Atmosphere–Ocean Interactions and Their Footprint on Heat Transport Variability in the Northern Hemisphere

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ABSTRACT

Interactions between the atmosphere and ocean play a crucial role in redistributing energy, thereby maintaining the energy balance of the climate system. Here, we examine the compensation between the atmosphere and ocean's heat transport variations. Motivated by previous studies with mostly numerical climate models, this so-called Bjerknes compensation is studied using reanalysis datasets. We find that atmospheric energy transport (AMET) and oceanic energy transport (OMET) variability generally agree well among the reanalysis datasets. With multiple reanalysis products, we show that Bjerknes compensation is present at almost all latitudes from 40° to 70°N in the Northern Hemisphere from interannual to decadal time scales. The compensation rates peak at different latitudes across different time scales, but they are always located in the subtropical and subpolar regions. Unlike some experiments with numerical climate models, which attribute the compensation to the variation of transient eddy transports in response to the changes of OMET at multidecadal time scales, we find that the response of mean flow to the OMET variability leads to the Bjerknes compensation, and thus the shift of the Ferrel cell at midlatitudes at decadal time scales in winter. This cell itself is driven by the eddy momentum flux. The oceanic response to AMET variations is primarily wind driven. In summer, there is hardly any compensation and the proposed mechanism is not applicable. Given the short historical records, we cannot determine whether the ocean drives the atmospheric variations or the reverse.

1. Introduction

Interactions between the atmosphere and ocean play a crucial role in redistributing energy, thereby maintaining the energy balance of the climate system. Here, we study the compensation between the atmosphere's and ocean's heat transport variations. It influences large-

scale dynamics by reshaping the energy budget. Many efforts have been made to determine heat transport. In the 1960s, Jacob Bjerknes suggested that energy transport in the atmosphere and ocean should compensate each other if the variations of heat fluxes at the top of the atmosphere (TOAFlux) and the ocean heat content (OHC) are small (Bjerknes 1964). This process has come to be known as Bjerknes compensation.

However, since the variations of total energy are small from the interannual time scales onward (Shaffrey and Sutton 2006), Bjerknes compensation is mostly studied using numerical climate models, which makes it possible

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to generate relatively long time series (Shaffrey and Sutton 2004, 2006; Van der Swaluw et al. 2007; Jungclaus and Koenigk 2010; Farneti and Vallis 2013; Van der Linden et al. 2016; Outten and Esau 2017). Shaffrey and Sutton (2004) reported an anticorrelation between the atmospheric meridional energy transport (AMET) and oceanic energy transport (OMET) variations in the North Atlantic Ocean at midlatitudes with the HadCM3. They found that variability of OMET in the North Atlantic is dominated by the wind-driven Ekman processes, while AMET in the Northern Hemisphere appears to be governed primarily by changes in stationary waves at midlatitudes across interannual time scales. In their following work with the same model, Shaffrey and Sutton (2006) further claimed that AMET and OMET anomalies are significantly anticorrelated in the extratropical Northern Hemisphere at decadal time scales. They concluded that positive OMET anomalies could reduce atmospheric baroclinicity and hence lead to weakened AMET transport, primarily by transient eddies. It also causes weakened stationary eddy transport, but this is less pronounced than the variations caused by transient eddies.

Those findings are further confirmed by Van der Swaluw et al. (2007) as they also found that compensation between atmosphere and ocean can be attributed to changes in transient eddies with varying OMET at multidecadal time scales in their numerical experiments with the HadCM3. They emphasized the coupling between OMET in the Atlantic Ocean and the Atlantic meridional overturning circulation (AMOC). Moreover, they argued that there is a link between compensation and the variability of the Arctic sea ice. Van der Linden et al. (2016) demonstrated that the link between sea ice cover variations in the Barents Sea and ocean-atmosphere heat exchanges originates from the low-frequency changes in the ocean, which supports the statement from Van der Swaluw et al. (2007) about the link between sea ice and Bjerknes compensation. Jungclaus and Koenigk (2010) investigated Bjerknes compensation at decadal time scales with long time series from simulations with the ECHAM5/MPI-OM. They also found that in their model the compensation peaks at 70°N. Both Van der Swaluw et al. (2007) and Jungclaus and Koenigk (2010) agree that the compensation reaches a maximum when the ocean leads the atmosphere. With the GFDL climate model, Farneti and Vallis (2013) detected a strong compensation between AMET and OMET in the North Atlantic at decadal time scales. They confirmed the crucial role of AMOC in generating OMET anomalies as well as the Bjerknes compensation.

Apart from studies using fully coupled general circulation models, some researchers also investigated the

mechanism of compensation using simplified mathematical models (Yang et al. 2013, 2018). By conducting experiments with the Fast Ocean Atmosphere Model, Yang et al. (2013) found that AMET and OMET compensate very well in the tropics and extratropics driven by the freshwater input in the high-latitude Atlantic. Similarly, Yang et al. (2018) found a stabilizing role of Bjerknes compensation caused by the freshwater influx in the North Atlantic using a simple box model.

Recently an intercomparison of Bjerknes compensation found in different numerical models shows that the results are model dependent (Outten et al. 2018). Given the differences in representing Bjerknes compensation in the numerical models, we are interested in the interaction between atmosphere and ocean, and Bjerknes feedback in the historical climate record. Serving as a good representation of the historical climate, reanalyses allow us to look for the Bjerknes compensation in a state that is close to reality. Note that these reanalysis products incorporate the numerical models that fail to capture the robust compensation response (Outten et al. 2018), but they are constrained by observations. Nowadays, the time span covered by most reanalysis products makes it possible to examine the response and feedback from atmosphere to ocean at interannual to decadal time scales (Dee et al. 2011; Ferry et al. 2012a; Balmaseda et al. 2013; Harada et al. 2016; Gelaro et al. 2017; Carton et al. 2018). But the time series are not long enough to account for the multidecadal variability of energy transports, which are of interest to many studies with numerical models.

In this paper, for the first time, Bjerknes compensation is studied using multiple atmospheric and oceanic reanalysis datasets. We quantify and intercompare meridional energy transport (MET) in the atmosphere and ocean with six state-of-the-art reanalysis products. With these reanalysis products, Bjerknes compensation is investigated from interannual to decadal time scales. Moreover, we are interested in the energy transport variability and their influence on the circulation of the atmosphere and ocean. The atmospheric response and potential feedback to the variability of the ocean are illustrated in detail. We examine the role of the ocean in air-sea interactions and explore the mechanism of Bjerknes compensation. The large-scale climate variability associated with the interactions between the atmosphere and ocean is also examined in this study.

The paper is organized as follows. The reanalysis datasets used in this study are listed in section 2, data and methodology. In this section, we elaborate on the methods for the quantification of AMET and OMET. A description of the statistical tests is also included in this section. The calculated AMET and OMET and an

TABLE 1. Basic specification of reanalysis products included in this study.

Type	Product name	Producer	Period	Temporal resolution	Spatial resolution/grid
Atmosphere	ERA-Interim	ECMWF	1979–2016	Daily and monthly	TL255, L60 up to 0.1 hPa
	MERRA-2	NASA	1980–2016	Daily and monthly	0.5° × 0.625°, L72 up to 0.01 hPa
	JRA-55	JMA	1979–2016	Daily and monthly	TL319, L60 up to 0.1 hPa
Ocean	ORAS4	ECMWF	1979–2016	Monthly	ORCA1
	GLORYS2V3	Mercator-Ocean	1993–2014	Monthly	ORCA025
	SODA3	University of Maryland	1980–2015	5-daily	MOM5

overview of Bjerknes compensation are shown in [section 3](#), results, together with our exploration of the physical mechanism of compensation and the atmospheric response and feedback to the ocean. In [section 4](#), conclusions and discussion, we summarize this study and provide suggestions and a perspective for future work.

