Ocean Heat Transport and Its Projected Change in CanESM2

Duo Yang and Oleg A. Saenko

Canadian Centre for Climate Modelling and Analysis, Environment Canada, Victoria, British Columbia, Canada

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ABSTRACT

The meridional ocean heat transport (MOHT), its seasonal variability, and projected changes simulated by the second generation Canadian Earth System Model (CanESM2) are presented. The global mean MOHT is within the uncertainty of the observational estimates. However, a correct simulation of the MOHT for individual ocean basins is more challenging, and the Atlantic MOHT south of 30°N is underestimated. The partitioning of the MOHT into the overturning and gyre components is generally consistent with such partitioning in an observationally optimized ocean model. At low latitudes, the time-mean MOHT is dominated by its overturning component, whereas in the Southern Ocean and, especially, in the subpolar North Atlantic, it is the gyre component that plays a more important role. In the projected warmer climates, CanESM2 simulates a weakening of the poleward MOHT essentially in both hemispheres. The projected MOHT changes are largely determined by the overturning component, except in the subpolar Atlantic where it is dominated by the gyre component. Consistent with (the limited number of) previous studies, the seasonal variability of the MOHT is large and is mostly driven by the seasonal variability of the meridional Ekman transport. In the simulated warmer climates, the seasonal cycle of the MOHT is projected to change, mostly in the tropics and also in the Southern Hemisphere midlatitudes. The eddy contribution to the MOHT is broadly consistent with that in the observationally optimized eddy-permitting model. However, in the tropics a significant fraction of the eddy energy is converted back to the mean circulation, and the heat transports due to the parameterized and permitted eddies differ.

1. Introduction

The poleward energy transport within the ocean–atmosphere system balances the surplus of radiation at low latitudes and its deficit at high latitudes. Early studies have argued that the ocean’s contribution to this net heat transport is comparable to that of the atmosphere (e.g., Hsiung 1985; Trenberth and Solomon 1994). However, more recent estimates suggest that, outside of low latitudes, the meridional ocean heat transport (MOHT) is relatively weak (Keith 1995; Trenberth and Caron 2001, hereafter TC2001) but, nevertheless, still has a significant effect on climate (Herweijer et al. 2005; Wunsch 2005; Saenko 2009). The time-mean MOHT has therefore long been a focal point, and its direct and indirect estimates have been converging (Bryden and Imawaki 2001; Trenberth and Solomon 1994; Keith 1995) with the peak values in the tropics of about ±2 PW. The direct estimates are primarily based on individual hydrographic sections or on observationally constrained inverse models (e.g., Ganachaud and Wunsch 2003, hereafter GW2003). The indirect estimates calculate the MOHT as a residual between the net ocean–atmosphere heat transport, which can be derived from the observed radiation at the top of the atmosphere, and the heat transport within the atmosphere, which is typically estimated from atmospheric data assimilation products (TC2001). Much like the direct estimates, the indirect heat transport estimates may have large uncertainties (Bryden and Imawaki 2001), depending on the models employed and data assimilation procedures used (TC2001; see also Figs. 1a,c,d).

The observational estimates of MOHT provide valuable information for evaluating climate system models. Our focus here is on the MOHT simulated by one such model—the second generation Canadian Earth System Model (CanESM2.) This is a much improved version of the coupled model described in Flato et al. (2000) (see section 2), which has been employed recently to conduct climate change simulations in support of the Fifth Assessment Report of the Intergovernmental Panel on Climate Change (IPCC AR5). The first goal is to compare the CanESM2 MOHT with some existing observational

 Corresponding author address: Duo Yang, CCCma, 3800 Finnerty Road, Victoria BC V8P 5C2, Canada.
 E-mail: duo.yang@ec.gc.ca

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estimates. However, the data from hydrographic sections are often too sparse to evaluate the eddy component of the MOHT. While CanESM2 does not resolve mesoscale variability in the ocean, it has the effect of mesoscale eddies (on the large-scale ocean transport) parameterized. As such, we will also compare the MOHT simulated by CanESM2 with the MOHT derived from an observationally optimized eddy-permitting simulation (see section 2).

In contrast to the time-mean MOHT, the seasonally varying MOHT has drawn less attention. The available estimates of the seasonal MOHT differ in magnitude, and even in sign at some latitudes. Depending on the method used, the peak-to-peak values in the tropics can vary from about 4 to 8 PW (e.g., Bryan and Lewis 1979; Carissimo et al. 1985; Levitus 1987; Hsiung et al. 1989; Adamec et al. 1993; Ghirardelli et al. 1995; Jayne and Marotzke 2001). However, it is possible that the seasonal cycle of the MOHT is driven by rather simple dynamics (Jayne and Marotzke 2001), which may help in reducing this uncertainty. In particular, based on some previous theoretical and modeling results (Anderson et al. 1979; Willebrand et al. 1980), Jayne and Marotzke argue that seasonal fluctuations in the upper-ocean Ekman flux are compensated by flows, which are largely barotropic. The associated seasonal overturning circulation, as well as the temperature difference between the shallow Ekman layer and the compensating deep flow, can give rise to a large seasonal cycle of the MOHT. Therefore, it seems reasonable to expect (Jayne and Marotzke 2001) that the seasonal variability of the MOHT is, to the leading order, captured by the so-called Ekman heat transport $Q_{Ek}$ (e.g., Kraus and Levitus 1986):

