Centennial Changes of the Global Water Cycle in CMIP5 Models

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ABSTRACT

The global water cycle is predicted to intensify under various greenhouse gas emissions scenarios. Here the nature and strength of the expected changes for the ocean in the coming century are assessed by examining the output of several CMIP5 model runs for the periods 1990–2000 and 2090–2100 and comparing them to a dataset built from modern observations. Key elements of the water cycle, such as the atmospheric vapor transport, the evaporation minus precipitation over the ocean, and the surface salinity, show significant changes over the coming century. The intensification of the water cycle leads to increased salinity contrasts in the ocean, both within and between basins. Regional projections for several areas important to large-scale ocean circulation are presented, including the export of atmospheric moisture across the tropical Americas from Atlantic to Pacific Ocean, the freshwater gain of high-latitude deep water formation sites, and the basin averaged evaporation minus precipitation with implications for interbasin mass transports.

1. Introduction

The movement of water between oceanic, atmospheric, and land reservoirs is a complex component of Earth’s climate collectively referred to as the water cycle. These flows move not only water but also heat throughout the climate system. While it is the small fraction of evaporation and precipitation that occurs over land that directly impacts human populations and industries, observational estimates indicate that 85% of the global evaporation and 77% of precipitation occur over the ocean (Durack 2015; Schanze et al. 2010; Trenberth et al. 2007). This net evaporation over the oceans is crucial for driving the relatively small amount of precipitation over land (Gimeno et al. 2010, 2011; van der Ent and Savenije 2013), and as such a full understanding of the water cycle requires a global perspective. Patterns of evaporation minus precipitation \((E - P)\) also have cascading impacts on oceanic circulations via buoyancy and mass forcing, and on atmospheric circulation through latent heat release and cloud feedbacks (Allan 2012).

The starting point for relating changes in global temperature to the water cycle is the increased mobility of water molecules at higher temperatures. According to the Clausius–Clapeyron (CC) relation for the saturation vapor pressure

\[
\frac{d \ln(e_s)}{dT_s} = \frac{L_v}{RT_s^2} \approx 0.07 \text{ K}^{-1}.
\]

Thus water vapor content in the atmosphere increases by 7% K\(^{-1}\) of warming at constant relative humidity for Earth’s current mean temperature \([L_v\text{ is the latent heat of vaporization (2.5 } \times 10^6 \text{ J kg}^{-1}), \text{ } R \text{ is the gas constant (461 J K}^{-1} \text{ kg}^{-1}), \text{ } T_s \text{ is the near-surface air temperature}].\) Current global climate models (GCMs) obey CC scaling quite closely for the increase in global atmospheric moisture \(W\) (Held and Soden 2006; Allan et al. 2014).

The water cycle is made up of all transports and fluxes of freshwater, and therefore “intensification” of the water cycle is generally used to describe, in addition to globally increased \(W\), an increase in total precipitation and evaporation, amplification of the existing \(E - P\)
pattern, and enhanced atmospheric vapor transport $Q$. There is robust theoretical (Chou and Neelin 2004; Muller and O’Gorman 2011), model (Held and Soden 2006; Chou et al. 2009), and even emerging observational (Huntington 2006; Durack et al. 2012) evidence that the water cycle will intensify in the aforementioned ways in a warming climate. However, the strength of the water cycle is governed by more complex processes than CC scaling. While there is a positive correlation between global temperature and precipitation, the sensitivity is thought to be less than CC (Stephens and Ellis 2008) due to energetic constraints governing tropospheric radiative cooling (O’Gorman et al. 2012; Allan et al. 2014). Analyses from phase 2 of the Coupled Model Inter-comparison Project (CMIP) (CMIP2; Allen and Ingram 2002), phase 3 of CMIP (CMIP3; Held and Soden 2006), and phase 5 of CMIP (CMIP5; Allan et al. 2014; Hegerl et al. 2015) have shown the increase in global precipitation to be $2\%–3\%$ K$^{-1}$ with significant model scatter.

