slightly lower levels, while \( F_{ov} \) together with its covariances goes to zero.

The intent of this study is to investigate the importance of atmospheric and oceanic heat flux contributions for predictions of the surface climate of the Arctic. Hence, we are interested only in the surface regime, which justifies the particular choice of 90 m as the lower boundary. Had we chosen a shallower lower boundary, qualitative conclusions elsewhere in this manuscript would have remained the same, with, however, increased importance of \( F_{ov} \) and decreased importance of \( F_{ol} \). Had we chosen a deeper lower boundary, conclusions about the contributions of oceanic and atmospheric heat fluxes would have been different, but at the same time these conclusions would have been less relevant for Arctic surface conditions.

b. Initial-condition dependence

The discussion in the previous section was concerned with the partitioning of the heat fluxes when averaged across ensembles with different initial conditions. However, it is possible to use Eq. (5) without averaging, which results in one decomposition for each ensemble and gives insight into the initial-condition dependence of the statements in the previous section.

To characterize this initial-condition dependence, we plot the middle tercile of the decompositions of the November ensembles at each lead time in Fig. 6a. Note that this is in complete analogy to Fig. 4b, only that we now show the spread instead of the mean of ensemble variances. Although there is a considerable amount of variation across the start years, the qualitative conclusions drawn in the previous section remain valid: for the first lead year, the surface heat flux dominates; after that, the oceanic heat flux catches up; and the negative covariance between the two is of the same order as the individual heat fluxes.

There is substantial initial-condition dependence in the magnitude of heat budget IFU. For example, at a lead time of 34 months for the ensembles started in November it ranges from 0.5 to \( 7 \times 10^5 \) J m\(^{-2}\) (Fig. 6b). However, years with large atmosphere and ocean heat flux IFU tend to also have strong negative covariance between the two. As a consequence, the location of the curves for each start year within the uncertainty bands depicted in Fig. 6a is not random, but curves tend to move away or toward the zero line in unison. Thus, despite significant initial-state dependence of the overall magnitude of IFU, the relative contributions of atmospheric and oceanic heat fluxes are a robust feature across different initial states.
c. Atmospheric controls on the ocean heat flux uncertainty

Here, we want to corroborate our statements from section 3a about the physical processes that might be responsible for the diagnosed contributions of atmospheric and ocean heat fluxes to heat budget anomalies. Correlations between monthly mean (not accumulated) quantities are a good diagnostic to do that. Hints about the physical processes involved can be derived from seasonal dependence and from the presence or absence of a change over lead time; while a (cyclo-)stationary correlation suggests fast processes that act on shorter-than-monthly time scales, a correlation that gradually changes over the lead time suggests that slow processes are involved, which take several months or even years to develop an effect.

Motivated by the dominance of the negative covariance between accumulated $F_{sfc}$ and $F_{ov}$, we start by looking at the correlation of their monthly mean values. Figure 7a shows that the correlation between monthly mean $F_{sfc}$ and $F_{ov}$ has a strong but stationary seasonal cycle. There is strong negative correlation already in the first winter months after the ensemble start date in November. With the beginning of the melt season, the correlation becomes zero and eventually slightly positive later in summer before it becomes strongly negative in the second winter at the same value as in the first winter. This seasonal and stationary signature is strongly suggestive of buoyancy forcing at the surface being the fast-acting physical process responsible. In winter, cooling at the surface leads to buoyancy loss, surface mixed layer deepening and entrainment of warm water from below, whereas in summer, heating at the surface leads to shallowing of the mixed layer so that heat exchange at 90-m depth is inhibited and the correlation between $F_{sfc}$ and $F_{ov}$ is weak. This supports the hypothesis that the variability in $F_{ov}$ is to a large part driven by atmospheric variability in $F_{sfc}$ during wintertime.

The second process we want to investigate is the ocean heat transport through the Arctic gateways, especially the Fram Strait and the Barents Sea Opening. It has often been argued that these transports are predominantly wind driven (e.g., Schlichtholz 2011). We follow the method of Koenigk and Brodeau (2014) and decompose the ocean heat transport anomalies into velocity and temperature contributions. A dominance of the former would suggest fast wind-driven dynamics, while the latter would point toward slow internal ocean dynamics.

