Meridional Energy Transport in the Coupled Atmosphere–Ocean System: Compensation and Partitioning

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ABSTRACT

The variability and compensation of the meridional energy transport in the atmosphere and ocean are examined with the state-of-the-art GFDL Climate Model, version 2.1 (CM2.1), and the GFDL Intermediate Complexity Coupled Model (ICCM). On decadal time scales, a high degree of compensation between the energy transport in the atmosphere (AHT) and ocean (OHT) is found in the North Atlantic. The variability of the total or planetary heat transport (PHT) is much smaller than the variability in either AHT or OHT alone, a feature referred to as ''Bjerknes compensation.'' Natural decadal variability stems from the Atlantic meridional overturning circulation (AMOC), which leads OHT variability. The PHT is positively correlated with the OHT, implying that the atmosphere is compensating, but imperfectly, for variations in ocean transport. Because of the fundamental role of the AMOC in generating the decadal OHT anomalies, Bjerknes compensation is expected to be active only in coupled models with a low-frequency AMOC spectral peak. The AHT and the transport in the oceanic gyres are positively correlated because the gyre transport responds to the atmospheric winds, thereby militating against long-term variability involving the wind-driven flow. Moisture and sensible heat transports in the atmosphere are also positively correlated at decadal time scales. The authors further explore the mechanisms and degree of compensation with a simple, diffusive, two-layer energy balance model. Taken together, these results suggest that compensation can be interpreted as arising from the highly efficient nature of the meridional energy transport in the atmosphere responding to ocean variability rather than any a priori need for the top-of-atmosphere radiation budget to be fixed.

1. Introduction

In this paper, we try to better understand the processes that determine the partitioning of meridional energy transport between the atmosphere and ocean and the possible compensation between the two. (The energy transport is nearly all in the form of static energy, such as potential and internal. It is common, albeit somewhat loose, practice to refer to this as ''heat transport.'') The issue is important at both a fundamental and a practical level, for variations in such transport may lead to variations in climate on decadal and longer time scales. Variability and predictability of the coupled ocean–atmosphere system at interannual, decadal, and longer time scales are usually associated with the extratropics, where the low-frequency oceanic variability has the potential to force the atmosphere. On the other hand, the great gyres of the ocean are usually considered to be forced by the atmosphere, and truly coupled ocean–atmosphere interactions give rise to climate variability in tropical and equatorial regions. Progress in understanding and modeling low-frequency variability of climate is hampered by the long periods in consideration and the lack of historical observations, especially in the ocean. The evolution of the Atlantic meridional overturning circulation (AMOC) is of particular interest for its potential predictability and impact on climate (Hurrell et al. 2010). However, it is still a matter of debate whether the AMOC varies on a decadal period and, most importantly, how its associated anomalies might affect the atmospheric state (Latif et al. 2010).

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The notion that there is a degree of compensation between the meridional energy transport in the atmosphere and in the ocean dates back at least to Bjerknes (1964), who first suggested quasi-constant total meridional heat fluxes and opposite fluctuations of its oceanic and atmospheric parts in the Northern Hemisphere Atlantic sector. Bjerknes also concluded, from an observational data analysis of the midlatitude Atlantic region, that the atmosphere is likely to drive the ocean at interannual time scales, while for decadal–multidecadal periods it is ocean dynamics that prevail. More generally, it is supposed that such compensation arises from the need to keep the top-of-atmosphere (TOA) radiative budget fixed on long time scales, and hence, fixed total meridional heat flux (Bjerknes 1964). If this is the case then any natural variation in one component of the climate system such as the ocean must be compensated by a countervariation in another component, such as the atmosphere. To express this argument a little more precisely, the atmospheric meridional energy flux (AHT) can be obtained by integrating the divergence of the zonally averaged surface and TOA fluxes (Peixoto and Oort 1992; Magnusdottir and Saravanan 1999; Fasullo and Trenberth 2008), namely

$$\nabla \cdot \mathrm{AHT} = F^{\mathrm{TOA}} - F^{\mathrm{SFC}}, \qquad (1)$$

where F^{SFC} and F^{TOA} are the surface and net TOA fluxes, respectively. In the ocean, the balance is between its heat content (OHC) tendency and the zonally integrated surface flux, so that the oceanic meridional heat transport (OHT) is given by

$$\nabla \cdot \mathrm{OHT} = F^{\mathrm{SFC}} - \partial_t \mathrm{OHC}. \qquad (2)$$

The total or planetary energy transport (PHT) is then

$$\nabla \cdot \mathrm{PHT} = \nabla \cdot \mathrm{OHT} + \nabla \cdot \mathrm{AHT} = F^{\mathrm{TOA}} - \partial_t \mathrm{OHC}. \qquad (3)$$

If we assume that on sufficiently long time scales the total ocean heat content does not vary, the PHT may be computed from net radiation at the top of the atmosphere; if this is constant, then one subsystem must compensate for variations in another. However, such an argument cannot literally hold, for it is known that the total meridional energy transport is sensitive to such things as the earth's rotation rate (Vallis and Farneti 2009) and is certainly not fixed a priori.

Nevertheless, and whatever the ultimate reason behind it, compensation has been found in a number of previous studies in numerical models, such as Shaffrey and Sutton (2006), van der Swaluw et al. (2007), and Vellinga and Wu (2008). These studies typically found that decadal-scale anomalous transports in the ocean are followed by anomalies of opposite sign in the atmosphere, which partially compensate in order to maintain a quasi-constant PHT. The compensation was most robustly found in the extratropics and for decadal time scales. In these cases, it is most straightforward to interpret the results as implying that the atmosphere is compensating for natural variations in ocean heat transport on decadal time scales. Farneti and Vallis (2011) showed that, in both the Geophysical Fluid Dynamics Laboratory (GFDL) Intermediate Complexity Climate Model (ICCM) and the GFDL Climate Model, version 2.1 (CM2.1), the coupled system undergoes sustained decadal-scale variability, associated with variations in the AMOC and concomitant variations in ocean heat transport. In cases with different coupled models, the nature of the variations in the ocean heat transport may differ, but the compensation by the atmosphere is expected to hold.

In this paper, we try to make further progress in the understanding of the compensation between meridional atmospheric and oceanic heat transports. In particular, we study the degree, nature, and time scale of compensation, as well as the origin of decadal OHT and AHT anomalies. We further decompose the transports into contributions from the wind-driven gyres (for the ocean) and moisture and sensible fluxes (for the atmosphere). We guide our analysis with the theoretical arguments proposed in Vallis and Farneti (2009), where scaling relations were proposed between the ocean and atmospheric heat transport. We focus on the mechanisms of compensation that could arise in the Atlantic Ocean, where the AMOC is responsible for a large fraction of the heat transport and variability. However, we show that our scaling for the wind-driven gyre and atmospheric transport is also valid for the Pacific Ocean. We address these problems using a hierarchy of coupled atmosphere–ocean climate models, namely a comprehensive coupled climate model (CM2.1; Delworth et al. 2006), an intermediate complexity coupled model (ICCM with both a dynamical ocean and dynamical atmosphere; Farneti and Vallis 2009), and a simple energy balance model. The paper is organized as follows. In section 2, we describe the coupled models used in this study and how we partition the energy transport in the ocean. In section 3, we present the results on the decadal anomalies in oceanic (both overturning and wind driven) and atmospheric transports. Section 4 provides a hypothesis for the invariance in the planetary meridional energy transport with the help of an energy balance model. Section 5 concludes with a summary and discussion of our findings.

2. Description of the coupled models and energy transports

The comprehensive climate model we use is GFDL CM2.1. The ocean model resolution is 1°, with a progressively finer meridional resolution equatorward of 30° reaching ⅓° at the equator, with 50 unevenly spaced vertical levels. The model employs the Gent and McWilliams (1990) scheme (hereafter the GM scheme) for parameterizing mesoscale eddies as implemented by Griffies (1998) and an along isopycnal diffusion of tracers (Griffies et al. 1998). The isopycnal diffusion coefficient κ_I is set to a constant value of $600\,\mathrm{m^2\,s^{-1}}$, whereas minimum and maximum values of κ_{gm} are set to 100 and $600\,\mathrm{m^2\,s^{-1}}$, respectively. The atmospheric model uses a finite volume dynamical core with 24 levels and a horizontal resolution of $2.5° \times 2°$. The formulation of the CM2.1 components, as well as preindustrial and climate change simulations, has been extensively documented in Delworth et al. (2006).