$$Q_{Ek} = - \int c \frac{\tau_f}{f} (T_{Ek} - \langle |\theta| \rangle) \, dx,$$  (1)
where the zonal integral is taken from coast to coast, \(c_p\) is the specific heat of seawater, \(f\) is the Coriolis parameter, \(\tau_z\) is the (time varying) zonal wind stress, \(T_{Ek}\) is the temperature of the Ekman layer, and \(\theta\) is the potential temperature of seawater; \(\langle \cdot \rangle\) represent zonal averaging and \([\cdot]\) represent vertical averaging. While there have been attempts to estimate the contribution of the Ekman heat transport to the time-mean MOHT (Bryan 1962; Bryden and Hall 1983; Kraus and Levitus 1986; Levitus 1987; Adamec et al. 1993; Ghirardelli et al. 1995), Jayne and Marotzke (2001) emphasized that Eq. (1) is applicable for estimating the MOHT only on seasonal (and shorter) time scales. On time scales comparable with (or larger than) the time it takes the first-mode baroclinic Rossby waves to cross the ocean, the nature of the flow compensating for the Ekman mass flux in Eq. (1) is unlikely to be barotropic. This follows from the theories and spinup calculations addressing oceanic response to perturbations in the wind field (e.g., Willebrand et al. 1980; Young 1981). Based on these studies, it can be concluded that the initial, essentially depth-independent, response is set by the fast barotropic mode. This is followed by a much slower adjustment associated with propagation of baroclinic modes. The phase (or group) speed of the first-mode baroclinic Rossby waves strongly depends on latitude (Gill 1982, p. 503), so it may take from years (at low latitudes) to decades (at midlatitudes) for these waves to propagate across an oceanic basin. Without changing the net vertically integrated mass transport, the baroclinic modes confine the interior circulation response to the upper ocean.

Observations do reveal relatively strong, time-varying currents in the deep ocean that appear to be driven by time-varying winds (Koblinsky and Niiler 1982; Niiler and Koblnsky 1985; Brink 1989; Niiler et al. 1993). Modeling results also suggest that the directly wind-forced currents can penetrate to large depths on seasonal time scales (Willebrand et al. 1980; Jayne and Marotzke 2001; Saenko 2008). This supports the suggestion that a reasonably good estimate of the seasonal MOHT can be obtained based on Eq. (1). It assumes that the interior transport that compensates for the Ekman transport in the upper ocean can penetrate to the very cold abyss. This, jointly with the fact that a basin-integrated meridional Ekman flux can be of the order of 100 Sv (Sv = \(10^8\) m\(^3\) s\(^{-1}\)), transporting waters with temperatures of some 20°C, supports the suggestion that the amplitude of the seasonal MOHT is likely to be much larger than that of the time-mean MOHT (Bryan and Lewis 1979; Bryan 1982b; Levitus 1987; Hsiung et al. 1989; Boning and Herrmann 1994; Nakano et al. 1999; Jayne and Marotzke 2001). Studies that are based on ocean-only models do support the notion that the Ekman heat transport, as given by Eq. (1), can represent seasonal variability of the MOHT reasonably well (Jayne and Marotzke 2001). However, such models are constrained by the specified surface forcing fields (Jayne and Marotzke 2001), often using monthly heat fluxes and/or restoring sea surface temperature to climatology. CanESM2, on the other hand, is a fully coupled earth system model. This means that, in particular, it can simulate the atmosphere–ocean feedbacks that are not present in ocean-only simulations. Hence, our second goal is to examine the applicability of the heat transport dynamics described by Jayne and Marotzke to a fully coupled model, such as CanESM2.

In a warmer climate, the thermal structure of the ocean and its circulation (including the seasonal component) could change. Hence, our third goal is to evaluate the changes in the time-mean and seasonal MOHT’s under some of the scenarios developed for IPCC AR5 (due for publication in 2013/14).

2. Models, observations, and simulations

CanESM2 is composed of ocean, sea ice, atmosphere, land, and carbon cycle models. The ocean component of CanESM2 is a version of the National Center for Atmospheric Research (NCAR) Community Ocean Model (Gent et al. 1998), which was developed from version 1 of the Geophysical Fluid Dynamics Laboratory Modular Ocean Model. It has a horizontal resolution of 1.41° × 0.94° (longitude × latitude), and there are 40 vertical levels with spacings ranging from 10 m near the surface to nearly 400 m in the deep ocean. The ocean model employs anisotropic viscosity (Large et al. 2001) and eddy transport (Gent and McWilliams 1990) parameterizations; the eddy transfer coefficient in the Gent and McWilliams (1990) scheme is set to \(10^3\) m\(^2\) s\(^{-1}\). The ocean model also accounts for the effect of the dissipation of internal tides on deep ocean vertical mixing, implemented in a fashion similar to that of Simmons et al. (2004). Vertical mixing driven by buoyancy and shear is parameterized by the K-profile parameterization (KPP) scheme (Large et al. 1994). The atmospheric model uses the spectral transform method with T63 resolution in the horizontal and has 35 vertical levels (von Salzen et al. 2012, manuscript submitted to Atmos.–Ocean). Wind stress is calculated using a wind-speed-dependent drag coefficient corrected for atmospheric stability (Abdella and McFarlane 1997). The sea ice component uses a cavitating-fluid rheology (Flato and Hibler 1992), combined with the thermodynamics described in McFarlane et al. (1992). Coupling with terrestrial ecosystem and ocean carbon models (see Christian et al. 2010, and references therein) enables some important biogeochemical
processes to be represented and feed back on the physical climate.