Detecting changes in the water cycle via observations of $E - P$ has proved challenging, as the small net flux is a difference between large and uncertain quantities (Huntington 2006; Schmitt 2008; Schanze et al. 2010; Josey et al. 2013). It has been suggested that a more reliable indicator of the water cycle is sea surface salinity (SSS), which is very sensitive to changes in the surface fluxes. Evidence from 50 years of SSS observations combined with CMIP3 scaling by Durack et al. (2012) suggests that the sensitivity of SSS could be much higher than CC and also more than predicted by climate models, as much as $16\%$ K$^{-1}$. Ocean mixing and advection work to smooth out the spatial and temporal variability of surface water fluxes, and hence SSS is also an effective integrator of long-term changes in the water cycle (Durack et al. 2012). Analysis of observational data has already revealed robust trends in global salinity in line with projections of anthropogenic climate change (Boyer et al. 2005; Stott et al. 2008; Durack and Wijffels 2010; Durack et al. 2012; Terray et al. 2012; Pierce et al. 2012). Preexisting interbasin salinity contrasts are driven by net imbalances in $E - P$ (Stommel 1980; Zhou et al. 2000), and so intensification of the water cycle tends to increase these basin contrasts.

Given the complexity of the water cycle, the inherent coupling between oceanic and atmospheric processes, and the sparseness of oceanic observations, GCMs are a useful tool for predictive study of the water cycle. Current-generation GCMs used in CMIP5 (Taylor et al. 2012) are generally effective at replicating global patterns of $E - P$ and atmospheric and oceanic transports of water. Liepert and Lo (2013) found that, with several exceptions, CMIP5 models maintain a balanced atmospheric water cycle without long-term drift in global water content.

Several caveats should be noted regarding the representation of the water cycle in GCMs. First, although model resolution has improved in recent generations, current grid sizes still require significant topographical smoothing. Therefore, steep mountain ranges such as the Andes are represented with reduced peak heights, and narrow oceanic throughflows like the Bering Strait and Strait of Gibraltar have greatly simplified geometry. A modeling study by Schmittner et al. (2011) demonstrated that alterations to global topography can substantially change water vapor fluxes, ocean circulation, and the climate system as a whole. The highest South American peak in models presented here ranges from 3916 to 4591 m, while the Andes reach an actual height of 6962 m. This discrepancy may result in excess moisture transport across ranges such as the Andes, as discussed by Richter and Xie (2010). To combat these problems, modelers sometimes enhance orography at the expense of increasing the mean height of continents, resulting in a dry bias of atmospheric vapor content over land (Gaffen et al. 1997). In general, however, GCMs display a slightly overactive water cycle, with greater total precipitation and greater rainfall frequency than observations (Stephens et al. 2010; Tian et al. 2013; Demory et al. 2014).

Nearly all current GCMs have well-documented issues with simulating tropical precipitation dynamics, particularly in the Pacific Ocean (Lin 2007; Pincus et al. 2008; Collins et al. 2011; Brown et al. 2013; Flato et al. 2013; Li and Xie 2014). A set of connected biases combine to produce an excessive cold tongue in the eastern Pacific, overly strong low-level trade winds, and an unrealistic double ITCZ flanking the equatorial Pacific (Lin 2007; Hwang and Frierson 2013).

While all GCMs solve water conservation equations and should ultimately produce an internally balanced water cycle, some CMIP3 and CMIP5 models display constant atmospheric moisture content despite having globally unbalanced evaporation and precipitation, suggesting an unphysical “ghost” source/sink of atmospheric moisture (Liepert and Previdi 2012; Liepert and Lo 2013). The ocean component of GCMs can also produce unrealistic freshwater balances, evidenced by a drift in global mean salinity (Durack et al. 2014). Therefore, water cycle consistency checks are important when analyzing GCM data.