2. Data and methodology

In this section, we introduce six reanalysis datasets used in this study. In addition, we elaborate on our methodology for the computation of AMET and OMET, and the statistical analysis performed in this work.

a. Reanalysis datasets

Unlike most of the previous studies that took output from climate models, we investigate Bjerknes compensation within reanalysis datasets. Six reanalysis products are included in this paper: three atmospheric reanalysis products [European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim); Modern-Era Retrospective Analysis for Research and Applications, version 2 (MERRA-2); and Japanese 55-year Reanalysis (JRA-55)] and three oceanic reanalysis products [Ocean Reanalysis System 4 (ORAS4); Global Eddy-Permitting Ocean Reanalysis, version 3 (GLORYS2V3); and Simple Ocean Data Assimilation (SODA), version 3 (SODA3)]. A synthesis of the basic specification of the reanalysis products used in this study is provided in [Table 1](#). We chose these six reanalysis products because they are the state-of-the-art atmospheric and oceanic reanalyses incorporating the latest weather forecast models and data assimilation schemes. Although there is overlap between the assimilated observations in the chosen products, the inter-comparison between them is still fruitful as most of them were built upon different numerical models and data assimilation techniques. For instance, ERA-Interim is based on the ECMWF's Integrated Forecast System (IFS) ([Dee et al. 2011](#)), while MERRA-2 is based on the Goddard Earth Observing System (GEOS) ([Gelaro et al. 2017](#)). JRA-55 generates data with an advanced weather prediction system from the Japan Meteorological

Agency (JMA; [Kobayashi et al. 2015](#); [Harada et al. 2016](#)). The numerical schemes and data assimilation systems are also different among the chosen oceanic reanalysis products. More details are provided in [Table 1](#) and the following introduction of reanalysis products.

Note that the latest atmospheric reanalysis, ERA5 from ECMWF, is not included here since the model-level data had not been opened to the public at the time this analysis was conducted ([ECMWF 2017](#)). Besides, with ERA5 the computation is too expensive to achieve time series long enough to account for the interannual to decadal variability of AMET transported by the eddies. The calculations are as much as possible based on data from model grids, so as to avoid interpolation errors and imbalances in the mass budget introduced by regridding. A brief introduction about the six reanalysis datasets is given below.

1) ERA-INTERIM

Serving as one of the most used global reanalysis products, ERA-Interim is a reanalysis dataset produced by ECMWF ([Dee et al. 2011](#)). It has been operating since the beginning of the “satellite era” starting in 1979, the implementation of satellites providing much more data than before. It is based on cycle 31r2 of ECMWF's IFS and generates data using four-dimensional variational analysis (4D-Var) assimilation with a T255 (~79 km) horizontal resolution on 60 vertical levels ([Berrisford et al. 2009](#)). The system includes a 4D-Var with a 12-h analysis window. We use the daily and monthly data on the model grid with a $0.75^\circ \times 0.75^\circ$ horizontal resolution spanning from 1979 to 2016.

2) MERRA-2

MERRA-2 ([Gelaro et al. 2017](#)) is an advanced reanalysis product operated by the Global Modeling and Assimilation Office (GMAO) of the National Aeronautics and Space Administration (NASA), which assimilates observational data with the GEOS model and analysis scheme ([Molod et al. 2015](#); [Gelaro et al. 2017](#)). The data are produced by 3D-Var assimilation and have a temporal coverage from 1980 to the present. We use the daily and

monthly data on the model grid with a resolution of $0.5^\circ \times 0.625^\circ$ from 1980 to 2016.

3) JRA-55

JRA-55 is the second reanalysis product made by the JMA (Kobayashi et al. 2015; Harada et al. 2016). It is based on the JMA operational numerical weather prediction (NWP) system and a 4D-Var assimilation system. Although the product has a wide coverage dating back to 1958, we only use the data from 1979 onward, as it is the “satellite era.” The daily and monthly mean fields with a horizontal resolution of $0.5625^\circ \times 0.5625^\circ$ from 1979 to 2015 are employed in this study.

4) ORAS4

ORAS4 is an oceanic reanalysis system generated by ECMWF (Balmaseda et al. 2013). It implements Nucleus for European Modelling of the Ocean (NEMO) as an ocean model (Madec 2008; Ferry et al. 2012b) and uses the NEMO variational data assimilation system (NEMOVAR) as the data assimilation system (Mogensen et al. 2012). The model is forced by atmosphere-derived daily surface fluxes, from ERA-40 from 1957 to 1989 and ERA-Interim-derived fluxes from 1989 onward. ORAS4 produces data with 3D-Var assimilation and spans 1958 to the present. However, due to the uncertainties reported by Balmaseda et al. (2013), we only use the data from 1979 onward. ORAS4 runs on the ORCA1 grid. To avoid interpolation error, we use the monthly mean fields on the native model grid with a horizontal resolution of about 1° on 42 vertical levels from 1979 to 2016.

5) GLORYS2V3

GLORYS2V3 is a global ocean and sea ice eddy-permitting reanalysis created by the Mercator Ocean, the Drakkar Consortium, and Coriolis Data Centre (Ferry et al. 2010, 2012a). It spans the altimeter and Argo eras from 1993 onward and is built on top of the NEMO ocean model with ORCA025 grid. The model is forced by surface fluxes using the ERA-Interim atmospheric near-surface parameters. A 3D-Var assimilation scheme is applied for assimilating temperature and salinity profiles from the CORA3.3 database (Ferry et al. 2012a). We use the monthly fields on the native model grid with a horizontal resolution of about 0.25° from 1993 to 2014.

6) SODA3

SODA3 is a reanalysis product made by the University of Maryland (Carton et al. 2018). It is in production since 1980. SODA3 is built on the Modular Ocean Model, version 5 (MOM5), ocean component of the Geophysical

Fluid Dynamics Laboratory CM2.5 coupled model (Delworth et al. 2012). To be consistent, we take the version forced by ERA-Interim, which is SODA3.4.1. The 5-daily data from 1980 to 2015 on the original model grid with a horizontal resolution of about 0.25° are used in this study.

More details about the chosen reanalysis datasets are given by Liu et al. (2020).

b. Methodology

The methods for the computation of AMET and OMET are given in this subsection, together with the statistical analysis performed in this study.

1) ENERGY TRANSPORT IN THE ATMOSPHERE AND EDDY DECOMPOSITION

Following Magnusdottir and Saravannan (1999), Shaffrey and Sutton (2006), and Van der Waluw et al. (2007), we calculate the zonal integral of AMET at each latitude as the residual between the zonally integrated net surface flux (SFlux) from the ocean into the atmosphere (F_{surface}), and the downward net radiation flux at the top of the atmosphere (F_{TOA}). The equation is shown below:

$$\frac{1}{a \cos \theta} \frac{\partial}{\partial \theta} (\cos \theta E) = \int_0^{2\pi} (F_{\text{surface}} + F_{\text{TOA}}) a \cos \theta d\phi, \quad (1)$$

where $\partial/\partial\theta(\cos\theta E)$ is the divergence of global AMET (PW), a is the Earth radius m , θ denotes the latitude, and ϕ indicates the longitude. The calculation of AMET was performed on monthly mean fields with chosen reanalysis products.