The Intermediate Complexity Climate Model is one step ''down'' in a hierarchy of climate models. It is simplified both in geometry and physical parameterizations but nevertheless is capable of reproducing a fairly realistic Atlantic-like climate in both the ocean and atmosphere and, because of its computational efficiency, it allows integrations of several millennia as well as broad parameter studies. It is described more fully in Farneti and Vallis (2011) so our description here is brief. The atmospheric component is a moist hydrostatic primitive equation model, using idealized physical parameterizations (Frierson et al. 2006) and configured as a 120° reentrant sector (Fig. 1). The atmosphere is based on the GFDL ''B grid'' dynamical core with a horizontal resolution of $3.75° \times 3°$ extending between 75°N and 75°S and seven vertical levels. The model utilizes a simplified Betts–Miller convective scheme and Monin–Obukhov boundary layer. Infrared radiation is calculated using a gray scheme with fixed emissivities, so that the upward and downward fluxes are functions of temperature only, and the atmosphere is taken to be transparent to solar radiation, which has neither a diurnal nor an annual cycle. The ocean component is the Modular Ocean Model, version 4 (MOM4; Griffies et al. 2005), which is a free surface primitive equation model in z coordinates. The main simplification is that there is a single basin, which in our control configuration is 60° wide, extending from 70°S to 70°N, representative of the Atlantic basin (Fig. 1). In the region from 65° to 51°S, the walls are removed and the basin is periodic in longitude, thus, recreating the topology of the Southern Ocean. The horizontal resolution is 2°, there are 24 levels, and the ocean is flat bottomed with a depth of 4000 m, except in

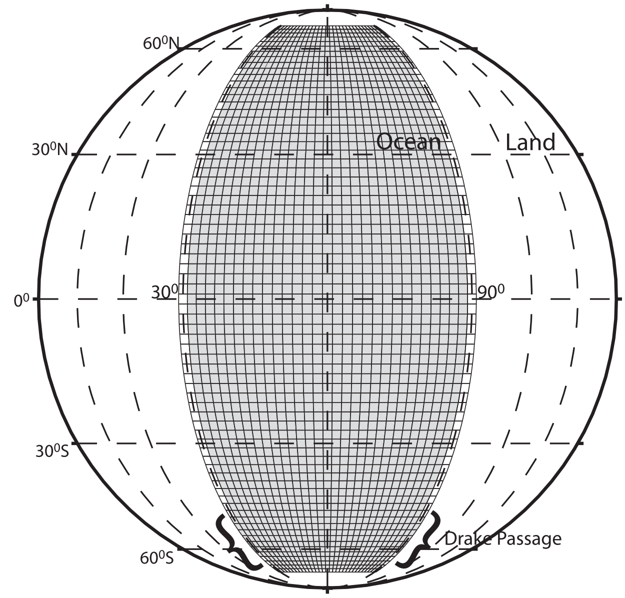

FIG. 1. The ICCM oceanic and atmospheric grid: the ocean basin sits in the center of the atmospheric sector, spanning 60° in longitude and extending from 70°S to 70°N. A cyclic channel extends from 65° to 51°S. The atmosphere is 120° wide with solid meridional walls at 84°S and 84°N.

the model Drake Passage where the total depth is reduced to 2500 m. The model uses a constant diapycnal diffusion with a value of $5 \times 10^{-5}\,\mathrm{m^2\,s^{-1}}$ and the GM scheme, with κ_I and κ_{gm} set to $800\,\mathrm{m^2\,s^{-1}}$. The ocean is coupled to the GFDL dynamical and thermodynamical sea ice model (Winton 2000) at the same horizontal resolution. Finally, a land model is implemented as a single bucket land, with constant values of water availability, heat capacity, roughness, and drag coefficients. Rivers redistribute the precipitated water back into the ocean at the nearest grid point. Farneti and Vallis (2011) showed that ICCM and CM2.1 exhibit very similar AMOC and heat transport decadal variability. We make use here of the 2000-yr-long preindustrial run of CM2.1 to compare its heat transport variability to that of ICCM and justify the conclusions that follow.

Energy transport partition in the ocean

A simple thermodynamic argument shows that the static energy flux in the ocean is given to a good approximation by the potential temperature flux, multiplied by the heat capacity of seawater evaluated at the surface of the ocean (Bacon and Fofonoff 1996; Warren 1999). (This is not the case with the atmosphere, in which work done by compression cannot be neglected and where a contribution due to the transfer of latent heat must also be included in a moist atmosphere.) The oceanic heat, or more precisely enthalpy, transport is nearly all advective and, apart from the Southern Ocean and western boundary currents where eddy heat fluxes are important, it is nearly all accomplished by the resolved flow

$$\mathrm{OHT} \approx \rho_o c_o \int v\theta \, dx \, dz \,, \tag{4}$$

where v is the meridional velocity, θ is the potential temperature, c_o is the heat capacity of seawater per unit mass, and ρ_o is the reference density used in the Boussinesq approximation. One may attempt to decompose this transport into an ''overturning'' circulation and a ''gyre'' circulation by vertically and/or horizontally averaging the flow. However, because the gyres also circulate vertically this partitioning ascribes some of the gyre flow to the overturning circulation and the interpretation of the results is ambiguous. Another partitioning uses θ itself as a coordinate (e.g., Klinger and Marotzke 2000) and the heat transport is then defined as (Vallis and Farneti 2009)

$$\mathrm{OHT} = \rho_o c_o \int_{\theta_1}^{\theta_2} \Psi \, d\theta \,, \tag{5}$$

where Ψ is the overturning streamfunction in temperature coordinates. We can now decompose the transport into the contribution given by different cells (delimited by some temperature class interval) and so effectively introduce a measure of overturning and gyral decomposition in isentropic coordinates. For our purpose, we chose $(\theta_1, \theta_2) = (20°, 25°)$ for the tropical cell, $(13°, 20°)$ for the midlatitude or intermediate cell, and $(0°, 13°)$ for the cold cell (Fig. 2). In this framework, the warm and intermediate cells are to be interpreted as being wind driven and the cold cell as being an overturning cell. Although the partitioning is a little arbitrary and model dependent, the contributions of the three cells are fairly clearly delineated. One could choose the dividing temperatures on more physical grounds by selecting them to correspond to latitudes where the mean-wind stress curl changes sign, but this makes little difference in our models. We also note that Ferrari and Ferreira (2011) present a somewhat different way of quantifying the meridional transport of heat by various branches of the ocean circulation, utilizing a three-dimensional ''heat function'' diagnostic. We have not explored the use of this here, as our methodology is more than adequate for our needs.

3. Energy transports in the coupled system

a. The atmosphere and the deep ocean

In the Atlantic Ocean, a large fraction of meridional heat transfer is achieved by the deep AMOC (Talley

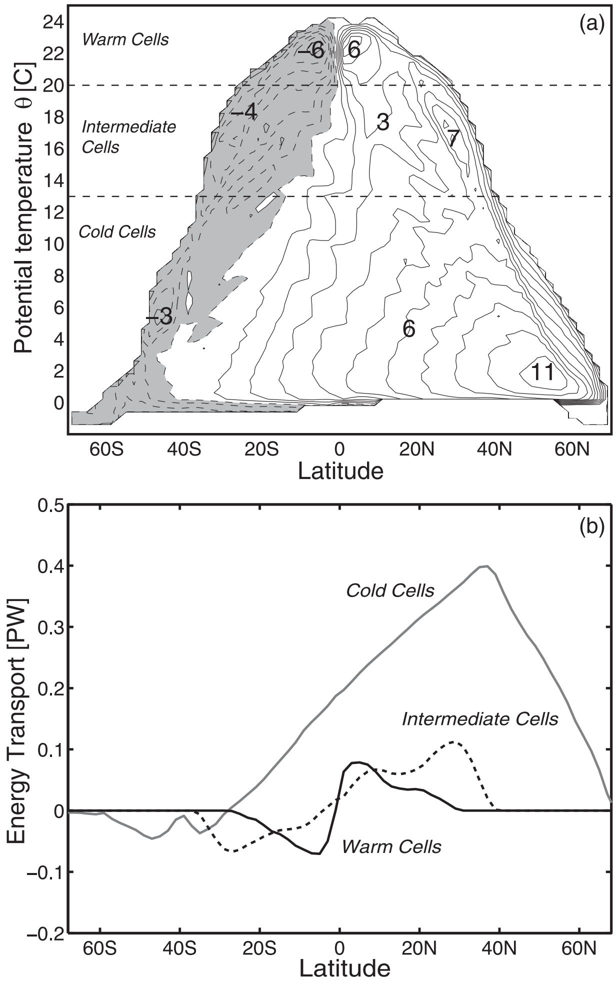

FIG. 2. (a) The oceanic overturning circulation (Sv) in potential temperature space for the ICCM, with circulations divided into temperature classes. (b) The ocean heat transport (in PW; 1 PW = 10^{15} W) for the different cells as derived from (a).