The CanESM2 historical simulations (1850–2005) serve as our base experiments. The five ensemble members are spawned from a long (>1000 yr) preindustrial control run at an interval of 50 years. Two groups of climate projections (2006–2100) were initiated from the end of each historical run. These are based on the newly developed representative concentration pathways (RCPs) of radiatively important greenhouse gases (http://www.pik-potsdam.de/~mmalte/rcps/index.htm), forced with specified concentrations following the medium mitigation scenario (RCP4.5) and the high emission scenario (RCP8.5). The design of the climate change experiments is part of the Coupled Model Intercomparison Project Phase 5 (details of which can be found online at http://cmip-pcmdi.llnl.gov/cmip5/). The analyses for RCP4.5 and RCP8.5 are based upon the last 10 years of the corresponding CanESM2 simulations (2091–2100), whereas the CanESM2 historical simulations are analyzed for the period of 1992–96. The former period ensures that the results, while being statistically robust, are close to the targeted radiative forcing at year 2100. The 1992–96 period, in turn, is to accommodate the observations that are based primarily upon the World Ocean Circulation Experiment (WOCE) (GW2003). It should be noted that the main conclusions remain essentially the same when 20-yr periods (1981–2000, 2081–2100) are chosen for the analysis of the CanESM2 simulations.

As part of the CanESM2 evaluation, we also employ ocean model data from the Estimating the Circulation and Climate of the Ocean project, phase II (ECCO2), which are available from 1992 onward. The ECCO2 data are obtained from a freely running and optimized high-resolution simulation based on the Massachusetts Institute of Technology general circulation model (Marshall et al. 1997). The model parameters have been optimized using the Green function approach (Menemenlis et al. 2005), with the baseline integration derived from a globally optimized simulation (Menemenlis et al. 2008). The surface forcing was derived from the Japanese 25-year Reanalysis Project (Onogi et al. 2007) and converted to wind stress and buoyancy forcing using a bulk formula. The ECCO2 model employs a cube-sphere grid projection with mean horizontal grid spacing ~18 km. The model data used here have been optimally interpolated (and, for vectors, rotated) onto a 0.25 × 0.25 latitude × longitude grid. To ensure mass conservation, we have also applied a uniform adjustment to the interpolated ECCO2 velocity; the computed MOHT was found to be insensitive to such an adjustment. ECCO2 has 50 vertical levels with spacings ranging from 10 m near the surface to 456 m near the bottom. The variables of interest (ocean meridional velocity and potential temperature) are available online (see http://ecco2.jpl.nasa.gov/products/).

Our focus here is on the advective MOHT, which is estimated according to

\[
\text{MOHT} = \int \rho_0 c_p \theta v \, dx \, dz,
\]

where \(v\) is the meridional velocity and \(\rho_0\) is the reference density. In the case of CanESM2, \(v\) includes the eddy-induced component via the Gent and McWilliams (1990) parameterization. In addition, we will decompose the MOHT into overturning and gyre components following the approach of Bryan (1982a). The overturning heat transport is calculated as the integral of the product \(\rho_0 c_p \theta \langle \nu \rangle\), whereas the gyre component is calculated as the integral of the product \(\rho_0 c_p \theta^* \nu^*\) (where \(\theta^* = \theta - \langle \theta \rangle\) and \(\nu^* = v - \langle v \rangle\)). In the case of CanESM2, monthly mean \(v\) and \(\theta\) are utilized (the use of more frequent datasets was found to be nonessential at the relatively coarse resolution employed in CanESM2). For the corresponding calculations at the eddy-permitting resolution (i.e., in the case of ECCO2) 3-day averages of \(v\) and \(\theta\) are used. The heat transports are calculated for the Atlantic–Arctic and Indo-Pacific basins, as well as for the global ocean.

3. Time-mean ocean heat transport

We begin by comparing the time-mean MOHT in CanESM2 with observations (Bryden 1993; Klein et al. 1995; Speer et al. 1996; Lavin et al. 1998; Macdonald 1998; Holfort and Siedler 2001; Hobbs and Willis 2012; TC2001; GW2003), as well as with the mean MOHT estimated from the ECCO2 data. This is followed by the analysis of the possible future changes in the MOHT projected by CanESM2. The simulated mean MOHT then serves as the reference for the time-varying (seasonal) MOHT, discussed in the next section. The ensemble standard deviation of the CanESM2 heat transport has a maximum value of 0.1 PW, found in the tropics. This is smaller than typical error bars in observational estimates. Unless stated otherwise, the analysis below is based on the ensemble mean MOHT.

a. Historical simulation

The time-mean MOHT simulated by CanESM2 for the global ocean is shown in Fig. 1a. It agrees with the MOHT estimated by GW2003 within the error bars. The peak values of the global northward heat transport in the CanESM2 ocean range from ~1.1 PW around 12°S to 1.6 PW around 21°N (Fig. 1a). Hence, the corresponding maximum heat divergence from the low-latitude ocean, which ultimately must balance the heat gained by the
tropical ocean at the surface, is 2.7 PW. Apparently, the corresponding value is lower in ECCO2 (Fig. 1b) during the period of interest here. This could be due to the constraint imposed on the surface heat flux in ECCO2. It could also result from the stronger heat convergence associated with the gyre component in the tropical ECCO2 as discussed in the next paragraph. In addition, it may be related to the strong eddy contribution to the MOHT in ECCO2, particularly at low latitudes where it tends to oppose the poleward heat transport by the mean currents. The eddy heat transport will be discussed in section 5.