For the eddy decomposition, we follow the mathematical definition given by Peixoto and Oort [1992; see chapter 4.1 and Eq. (4.9)]. We compute the transient eddies and stationary eddies on selected pressure levels with daily mean fields. The eddy decomposition is performed on each component of atmospheric heat transport separately in order to calculate the total heat transport by eddies. The total meridional heat transport in the atmosphere consists of four components: internal energy transport I , latent heat transport H , geopotential energy transport Φ , and kinetic energy transport k . They are defined as

$$\begin{aligned} I &= c_p T v \\ H &= L_v q v \\ \Phi &= g z v \\ k &= \frac{1}{2} \mathbf{v} \cdot \mathbf{v} v, \end{aligned} \quad (2)$$

where c_p is the specific heat capacity of dry air at a constant pressure ($\text{J kg}^{-1} \text{K}^{-1}$), T is the absolute temperature (K), L_v is the specific heat of condensation (J kg^{-1}), q is the specific humidity kg kg^{-1} , g is the gravitational acceleration $\text{kg m}^{-1} \text{s}^{-2}$, z is the altitude (m), $\mathbf{v}$ is the zonal–meridional wind velocity (m s^{-1}), and v is the meridional wind velocity (m s^{-1}). The northward propagation is positive. A constant value of $c_p = 1004.64 \text{ J kg}^{-1} \text{K}^{-1}$ was used to compute the AMET.

Taking the internal energy transport as an example, the decomposition is as follows (Peixoto and Oort 1992; see Eq. (4.10)):

$$c_p [\overline{vT}] = c_p [\overline{v}][\overline{T}] + c_p [\overline{v}^*][\overline{T}^*] + c_p [\overline{v'T'}], \quad (3)$$

where $[v]$ is the zonal average of meridional wind velocity, $\overline{v}$ is the temporal average, v^* is the departure from zonal mean, and v' is the departure from temporal mean; similar notation is used for temperature T . The total internal energy is now decomposed into three components: temperature transport by 1) steady mean circulation $c_p [\overline{v}][\overline{T}]$, 2) stationary eddies $c_p [\overline{v}^*][\overline{T}^*]$, and 3) transient eddies $c_p [\overline{v'T'}]$. In this case, we assume that cross terms, which indicate the temperature transport carried by instantaneous and asymmetric flow, are small compared to the other components.

After applying the decomposition to all the components in AMET, we can calculate the total heat transport by transient eddies and stationary eddies, respectively. Note that the calculation suffers from mass imbalance reported by previous studies (e.g., Trenberth 1991; Graversen 2006; Liu et al. 2020). We have examined the mass budget correction methods in a previous paper (Liu et al. 2020). In general, barotropic mass correction is the most practical method used by many studies to calculate mass residuals (Trenberth 1991; Graversen 2006; Fasullo and Trenberth 2008; Mayer and Haimberger 2012; Liu et al. 2020). Unfortunately, such barotropic correction is a rough mass budget correction. It is assumed that the barotropic wind causes the mass imbalance and introduces corrections to barotropic wind fields, which only works for the vertical integral of an air column. It cannot quantify reliably the mass residuals on each pressure level. Liang et al. (2018) proposed a new mass budget correction method for the variability of the total heat transport at short time scales. However, it cannot be applied to a decomposition of heat transport driven by transient or stationary eddies.

It is still necessary to have an estimate of the energy transported by mass residuals. Therefore, we performed barotropic mass correction and show the energy transported by mass residuals. For instance, the zonal integrals of total energy transported by mass residuals at

60°N in the chosen atmospheric reanalysis products are shown in Fig. S2 in the online supplemental material. The values are smaller than the total energy transport at that latitude, but much larger than the anomalies. Although the impact of mass residuals on each layer should be much smaller than the total integral, the results should be interpreted with care.

2) ENERGY TRANSPORT IN THE OCEAN

Different from the atmosphere, we calculate the total energy transport in the ocean by integrating the divergence of OMET over the entire water column. The mathematical expression of the calculation of OMET at each latitude θ_i is given below (Hall and Bryden 1982):

$$E = \oint_{\theta=\theta_i} \int_{z_b}^{z_0} \rho_0 c_{p_0} \theta_p v dz d\theta, \quad (4)$$

where ρ_0 is the seawater density (kg m^{-3}), c_{p_0} is the specific heat capacity of seawater ($\text{J kg}^{-1} \text{°C}^{-1}$), θ_p is the potential temperature (°C), v is the meridional current velocity (m s^{-1}), and z_0 and z_b are sea surface and the depth to the bottom (m), respectively. A constant value of $c_{p_0} = 3987 \text{ J kg}^{-1} \text{°C}^{-1}$ was used in all the calculations of OMET.

Note that due to the sources and sinks in oceanic reanalyses, OMET can suffer from mass budget imbalances. It is very difficult to correct for the imbalance in sources and sinks in oceanic reanalyses as we cannot perform barotropic correction plausibly for the ocean due to varying sea surface heights. In the ocean, with its intense boundary circulations, even the smallest imbalance can lead to significant errors in the heat flux. So far, there is no practical way to correct the mass budget for the oceanic reanalyses. In addition, the calculation above cannot account for recirculations (Zheng and Giese 2009). In oceanographic literature, it is common to use a reference temperature when calculating OMET in both observations and model diagnostics (e.g., Hall and Bryden 1982; Johns et al. 2011). To deal with these issues, we take a reference temperature θ_r (°C) and update Eq. (4) as

$$E = \oint_{\theta=\theta_i} \int_{z_b}^{z_0} \rho_0 c_{p_0} (\theta_p - \theta_r) v dz d\theta. \quad (5)$$

Here we take θ_r equal to 0°C . Unlike atmospheric reanalyses, oceanic reanalyses often use multipole grids to overcome the North Pole singularity (Madec and Imbard 1996). These grids are always nonuniform, which complicates the definition of zonal values. To work with the data on curvilinear grids, we employ a zigzag setup that is elaborated upon in detail by Outten

et al. (2018) (see their Fig. 2). In short, we work on the original multipole grid and follow the native zonal directions when performing numerical operations.

3) STATISTICAL ANALYSIS

Intending to explore the relations between different variables, we performed linear regressions on various fields. We implemented a Student's t test to check the significance of the regressions. Before performing the linear regression, all the time series were detrended. We determined the trends by applying a polynomial fitting to the time series. We found that the second-order polynomial fitting can capture the multidecadal trend without losing variations at decadal time scales, especially for OMET (not shown). The t tests were performed on detrended monthly fields without a low-pass filter.

Note that some of the results discussed in the following sections are below the 99% confidence level. This is mainly due to the short time series provided by the reanalysis data (no more than 456 months at monthly time scales, see Table 1). However, compared to the outputs from numerical models, we believe the results are still insightful as these reanalyses can represent AMET and OMET in the real world.

3. Results

The results shown in this section are based on monthly mean fields. Results across larger time scales are obtained with low-pass filters (running mean from 1 to 10 years) applied to the monthly mean fields.

a. Meridional energy transport

To investigate the compensation between the atmosphere and ocean, we start with the quantification of AMET and OMET. Estimated from the chosen reanalysis products, the mean AMET and OMET as a function of latitude in the Northern Hemisphere are shown in Fig. 1. Generally, AMET between the chosen atmospheric products agrees well. Starting from 20°N, AMET increases with latitude and peaks at 41°N. The peak of AMET in ERA-Interim is about 5 PW, which is consistent with previous studies (e.g., Trenberth and Caron 2001; Fasullo and Trenberth 2008). The AMET from MERRA-2 and JRA-55 are a bit lower than that in ERA-Interim. The meridional variations of mean AMET agree well with literature (Trenberth and Caron 2001; Fasullo and Trenberth 2008).

Furthermore, we inspect the temporal evolution of AMET at specific latitudes. Our interest is in meridional variations in energy transport associated with midlatitude–Arctic interactions. The time series of

FIG. 1. Mean AMET and OMET over the entire time span of each product as a function of latitude in the Northern Hemisphere. AMET is shown with solid lines and OMET with dashed lines. The shaded regions represent the full range of MET across the entire time series at each latitude. The time span of each product used in this study is given in Table 1.