The ocean heat transport OHT is the product of the ocean velocity $V$ and the temperature departure $T$ from the reference temperature: $OHT = VT$. For any given ensemble member, the departure of ocean heat transport from the ensemble mean, given by $OHT' = OHT - \langle OHT \rangle$, can be decomposed into a contribution from temperature anomalies $T' = T - \langle T \rangle$, velocity anomalies $V' = V - \langle V \rangle$, and an inseparable product of the two, $V'T'$:

$$OHT' = V'\langle T \rangle + \langle V \rangle T' + V'T' - \langle V'T' \rangle,$$  

where $\langle \cdot \rangle$ denotes the ensemble mean. We find that the product component $V'T' - \langle V'T' \rangle$ is always small with respect to the other two components in our simulations, so we do not discuss it further. To attribute IFU of ocean heat transport to either temperature anomalies or velocity anomalies, one can then write the following:

$$\text{var}(OHT') = \text{var}(V'\langle T \rangle) + \text{var}(\langle V \rangle T') + 2\text{cov}(V'\langle T \rangle, \langle V \rangle T'),$$  

where

\begin{align*}
\text{var}(OHT') &= \text{var}(V'\langle T \rangle) + \text{var}(\langle V \rangle T') + 2\text{cov}(V'\langle T \rangle, \langle V \rangle T'), \\
&= \text{var}(V'\langle T \rangle) + \text{var}(\langle V \rangle T'),
\end{align*}
The relative contributions of the three terms on the right-hand side of Eq. (7) are shown in Figs. 7b,c. For the Barents Sea Opening (Fig. 7b), velocity anomalies dominate, explaining all of the ocean heat transport variance initially and then levelling off at explaining about 80% of the variance throughout the lead time. Correlation of Barents Sea Opening ocean heat transport anomalies with the NAO index (Fig. 7a) suggests that the velocity anomalies are indeed wind driven.

For the Fram Strait (Fig. 7c), the situation is different. Velocity anomalies are still the largest contribution to the ocean heat transport anomaly, but temperature anomalies do play a growing role over lead time, together with a strong negative covariance term. This suggests that several factors, not only the local winds, play a role in setting up the Fram Strait heat transport. This is plausible given the fact that in the Fram Strait, other than in the Barents Sea Opening, there is a transport component out of the Arctic (the East Greenland Current), and sea ice shields the water from wind stress for part of the year.

Further evidence that—on seasonal time scales—the large-scale atmospheric surface circulation regime is important for Arctic heat flux anomalies can be provided by regressions of sea level pressure on the heat fluxes into the Arctic Ocean domain. These regressions are shown in Fig. 8 for the first winter season [December–February (DJF)] of prediction ensembles started in November. Figure 8a indicates that reduced heat loss of the ocean through the surface \( F_{\text{sc}} \) is associated with low SLP over the Arctic Ocean and Nordic Seas. With this SLP anomaly pattern, the prevailing surface winds are northerly over the North Atlantic and westerly along the Siberian Arctic Ocean coast. This is consistent with warm Arctic Ocean surface air conditions because warm air over the North Atlantic is advected into and circulated around in the domain. At the same time, low pressure favors cloudy conditions leading to reduced longwave heat loss at the surface.

The SLP regression on the vertical ocean heat flux at depth \( F_{\text{ov}} \) has a pattern that is similar to the \( F_{\text{sc}} \) regression but has the opposite sign (Fig. 8b). This corroborates the hypothesis that the strong negative covariance between \( F_{\text{sc}} \) and \( F_{\text{ov}} \) in winter is tightly linked to the prevailing atmospheric surface circulation; warm surface conditions will lead to less buoyancy loss and hence less vertical mixing in the upper ocean. Additionally, Ekman pumping might play a role, as suggested by Yang (2009); the zonal wind anomalies along the Siberian and North American coasts of the Arctic Ocean tend to be stronger in the SLP/\( F_{\text{ov}} \) regression than they are in the SLP/\( F_{\text{sc}} \) regression. In the case of anomalous high pressure over the Arctic Ocean, the increased easterly winds push surface water and ice away from the coast and bring more warm deep water close to the surface by Ekman pumping, which increases \( F_{\text{ov}} \).

Finally, anomalously high ocean heat transport through the Barents Sea Opening \( F_{\text{BSO}} \) is associated with strong southerly surface winds enhancing the Ekman transport across the Opening (Fig. 8c). Thus, in our simulations, Barents Sea Opening heat transport anomalies are mainly driven by ocean current velocity.
anomalies (Fig. 7b), which in turn are caused by anomalous surface winds.

d. Regional contrasts

Heat fluxes and heat content changes across the whole Arctic domain are of scientific interest, but might be of little relevance to users of climate forecasts who require information for a specific location. Here, we provide spatial maps of heat budget terms at a few selected lead times as well as the temporal evolution of heat budget terms over the full lead time for a few selected representative locations. This also allows a better understanding of the physical processes since spatial averaging over regions with different and unrelated characteristic processes is avoided.