2. Theory and methods

a. Atmosphere

In a steady-state atmosphere, the vertically integrated water vapor budget can be written as a simple balance
between horizontal transport divergence and the surface flux \(E - P\) (following Peixóto and Oort 1983; Zaucker and Broecker 1992; Cohen et al. 2000; Sohn et al. 2004; Richter and Xie 2010):

\[
\mathbf{v} \cdot \mathbf{Q} = E - P,
\]

where \(\mathbf{Q}\) is the vertically integrated horizontal water vapor transport, given by

\[
\mathbf{Q} = \frac{1}{g} \int_{\text{surface}}^{\text{top}} q \mathbf{v} \, dp,
\]

where \(q\) is the specific humidity, \(\mathbf{v}\) is the horizontal wind vector, \(p\) is the pressure, and \(g\) is the gravitational constant.

Over any area, water gained or lost by \(E - P\) must equal the transport through the boundaries:

\[
\int (\mathbf{Q} \cdot \mathbf{n}) \, dl = \int E - P \, dA,
\]

where \(dl\) is a line segment along the boundary and \(dA\) is an element of surface area within the boundary. Note that \(\mathbf{Q}\) can be separated into mean and eddy components via Reynolds decomposition:

\[
\mathbf{Q} = \bar{\mathbf{Q}} + \mathbf{Q}'.
\]

For the zonal transport, particularly in the tropics where winds are steadiest, \(Q'\) is small (Wang et al. 2013). However, in the meridional flux \(Q'\) and \(\bar{\mathbf{Q}}\) are of similar magnitude. To fully capture the meridional transport then, the data used for this calculation must be sampled at higher resolution than typical time scales of synoptic weather systems. In this study, 6-hourly model outputs have been used.

b. Ocean

In the ocean, any imbalance in the surface fluxes (evaporation, precipitation, runoff, and ice flow) induces a barotropic flow known as the Goldsborough circulation (Goldsborough 1933; Huang and Schmitt 1993; Huang 2005). These circulations sometimes oppose the wind-driven gyres, such as in the North Atlantic, where net evaporation drives a northward interior flow and a southward western boundary current (Huang and Schmitt 1993). Applying steady-state mass conservation [following Wijffels et al. (1992), Stammer et al. (2003), Talley (2008), and Schanze et al. (2010)],

\[
\mathbf{v} \cdot \rho \mathbf{v} = E - P - R,
\]

and integrating through successive zonal sections, we can obtain the meridional transport in each basin dictated by the surface fluxes to within an integration constant \(T_b\):

\[
\iint (E - P - R) \, dx \, dy = \iint \rho \mathbf{v} \, dx \, dz + T_b,
\]

where \(T_b\) is the interbasin transport, applied at the meridional boundaries of each basin; \(R\) is runoff into the ocean; \(\rho\) is the ocean density; \(\mathbf{v}\) is the meridional ocean velocity; and \(x\), \(y\), and \(z\) are the zonal, meridional, and vertical coordinates respectively. As discussed by Wijffels et al. (1992), the Bering Strait (BS) is a particularly useful point of reference, as flows within the Pacific and Atlantic are zonally bounded from this point until they reach the Pacific Indonesian Throughflow (PIT) and Southern Ocean. For case of comparison to previous observational estimates (Wijffels et al. 1992; Schanze et al. 2010), the BS transport has been used along with the net \(E - P - R\) over the Arctic to impose integration constants for the Atlantic and Pacific, and the effect of the PIT has been neglected.

In steady state, the global net precipitation over land exactly equals the runoff into the ocean. Therefore, a closed budget can be derived for the ocean climatology using only \(E - P\) data, but the precise location of runoff sources will not be correctly represented, shifting the meridional transport curve (Dai and Trenberth 2002). A calculation that utilizes river and ice fluxes gives the most accurate representation of ocean freshwater transports. In the meridional integrations presented here the water flux into ocean (wfo) model output has been used, which combines all freshwater fluxes into a single field on the native ocean grid.