AMET and their anomalies at decadal time scales at 60°N are given in Fig. 2. The monthly averaged AMET provided by each atmospheric reanalysis product can be found in Fig. 2a. The variations of AMET are similar among the chosen products, but they differ slightly in amplitudes. The average of AMET in ERA-Interim is 3.06 ± 1.33 PW, while in MERRA-2 it is 2.88 ± 1.20 PW, and in JRA-55 it is 2.90 ± 1.20 PW. After removing the seasonal cycle and implementing a low-pass filter of a 10-yr running mean to each time series, we obtain the anomalies of AMET from each reanalysis (Fig. 2b). MERRA-2 and JRA-55 have similar AMET anomalies in terms of the variations and amplitudes, while ERA-Interim differs from them in amplitude. This is also reflected in Fig. 2a, as ERA-Interim gives a higher mean AMET at each latitude. In addition, ERA-Interim has a larger standard deviation (std) compared to the other two products.

For the ocean, the mean OMET at each latitude estimated from chosen oceanic reanalyses are shown in Fig. 1. Compared to AMET, OMET is relatively small. The mean OMET in each oceanic reanalysis agrees well at almost all latitudes except for OMET between 30° and 40°N. This could be due to the difference in the model's capability of representing mesoscale eddies in the Gulf Streams among the chosen oceanic reanalyses. We have explored this in our previous work (Liu et al. 2020). Both GLORYS2V3 and SODA3 have a high spatial resolution, employ eddy-permitting models, and simulate the large-scale eddy variability, but this is not the case for ORAS4 (Ferry et al. 2012a,b; Balmaseda et al. 2013; Carton et al. 2018).

FIG. 2. Time series of the zonal integral of AMET at 60°N. (a) The original time series (i.e., no filter) and (b) the anomalies with a low-pass filter after detrending. For the detrended low-pass-filtered ones, we take a running mean of 10 years. The standard deviations σ and the means μ of the entire time series are also shown.

Moreover, we are interested in the temporal evolution of OMET. The time series of OMET and their anomalies at decadal time scales at 60°N after detrending are given in Fig. 3. Differences in both the amplitude and variations can be observed in all three time series (Fig. 3a). The differences are consistent with several recent studies. Uotila et al. (2019) demonstrate the large difference in ocean heat transport between several different oceanic reanalyses in their Polar Ocean Reanalyses (ORA) Intercomparison Project, especially for the Arctic region. Karspeck et al. (2017) report that with the same surface forcing (CORE-II forcing), ocean models can behave very differently without further constraints from observations. They find large differences in volume transports between six reanalysis products (also including ORAS4 and SODA) in their tests. The OMET time series considered here do not differ much from each other statistically. The mean OMET at 60°N in ORAS4 is 0.47 ± 0.06 PW, while in GLORYS2V3 it is 0.44 ± 0.07

PW, and in SODA3 it is 0.46 ± 0.07 PW. After removing the climatology of OMET and implementing a low-pass filter of 10 years, the differences become clear (e.g., Fig. 3b). It can be noticed that ORAS4 resembles SODA3 but differs much from GLORYS2V3. It was explained in our previous work that the first 10 years in GLORYS2V3 are not reliable (Liu et al. 2020). Given the absolute value of OMET anomalies (~ 0.01 PW), the standard deviations of OMET anomalies (~ 0.01 PW) from these three oceanic reanalyses are large.

To conclude, all the chosen atmospheric reanalyses agree on mean AMET at almost every latitude in the Northern Hemisphere. The low-frequency AMET anomalies in ERA-Interim differ a bit from those in MERRA-2 and JRA-55, but the differences in variability are relatively small compared to the differences between low-frequency AMET anomalies calculated through integration of energy transports over the entire air column (Liu et al. 2020). Similarly, the mean OMET

FIG. 3. Time series of the zonal integral of OMET at 60°N. (a) The original time series and (b) the anomalies with a low-pass filter after detrending. For the detrended low-pass-filtered ones, we take a running mean of 10 years. The standard deviations σ and the means μ of the entire time series are also shown.

given by all three oceanic reanalyses agree at almost all latitudes in the Northern Hemisphere except for around the Gulf Stream. The OMET anomalies given by ORAS4 resembles that provided by SODA3, especially for the OMET variability. However, in GLORYS2V3 the OMET anomalies are not so reliable. Liu et al. (2020) found that ORAS4 is more consistent with the direct heat transport observations in the Atlantic Ocean [Overturning in the Subpolar North Atlantic Program (OSNAP) and RAPID array] than SODA3 and GLORYS2V3 with aspect to OMET. Given the similarities in the variability of OMET anomalies in all the chosen oceanic reanalysis products excluding GLORYS2V3, and the consistency of OMET given by ORAS4 and the observations (Liu et al. 2020), we further investigate the Bjerknes compensation with ORAS4 only. Three atmospheric reanalyses are all included in the following analysis to test the robustness of our study among atmospheric reanalysis products.

b. Bjerknes compensation and atmospheric response to ocean variability

According to Bjerknes (1964), compensation between atmosphere and ocean is expected to take place when the net solar radiation flux at TOA and the heat content in the ocean remain relatively stable. This suggests that the Bjerknes compensation should be investigated across large time scales, at least at interannual (~ 5 yr) time scales. In search of the Bjerknes compensation, we illustrate the correlation between AMET and OMET with time lags up to 10 years from interannual to decadal time scales in Fig. 4. All months are included in the lag regressions. We also investigated the compensation between AMET and OMET anomalies by season, and they will be shown in the following analyses of the mechanism of compensation. Results at annual time scales are included for comparison. At annual time scales, the anticorrelation is small in the chosen atmospheric reanalysis

FIG. 4. Lag regression of OMET anomalies on AMET anomalies at different latitudes in the Northern Hemisphere at (left) annual, (center) interannual, and (right) decadal time scales. The correlation coefficients are shown by contour lines. Positive time lag indicates that ocean leads the atmosphere and vice versa. The regression was performed on monthly mean series of AMET from (a)–(c) ERA-Interim, (d)–(f) MERRA-2, and (g)–(i) JRA-55 and OMET integrated over each latitude after taking 1-, 5-, and 10-yr running means, including all months. The stippling indicates a significance level of 95% based on a t test with unfiltered time series.

products, which is consistent with results shown by [Shaffrey and Sutton \(2004\)](#) using a numerical climate model. At longer time scales, compensation between AMET and OMET is found at almost all latitudes from 40° to 70°N. The maximal compensation rates are reached when the ocean leads the atmosphere for the chosen atmospheric reanalyses from interannual to decadal time scales. However, the time lag between OMET and AMET for the maximum compensation varies across time scales at different latitudes among chosen atmospheric products. At interannual time scales, compensation reaches a maximum when there is no lag at midlatitudes around 40°N in ERA-Interim and JRA-55, while close to 70°N the maximum compensation occurs when the ocean leads by 2.5 years. MERRA-2 differs slightly. The compensation peaks when the ocean leads by 2.5 years around 40°N, while between 65° and 70°N the compensation reaches the maximum when the ocean leads the atmosphere by several months. Similarly, in ERA-Interim and JRA-55, at decadal time scales, the maximal compensation rate is found at around 40° to 50°N and 60° to 65°N when the ocean leads by 2 months and 2.5 years, respectively, whereas in

MERRA-2 the compensation peaks around the same latitudes but always with ocean leading by 3 years.