2003). Also, model simulations show that the Atlantic meridional overturning circulation (MOC) may vary on decadal–centennial time scales, suggesting a potential for decadal predictability (Latif et al. 2010; Hurrell et al. 2010). Unfortunately, the AMOC is characterized by different periods, strengths of the oscillation, and even mechanisms depending on the model. It follows that a large uncertainty also exists in the variability of the OHT associated with the AMOC.

We begin by looking at the time series of the heat transport anomalies of the ocean and atmosphere in a 1000-yr segment of the preindustrial run of CM2.1 (Fig. 3). The series have been low-pass filtered to show variability on time scales of 10 years and longer, and we present results for the Northern Hemisphere only, averaged between 20° and 70°N. In CM2.1, the AMOC presents a 20-yr spectral peak. We see in Fig. 3 that large

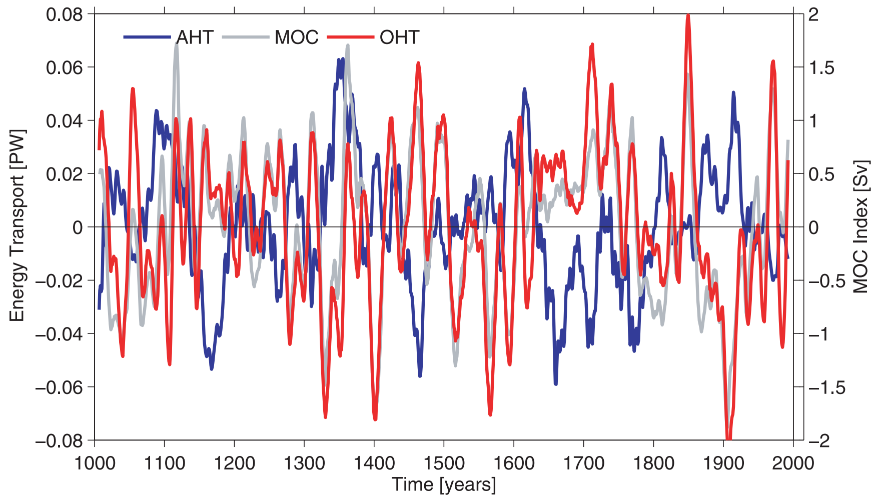

FIG. 3. Time series of decadal anomalies in extratropical (20°–70°N) Atlantic OHT (red), AHT (blue), and Atlantic MOC (gray) in a 1000-yr segment of the GFDL CM2.1 control simulation. The correlation between MOC and OHT is 0.70, and OHT and AHT are anticorrelated (−0.52). Note the different vertical scale for the energy transport and MOC.

amplitude (~1.5 Sv; 1 Sv $\equiv 10^6\,m^3\,s^{-1}$) anomalies in AMOC lead decadal anomalies in OHT, and the two are highly correlated at those time scales (0.70). Further, AHT anomalies are anticorrelated with OHT (−0.52) and are of similar magnitude. Evidently, there is a high degree of compensation between the CM2.1 ocean and atmosphere heat fluxes in the Northern Hemisphere. Also, the variability of the total heat transport is much smaller than the variability in either the atmosphere or ocean alone (not shown), and PHT is positively correlated with OHT, implying that the atmosphere is compensating, albeit imperfectly, for variations in the ocean transport. Opposite fluctuations in energy fluxes by the atmosphere and ocean—Bjerknes compensation—are thus present in CM2.1 at decadal time scales and in the extratropics. Very similar results were obtained by Shaffrey and Sutton (2006) in their analysis of the Hadley Centre Coupled Model, version 3 (HadCM3), with similar correlation coefficients. Partial compensation between energy transports might then be a robust mechanism in coupled model simulations, provided that significant AMOC oscillations are present.

We now turn to the ICCM simulation. Figure 4 shows the time series of ocean and atmosphere anomalous fluxes of heat in ICCM, together with MOC anomalies, for a 500-yr period. Both Northern and Southern Hemisphere results are presented. As in CM2.1, for the Northern Hemisphere and decadal periods, there is a strong correlation between MOC and OHT (0.65); the latter is strongly anticorrelated with AHT (−0.71), and PHT anomalies are positively correlated with those of OHT (0.51). To further quantify the Bjerknes compensation in our models, we compute the compensation rate following van der Swaluw et al. (2007). For the decadal time scales considered, the time-averaged compensation rate in the northern extratropics is 29% for ICCM and 24% for CM2.1. Both rates agree reasonably well with previous estimates where the compensation rate was shown to increase from tropical to subpolar regions, reaching values up to 50% (van der Swaluw et al. 2007).

Farneti and Vallis (2011) extensively described the mechanism for AMOC interdecadal variability and associated atmospheric response in ICCM. Bjerknes compensation, at decadal time scales and in the extratropics, may be explained as follows. The oceanic oscillation, of about 20 years, is causing meridional heat flux anomalies that result in an SST/sea surface salinity (SSS) dipole, approximately across 60°N. In the positive phase of the oscillation, SST will be raised in the north and reduced in the south by the anomalously strong heat flux. This oceanic condition reduces the meridional temperature gradient in the atmosphere resulting in reduced baroclinicity and poleward eddy heat transport. Thus, the atmosphere responds with an anomalously weak meridional heat transport in an attempt to compensate for the decadal-scale oceanic variability. A similar mechanism for partially compensating anomalies in the atmosphere and Atlantic Ocean was also invoked by Shaffrey and Sutton (2006). We now provide evidence of the ocean circulation driving the AHT anomalies at the decadal time scale in the Northern Hemisphere in our coupled models, as suggested by the significant anticorrelation between the two (−0.46; Fig. 4a). Variations in the strength of the MOC tend to lead the variability in OHT by a few years

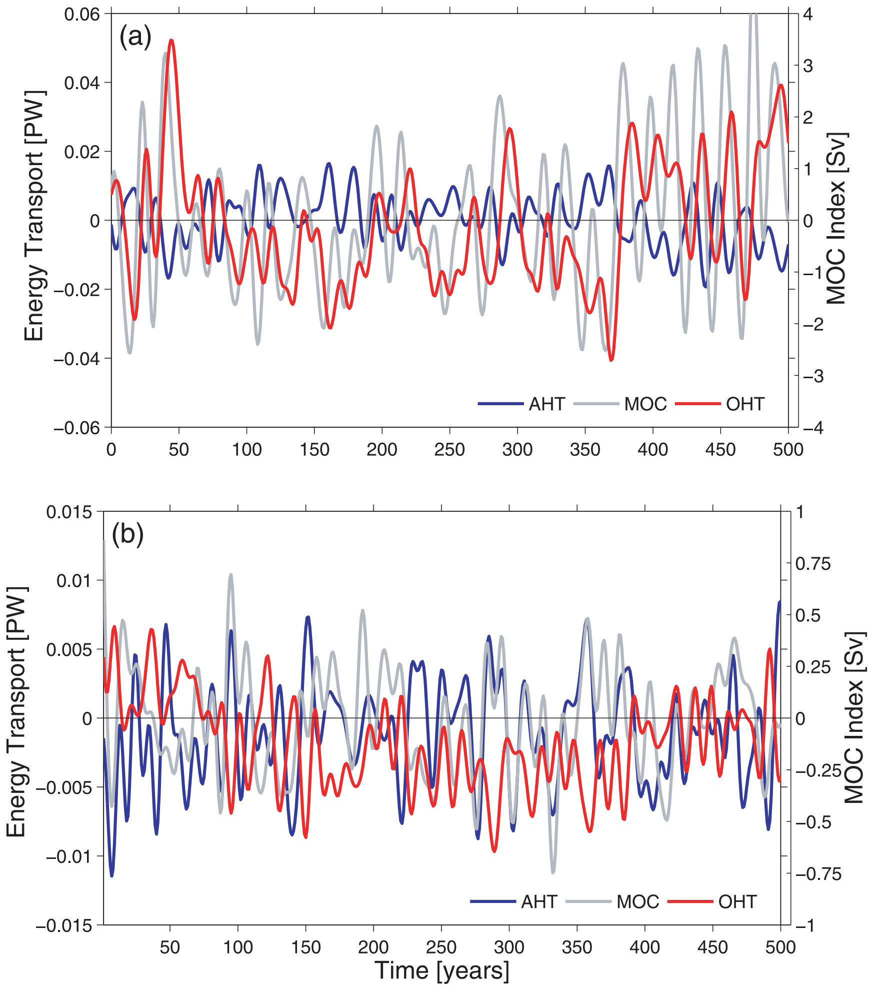

FIG. 4. (a) As in Fig. 3, but for the ICCM, averaged from 20° to 60°N. The correlation between MOC and OHT is 0.65, and OHT and AHT are anticorrelated (−0.71). (b) As in (a), but for the Southern Hemisphere where no significant correlations are found. Note the different vertical scale for the energy transport and MOC. Also note the smaller scale in transports in (b) compared to (a).