c. Latent heat transport

Because of water’s high latent heat of vaporization, the water cycle carries a significant amount of the poleward heat flux. The latent heat flux is usually calculated from the atmospheric moisture transport (Huang 2005):

\[
H_L = L_v \mathbf{Q},
\]

here \(\mathbf{Q}\) is a mass flux (in kg s\(^{-1}\)) of freshwater. The resulting heat flux associated with the meridional vapor transport is as large as 2.5 PW near 45°S or about half of the total energy transport there (Trenberth and Caron 2001; Huang 2005; also see Fig. 6). While this flux is often attributed to the atmospheric moisture transport, it is really a product of the coupled ocean–atmosphere water cycle, as the ocean must carry the compensating return flow of liquid water. Therefore, while the ocean mass transports induced by the surface flux \(E - P - R\) are an order of magnitude smaller than the wind-driven circulation, they are of great interest to the global energy budget.
For the projections presented here, all CMIP5 models available on a local server (Table 1) were used for global projections of temperature, vertically integrated water vapor, evaporation, precipitation, and ocean salinity using monthly means (Figs. 1–4). A total of six models have been used for the atmospheric and oceanic budget and transport calculations (Figs. 5–7). The limited number of models utilized in the transport calculations is due to the large volume of data required when processing 6-hourly fields, which are required to adequately...
capture the eddy transports in the atmosphere. Only models found by Liepert and Lo (2013) to have realistically balanced water cycles were selected for the budget and transport calculations. Models with a range of horizontal grid sizes were intentionally chosen to explore the effect of topographical smoothing on the atmospheric transport. All multimodel means (MMM) use a single ensemble run (r1i1p1) from each model.

Ten-year averages from the historical (1990–2000) and RCP8.5 (2090–2100) scenarios were processed to quantify centennial-scale changes in the water cycle under global warming. All “projections” described in the results section are taken to be MMM changes between these two time periods. Representative concentration pathway 8.5 (RCP8.5) is a high-emissions scenario intended to produce an additional radiative forcing of 8.5 W m$^{-2}$ by the year 2100, although the effective forcing varies between models and was found by Forster et al. (2013) to be only 7.4 W m$^{-2}$ in the MMM. RCP8.5 produces a rise in MMM surface air temperature of 4.0°C for the 100 years analyzed. It is used here to represent the potential for changes in the water cycle but should not be interpreted as a quantitatively accurate forecast for the coming century. It should be noted that the CC scaling associated with this temperature rise would predict a 28% increase in surface specific humidity.

Additionally, a water budget has been calculated from state-of-the-art observationally based products from the satellite era (1987–2014) as a comparison for the CMIP5 historical simulations. The atmospheric moisture transport $Q$ was calculated from ERA-Interim (Dee et al. 2011) reanalysis output, and the estimate of surface flux into the ocean $E - P - R$ was compiled from GPCP precipitation (Adler et al. 2003), OAFlux evaporation (Yu and Weller 2007), and runoff estimates from Dai and Trenberth (2002) [see Schanzle et al. (2010) for selection]. This $E - P - R$ estimate for the ocean gives total $P$ of 12.6 Sv, $E$ of 13.1 Sv, and $R$ of 1.25 Sv, with an imbalance of only 0.13 Sv (1 Sv = 10$^6$ m$^3$ s$^{-1}$).

The $Q$ calculation was performed using model outputs at 6-hourly time steps for wind and humidity in order to fully capture the eddy transport. The outputs were processed in the native hybrid-sigma pressure coordinate, which conforms to topography at low altitudes, and a trapezoidal scheme was used for the discrete vertical integral. The horizontal output was interpolated onto a standardized $1^\circ \times 1^\circ$ grid for each model. ERA-Interim offers a direct monthly output of $Q$.

Ocean freshwater fluxes for the meridional integration were taken from the combined $E - P - R$ flux into the ocean of each model and first integrated on the native ocean grid, in order to avoid introducing interpolation errors at point source river mouths.

### 3. Results and projections

RCP8.5 produces a rise in global mean surface air temperature of 4.0°C and an increase in atmospheric water content $W$ of 32%, versus the predicted 28% for CC scaling (Fig. 1). Model projections for $E - P$ are presented in Fig. 2. Total global evaporation (equal to global precipitation) increases from 17.3 to 18.6 Sv in the multimodel 100-yr projection, an 8% increase, or about 2% K$^{-1}$. Best observational estimates based on satellite data from GPCP (Trenberth et al. 2007) estimate global precipitation at 15.4 Sv. The net water transport from ocean to land ($E - P$ over land, equivalent to $R$) increases by 12% from 1.17 to 1.31 Sv. Total precipitation increases by 10% over land and just 6% over the ocean. A greater acceleration of the water cycle over land is expected due to slow thermal adjustment of the oceans (Dommenget 2009) and subsequently greater warming over land in the 100-yr period analyzed here.