Figure 9 shows the same heat budget decomposition as in Fig. 4 but on a gridpoint level and for fixed lead times. For reference, we can convert typical heat content changes to equivalent changes in temperature and ice thickness with a simple back-of-the-envelope calculation: an ensemble variance of $9 \times 10^{16} \text{J} \cdot \text{m}^{-4}$ corresponds to an IFU of 1 m in sea ice thickness or 0.8 K in ocean temperature in the upper 90 m. Note that here the lateral ocean heat flux $F_{\text{ol}}$ is not the sum of latent and sensible heat transport through the Arctic gateways as before but instead a local heat transport convergence caused by the local advection of snow and ice and local ocean currents in the upper 90 m. Because of limitations in the available diagnostics, we are able to produce a map for only the total ocean heat flux $F_{\text{ocn}} = F_{\text{ol}} + F_{\text{ov}}$, not for the components $F_{\text{ol}}$ and $F_{\text{ov}}$ individually.

The conclusions from the domainwide numbers (Fig. 4) are broadly confirmed by the local heat budget decomposition shown in Fig. 9: contributions from the atmospheric surface heat flux $F_{\text{sfc}}$ and oceanic heat flux $F_{\text{ocn}}$ to IFU are of similar magnitude, and their covariance is negative and strong. However, there are clear exceptions for some regions and lead times. For instance, the heat budget of the coastal region of the Beaufort Sea from November to March is clearly dominated by $F_{\text{ocn}}$, and there is a positive covariance between $F_{\text{sfc}}$ and $F_{\text{ocn}}$ in July over large parts of the Arctic Ocean.

Everywhere but in the central Arctic, the IFU of the local (gridpoint) heat budget is an order of magnitude larger than the IFU of the domainwide heat budget. This points to the important role of local ocean currents and ice advection in redistributing heat within the Arctic domain, a process that does not contribute to the domainwide heat budget. The central Arctic with its perennial ice cover stands out as the region with the smallest IFU. In the seasonal sea ice zone represented by the Beaufort, Chukchi, East Siberian, and Laptev Seas, IFU is elevated. Finally, the Barents and Kara Seas as well as the region northeast of the Fram Strait exhibit IFU that is up to two orders of magnitude larger than in the central Arctic.

These strong regional contrasts within the Arctic Ocean domain are related to the different processes that characterize the regions: the Fram Strait, Barents Sea, and Kara Sea regions are heavily influenced by warm Atlantic water inflow; the ocean mixed layer is typically deeper than the lower domain boundary at 90-m depth; and large parts of these regions are open water even in winter. This means all three heat flux components can vary much more than in the central Arctic, where a
perennial sea ice cover prevents large surface heat fluxes, a shallow ocean mixed layer prevents vertical heat exchange at 90-m depth, and spatially homogeneous surface conditions and sluggish ocean currents keep lateral heat fluxes small. Finally, intermediate levels of IFU in the marginal sea ice zones along the shelf seas of the Arctic Ocean are to be expected since these regions experience a mixture between ice-covered periods with inhibited heat fluxes and ice-free periods with enhanced heat fluxes.

The temporal evolution of the heat budget for a selection of representative locations is shown in Fig. 10. Note the very different scales in the vertical axes, and how they compare to the scales in Fig. 4. For all regions shown here except for the Barents Sea, the ocean contribution is now larger than the atmospheric surface contribution at all lead times. This is in contrast to the domainwide result, where the surface heat flux tends to be larger, especially at subannual lead times. In all marginal Arctic seas except for the Barents and Kara Seas, there is strong seasonality of the heat content IFU; it is high in summer and low in winter. This is consistent with the positive ice–albedo effect amplifying differences in the initial state at the beginning of summer, whereas with the onset of freezing negative feedbacks dominate and reduce IFU again (see Tietsche...
et al. 2011). However, in the Laptev and Beaufort Seas (Figs. 10c,e) the increased $F_{\text{sfc}}$ in summer associated with the ice–albedo feedback is only part of the reason for elevated levels of IFU. There is a contribution from the strong seasonality in the covariance between ocean and atmosphere heat fluxes. As suggested already in Fig. 9, the marginal Arctic seas except for the Barents and Kara Seas have a positive covariance between $F_{\text{sfc}}$ and $F_{\text{ocn}}$ during the melt season, in contrast to the overall strong negative covariance. This fact might explain why predictions of Arctic summer sea ice are particularly challenging: a forecast error at the surface will be reinforced by internal oceanic heat fluxes.