In both CanESM2 and ECCO2, most of the surface heat gain by the low-latitude ocean is diverged poleward by the overturning component of the MOHT (Figs. 1a,b). The gyre component opposes this process between about 12°S and 20°N in CanESM2 and between about 20°S and 20°N in ECCO2 (Figs. 1a,b), mostly carrying heat toward the tropics. This means, according to the gyre heat transport definition, that the equatorward (poleward) circulation anomaly (relative to the zonal mean velocity) typically carries positive (negative) temperature anomaly in the tropics. The maximum heat convergence into the tropical ocean by the gyre component (i.e., the corresponding peak-to-peak difference) is somewhat larger in ECCO2 compared to that in CanESM2. It is also interesting to note that in both models the gyre component dominates the net poleward MOHT in the Southern Ocean and, especially, around 50°N. We shall touch on the latter below when addressing the heat transport components in the Atlantic.

While the global ocean MOHT simulated by CanESM2 is generally within the GW2003 error bars, it is of interest to partition it into the contributions from the Atlantic–Arctic and Indo-Pacific basins. The corresponding MOHTs are compared in Figs. 1c,d with the MOHT estimates based on the WOCE and some pre-WOCE hydrography; the estimates from the recent continuous observations derived from the Rapid Climate Change (RAPID) array (Johns et al. 2011), and the Argo data (Hobbs and Willis 2012) are also presented. We note that GW2003 have documented the data consistency between pre-WOCE and the WOCE periods. It should also be pointed out that over the last one to two decades, there seem to have been only weak statistically significant trends in the large-scale ocean circulation and heat transport (e.g., Wunsch and Heimbach 2006, 2009).

In agreement with the observations, the Atlantic–Arctic heat transport in CanESM2 is directed northward at all latitudes (Fig. 1c). This appears to be due to the circulation associated with the formation of North Atlantic Deep Water (NADW) (e.g., Bryden and Imawaki 2001; TC2001). However, our analysis indicates that, while the global ocean heat transport can be simulated reasonably well, a correct simulation of the MOHT for individual ocean basins may present a challenge (Figs. 1c,d). Essentially, CanESM2 underestimates the observed northward heat transport through most of the Atlantic–Arctic basin, particularly south of about 30°N (Fig. 1c). This seems to be related to the simulated NADW being somewhat too weak and too shallow (Fig. 2a), as compared to observational estimates (e.g., see Fig. 2b in Talley et al. 2003). The associated meridional overturning circulation operates on the weaker than observed temperature contrast in the upper 2 km of the ocean (Figs. 2a,b). In addition, in the Atlantic abyss (below about 2 km) the recirculation of Antarctic Bottom Water operates on the stronger than observed temperature contrast (Figs. 2a,b). This also contributes to the bias in the CanESM2 Atlantic heat transport. In the Indo-Pacific, the model tends to underestimate the southward heat transport around 20°S (Fig. 1d) (although the corresponding estimate is still within the error bars). Thus, the correspondence between the simulated and observed global ocean MOHT in the Southern Hemisphere (Fig. 1a) may, in part, be a result of the biases in the MOHTs simulated for the individual ocean basins that tend to cancel each other in the global mean (Figs. 1c,d). However, it should be noted that the uncertainties in the observational estimates are large, particularly in the Southern Hemisphere.

In the Northern Hemisphere, the situation is somewhat more encouraging. In particular, at 24°N in the Atlantic, which has perhaps the most abundant observations, the CanESM2 MOHT is in good agreement with the observational estimates of Macdonald (1998) and Bryden (1993). It is, however, somewhat weaker than the MOHT estimated for this latitude by GW2003 and Lavin et al. (1998). Around 50°N in the Atlantic–Arctic basin (and also in the global mean) there is suddenly an increase in the CanESM2 MOHT (mainly due to the gyre). This is consistent with GW2003, whereas the indirect estimates of TC2001 do not show this increase. In the North Pacific, the simulated MOHT is in good agreement with the available observational estimates (Fig. 1d), although the density of observations is low compared to the North Atlantic.

Much as for the global ocean, it is of interest to partition the MOHT into the overturning and gyre components for the individual basins. In the Atlantic the overturning component transports about 80% of the heat between the equator and 40°N, reaching a maximum of 1.0 PW at 14°N. While this is somewhat larger than the modeled value in Gulev et al. (2003), our Atlantic overturning-to-gyre heat transport ratio in CanESM2 appears to be lower than that estimated by Johns et al. (2011). In particular, using continuous array data, Johns
et al. report that about 90% of the MOHT (1.33 ± 0.14 PW) across 26.5°N in the Atlantic could be due to the overturning component. In the South Atlantic the gyre component becomes important around 10°S, whereas at 30°S it is the overturning component that dominates the northward MOHT (cf. GW2003, their Fig. 8).