While evaporation and precipitation both increase globally, planetary-scale patterns of $E - P$ also show coherent changes. There is high model agreement (>90%) that ocean locations poleward of 45°N and 45°S will experience increased precipitation. The five
subtropical ocean gyres display a decrease in precipitation albeit with weaker model agreement. There is high agreement for an increase in evaporation over almost all of the oceans except the Southern Ocean, which also shows reduced warming relative to the global mean. Projections for land areas show a modest increase in net precipitation but with poor model agreement. The model average net precipitation increases significantly in the tropical rain belt of the Pacific and Indian Oceans but with large uncertainty among models. Integrated over each basin, the Atlantic becomes 0.20 Sv more evaporative, the Pacific 0.16 Sv more precipitative, the Indian Ocean 0.08 Sv more evaporative, and the Arctic Ocean 0.05 Sv more precipitative.

SSS is strongly forced by the surface flux \( E - P - R \), and hence would be expected to display similar changes. The meridional patterns of increased subtropical evaporation and increased tropical and high-latitude precipitation are evident in the SSS projection presented in Fig. 3, but are dwarfed by a large zonal contrast between subtropical Atlantic and Pacific. With the exception of modest salinification in the southern subtropical Pacific, all other areas of the Pacific display surface freshening. The weak signature in the northern subtropical Pacific may partly be a function of weakened trade winds and evaporation (Laimé et al. 2014), as well as advection of freshened water from the equatorial Pacific. In the Atlantic, all subtropical regions show surface salinification, as high as 0.8. As the area of the low-latitude Atlantic region defined here is less than half of its Pacific counterpart, it is expected that the surface salinity forcing would be felt more strongly in the Atlantic.

Following Durack et al. (2012), an MMM linear regression of zonal mean SSS change onto its modeled
historical values gives a pattern amplification of 5.9% K$^{-1}$, slightly lower than their values for CMIP3. The same calculation for $E - P$ gives a pattern amplification of 5.2% K$^{-1}$, in line with previous values (Durack et al. 2012). The SSS regression is particularly sensitive to the use of individual model climatologies versus MMM climatology, as performing the regression on the MMM smooths out covariances and gives only 3.1% K$^{-1}$.

While the projected $E - P$ changes do not at first glance display this zonal asymmetry, salinity is an effective spatial integrator of surface forcing as ocean currents and mixing tend to smear out smaller-scale variability. It is therefore useful to assume a basinwide perspective of the atmospheric transports and their associated freshwater fluxes into and out of the oceans. Figure 4 summarizes this perspective, with drainage basins defined by topographical maxima. Low- and high-latitude basins are also divided at 30°N and 30°S. The transport from tropical Atlantic to Pacific has been broken into three components: South America (30°S–8°N), Central America (8°–20°N), and Mexico (20°–30°N).

The atmospheric water budget produced by the CMIP5 historical simulations (Fig. 4b) shows many similarities to the observational estimate (Fig. 4a). Both capture the export of water vapor from low to high latitudes, zonal convergence in the Pacific, and divergence in the Atlantic and Indian Ocean. The models and observations produce the opposite sign for the small net flux across the Maritime Continent. The largest discrepancies seen in the fluxes are across South America, with the CMIP5 models producing a much higher transport estimate there. There is also a significant discrepancy in the GPCP/OAFlux and ERA-Interim budget for the low-latitude Pacific with a residual of 0.28 Sv, making this region poorly constrained for comparison with models. An imbalance of 0.27 Sv is also present in the Southern Ocean region. There are small imbalances between the CMIP5 divergences and surface fluxes as well, but all are less than 0.1 Sv. These model imbalances are likely related to the regidding scheme used here, or the relatively short 10-yr integration which may allow some reservoir storage.