Focusing on multidecadal time scales, previous studies with numerical climate models can provide more insight into the compensation because of longer time series. After analyzing 93 decades of model output from HadCM3, [Shaffrey and Sutton \(2006\)](#) found that at decadal time scales, compensation between atmosphere and ocean occurs in the Northern Hemisphere and reaches maximum only around 70°N. Similarly, [Van der Swaluw et al. \(2007\)](#) investigated Bjerknes compensation with 341 years of data from a preindustrial run of HadCM3. They found a strong anticorrelation between AMET and OMET at around 70°N at interdecadal time scales and the compensation rate peaks when the ocean leads the atmosphere by 1 year. Using 505 years of the IPCC AR4 preindustrial experiment with the ECHAM5/MPI-OM model, [Jungclaus and Koenig \(2010\)](#) found Bjerknes compensation around 70°N, and the compensation became stronger when the ocean leads by 1–2 years. In summary, these studies using numerical climate models all suggest that the compensation between atmosphere and ocean occurs and peaks around 70°N

FIG. 5. Regression of net surface turbulent and radiation flux anomalies on vertically integrated OMET anomalies at 60°N at decadal time scales. The regression was performed on monthly fields from (a) ERA-Interim, (b) MERRA-2, and (c) JRA-55 after taking a 10-yr running mean, including all seasons. The regression coefficients (TW/PW) are shown spatially, with a positive sign indicating fluxes directed toward the ocean. The gray contour lines indicate a significance level of 99% based on a t test with unfiltered time series.

when the ocean leads at decadal and multidecadal time scales.

With reanalysis data, we find that at decadal time scales the Bjerknes compensation occurs and peaks not only at subpolar latitudes but also at middle to low latitudes. Specifically, we notice the highest compensation rate is found at 40° and 60°N at decadal time scales when the ocean leads. However, we should note that our time series are very short. Recently, [Outten and Esau \(2017\)](#) examined the output from the Bergen Climate Model at decadal time scales using 600 years of data and found the strongest compensation at a 0-yr lag between the oceans and the atmosphere. Their results are supported by other CMIP5 models examined in [Outten et al. \(2018\)](#), which is a synthesis of the diagnostic results using model outputs from 15 CMIP5 models. The finding that Bjerknes compensation appears at midlatitudes, as well as high latitudes, was put forward in [Outten et al. \(2018\)](#). They found that peak compensation occurs around the latitude 70°N (near the sea ice edge) and around 45°N (near the tracks of transient eddies). Those results are similar to our findings with reanalysis data. However, [Outten et al. \(2018\)](#) also show that numerical climate models do not agree on the locations and magnitudes of Bjerknes compensation. Given the large uncertainties and differences between the presence of the compensation in numerical models and reanalysis datasets, we further explore the physical mechanism of the compensation between the atmosphere and ocean so as to obtain a physically consistent picture of Bjerknes compensation.

Since the compensation between the atmosphere and ocean becomes more persistent at decadal time scales in

three atmospheric reanalyses and ORAS4, for the rest of this paper we focus on decadal time scales and apply a low-pass filter of 10 years to the monthly fields, unless specifically noted. To compare the results given by reanalyses and models from literature, we emphasize on the compensation and its physical mechanism in the subpolar region. Consequently, most of the regressions shown below were conducted on the AMET and OMET anomalies at 60°N. We do realize that the time series are very short for this, but we are looking for various strands of evidence for Bjerknes compensation.

A first impression of the physical picture about the compensation between ocean and atmosphere can be obtained by looking at the surface flux. The SFlux anomalies, which consist of the net turbulent flux and net radiation flux at the surface, regressed on OMET anomalies at 60°N at decadal time scales, are shown in [Fig. 5](#). There is no time lag between these two variables. Positive values indicate SFlux toward the ocean, and vice versa. The chosen atmospheric reanalysis products provide similar results. It can be observed that the atmosphere provides energy to the ocean in subtropical regions in the Atlantic Ocean. In the Pacific Ocean close to the Asian continent, SFlux variations associated with AMET are also directed from the atmosphere to the ocean. This means that in the places where the strongest northward flow is situated in the Atlantic (the Gulf Stream) and in the Pacific (the Kuroshio), the ocean always receives heat from the atmosphere. In the subpolar Atlantic, especially in the Labrador Sea, the ocean releases energy back to the atmosphere. [Jungclauss and Koenig \(2010\)](#) found similar results, but the locations where the air–sea interaction is most active are different.

FIG. 6. Regression of (a),(d),(g) AMET anomalies, (b),(e),(h) Stokes streamfunction anomalies, and (c),(f),(i) du/dz anomalies on each pressure level on vertically integrated OMET anomalies at 60°N in winter (DJF) at decadal time scales when OMET leads by 1 month. The climatology of the Stokes streamfunction of the atmosphere is placed on top of the shading as contour lines with a unit of $10^{10} \text{ kg s}^{-1}$. du/dz is the change of zonal wind velocity divided by the change of height, which indicates the vertical shear of zonal wind, with positive signs indicating enhancement of baroclinicity. The regression was performed on monthly fields from (a)–(c) ERA-Interim, (d)–(f) MERRA-2, and (g)–(i) JRA-55 after taking a 10-yr running mean. The stippling indicates a significance level of 99% based on a t test with unfiltered time series.

Our results suggest that the change of OMET at 60°N is linked to the changes of SFlux from subtropical to subpolar regions. The cause–effect relation between the atmosphere and ocean, in general, cannot be resolved due to the short time series of the analyzed reanalysis products. But the strong correlation between SFlux and OMET at subtropical and subpolar Atlantic is consistent with the anticorrelation in Figs. 4c, 4f, and 4i at decadal time scales.

A further look into the mechanism of compensation requires more insight into the physical processes in the atmosphere. We investigated the air–sea interactions in both summer and winter. Since the atmospheric dynamic processes are more dominant in winter than in summer (e.g., Gastineau and Frankignoul 2015; Van der Linden et al. 2016), we found clearer signals of compensation and more active interaction between ocean and atmosphere in winter than in summer. Therefore, we mainly elaborate on processes in winter and will

briefly mention the results from summer. The correlations between OMET anomalies at 60°N and AMET anomalies on each pressure level in winter are shown in Figs. 6a, 6d, and 6g. In this case, the ocean leads by 1 month. The mean meridional mass streamfunction of the atmosphere is also shown. At lower layers, a strong anticorrelation between AMET and OMET in the subtropical region and a strong correlation at the subpolar region is consistent with the air–sea interaction shown in Fig. 5. However, it is not consistent with Bjerknes compensation. A strong compensation around 60°N at high altitudes is observed, and this should be responsible for the strong compensation illustrated in Fig. 4c around 60°N . Given the mean meridional atmospheric circulation, the region contributing to the compensation is located at the ascending branch and overturning part of the Ferrel cell. This motivates us to investigate the relation between the Ferrel cell and OMET variability. We regressed the anomalies of the

FIG. 7. Regression of meridional component of eddy momentum anomalies on vertically integrated OMET anomalies at 60°N in winter (DJF) at decadal time scales when OMET leads by 1 month. The eddy momentum $\mathbf{v} \cdot \mathbf{v}$ was decomposed into (left) transient eddies and (right) standing eddies. The regression was performed on monthly fields from (a),(b) ERA-Interim, (c),(d) MERRA-2, and (e),(f) JRA-55 after taking a 10-yr running mean, with positive signs indicating enhancement of eddy transport. The regression coefficients are shown with a unit of $\text{m}^2 \text{s}^{-2} \text{PW}^{-1}$.

Stokes streamfunction on OMET at 60°N in winter (DJF) when ocean leads by 1 month in Figs. 6b, 6e, and 6h. A strong anticorrelation is observed between 50° and 70°N . Together with the climatology of the Stokes streamfunction shown in this figure, we observe that the changes in meridional mean flow is primarily due to a shift rather than a change in intensity of the Ferrel cell. The result is consistent with the observed anticorrelations provided in Figs. 6a, 6d, and 6g. Although the magnitudes of regression coefficients are different among the chosen atmospheric reanalysis products in Fig. 6, the patterns are quite robust among these datasets. In addition, we also notice a strong coupling between the Hadley cell and OMET (not shown). This suggests that a low-latitude driver may contribute to the compensation at middle to high latitudes. However, such a teleconnection is beyond the scope of this paper.