However, in the subpolar Atlantic–Arctic, it is the gyre component that takes the leading role and contributes more than 85% to the net northward heat transport in CanESM2 (Fig. 1c). This is in general agreement with the corresponding model-based estimates of Gulev et al. (2003, see their Fig. 2b). One reason for the dominance of the gyre component over the overturning component in the subpolar Atlantic is that the North Atlantic Current carries relatively warm waters northeastward. Upon giving their heat to the atmosphere, some of these waters return southward along the western boundary, thereby contributing to the gyrelike heat transport in the subpolar Atlantic.

In the subpolar Indo-Pacific basin, both gyre and overturning heat transports simulated by CanESM2 are relatively weak (Fig. 1d). As a result, the net MOHT north of 40°N is also weak in this basin, in agreement with the (limited number of) observational estimates. In the low-latitude Indo-Pacific, the partitioning between the gyre and the overturning components of the MOHT follows that in the global ocean (Figs. 1d,a). This is expected given the size of the Indo-Pacific basin. The net MOHT in the Indo-Pacific is within the GW2003 error bars at 24°N, 18°, and 30°S (Fig. 1d). We also note a reasonably good agreement between the simulated and observed MOHTs at 24°N in the Pacific. This typically requires a well-simulated Kuroshio (Wilkin et al. 1995), which drives the dominant gyre component. Indeed, replacing the model-simulated temperature with the Levitus climatology gives a similar estimate of the MOHT at 24°N.

b. Projected changes

The evaluation using observations suggests that CanESM2 can be a useful tool for projecting the MOHT under climate change scenarios. Examples of such projections, derived for the RCP4.5 and RCP8.5 scenarios, are presented in Fig. 3. However, when making such projections one should keep in mind the accuracy of the MOHT simulated by the CanESM2. The analysis in the previous subsection shows that, for some latitudes, this
accuracy may not be satisfactory. This applies to the MOHTs simulated for the individual ocean basins, as well as for the global ocean.

Essentially, CanESM2 projects a weakening of the poleward heat transport in the global ocean in both hemispheres and across most latitudes (Figs. 3a and 4a,b). Under the two RCP scenarios, the MOHT changes are qualitatively similar (Figs. 4a,b). However, their magnitude does depend on the scenario. It is generally stronger in the RCP8.5 than in the RCP4.5 (Figs. 4a,b).

In terms of the relative contribution of the gyre and overturning components, while it varies between the basins and RCPs (Figs. 4c–f), in the low-latitude global ocean the projected MOHT anomaly is largely set by the overturning component, particularly under the RCP8.5 scenario (Fig. 4b). The same applies to the midlatitude ocean in the Southern Hemisphere where the strongest decrease (of the southward MOHT) is simulated around 50°S, that is, just north of the Drake Passage latitudes. This decrease reaches up to 0.2 PW in RCP4.5 (Fig. 4a) and 0.4 PW in RCP8.5 (Fig. 4b). It is related, at least in part, to the projected strengthening of the Southern Hemisphere midlatitude westerlies (Fyfe and Saenko 2005) and to the associated increase in the northward Ekman mass flux. In contrast, in the Northern Hemisphere midlatitudes (i.e., around 50°N), the MOHT anomaly is essentially maintained by the gyre component (Figs. 4a,b), mostly attributable to the anomaly in the gyre strength in the subpolar North Atlantic. North of about 65°N, the model projects an increase in the northward heat transport. This is largely due to the increase in the temperature contrast between the relatively warm northeastward flow of the Norwegian Current and the southward flow of the East Greenland Current.

It should be noted that had the CanESM2 resolved/permitted mesoscale eddies, the projected MOHT anomalies might have been somewhat different. However, the role of eddies in meridional heat transport, while certainly important in some ocean regions (see section 5), is still under debate, including in the Southern Ocean (Volkov et al. 2010).

Several interesting points can be made by considering the changes in the MOHT projected for the individual ocean basins. In particular, under the RCP4.5 scenario, the northward MOHT does not change much in the Indo-Pacific, particularly north of the equator, but does decrease considerably in the Atlantic–Arctic (Figs. 3b,c). Under the RCP8.5 scenario, however, the MOHT changes in the Indo-Pacific and Atlantic–Arctic basins are of comparable magnitude (Figs. 3b,c). This may seem surprising, given the difference in the zonal extent of the two ocean basins. However, some of the changes in the Atlantic–Arctic MOHT are related to the projected changes in the rate of the NADW formation, whereas deep-water formation is essentially absent in the North Pacific. By the end of the twenty-first century, CanESM2 projects a weakening of the Atlantic MOC (Figs. 2c,d), followed by its recovery once the concentration of greenhouse gases is held fixed at year 2100 (not shown). The weakening may reach up to 6 Sv at the latitudes of maximum overturning (~37°N in
CanESM2; Figs. 2a,c,d); however, it is only 1–2 Sv in the region of NADW outflow (between 20° and 30°S), which is within one standard deviation of its simulated variability.