The RCP8.5 projection for basin to basin transports reveals the larger-scale pattern that explains the zonal SSS trend in Fig. 3. Relative to the historical scenario, RCP8.5 shows an increased westward transport by the trade winds at low latitudes, and an increased poleward transport from low to high latitudes driven primarily by the synoptic eddy field. Although the zonal transport across all low-latitude divides increases, the preexisting differences in magnitude are amplified by such a pattern. Therefore the net evaporation in the low latitudes gets larger in the Atlantic and smaller in the Pacific. Additionally, there is an increase in net precipitation for all extratropical regions.

Among the six models analyzed here, there is 100% agreement for the sign of the change in net $E - P - R$ for all basins except the tropical Indian Ocean (5/6 in agreement). There is similarly robust agreement for the changes in atmospheric transport across drainage divides, with many segments below displaying 100% agreement in the sign of the change. The components with some model disagreement are the northward export out of the tropical Atlantic (4/6 in agreement), the southward export out of the tropical Pacific (4/6 in agreement), the tropical Pacific to tropical Indian Ocean (5/6 in agreement), and across the Himalayas (3/6 in agreement).

The MMM of zonally averaged transports (Fig. 5) shows an increase in atmospheric freshwater fluxes everywhere. The intermodel spread is displayed for an idea of the model uncertainty in moisture transports. The change in the zonally integrated meridional transport (Fig. 6) is as large as 0.3 Sv near 10°N and 40°S, which corresponds to nearly 0.7 PW of additional latent heat transport. The meridional ocean transports separated into basins (Fig. 7) show generally weaker southward transport through the Atlantic and weaker northward transport through the Pacific. All transport curves in Figs. 5–7 match qualitatively with the observational comparison.

4. Discussion

a. Zonal atmospheric transports

To first order, $Q$ scales similarly to CC, which would be the expectation if warming caused no dynamical response in the circulation (Held and Soden 2006). The global mean zonal transport $Q_x$ increases 35%, and the meridional transport $Q_y$ increases 30%, noting again that the increase in $W$ is 32%. A robust consequence of this thermodynamic scaling is that regional patterns of moisture convergence and divergence are amplified in a warming climate even without circulation changes. The preexisting zonal pattern in the subtropics is a useful example. There is a large moisture transport across Central and South America relative to the transport across Africa and the Maritime Continent. If the increase in temperature and thus specific humidity is uniform, then a constant fractional increase of each interbasin transport causes an increased volume convergence in the Pacific.

A slight increase in the strength of the trades over South America causes the change in $Q$ to be higher than CC scaling. Conversely, a slight decrease in the trades across Africa gives a lower scaling there. In the models analyzed here, there is not a significant correlation
FIG. 4. Summary of the atmospheric freshwater budget showing $Q$ and $E - P$ for (a) ERA-Interim winds and humidity with GPCP/OAFlux fluxes for 1979–2014, and multimodel means of the CMIP5 (b) historical (1990–2000) and (c) RCP8.5 (2090–2100) simulations. For the observational dataset, $E - P - R$ into the ocean has been used rather than $E - P$ over combined land and ocean areas. The map has been divided (thick black lines) by topographical drainage basins between the Atlantic, Pacific, and Indian Oceans, and into low- and high-latitude regions separated at 30°N and 30°S, corresponding to the approximate center of the Hadley cells. Colored values are $E - P$ integrated over each region, with red indicating net evaporation and blue net precipitation (Sv). Filled contours are the global $E - P$ (m yr$^{-1}$). Small arrows indicate the vertically integrated atmospheric moisture flux, and thick arrows show the fluxes across drainage divides (Sv). The CMIP5 transports are an average of the six models listed in Table 1.
between horizontal resolution and transport over the Andes, although all display much greater transports than the ERA-Interim value of 0.16 Sv. Other reanalysis products show greater Andean transports closer to the CMIP5 values (Richter and Xie 2010), particularly along the northern segment, so it is unclear how topographical smoothing affects this transport. Xu et al. (2005) showed that, in a high-resolution model, transport across Central America was insensitive to the fine structure of the orography, as intense jets around mountains or diffuse flow over smooth topography produced nearly the same transport.