(Fig. 5a) and the latter leads AHT anomalies by 1–2 yr (Fig. 5b). Similar lead–lag correlations are found for CM2.1. Decadal climate variability thus stems from variability in the ocean, but dependency on significant AMOC variability makes this result somewhat model dependent, although we might expect compensation to occur provided there is sufficient variability on sufficiently long time scales—several years or longer—to give the atmosphere time to respond coherently. Observational results (e.g., Bjerknes 1964; Czaja and Marshall 2001) seem to suggest this might be a viable mechanism.

In the Southern Hemisphere no significant anticorrelation is found between OHT and AHT anomalies (−0.14; Fig. 4b), consistent with the absence of decadal-scale variability in the ocean and, hence, compensation at those periods. The same result applies to the CM2.1 integration (not shown). The meridional OHT in the Southern Hemisphere is weak, especially in the Atlantic (Fig. 2), and variations in the heat transport are correspondingly small—too small to substantially affect the atmosphere and allow it to meaningfully compensate for any oceanic changes. Also, both the atmospheric and oceanic transports are positively correlated with the total PHT, strongly for AHT (0.66) and weakly for OHT (0.36). The variations in the PHT are then essentially residuals of interannual variations that arise primarily from the atmosphere, with the ocean varying very roughly in unison with the atmosphere.

Whether the atmosphere or ocean should be considered the driver of climate variability depends on the time scale under consideration. For example, on a weekly time scale the ocean may be considered to be steady and the

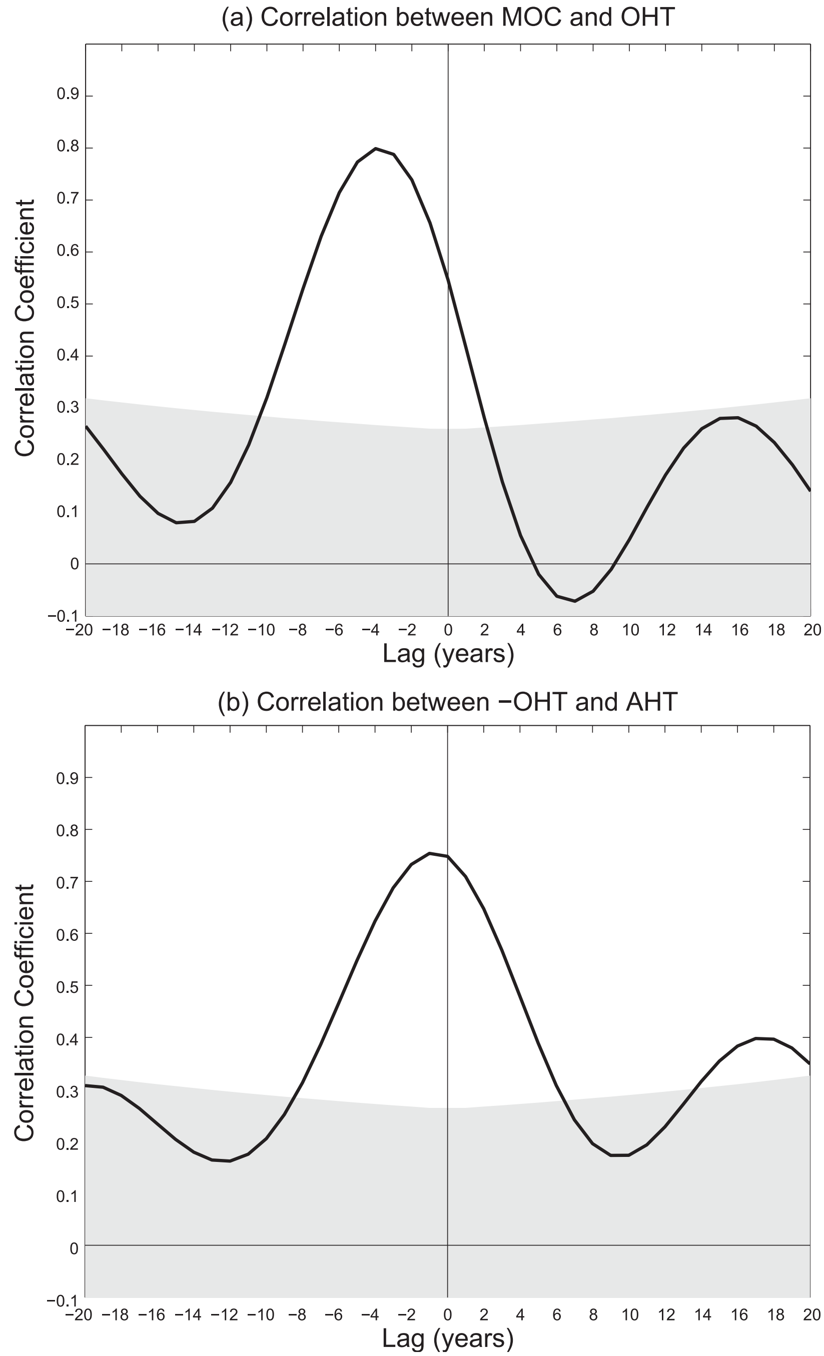

FIG. 5. (a) Cross correlation between the low-passed MOC and OHT in ICCM. The MOC is leading for negative lags. (b) As in (a), but for the correlation between −OHT and AHT. OHT is leading for negative lags. The shading area is the 10% confidence.

variability comes almost entirely from the atmosphere, whereas at longer time scales the ocean should be considered the regulator. Shaffrey and Sutton (2004, 2006) suggest that Bjerknes compensation requires decadal periods and works only in extratropical latitudes because the variations in the oceanic heat storage and TOA fluxes were found to be small compared to the variations in the oceanic and atmospheric energy transports for those time scales and latitudes. Figure 6 investigates this point through the results of both ICCM and CM2.1. The figure shows the correlation between OHT and AHT as a function of latitude and time scale. The thick red curves show

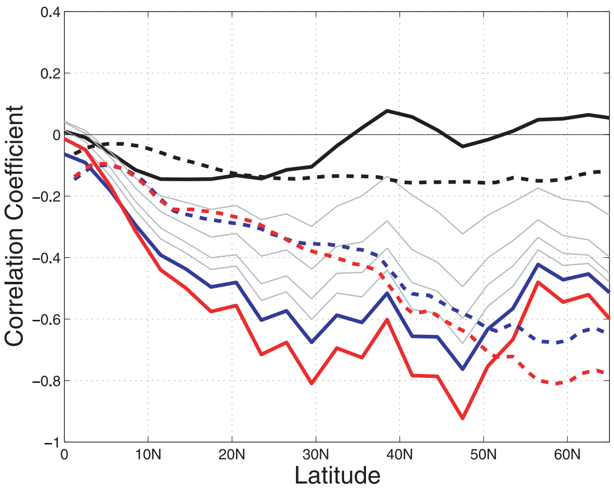

FIG. 6. Time scales of correlations in ICCM (solid lines) and CM2.1 (dashed lines) between OHT and AHT. The red lines show the 20-yr low-passed time series correlation, the blue lines the correlation for the 10-yr low-passed time series, and the black lines the correlation for the 1-yr time series. Gray lines show the correlations for the 2-, 4-, 6-, and 8-yr low-passed time series in ICCM.

the correlation for the 20-yr low-passed time series (the period of the MOC spectral peak in both models), the thick blue curves show the correlation for the 10-yr low-passed time series, and the thick black curves show the correlation for the 1-yr time series. Solid lines are for ICCM, while dotted lines are for CM2.1. Further, thin gray curves show the progressive correlations for the 2-, 4-, 6-, and 8-yr low-passed time series in ICCM.