Despite the projected changes in the Atlantic overturning circulation, the changes in the northern North Atlantic MOHT (~50°N) are dominated by the gyre component (Figs. 4e,f). Thus, while between about 0° and 40°N the dominance mostly belongs to the overturning component, it is the gyre component that maintains the projected MOHT change in the subpolar Atlantic. Furthermore, at some latitudes the magnitude of the MOHT overturning changes is larger in the Indo-Pacific than in the Atlantic–Arctic, particularly under the RCP8.5 scenario (Figs. 4d,f). Thus, despite its usefulness, it should be kept in mind that the decomposition of MOHT into the gyre and overturning components is purely geometric.

4. Seasonal cycle of the ocean heat transport

We now consider the seasonal variations in the MOHT, following some of the ideas and approaches presented in Jayne and Marotzke (2001) and applying them to the projected climate change. In addition to the estimates based on ECCO2, the seasonal cycle of the MOHT
simulated by CanESM2 is compared with two continuous observations in the Atlantic (Johns et al. 2011; Hobbs and Willis 2012). As we shall see, the amplitude of the seasonal MOHT is, generally, much larger than that of the time-mean MOHT. The dynamical reasons for this have been summarized by Jayne and Marotzke. Essentially they argue, based on the work of Willebrand et al. (1980) and some others, that for a basinwide forcing with periods from one day to one year, the vertical scale of the ocean’s response can be comparable to (and even larger than) the mean depth of the ocean. As a result, on seasonal time scales the wind-driven meridional overturning, including that in the tropical ocean, can penetrate to the abyss. This, combined with the large temperature contrast such an overturning operates on, leads to a large seasonal MOHT.

### a. Historical simulation

The seasonal cycle of the MOHT simulated by CanESM2, expressed as the difference between December–February (DJF) and June–August (JJA), is displayed in Fig. 5a. It has the largest amplitude in the tropical ocean, roughly between 20°S and 20°N, with the maximum reaching 9.9 PW. The MOHT in winter is typically of opposite sign to that in summer, directed from the summer to the winter hemisphere (the corresponding anomalies of meridional overturning circulation are shown in Fig. 6). This amplifies the seasonal cycle of the ocean heat transport in the tropics. Since the absolute value of the MOHT in boreal winter is larger than in boreal summer (austral winter), the corresponding peak north of the equator is somewhat larger than that south of it. In the subtropics, the seasonal MOHT is weaker than and of opposite sign to that in the tropics (Fig. 5a, see also Fig. 6). Most of the seasonal MOHT in CanESM2 is due to its overturning component (Fig. 5a) with the gyre component accounting for a relatively small fraction (except poleward of 45° where the seasonal variations of MOHT are small). This is in agreement with Bryan and Lewis (1979, see their Fig. 14b).

Essentially, all of these features are consistent with the seasonal MOHT derived from the ECCO2 data (Fig. 5b) and with the analysis presented in Jayne and Marotzke (2001, see their Fig. 10). However, the seasonal variations of the MOHT in the tropics simulated by CanESM2 are larger (by up to 40%) than those derived from ECCO2 (Figs. 5a,b), as well as those simulated by Jayne and Marotzke (2001). Perhaps the main reason for this is related to the difference in amplitude of the seasonal cycle.
of zonal wind stress (discussed below), as well as to the
effect of eddies in ECCO2. It could also be in part because
CanESM2 is a fully coupled model, whereas the ECCO2
and the Jayne and Marotzke (2001) results are based on
ocean-only simulations. In particular, the ocean model
employed by Jayne and Marotzke restores surface tem-
peratures to climatology and uses monthly surface
heat fluxes from a reanalysis product. This can affect
variation in the seasonal heat content in the ocean and
may have some implications for the seasonal MOHT.
We note, however, that, owing to the lack of adequate
observations, the amplitude of the seasonal MOHT is
quite uncertain (Jayne and Marotzke 2001, their Fig. 15).

Away from the equator (i.e., away from the region
of zero or near-zero Coriolis parameter), the seasonal
cycle of the MOHT in both CanESM2 and ECCO2
closely resembles the corresponding meridional Ekman
heat transport given by Eq. (1) (Figs. 5a,b). As we noted
above, a detailed dynamical explanation for this is pro-
vided by Jayne and Marotzke (2001). Essentially, the
close similarity between the seasonal MOHT and the cor-
responding Ekman heat transport emphasizes the driving
role played by the seasonal variations in the zonal wind
stress (Fig. 5c, see also Jayne and Marotzke 2001). Fur-
thermore, the smallness of the Coriolis parameter near
the equator significantly amplifies the corresponding me-
ridional Ekman volume transport (Fig. 5d). As a result,
the seasonal MOHT in the tropics is larger than else-
where in the ocean (Fig. 5a). It also follows from Fig. 5d
that the seasonal Ekman transport within 20° of the

FIG. 6. Anomaly (relative to the annual mean) of the global ocean meridional overturning
streamfunction (Sv) from the CanESM2 historical simulation (1992–96; the first ensemble
member) for (a) DJF and (b) JJA.
equator is larger in CanESM2 than in ECCO2. As noted above, this could be the key reason for the difference between the amplitudes of seasonal MOHTs simulated by the two models (Figs. 5a,b). The stronger seasonal Ekman volume transport at 5°–10°N, as compared to that at 5°–10°S (Fig. 5d), apparently contributes to the corresponding asymmetry in the seasonal MOHT (Fig. 5a) with respect to the equator.