Further research with more explicit diagnostics for the vertical ocean heat flux at depth would be necessary to better understand the processes responsible for this positive summer covariance. Our present hypothesis is that cyclonic activity in summer has a double cooling effect on the surface waters: cool and overcast conditions reduce surface heating, and at the same time strong winds can bring water from depth to the surface through wind-driven mixing and Ekman pumping. Because in summer the water at the ice-free surface becomes warmer than water at depth, vertical exchange leads to a cooling of the surface. This would then lead to the diagnosed positive covariance between $F_{\text{sfc}}$ and $F_{\text{ocn}}$.

In summary, the IFU of heat content change on a gridpoint level exhibits characteristic regional differences within the Arctic domain. Nevertheless, its decomposition into heat fluxes is similar to that for the whole domain: relative contributions of surface atmospheric and internal ocean heat flux are of comparable magnitude, with strong negative covariance between the two. For the local heat budget, the internal ocean heat flux

![Graphs showing decomposition of IFU into atmospheric and oceanic heat fluxes](image-url)
flux tends to be more important than the surface atmospheric heat flux at all lead times.

4. Conclusions

Here, we used idealized ensemble predictions with the Earth system model MPI-ESM to determine where irreducible forecast uncertainties in Arctic surface climate originate. By constructing and closing the heat budget of the tightly coupled sea ice and upper-ocean waters, we can attribute unpredictable variations either to the atmospheric surface heat flux or to lateral and vertical ocean heat fluxes. The results presented here depend to some degree on the model used, so we would like to encourage similar analyses with different Earth system models.

We find that irreducible forecast uncertainty of Arctic surface climate is equally attributable to the uncertainty in predicting the atmospheric surface heat flux and the oceanic heat flux. Hence, the poorly observed state of the subsurface Arctic Ocean might be more important for seasonal to interannual predictions of Arctic surface climate than previously thought, even despite strong stratification and slow surface currents. This might explain why current forecasting systems perform poorly and inconsistently in the Arctic.

Unpredictable variations in atmospheric surface heat flux are strongly anticorrelated with vertical ocean heat flux anomalies at 90-m depth. This is probably due to entrainment of warm water from below during mixed layer deepening and Ekman pumping. Hence, a large fraction of surface heat flux anomalies does not contribute to heat content anomalies at the surface but is deposited in deeper ocean layers. To avoid “getting the right results for the wrong reasons,” current forecasting systems likely need a better representation of vertical heat exchange in the upper Arctic Ocean to improve forecast quality further.

Likewise, the appreciable fraction of irreducible forecast uncertainty from mainly wind-driven heat transport anomalies through the Arctic gateways indicates that initialization and realistic advection of surface heat content anomalies in the North Pacific and North Atlantic are important for predictions on longer-than-seasonal time scales.

The relative contributions of atmospheric and oceanic heat flux are modulated by season, lead time, and location within the Arctic Ocean domain. For Arctic-wide surface heat flux changes, the atmospheric surface heat flux uncertainty tends to dominate for seasonal lead times below a year, but oceanic heat flux uncertainty becomes more important on interannual lead times. For the local heat budget, ocean heat flux uncertainty dominates at all lead times in most places. Hence, when moving toward more refined, user-oriented Arctic predictions, it is important to appreciate that each season, lead time, and location will have its own characteristic pathways through which predictability is lost, and hence the physical processes requiring attention will be different accordingly.

Finally, there is substantial variation of irreducible forecast uncertainty across different initial states of the ensembles. This points to the important role of ocean and sea ice initial conditions for seasonal to interannual predictions of Arctic surface climate and hence the compelling need for better sea-ice and subsurface observations in the Arctic Ocean.

Acknowledgments. This research was supported by the APPOSITE project (Grant NE/I029447/1) funded by the U.K. Natural Environment Research Council. Model output from the simulations is openly available from the British Atmospheric Data Centre. We thank Jon Robson, David Ferreira, Camille Lique, and Clara Deser for valuable discussions. All simulations were performed at the German Climate Computing Center (DKRZ) in Hamburg, Germany, with the kind support of Johanna Baehr and Daniela Matei.

REFERENCES