Since the mean meridional cells are primarily driven by the eddy forcing at midlatitudes (e.g., Phillips 1956; Lorenz 1967; Holton 1973; Andrews et al. 1987; Trenberth and Stepaniak 2003), we expect the shift of the Ferrel cell to be related with changes of transient eddy momentum fluxes and thus the baroclinicity of the atmosphere. To inspect the baroclinicity, we regress du/dz anomalies on OMET anomalies at 60°N in winter (DJF) at decadal time scales when ocean leads by 1 month in Figs. 6c, 6f, and 6i. The results are very similar for different atmospheric reanalysis products. A direct look at the momentum transported by eddies at multiple pressure levels is given in Fig. 7. Around 50°N from the surface to 400 hPa and 70°N above 300 hPa, with increased OMET we find increased baroclinicity in the atmosphere, which implies that more transient eddy activity can be anticipated. Following Peixoto and Oort (1992)

we decomposed the total meridional eddy momentum transport anomalies into transient eddy momentum transport and standing eddy momentum transport on multiple pressure levels, and then regressed them on OMET anomalies at 60°N in winter (DJF). The OMET leads the atmosphere by 1 month. The results are similar among the chosen atmospheric reanalysis products. Between 50° and 70°N at 200 hPa, transient eddies become more active when OMET increases. The standing eddies are correlated with OMET from 50° and 70°N at almost all the chosen pressure levels. Physically, it seems heat converges due to increased OMET and the convergence could lead to increased baroclinicity. A change in gradient will occur as a consequence and hence stronger transient eddy activity in the upper tropopause. The responses of transient eddies and standing eddies are comparable, in terms of the magnitudes of eddy momentum in response to the variability of OMET (Fig. 7). The mechanisms for changes in standing eddies are unclear. It might be that the processes associated with the increased baroclinicity involve large-scale orographic forcing of the stationary waves (Williams et al. 2007). Due to an interaction with transient eddies and changes in the vertical structure of the atmosphere, stronger standing eddy activity is therefore expected (Held et al. 2002). The result is consistent with the strong correlation in Figs. 6c, 6f, and 6i between 50° and 70°N. Consequently, with increased OMET the shift in the Ferrel cell is likely to be driven by eddy momentum fluxes due to changes in baroclinicity, which explains the anticorrelation between OMET and the Ferrel cell around 60° in Figs. 6b, 6e, and 6h.

In summer, compensation is much weaker than that in winter. Similar to our study in winter months, we examined the compensation and its mechanism with the same methods. The correlations between OMET anomalies at 60°N and AMET anomalies on each pressure level in summer (JJA) are shown in Figs. S3a, S3d, and S3g in the online supplement. To highlight the difference between results in winter and summer, the same ranges are taken for these plots. It can be observed that there is a weak compensation at the surface around 60°N. We regressed the anomalies of the Stokes streamfunction on OMET at 60°N in summer (JJA) when ocean leads by 1 month in Figs. S3b, S3e, and S3h. No strong correlation between OMET and the mean flow is found. Similarly, we regress du/dz anomalies on OMET anomalies at 60°N in summer (JJA) at decadal time scales when ocean leads by 1 month (Figs. S3c,f,i). There is no sign about the coupling between OMET and baroclinicity in subpolar regions. The patterns given by ERA-Interim, MERRA-2, and JRA-55 are slightly different in Fig. S3. In summary, the air–sea interaction

is weak in summer and it hardly leads to Bjerknes compensation. Given the fact that both the mean flow and the baroclinicity of the atmosphere are not strongly linked to the variability of OMET in summer, the physical processes are also very different than those in winter.

Apart from the zonal mean response, the atmosphere also shows spatial patterns in the horizontal plane in response to OMET variations. The atmospheric response to the changes of OMET in winter at decadal time scales is given in Fig. 8. It illustrates the relation between geopotential height anomalies at 500 hPa and OMET anomalies at 60°N when OMET leads by 1 month. Both the patterns and magnitudes are very consistent among three atmospheric reanalysis datasets. The variations of OMET are correlated with a low pressure system over Greenland, the subpolar North Atlantic, and the central Arctic Ocean and a high pressure system over the subtropical Atlantic. In other words, the atmospheric response to positive OMET anomalies is characterized by an AO/NAO-like pattern in the North Atlantic and the polar low pressure system is shifted southward. Such an AO/NAO-like pattern was found by Jungclauss and Koenigk (2010) too. In addition, they also noticed a high pressure center over the North Pacific (see their Fig. 6a) and they claimed that the Bjerknes compensation is likely to be influenced by the changes in the North Pacific sector after inspecting the longitudinal variations of AMET (see their Fig. 7). On the other hand, Van der Swaluw et al. (2007) reported a low pressure center driven by positive OMET anomalies at the Greenland Sea and no AO/NAO-like pattern was observed at the North Atlantic. Although the chosen reanalyses and models do not agree well on the atmospheric response to OMET anomalies in the North Pacific and the subtropical North Atlantic regions, a strong polar low is found by all of them. This also implies that OMET has a strong impact on the Arctic climate, which has been reported by many studies (Van der Swaluw et al. 2007; Jungclauss and Koenigk 2010; Koenigk and Brodeau 2014; Van der Linden et al. 2016).

To conclude, the atmosphere responds to OMET variations with a shift in the Ferrel cell in winter. This shift is likely to be driven by eddy momentum fluxes due to changes in baroclinicity. Horizontally, this shift shows as an AO/NAO-like response.

So far, we discussed the atmospheric response to the variations of OMET. We demonstrated that the atmosphere responds to the OMET variations with a shift of mean flow driven by eddy momentum transport in winter. This was also shown by Outten and Esau (2017) with Bergen Climate Model, where they regressed the

FIG. 8. Regression of geopotential height anomalies at 500 hPa on vertically integrated OMET anomalies at 60°N in winter (DJF) at decadal time scales when OMET leads by 1 month. The regression was performed on monthly fields from (a) ERA-Interim, (b) MERRA-2, and (c) JRA-55 after taking a 10-yr running mean. The gray contour lines indicate a significance level of 99% based on a t test with unfiltered time series.

OMET variations onto surface fluxes and showed a shift in the transient eddy storm tracks over the North Atlantic and Pacific, located over the Gulf Stream and Kuroshio. The corresponding AMET variations, which links to the compensation directly, contain additional information about the occurrence of Bjerknes compensation as shown in Figs. 4c, 4f, 4i, 6a, 6d, and 6g. Regarding the AMET response to the variations of OMET in the subpolar Atlantic, there are multiple explanations provided by early studies with numerical climate models. Shaffrey and Sutton (2006) found that a stronger OMET in the North Atlantic Ocean leads to a weakened atmospheric transient energy transport, and thus the Bjerknes compensation. Van der Swaluw et al. (2007) also claimed that the compensation is due to the variations of eddy components in their experiments, especially for the transient synoptic eddies. They found that the vertical shear of zonal wind decreases when OMET increases (see Figs. 7a,b in their paper), and this implies a decrease of baroclinicity. Consequently, the decreasing baroclinicity results in less eddy activity and this will lead to the changes in AMET. However, our findings about the changes of eddies are opposite to those from Van der Swaluw et al. (2007) (see Fig. 7). Moreover, we also notice that there is a shift of Ferrel cell due to the OMET variations. This encourages us to perform a decomposition of eddy components with AMET and compare them with the energy transported by steady mean flow. The regression of AMET anomalies transported by mean flow and eddy components on OMET on multiple pressure levels (from 850 to 200 hPa) at 60°N in winter (DJF) is given in Fig. 9. Note that the regression was performed at decadal time scales and the

ocean leads by 1 month. In general, the chosen reanalyses products agree well on the energy transport by eddies, especially around 60°N. The magnitudes of regression coefficients differ from product to product, though. Note that in MERRA-2 at lower levels (850 hPa) there is strong latitudinal variations in Figs. 9e and 9f. This is due to the fact that GEOS data assimilation system used to produce MERRA-2 does not extrapolate data to pressure levels greater than the surface pressure (Gelaro et al. 2017). Similarly, the latitudinal variations of steady mean energy transport are consistent in the chosen atmospheric reanalyses around 60°N. The magnitudes of regression coefficients are different, though.