It is also interesting to note that between roughly 45° and 60°S the seasonal MOHT and the Ekman volume transport, while small, are of opposite sign (Figs. 5a,d). This indicates—consistent with Jayne and Marotzke (2001)—the importance of the thermal forcing in setting the seasonal MOHT in this region. By decomposing the MOHT into the contributions from the time-varying velocity advecting time-mean temperature and time-mean velocity acting on the variations in temperature, etc., Jayne and Marotzke conclude that the velocity variations alone account for a majority (70% or more) of the MOHT variability over most latitudes. However, they found that temperature variations may contribute up to 80% of the MOHT variability at high latitudes, in particular between 45° and 60°S. Thus, the seasonal cycle of the thermal forcing could be an important factor in determining the seasonal cycle of the MOHT at high latitudes, especially in the Southern Ocean. However, within roughly 45° of the equator, except near 25°N and 25°S where the seasonal cycle of zonal wind stress is relatively weak (Fig. 5c), it is the seasonal cycle of wind stress that is the key driving force of the large-amplitude MOHT variability in both CanESM2 and ECCO2, again consistent with Jayne and Marotzke (2001).

For the Atlantic–Arctic basin, which has a higher observed data density than the Indo-Pacific, the annual cycles of the MOHT simulated by CanESM2 can be compared with the corresponding observational estimates (Johns et al. 2011; Hobbs and Willis 2012) at 26.5° and 41°N (Fig. 7). In general, the model simulates the shape of the cycle reasonably well but somewhat underestimates its amplitude, particularly in winter. It also follows that the seasonal MOHT in the Indo-Pacific basin is much stronger than in the Atlantic–Arctic basin (Figs. 8a,b). This is expected given the difference in the width between the two basins as well as the somewhat stronger seasonal cycle of the wind stress in the former. As a result, the dependence of the seasonal MOHT in the Indo-Pacific basin on latitude follows that in the global ocean (Figs. 8a,b).

b. Projected changes

In the warmer climates, CanESM2 projects the largest changes in the seasonal MOHT in the tropical ocean, as well as in the Southern Hemisphere midlatitude ocean (Fig. 8a). Most of these changes can be explained by the changes in the seasonal cycle of zonal wind stress (Fig. 8c) or by the corresponding changes in the meridional Ekman volume transport (Fig. 8d). In particular, the peak in the seasonal Ekman volume transport at 5°–10°N weakens (Fig. 8d), and so does the corresponding peak in the seasonal MOHT (Fig. 8a). The same applies to the region around 40°S, where the projected changes in the seasonal MOHT essentially follow changes in the seasonal variability of the zonal wind stress and, hence, the changes in the meridional Ekman volume transport (Figs. 8a,c,d).

At 5°–10°S, however, the seasonal Ekman volume transport is projected to decrease less than that at 5°–10°N (Fig. 8d), whereas the seasonal changes of the MOHT are larger at 5°–10°S than at 5°–10°N (Figs. 8a,d). The asymmetric response in MOHT is attributable to the overturning component. This suggests that in warmer climates, the changes in ocean thermal structure can play...
quantitatively, the model simulates that, at 5°–10°N, the seasonal cycle of MOHT decreases by 5% in RCP4.5 and by 9% in RCP8.5, whereas at 5°–10°S it increases by 14% and 23%, respectively. As expected, it is the Indo-Pacific basin that contributes most to the changes in the global ocean seasonal MOHT (Figs. 8a,b). At some latitudes, most notably around 40°S, the model simulates both changes in magnitude of the local MOHT peak and a shift (by several degrees) in its mean position. The latter is apparently maintained by the shift in the seasonal wind stress and by the corresponding shift in meridional Ekman flux (Figs. 8a,c,d).

5. Eddy heat transport

The horizontal resolution in the CanESM2 ocean is not high enough to resolve mesoscale eddies. Instead, their effect on the large-scale circulation is parameterized (see section 2). Since our focus is on the heat transport, it is important to evaluate how closely the eddy heat transport simulated by CanESM2 agrees with the eddy-permitting models. Figure 9 shows the contribution of eddies to the simulated MOHT in CanESM2 and ECCO2 (cf. Fig. 2 in Jayne and Marotzke 2002). The midlatitude eddies, permitted in ECCO2 and parameterized in CanESM2, transport heat poleward at a comparable rate. It should be noted, however, that at midlatitudes the baroclinic Rossby radius is comparable to the ECCO2 resolution. Therefore, the eddy heat transport simulated by ECCO2 at those latitudes may represent only a lower limit. In the tropics, however, the ECCO2 eddies transport heat equatorward, and the magnitude of this transport is much larger than that in CanESM2. This apparently arises due to the correlation between $v'$ and $\theta'$ that favors the eddy heat convergence toward the tropics (Fig. 9), thereby contributing to the somewhat weaker net poleward heat divergence in the ECCO2 tropical ocean than in the CanESM2 ocean (Figs. 1a,b).