Both models have correlations close to zero, at all latitudes, for the 1-yr time scale (black thick curves), indicating that compensation in meridional transports is not happening at the annual period (and shorter, not shown). Significant negative correlations do exist for the 10-yr low-passed time series in both models, increasing with latitudes until 50°N (60°N) for ICCM (CM2.1), suggesting partial midlatitude compensation. The consistency across models gives confidence on these results. We note, however, that Shaffrey and Sutton (2006) and van der Swaluw et al. (2007) have found maximum negative correlations at 70°N, so the precise latitude is model dependent and probably a function of the variance of OHT. In any case, we certainly cannot expect CM2.1 and ICCM to agree on the latitude of maximum covariability; as in ICCM, the ocean geometry is highly simplified and the Arctic Ocean is missing. We find even higher anticorrelations for the period of the MOC oscillations (20-yr low-passed; red thick curves) in the extratropics, where correlation coefficients stay close to these values for longer periods (not shown). The increasing strength of compensation with time scale is evidenced by the gray curves in Fig. 6, which, for ICCM, are approaching the decadal ones. This result suggests that the MOC is the dominant factor in setting the time scale of compensation through its associated heat flux anomalies. For any of the time scales considered in Fig. 6, the correlation between OHT and PHT never becomes significantly negative (not shown), suggesting that it is the atmosphere that is compensating for the ocean rather than the converse. However, the shorter the time scale, the greater the correlation between the total transport and that of the atmosphere. Finally, for even shorter time scales (less than a year) we would expect the correlation between AHT and PHT to become almost unity and the correlation between OHT and PHT to stay close to zero.

Farneti and Vallis (2011) explored a large oceanic parameter regime with ICCM and concluded that the ocean is able to drive the atmosphere at interdecadal periods only when significant MOC oscillations are present. It was also found that the period of the MOC oscillations, of about 20 years and similar to the full complexity CM2.1, depends strongly on the oceanic mean state. When sustained decadal MOC oscillations are disrupted, in the case of Farneti and Vallis (2011) through a change in background vertical diffusivity, strong coherent meridional oceanic heat flux anomalies are not generated and the ocean is not able to force the atmosphere to covary with opposite anomalies. The correlation of decadal Atlantic Ocean and atmospheric energy transports is not significant in those cases, and we can thus expect Bjerknes compensation to be found only in coupled climate models with a multidecadal AMOC spectral peak.

b. The atmosphere and wind-driven gyres

The partitioning of the mean energy transports in the tropics was discussed by Held (2001). Held pointed out that because the Ekman transport in the atmosphere and ocean are proportional to each other, the transport ratio in the two systems is given by the ratio of the gross stability of the atmosphere to that of the ocean multiplied by the fraction of the planet covered by the ocean. Both media see the same meridional temperature gradient, but the lower atmosphere is nearly saturated and the atmosphere and ocean have different heat capacities. Given all this, the ratio of the amplitudes of their static energy transports is of order unity. In this picture the energy flux of the atmosphere and ocean change in unison, essentially because the ocean circulation is forced by the atmosphere. Absence of compensation in the tropics was already suggested by Shaffrey and Sutton (2006), where large changes in TOA fluxes were found, so preventing Bjerknes compensation. Held's theory implies that the mean energy transports in the two media are strongly coupled. We can thus suppose that an

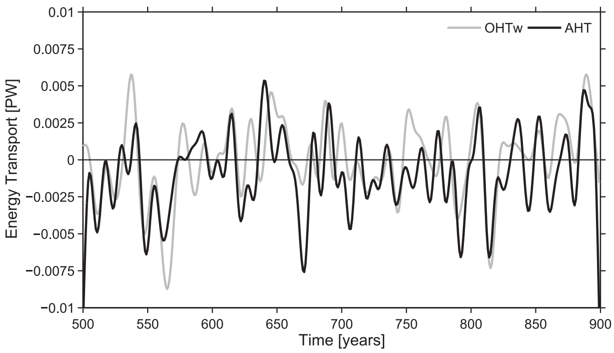

FIG. 7. Decadal anomalies of the oceanic warm and intermediate cells (OHT_w) and AHT in the ICCM averaged over the region 10°–40°N. The correlation between the two is 0.61.

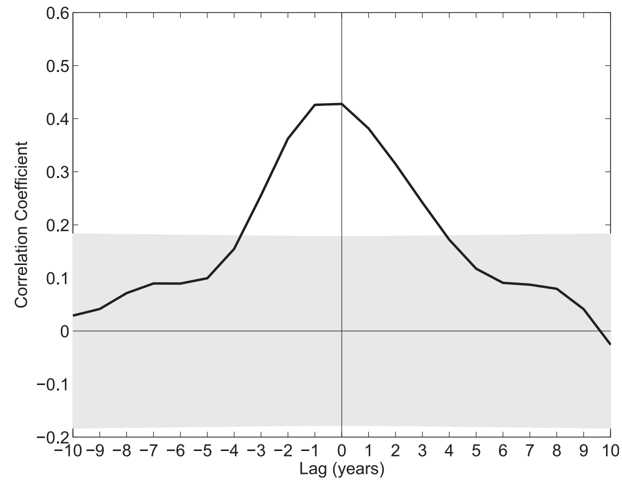

FIG. 8. Lag correlation between the AHT and the OHT by the wind-driven gyres (OHT_w) in the ICCM averaged over the region 10°–40°N. AHT is leading for negative lags. Shading indicates the 10% confidence.

increase in the atmospheric energy transport, because of a strengthening of the Hadley cell, can be assumed to be accompanied by a similar increase in the ocean's energy transport, ruling out the possibility of compensation in the tropics.

We argue in this section that a similar situation holds in the wind-driven gyres, especially in the subtropics. The subpolar gyres are somewhat weaker and the SST is more affected by variations in the MOC, hence, our discussion will focus on the subtropical gyres. The heat transport in the subtropical gyres depends directly on the surface winds and, thus, directly on the intensity of the circulation in the atmosphere. Vallis and Farneti (2009) argue that the gyral heat transport will thus vary in the same way as the atmospheric heat transport. To see this, note that the heat transport in the subtropical wind-driven gyres OHT_w scales as

$$\mathrm{OHT}_w = c_o\rho_o\int v\theta\,dx\,dz \sim c_o\rho_o V\Delta\theta_o L_x^o H, \qquad (6)$$

where L_x^o is the zonal extent of the ocean basin, H its depth, and $\Delta\theta_o$ the temperature difference between poleward and equatorward flowing water. Let us assume that the total transport in the subtropical gyre is given by Sverdrup balance, $\beta\rho_o[v] = \nabla \times \tau$, where $[\cdot]$ denotes a vertical integral, τ is the wind stress, and $\rho_o[v]$ is the vertically integrated meridional mass flux. The implied scaling for the meridional velocity is $V \sim \tau^x/(L_y\beta\rho_o H)$, where H is the average depth of the circulation and we assumed the wind stress to be predominantly zonal. A crude scaling for the energy flux OHT_w is then

$$\mathrm{OHT}_w \sim \frac{\tau^x L_x^o}{\beta L_y} c_o \Delta\theta_o. \qquad (7)$$

The estimate Eq. (7) does not depend on the depth of the wind-driven circulation because in Sverdrup theory the total transport is given directly by the wind stress. Nevertheless, the vertical structure of the temperature does, of course, quantitatively affect the heat transport and Eq. (7) does not account for that.

The meridional energy transport in the atmosphere is given by

$$\mathrm{AHT} = \int (C_p T + \Phi + Lq)v\,dp/g, \qquad (8)$$

where v is the meridional component of the total velocity, C_p is the specific heat capacity of the atmosphere, and T its temperature, $\Phi = gz$ is the geopotential, L is the latent heat of condensation, and q is the specific humidity. The total flux, or moist static energy (MSE), can be decomposed into a dry static energy flux DSE = $C_pT + \Phi$ and latent energy LE = Lq associated with water vapor.

In the midlatitudes, if we suppose that meridional fluxes may be parameterized diffusively then the DSE flux is approximately given by (Vallis and Farneti 2009)

$$\mathrm{DSE} \approx \rho_a C_p H_a \frac{L_x^a}{L_y}\kappa_a \Delta\theta_a, \qquad (9)$$

where κ_a is the coefficient of eddy diffusivity, H_a is a vertical scale for the atmosphere, L_x^a is now the zonal extent of the atmosphere, and $\Delta\theta_a$ is the temperature gradient across midlatitudes. Assuming also diffusive approximation for the surface stress, Vallis and Farneti (2009) arrived at the following ratio for the transport in atmosphere and ocean:

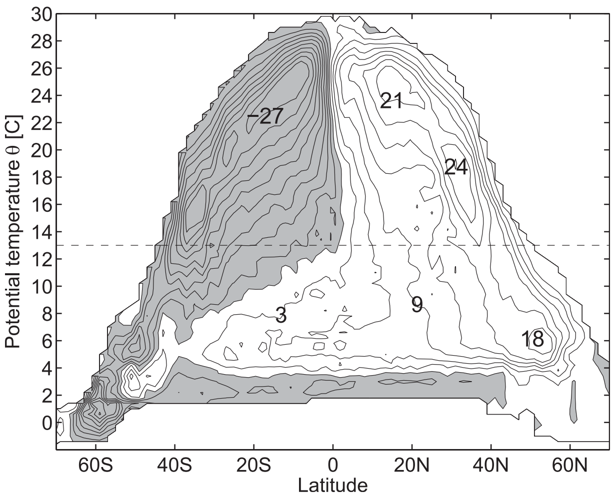

FIG. 9. As in Fig. 2a, but for GFDL CM2.1.

$$\frac{\mathrm{DSE}}{\mathrm{OHT}_w} \sim \frac{C_p \Delta\theta_a}{c_o \gamma \Delta\theta_o}, \tag{10}$$

where γ is the ratio of the zonal dimensions of the ocean and atmosphere (and so about 0.7). Including the effects of water vapor in the atmospheric transport, and so considering MSE instead, the ratio of the transports scales as the ratio of their respective gross stabilities. In midlatitudes, the wind-driven ocean heat transport and the atmospheric heat transport will not be equal, but they scale approximately the same way as external parameters change, as shown in Vallis and Farneti (2009). The above scaling thus suggests that AHT and OHT_w should vary in unison on long (e.g., decadal) time scales. The key assumption is that the atmospheric heat transport is related to the surface wind that drives the ocean gyres. Note also that although the atmospheric heat transport is assumed to be diffusive, the diffusivity itself does not appear in the ratio above.