To conclude, across interannual to decadal time scales we find that Bjerknes compensation occurs in both subtropical and subpolar regions in reanalysis data. We also investigated the mechanism behind the compensation in the subpolar region. Unlike the numerical experiments that attribute the Bjerknes compensation to the adjustment by the eddy heat transport components (Shaffrey and Sutton 2006; Van der Swaluw et al. 2007), we find that the compensation is primarily achieved by the changes of mean flow in the atmosphere in response to the OMET variability, thus a shift of the Ferrel cell, driven itself by eddy momentum fluxes around 60°N due to changes in baroclinicity in winter. Horizontally, the shift of the Ferrel cell leads to a shift of AO/NAO-like pattern, which confirms the early results with numerical model by Jungclaus and Koenigk (2010), as they also found a shift of patterns in atmospheric circulations due to compensation. In summer, there is hardly any compensation between the atmosphere and ocean.

FIG. 9. Regression of meridional energy anomalies transported by (a),(d),(g) steady mean flow, (b),(e),(h) transient eddies, and (c),(f),(i) standing eddies on multiple pressure levels on vertically integrated OMET anomalies at 60°N in winter (DJF) at decadal time scales when OMET leads by 1 month. The regression was performed on monthly fields from (a)–(c) ERA-Interim, (d)–(f) MERRA-2, and (g)–(i) JRA-55 after taking a 10-yr running mean, with a positive sign indicating a northward flow.

c. Drivers for the OMET variations

In the last section, we explored the atmospheric response to the OMET variations with respect to atmospheric circulation and heat transport variations. Now we turn to the drivers of OMET variations. In this section, we study wind and buoyancy forcing of the ocean by the atmosphere. It is well established that the atmosphere can force the ocean at the surface by wind stress and buoyancy flux, and the latter is related to the thermohaline circulation, which is closely linked to variations of AMOC (e.g., Kuhlbrodt et al. 2007; Thomas et al. 2014). At short time scales, a shallow upper-ocean layer adjusts quickly to wind stress variations, which causes the anomalous wind-driven Ekman transports. The relation between variations of meridional mass transport in this Ekman layer and changes of OMET anomalies at 60°N in winter (DJF) is illustrated in Fig. 10. The atmosphere (i.e., surface wind stress) leads by 1 month. Again, similar results are obtained with the chosen atmospheric reanalysis products, the surface wind stress is larger in JRA-55 than the other two products, though. Around 60°N in the North Atlantic, where the subpolar gyre resides, OMET is correlated to

the Ekman transports as expected. It indicates that atmosphere variability can influence the OMET variations through wind stress anomalies. This is consistent with Shaffrey and Sutton (2004), as they also found the variability of OMET in the North Atlantic is dominated by Ekman processes. Outten and Esau (2017) noticed that the variations in the strength of the subpolar gyre were well correlated to AMET but not well correlated to OMET, despite AMET and OMET being well correlated to one another. This indicates that the atmosphere could influence the ocean at decadal time scales by causing a spinup or spindown of the subpolar gyre. However, this relationship is not robust when they examined it in many CMIP5 models (Outten et al. 2018).

The atmosphere can further affect OMET through modifying the intensity of gyres or initiating a shift of the location of gyres and the compensating western boundary current strength through wind stress curl variations (e.g., Sverdrup et al. 1942; Thomas et al. 2014). Since the impact of wind stress curl on gyres are reflected in sea surface height (SSH), by exploring the relation between SSH anomalies and OMET variations, we can further investigate the drivers for wind-driven OMET variations. We performed regression of SSH

FIG. 10. Regression of meridional Ekman transport anomalies on vertically integrated OMET anomalies at 60°N in winter (DJF) at decadal time scales when atmosphere leads by 1 month. The meridional Ekman transport anomalies are calculated as $-\tau_x/f$, with τ_x the zonal wind stress and f the Coriolis parameter. The regression was performed on monthly fields from (a) ERA-Interim, (b) MERRA-2, and (c) JRA-55 after taking a running mean of 1 year. The gray contour lines indicate a significance level of 99% based on a t test with unfiltered time series.

anomalies on OMET anomalies at 60°N (Fig. 11). A correlation pattern and an anticorrelation pattern are observed at the Gulf Stream extension and the Labrador Sea, respectively. This is consistent with Fig. 10. We can conclude that the atmospheric winds influence OMET variations through anomalous Ekman transport and variations in the strength of gyres in the North Atlantic, which is consistent with early studies (e.g., Kuhlbrodt et al. 2007).

Moreover, the atmosphere can influence the thermohaline circulation of the ocean through surface buoyancy fluxes. The thermohaline processes affect the AMOC, and thus the variations of OMET (e.g., Kuhlbrodt et al. 2007; Smeed et al. 2014). The total buoyancy flux consists of a thermal flux and a haline flux. We have already elucidated the influence of SFlux on OMET in Fig. 5. To provide a complete picture, here we investigate the linkage between anomalous buoyancy flux variations and OMET anomalies. The regression of buoyancy flux anomalies on OMET anomalies in winter (DJF) at decadal time scales is shown in Fig. 12. Among the chosen atmospheric reanalysis products, a strong coupling between OMET and buoyancy forcing is observed in the subtropical Atlantic, while an anticorrelation between them is found in the subpolar Atlantic. The patterns resemble the correlation map in Fig. 5, which indicates that the buoyancy forcing is dominated by its thermal component. The anticorrelation between OMET and buoyancy forcing in the Labrador Sea and the Irminger Sea indicates that weaker buoyancy fluxes are associated with stronger OMET in these regions. This means that it is unlikely that the OMET variations are driven by the

AMOC associated with convection in this region at these short time scales. In that case, one would have expected the reverse sign. The patterns seem consistent with the wind variations associated with the OMET variability and therefore the confounded Ekman-driven OMET variations may cause the high correlations. However, it should be noted that large uncertainties have been found by Karspeck et al. (2017) regarding the location and intensity of deep convection in different

FIG. 11. Regression of SSH anomalies on vertically integrated OMET anomalies at 60°N at decadal time scales. The regression was performed on monthly fields from ORAS4 after taking a 10-yr running mean, including all seasons. The mean SSH is illustrated by contour lines with a unit of meters. The stippling indicates a significance level of 99% based on a t test with unfiltered time series.

FIG. 12. Regression of buoyancy forcing anomalies on vertically integrated OMET anomalies at 60°N at decadal time scales without time lags. The buoyancy forcing is calculated as $(g\alpha/c_p)Q_s + g\beta(P - E)$, with g the gravitational acceleration, c_p the specific heat capacity of seawater, Q_s the net sea surface heat flux, S the surface-layer salinity, α and β the thermal expansion coefficient and saline expansion coefficient, P the precipitation, and E the evaporation. For simplification, we use constant $c_p = 3987 \text{ J kg}^{-1} \text{ °C}^{-1}$, $S = 35 \text{ psu}$, $\alpha = 250 \times 10^{-6} \text{ K}^{-1}$, and $\beta = 7.5 \times 10^{-4} \text{ psu}^{-1}$. The regression was performed on monthly fields from (a) ERA-Interim, (b) MERRA-2, and (c) JRA-55 after taking a 10-yr running mean. The regression coefficients with a unit of $\text{N s}^{-1} \text{ PW}^{-1}$ are shown spatially. The gray contour lines indicate a significance level of 99% based on a t test with unfiltered time series.

reanalysis products. Also, advective feedbacks associated with anomalous temperature and salinity transports in the ocean can drive AMOC and OMET variations (e.g., [Jackson 2013](#)).