A possible reason for the difference between the heat transports due to the parameterized and permitted eddies in the low-latitude ocean could be related to the fact that the Gent and McWilliams (1990) scheme implies a transfer of available potential energy from mean flows to fluctuations, including locally. While mesoscale
Eddies are expected to remove potential energy from the large-scale circulation on global mean, this may not always hold locally, particularly in the tropics (Jayne and Marotzke 2002). To further test this idea, Fig. 10 shows the vertically integrated baroclinic energy transfer (BCET) in the low-latitude ECCO2 ocean calculated according to (e.g., Shore et al. 2008)

$$\text{BCET} = -\rho_0 L^2 (\mathbf{v} \cdot \mathbf{V_H})$$,  \hspace{1cm} (3)

where $L = \frac{g}{(\rho_0 N)}$ [with \(g\) and \(N(x, y, z)\) being, respectively, the acceleration due to gravity and time-mean buoyancy frequency], $\mathbf{v} = (u, v)$ is the vector of horizontal velocity, $\mathbf{V_H}$ is the horizontal component of the gradient operator, $\rho$ is the potential density, and the overbar and prime denote, respectively, averaging over three years (1992–94) and the deviation from the corresponding time-mean quantity.

It can be seen from Fig. 10 that, while in most places the depth-integrated BCET is positive (i.e., from the mean flows to the eddies), there are large regions in the tropical Atlantic, Pacific, and Indian Oceans where this energy transfer is negative. This means that in such regions the ECCO2 eddies transfer their potential energy to the large-scale currents. It is a challenge to represent such an effect in coarse-resolution ocean–climate models.

6. Discussion and conclusions

This study has three main goals: 1) to assess the MOHT simulated by CanESM2 against observations, 2) to quantify the contribution of the Ekman heat transport to seasonal variability of the MOHT, and 3) to project changes in the time-mean and seasonal MOHT in the twenty-first century. An additional goal is to estimate the contribution of the ocean eddies parameterized in CanESM2 to the net MOHT and how this compares...
with eddy heat transport simulated by eddy-permitting models, such as ECCO2.

Evaluation of the model MOHT is important before using the model for MOHT projections. We evaluated the MOHT simulated by CanESM2 using observations, as well as data from the observationally optimized ECCO2. We found that the global ocean MOHT in CanESM2 is within the uncertainty of observational estimates. However, accurate simulation of the MOHT for individual ocean basins presents a challenge. In particular, CanESM2 underestimates the observed time-mean MOHT in the Atlantic–Arctic basin, especially south of 30°N. A decomposition of the MOHT into the gyre and overturning components suggests that, in most regions, the overturning component dominates. However, in the Southern Ocean and, especially, in the subpolar North Atlantic it is the gyre component that contributes most to the MOHT, both in CanESM2 and in ECCO2.

Second, we considered seasonal variability of the simulated MOHT. To our knowledge this is the first time that a coupled climate model has been used to address this question systematically. We found, consistent with the ocean-only modeling results of Jayne and Marotzke (2001), that much of the seasonal variation in MOHT in our model is driven by the seasonal changes in zonal wind stress (through the associated variability of the meridional Ekman volume transport). Consistent with the limited number of observational studies, the amplitude of the seasonal MOHT is largest in the tropical ocean. It reaches almost 10 PW in CanESM2 and is dominated by the overturning component. Thus, at most latitudes the annual-mean MOHT is essentially a small residual of the corresponding time-varying MOHT. We further emphasize that, in accordance with Jayne and Marotzke (2001), the Ekman heat transport given by Eq. (1) is useful for reproducing seasonal variability of the MOHT. This is related to the fact that the so-called vertical trapping scale of the ocean’s response to perturbations is a strong function of forcing frequency and wavenumber (Willebrand et al. 1980; Jayne and Marotzke 2001). For a seasonal, large-scale forcing this scale can be comparable to, and even larger than, the mean depth of the ocean.

Last, we have found that in the simulations of future climate following the two RCP scenarios, CanESM2 projects a weakening of the net poleward MOHT in both hemispheres and essentially across all latitudes. In the low-latitude ocean, in the global mean, and in the Southern Ocean, the projected MOHT changes are largely determined by the changes in the overturning component, particularly under RCP8.5. In the subpolar Atlantic, the vertical temperature contrast is relatively weak, and the projected MOHT changes are dominated by the gyre component. The seasonal MOHT is projected to change in the tropical ocean (by up to 23% locally), as well as in the Southern Hemisphere midlatitude ocean. Some of the changes in the seasonal MOHT can be explained by the changes in the time-varying meridional Ekman transport. The important question as to what drives the associated changes in the time-varying zonal wind stress is left for future research.

In addition, we also briefly addressed the eddy contribution to the meridional heat transport in the ocean. It is found that in the midlatitude oceans the heat transports, due to both the permitted (ECCO2) and parameterized (CanESM2) eddies, are directed poleward and are of comparable magnitude. However, in the tropical ocean the permitted eddies transport heat equatorward, and this transport is much stronger than that due to the parameterized eddies. One plausible reason for this is that the tropical oceans appear to be the regions where a significant fraction of the eddy potential energy is converted back to the mean flows. It is a challenge to represent such an upgradient flux of available potential energy in coarse-resolution ocean–climate models, such as CanESM2.

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