To test Eq. (10), we looked at the time series of the ocean transport by the wind-driven gyres only (warm and intermediate cells of Fig. 2) and AHT in ICCM. A focus is given to the subtropical latitudes and anomalies are averaged between 10° and 40°N. Results are presented in Fig. 7 for the 10-yr low-passed transports, where a strong positive correlation (0.61) emerges between the two fluxes, with anomalies of similar magnitude consistent with the prediction of the scaling above. It is now the atmosphere driving the ocean gyral anomalies, from annual to decadal periods (Fig. 8). We should stress, however, that the correlation between the *total* OHT and AHT is negative on those time scales, as the atmosphere is compensating for the slow natural variability in the deep ocean controlled by the mechanically and diffusively driven MOC (Fig. 4).

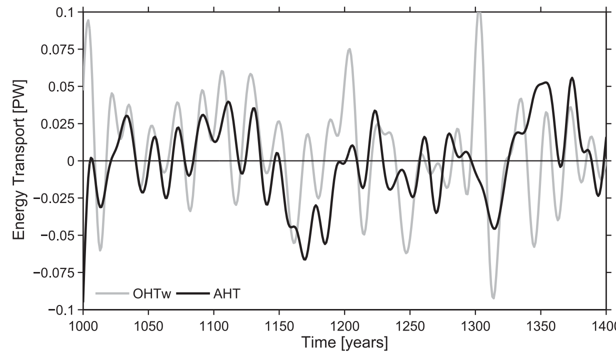

FIG. 10. As in Fig. 7, but for GFDL CM2.1. The correlation between the two transports is 0.46.

The geometry of the ICCM ocean is very idealized with no bathymetry, so that the ocean interior can be considered in Sverdrup balance. This is precisely one of the assumptions made by Vallis and Farneti (2009) in order to achieve the scaling given by Eq. (7). To further verify our proposed relation, we thus turn to the full complexity CM2.1. Similarly to ICCM, we compute the overturning streamfunction in potential temperature space and separate the warm and intermediate cells to the cold cell with the same temperature cutoff ($\theta = 13°$). Figure 9 shows stronger wind-driven subtropical cells, as the transport shown is now for the global ocean. In contrast, the cold overturning cell is of similar magnitude as in ICCM since this is almost entirely found in the Atlantic sector. Decadal anomalies for the CM2.1 OHT_w and AHT are given in Fig. 10, where the proposed scaling implying covariability still holds, albeit with a weaker positive correlation (0.46). Both anomalies are decidedly larger in CM2.1, and this is due to the inclusion of the Pacific subtropical gyre. The strength of the covariability might be a function of the chosen period, but the significant positive correlation for the heat fluxes in oceanic gyres and the atmosphere in the subtropics seems a robust feature of both models.

From our coupled model results and theoretical considerations, we can now argue that there is no mechanism for compensation of energy transports between the atmosphere and the wind-driven oceanic gyres. A tight coupling between the wind-driven ocean circulation and the atmosphere generally militates against long-term variability involving the wind-driven flow. To see this, suppose that the wind-driven gyres were to fluctuate, producing variations in energy transport on a time period long enough for the atmosphere to remain in a quasi-equilibrium—for example, on a decadal time scale. An increased oceanic energy transport would lead to a reduced meridional temperature gradient and a reduced atmospheric energy transport and hence reduced

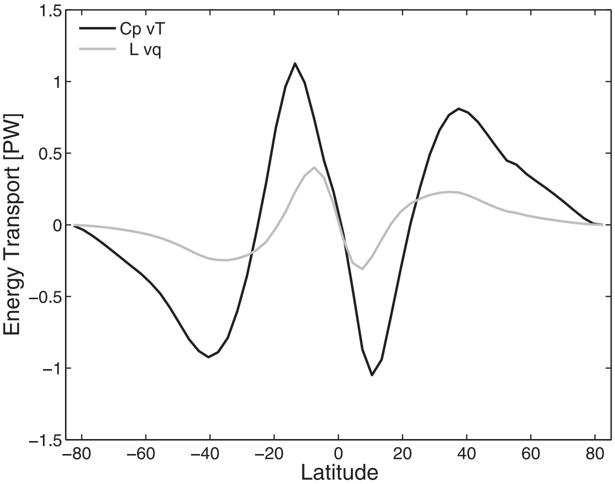

FIG. 11. Time-mean sensible heat flux $C_p\overline{vT}$ and latent energy flux $L\overline{vq}$ in ICCM.

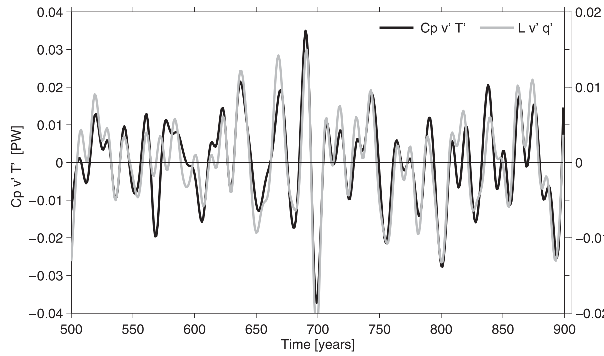

FIG. 12. Decadal anomalies of sensible heat and latent energy fluxes in the ICCM averaged over the region 30°–60°N. The correlation between the two is 0.88. Note the different vertical scales.

baroclinic activity and reduced surface winds. Thus, the wind forcing of the gyres would fall, reducing their energy transport; that is to say, any natural variation in the wind-driven component of the ocean circulation would be naturally damped by the atmospheric feedback. Of course, because there may be delays in the response of one component of the system, variability involving a delayed oscillator might arise (e.g., Cessi 2000).

c. Atmospheric moisture and sensible heat fluxes

We focused on the DSE flux imparted by midlatitude eddies, but those eddies are also responsible for the poleward transfer of water vapor (LE). A time mean of both fluxes in ICCM is shown in Fig. 11. Again, using the relation between potential temperature and DSE fluxes, Vallis and Farneti (2009) suggested that moisture and sensible heat transports are related by the following expression:

$$\overline{v'q'} \approx r\frac{\partial q_s}{\partial T}\overline{v'T'} \approx r\frac{\partial q_s}{\partial T}(T/\theta)\overline{v'\theta'}. \tag{11}$$

The approximation above implies that DSE and LE flux anomalies should covary in the extratropics; for example, in the case of low-frequency natural variability. This prediction is tested and confirmed with the ICCM integration by computing the decadal anomalies of the sensible heat ($v'T'$) and moisture ($v'q'$) fluxes, averaged between 30° and 60°N, which are shown in Fig. 12. The extremely high positive correlation (0.88) in the anomalous transports indicates that Eq. (11) holds in this model with a pronounced oceanic decadal variability. Trenberth and Stepaniak (2003) noted from reanalyses data that latent and dry static energy transients reinforce one another in the extratropics, where both transports are directed poleward and LE anomalies are the weakest. Instead, in the tropical regions anomalies tend to be similar but opposite in sign. Trenberth and Stepaniak (2003) concluded that DSE and LE transports are positively correlated for annual and interannual time scales, being both carried out by baroclinic disturbances, which seems consistent with our findings.