In summary, the oceanic response to the variations in the atmosphere at the time scales considered here is primarily wind driven. Together with the atmospheric response to the variations in the ocean, they form a physical picture of the air–sea interaction at subpolar Atlantic that can lead to Bjerknes compensation in reanalysis datasets. It should be noted that remote processes that could affect the ocean have not been considered here.

4. Conclusions and discussion

In this paper, we studied the interactions between atmosphere and ocean, and their relation to meridional energy transport. We quantified AMET and OMET using six reanalysis products and explored the Bjerknes compensation within the chosen reanalysis datasets. Moreover, we wanted to understand the physical mechanism of the compensation and we revisit the roles of atmosphere and ocean in terms of their forcing and response. Our work is motivated by previous studies on Bjerknes compensation based in numerical climate models only.

We find that the chosen reanalysis datasets agree well on the mean AMET and OMET in the Northern Hemisphere. Our findings are consistent with the results provided by previous studies (e.g., [Trenberth and Caron 2001](#); [Fasullo and Trenberth 2008](#)). The differences

between AMET anomalies given by the chosen atmospheric reanalyses are small. This is the same for OMET anomalies in the chosen oceanic reanalysis products, except for GLORYS2V3. Consequently, we chose ERA-Interim, MERRA-2, JRA-55, and ORAS4 as our benchmark to further investigate the interaction between atmosphere and ocean. From interannual to decadal time scales, we demonstrate that the Bjerknes compensation occurs in the Northern Hemisphere at almost all the latitudes from 40° to 70°N in reanalysis data. The intensity of compensation correlates to the length of time scale, which is consistent with [Shaffrey and Sutton \(2006\)](#). The compensation shows clearly in the surface flux, which is necessary because net radiation flux variations at TOA are small.

Further inspection of the mechanism of Bjerknes compensation at decadal time scales in winter shows that the atmosphere responds to OMET variations with a shift in the Ferrel cell. This shift is likely to be driven by eddy momentum fluxes due to changes in baroclinicity. Horizontally, this shift shows an AO/NAO-like pattern, which is also found by [Jungclauss and Koenig \(2010\)](#) in their numerical model. The corresponding AMET variations are primarily found in the zonal mean part and less in the eddy heat transport part. This is different from previous studies with numerical models ([Shaffrey and Sutton 2006](#); [Van der Swaluw et al. 2007](#)), but they address larger time scales. This proposed mechanism is robust across the different atmospheric reanalysis datasets used in this study. It is noteworthy that this mechanism does not apply to summer, as in

summer there is hardly any compensation between the atmosphere and ocean meridional heat transports. It should be noted that there is a big caveat due to mass budget imbalances (Trenberth 1991; Graverson 2006; Trenberth et al. 2014).

Moreover, the oceanic response to the changes in the atmosphere is illustrated. The atmosphere forces the surface ocean through wind stress and buoyancy fluxes (e.g., Kuhlbrodt et al. 2007; Gregory and Tailleux 2011; Thomas et al. 2014). The surface wind anomalies, which leads to the Ekman transport and Sverdrup transport variations, modify the OMET through an intensification of gyre transports and anomalous Ekman transport. On the other hand, the thermal flux and the saline flux, which constitute a buoyancy flux, could influence the thermohaline processes in the ocean and their influence is then reflected in the variations of AMOC (Timmermann and Goosse 2004; Gregory and Tailleux 2011; Yang et al. 2016). However, at decadal time scales the oceanic response to the variations in the atmosphere appears to be primarily wind driven. It should be noted that the time series are too short to unequivocally determine forcing–response relations. With our analysis we cannot decide whether the ocean drives the atmospheric MET variations or the reverse. The question of causality is beyond the scope of this study.

Note that different methods for the quantification of AMET can potentially lead to different results (Armour et al. 2019). We are aware of the two practical ways to calculate AMET with atmospheric reanalysis datasets. Apart from the implied method, which takes AMET as the residual between SFlux and TOAFlux as shown in this paper, a direct method that is used by many studies is to integrate the divergence of AMET over the entire air column (Trenberth 1991; Graverson 2006; Fasullo and Trenberth 2008; Mayer and Haimberger 2012). Mayer and Haimberger (2012) and Armour et al. (2019) elaborate on the difference between AMET computed using these two different methods. The method based on vertical integration suffers from problems like mass imbalance, unrealistic moisture budget, and sparseness of observations (Trenberth 1991). We also computed AMET as the vertical integral of AMET divergence over the air column in a previous paper (Liu et al. 2020), but the variability of AMET anomalies given by these three atmospheric reanalysis products are very different, which indicates a lack of constraints on AMET in reanalysis products (Liu et al. 2020). A comparison of low-frequency AMET anomalies at 60°N with implied and direct methods using all three chosen atmospheric reanalysis products is provided in Fig. S1 in the supplemental material. With the direct method, the correlation coefficient of low-frequency AMET anomalies at 60°N

between ERA-Interim and JRA-55 is 0.82 and between ERA-Interim and MERRA-2 it is -0.53 , while those correlation coefficients are 0.94 and 0.97 with implied methods. Moreover, with the AMET estimated as vertical integrals, a consistent picture of Bjerknes compensation was not observed and the energy flow between atmosphere and ocean even contradicts the surface and TOA energy budget (not shown). Hence we chose to compute AMET as the residual between SFlux and TOAFlux as in this way we can obtain a physically consistent picture.

Nevertheless, uncertainties are large in the SFlux and TOAFlux fields. Huang et al. (2017) compared the radiation flux at TOA and surface given by ERA-Interim, MERRA-2, and JRA-55 with NASA CERES. They found that the radiation fluxes at TOA given by the chosen datasets agree well in winter but deviate against the observation in summer. For the surface flux, short-wave radiation differs more in summer while longwave radiation differs more in winter. It seems the energy fluxes at TOA are more trustworthy than the energy fluxes at the surface (Huang et al. 2017). In addition, due to a lack of turbulent flux observations with global coverage, currently reanalysis products cannot assimilate turbulent fluxes at the surface. This also leads to large uncertainties in SFlux in reanalyses, which is confirmed by Liu et al. (2017) in their study.

In this paper, we demonstrate that the response of the Ferrel cell contributes to the Bjerknes compensation in reanalysis datasets at middle to high latitudes in the atmosphere. It is also interesting to investigate the other processes that are related to this change. For instance, the coupling between the Ferrel cell and the Hadley cell suggests that there might be a low-latitude driver for the changes of mean flow in the Ferrel cell. Also, variations can be driven by the tropical ocean or the AMOC (Trenberth and Stepaniak 2003). In this case, we need to explore the teleconnections between low-latitude drivers (e.g., ENSO) and middle- to high-latitude weather and climate. Much longer time series are needed to obtain statistically significant results and avoid uncertainties. Here, we find that at large time scales the Bjerknes compensation occurs and peaks not only at subpolar latitudes but also at middle to low latitudes. Specifically, we notice the highest compensation rate is present at 40°N at decadal time scales at the subtropics when the ocean leads, which is worth a visit for future work. Also, analysis can be extended to the Southern Hemisphere, and the relation between global warming and variability of Bjerknes compensation is of interest for further studies.

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