4. Compensation within an energy balance model

The robust but imperfect compensation by the atmosphere to variations in ocean total heat transport, noted in our coupled models and previous studies (Shaffrey and Sutton 2006; van der Swaluw et al. 2007), raises the question as to whether the total meridional transport in the atmosphere–ocean system might be nearly constant because the TOA radiative fluxes are to be considered constrained on long time scales, as postulated by Bjerknes (1964), or whether the atmospheric dynamics simply transports more (less) energy poleward as its temperature gradient increases (decreases)—as would any diffusive system—so providing a degree of compensation [see also the discussion in Stone (1978b)]. In the latter case, the near invariance of the TOA radiative fluxes arises as a consequence of the efficiency of the atmospheric dynamics and not as the cause for compensation. The fact that the total energy transport diminishes as rotation rate increases, as noted in Vallis and Farneti (2009), suggests that the radiative balance is not fundamentally constrained. We further examine this issue with a two-level zonally averaged energy balance model (EBM) with lateral heat transport in both the atmosphere and ocean.

The model is schematically illustrated in Fig. 13, and its evolution equations are

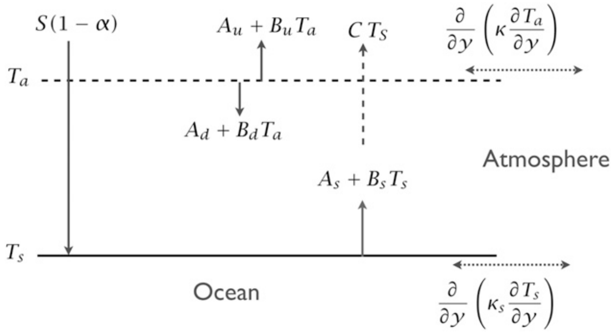

FIG. 13. Schema of the two-level energy balance model showing the incoming solar radiation, $S(1-\alpha)$, the various IR radiative and vertical energy transfer terms (involving the A, B, and C parameters; see Table 1), and the lateral energy transport.

$$C_s\dot{T}_s = S(1-\alpha) + (A_d + B_d T_a) - (A_s + B_s T_s) + \kappa_s \nabla^2 T_s \quad \text{and} \tag{12}$$

$$C_a\dot{T}_a = A_s + (B_s - C)T_s - (A_d + B_d T_a) - (A_u + B_u T_a) + \kappa_a \nabla^2 T_a, \tag{13}$$

where T_a and T_s are a midtropospheric and surface temperature, respectively. The physical parameterizations include the net incoming solar radiation $S(1-\alpha)$, the energy flux from surface to atmosphere $A_s + B_sT_s$, the energy flux from atmosphere to surface $A_d + B_dT_a$, the energy flux (IR) from atmosphere to space $A_u + B_uT_a$, the IR from surface escaping to space CT_s, and finally the lateral energy transport $\kappa_a\nabla^2 T_a$ for the atmosphere and $\kappa_s\nabla^2 T_s$ for the ocean. The model is written in spherical coordinates and the Laplacian is one dimensional, varying only in latitude. The physical parameterizations of this model are quite standard, and although the presence of a diffusive transfer to represent ocean fluxes cannot be regarded as realistic, it serves our purposes as a device for simulating lateral heat transport by the ocean. The diffusive approximation for the atmosphere is meant to account for the total atmospheric heat transport, but it plainly cannot differentiate the stationary and transient components and, in fact, is probably a better approximation for the transient component. Nor is a diffusive transport likely to be a good approximation at low latitudes. Given all these limitations, the applicability of the EBM results to the real system is, of course, limited.

Solutions are found by time stepping the model to equilibrium. A set of parameters that gives reasonable solutions for T_a and T_s is given in Table 1, and the incoming solar radiation distribution is the same as that used in the primitive equation ICCM.

To examine the issue of compensation in this model, a series of integrations are performed with differing amounts of oceanic heat transport affected by changing the oceanic lateral diffusivity κ_s. For each value of ocean diffusivity, the model was integrated to equilibrium, and the results are shown in Fig. 14. There is evidently a good degree of compensation in the total (atmospheric plus oceanic) heat flux PHT, although it is not perfect. Most theories of heat transport in the atmosphere suggest that the lateral heat transport will increase more rapidly than linearly with meridional temperature gradient (Stone 1978a; Held and Larichev 1996), and we incorporate this effect in the EBM by making atmospheric diffusivity proportional to the square of the temperature gradient between the equator and pole. The results from a corresponding sequence of experiments are shown in the lower two panels of Fig. 14, and we see that the compensation is now still better, and the atmospheric temperature (which largely determines the outgoing IR flux and so the total meridional heat transport) is almost constant over the entire range of ocean heat transports. The underlying reason for compensation in this framework comes from the efficiency of the AHT, which maintains a fairly flat temperature profile in the atmosphere. It is in this limit that Stone's arguments regarding the insensitivity of the total meridional energy flux to parameter variations elsewhere in system can be expected to hold.

TABLE 1. Values of radiative parameters for the energy balance model of Fig. 13. The units of the B and C parameters are $\mathrm{W\,m^{-2}\,K^{-1}}$ and the units of C_a and C_s are $\mathrm{J\,m^{-2}\,K^{-1}}$. Note that C_s corresponds to a 60-m-deep slab.

Parameter	B_d	B_u	B_s	C	C_a	C_s
Value	11.3	2.83	10.4	0.54	1.2×10^7	3.68×10^8

However, the PHT does change if the atmospheric diffusivity itself changes and the ocean is not as well able to compensate. This is seen in Fig. 15, which shows the results from a similar sequence of experiments but now with the atmospheric diffusivity changing and the ocean diffusivity held fixed. Although OHT increases as the atmospheric temperature falls, it does not do so sufficiently rapidly to fully compensate for the change in AHT, even though for most of the experiments in the sequence shown the oceanic transport is larger than that of the atmosphere. This is due to the fact that the outgoing longwave radiation is determined mainly by the atmospheric temperature because most of the longwave radiation emitted at the surface is absorbed by the atmosphere. Changes in OHT affect the ocean temperature and the atmospheric temperatures only indirectly, and changes in OHT do not immediately affect the energy balance. If both atmospheric and oceanic diffusivity are increased, the meridional energy transport also increases, albeit more slowly than linearly.

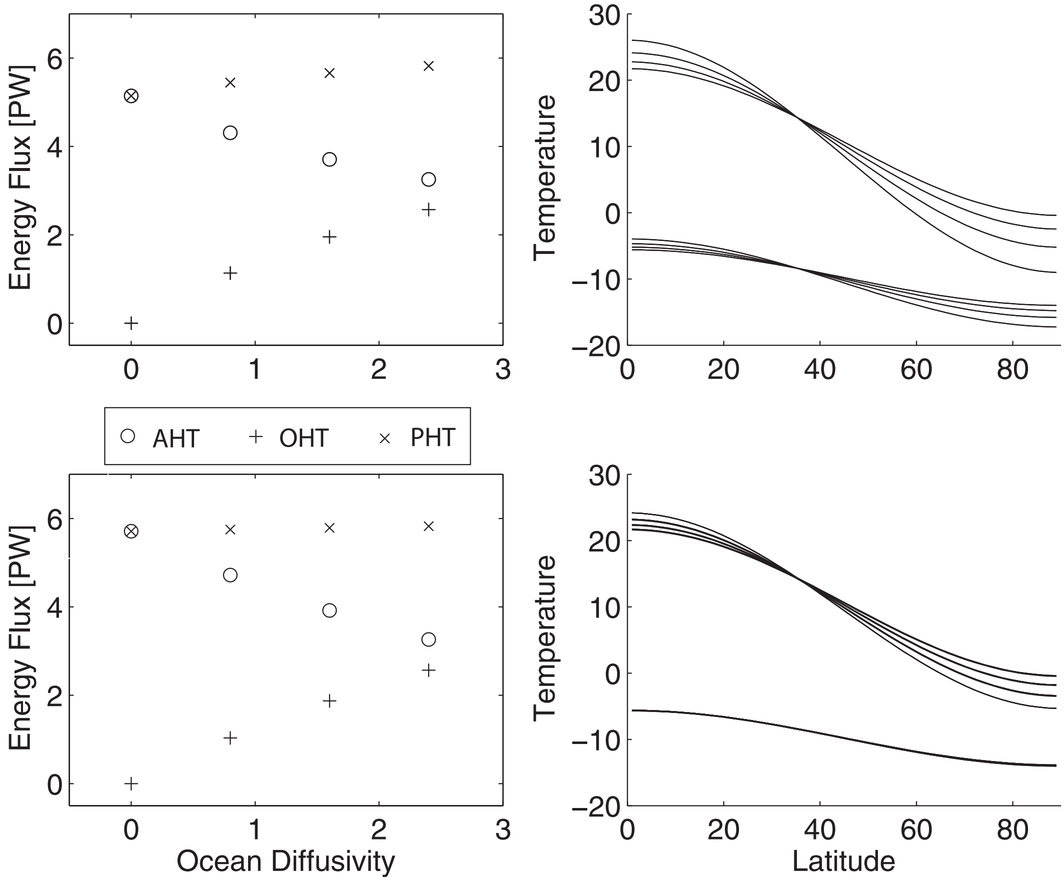

FIG. 14. Results from a two-level EBM. (top left) Atmospheric, oceanic, and total heat transport as a function of ocean diffusivity ($\kappa_s = 10^{13} \times$ axis value; $kg^2\,s^{-1}$) with constant atmospheric diffusivity of 8×10^{13}. (top right) Surface (upper set of curves) and atmospheric (lower curves) temperatures for each integration with the largest pole–equator temperature gradient corresponding to the smallest diffusivity. (bottom left),(bottom right) As in (top left) and (top right), respectively, but for where the atmospheric diffusivity varies with the square of the pole–equator temperature difference.

Moreover, variations in meridional energy transport typically occur on the daily-to-yearly time scales through changes in the atmospheric circulation and on the yearly-to-centennial (and possibly longer) time scales mainly through changes in the oceanic circulation. Whereas the atmosphere is able to reach a new quasi-equilibrium in a year or less in response to oceanic changes, we certainly cannot expect the ocean to respond quickly enough to compensate for the fast atmospheric variations. Thus, on the decadal scale, we expect that variations in the total energy transport will be positively correlated with, but smaller than, variations in oceanic energy transport, and variations in atmospheric transport will be negatively correlated with both oceanic and total energy transport variations, as discussed in the previous sections.

Finally, we should note that, given the simplifications of the EBM, its relevance to the real atmosphere–ocean system is limited and the results interpreted with care. The results demonstrate that, given a diffusive atmosphere and simple radiation scheme, compensation readily occurs (as might be expected). Nevertheless, we *can* change the TOA radiative balance by changing model parameters (as we can with a dynamical model), suggesting that an invariance of the TOA radiative balance is not a necessary condition for compensation. Drawing any stronger conclusions about the real atmosphere would not be warranted.

5. Discussion and conclusions

In a seminal paper, Bjerknes (1964) argued that thermal equilibrium at each latitude belt can be assumed for interannual and longer periods. This assumption requires both that anomalies in the incoming and outgoing radiation should balance and that heat storage does not vary. It follows that, for a constant top-of-atmosphere (TOA) radiative flux, the total meridional energy transfer (PHT) in the coupled atmosphere–ocean system is also constant,

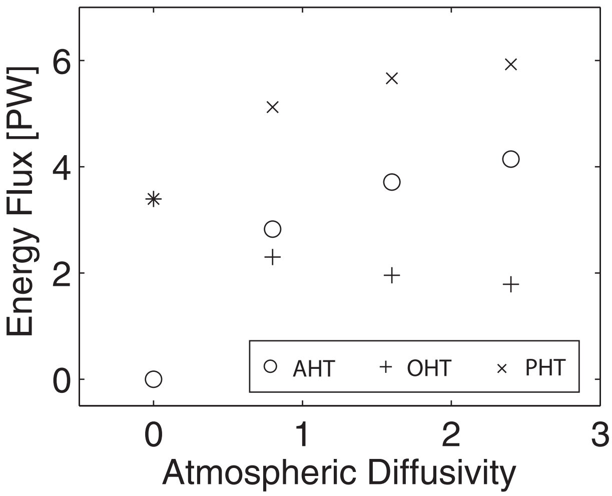

FIG. 15. As in Fig. 14, but as a function of atmospheric diffusivity ($\kappa_a = 5 \times 10^{13} \times$ axis value; $kg^2 s^{-1}$) with constant ocean diffusivity of 1.6×10^{13}.

and any internal natural fluctuation in atmospheric (AHT) and oceanic (OHT) heat transport will be equal and of opposite sign. These two necessary conditions, a constant TOA flux and small ocean heat storage, were taken as prerequisites for Bjerknes compensation. Anticorrelated meridional energy fluxes between ocean and atmosphere have already been described for the Atlantic sector in coupled climate models (Shaffrey and Sutton 2006; van der Swaluw et al. 2007) and are associated with fairly constant TOA radiative fluxes and oceanic heat storage.

The aim of this paper was to make further progress in the understanding of the degree, nature, and time scale of compensation between meridional atmospheric and oceanic heat transports, as well as the origin of the decadal-scale anomalies. We have investigated this problem using two coupled climate models of different complexity, the GFDL CM2.1 and the Intermediate Complexity Coupled Model (ICCM), as well as a simple energy balance model (EBM). We partitioned the oceanic contribution into a *warm* and a *cold* cell transport based on a potential temperature class definition. This partitioning roughly separates dynamically the circulation into wind-driven gyres and the deep meridional overturning circulation. We also separated the atmospheric transport into latent and dry static energy fluxes. Our main conclusions can be summarized as follows:

(i) As in previous modeling studies, at decadal time scales and for the extratropical Northern Hemisphere, opposite fluctuations in energy fluxes of the atmosphere and ocean are present in both CM2.1 and ICCM, such that the atmosphere is compensating, but imperfectly, for variations in the oceanic transport. In both comprehensive and intermediate complexity coupled models, climate variability on decadal and longer time scales stems from variability in the AMOC. The ocean overturning circulation in our simulations is mildly unstable, producing variations in heat transport with a period of about 20 years. The atmosphere is able to compensate fairly well for this, so allowing the atmosphere and ocean to oscillate with opposite phases, keeping the overall equator to pole heat transport approximately (but not exactly) constant. The variability of PHT is positively correlated with OHT and is much smaller than the variability in either the atmosphere or ocean alone. As also shown in Farneti and Vallis (2011), the ocean may be thus considered to force the atmospheric variability at interannual and longer time scales, provided that significant AMOC oscillations, responsible for strong meridional heat flux anomalies, are present. The compensation, and hence negative correlation, between energy transports depends on the time scale considered. Compensation is not significant at the annual period, but it increases steadily, becoming significant for interannual to decadal time scales and finally saturating at the preferred period of variability in the ocean. Both models used in this study present a significant MOC variability with a period of 20 years; hence, they both saturate at that time scale. Models with a different spectral peak are thus expected to saturate at different periods. There is obviously some uncertainty when trying to quantitatively extrapolate these results to the real climate system since the frequency and amplitude of the low-frequency variability of the AMOC are still largely unknown, as is the strength of the atmospheric response.

(ii) Results from our models also show that compensation does not hold if the variability in the heat transport in the ocean is driven by the surface winds. This is the case for the oceanic heat transport in the wind-driven gyres, which was found to vary in unison with the atmospheric transport. Large oscillations in either system are then harder to maintain because the radiative balance of the system would then have to covary with the oscillations. Theoretical arguments presented in Vallis and Farneti (2009) suggested that the energy transports in the atmosphere and wind-driven gyres ocean should in fact scale in similar ways. A consequence of this relation is that AHT and the OHT by the wind-driven gyres will covary on interannual and longer time scales, as found in both CM2.1 and ICCM. Moisture and sensible heat transports in the atmosphere are also positively correlated at decadal time scales in our models, consistent with observations.

(iii) Finally, and aided by the use of an EBM, we suggest that compensation between AHT and OHT decadal anomalies may be interpreted to be a result of an efficient (e.g., diffusive or superdiffusive) energy transport in the atmosphere that is able to constrain the meridional temperature distribution that at least in part determines the outgoing infrared radiation. Thus, if the ocean heat transport were to increase then the meridional temperature gradient of the sea surface temperature would fall and the atmosphere would respond to that, reducing its own heat transport. The degree of compensation would depend on how efficiently the atmosphere responds to changed surface temperatures and, given the possible changes in atmospheric vertical structure, is unlikely to be perfect.

The EBM, of course, is but a crude simplification of the true physical and dynamical ocean–atmosphere system. Many aspects of the circulation are poorly modeled, with cloud and radiative properties and the low-latitude circulation being among the most egregious. In such a simple diffusive framework, compensation can hardly fail to result, and the EBM results should not be overinterpreted. Still, these results and those obtained with the dynamical models suggest that the approximate invariance in TOA radiative fluxes is a consequence of the efficiency of the atmospheric dynamics rather than being the root cause of the compensation. Indeed, previous work has shown that the TOA radiative fluxes can be substantially changed by changing parameters in a dynamical coupled atmosphere–ocean model so that radiative invariance cannot be a truly fundamental constraint (Vallis and Farneti 2009). It is for these reasons that we suggest that compensation results from the efficient meridional transport of heat in the atmosphere, rather than from an a priori need for a constant TOA radiative balance.

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