There are a number of dynamical factors affecting tropical transports in a warming climate, although circulation changes are generally small compared to thermal affects (Held and Soden 2006). A common feature of GCM warming simulations is a decrease in the Walker circulation (Vecchi et al. 2006; DiNezio et al. 2013; Ma and Xie 2013). This is sometimes explained using the differential response of atmospheric water vapor (7% K\textsuperscript{1}) and precipitation (2%–3% K\textsuperscript{1}) to warming. Because the convective supply of water cannot exceed the loss of water from precipitation, the balance equations require a weakening of the upward mass flux and therefore the overturning circulation (Held and Soden 2006; Ma et al. 2012). However, observational evidence for changes in the Walker circulation is inconclusive (Tokinaga et al. 2012).
Regional patterns of SST increase may also play a role (Knutson and Manabe 1995; Tokinaga et al. 2012). In the tropics, a distinct El Niño–like warming mode is produced by CMIP5 models (Cai et al. 2014; Fig. 1a). Warm SST in the eastern tropical Pacific induces anomalous low pressure there, and an increase in trade wind strength across South America follows.

The change in $Q$ across the Maritime Continent (88%) is large in a relative sense, but remains small in terms of volume flux because the net transport of the seasonal monsoonal cycle is small. One argument regarding the monsoon circulation of the Indian Ocean basin predicts an increase in summer monsoon intensity due to the transient increase in land–sea temperature contrast under warming (Turner and Annamalai 2012; Lee and Wang 2014). Such an increase would bring additional moisture from the Pacific to the Indian Ocean during summer, but monsoon projections have large intermodel spread. The change in $Q$ is also large (44%) for the extratropical transport across North America from the Pacific to the Atlantic, dominated by anomalous northeasterly transport over Alaska, but this is primarily a function of amplified Arctic warming and therefore larger CC scaling.

Based on these trends in interbasin transport, a general enhancement of the zonal pattern of $E - P$ and SSS is projected under anthropogenic warming. The Atlantic becomes more evaporation and saline, while the Pacific becomes more precipitative and fresher. These trends will lead to halosteric effects on sea level, diminishing sea level rise in the Atlantic while enhancing it in the Pacific (Durack et al. 2014). Despite an increase in net evaporation over the Indian Ocean, a slight freshening is projected for all but the most western portions of the basin. This is consistent with the substantial advection of Pacific water westward through the PIT, water that is projected to freshen because of increased precipitation in the western Pacific.

Increased moisture export from the Atlantic to the Pacific and subsequent salinification of Atlantic surface waters has previously been discussed as a positive feedback on North Atlantic Deep Water (NADW) formation and the meridional overturning circulation (Broecker et al. 1990; Zaucker et al. 1994; Lohmann 2003; Richter and Xie 2010). Increased freshening and warming of the northern Atlantic and Arctic surface water tends to increase static stability of the water column and decrease deep convection (Aagaard and Carmack 1989; Hasumi 2002; Hu and Meehl 2005), and CMIP5 models generally do show a reduction in the Atlantic overturning. However, these simulations also confirm that warming increases surface salinities in the subtropical Atlantic, which could counteract the positive buoyancy forcing from increased freshwater input at higher latitudes.

b. Meridional atmospheric transports

The projected increase in meridional transport across 30°N and 30°S is somewhat less than CC scaling, and also shows large zonal variation. The synoptic eddy field plays a much greater role in determining meridional fluxes compared to zonal fluxes, and so the meridional transports may be more subject to dynamical changes and less directly tied to CC scaling. Poleward transport out of the North Pacific increases by 31%, while transport out of the North Atlantic shows nearly no change,
as increases in $W$ are offset by a weakening of the prevailing southwesterly transport there. In the Southern Hemisphere, the poleward transports out of the Atlantic and Indian Oceans are near CC scaling. Transport out of the South Pacific, however, shows almost no change. Here CC is counteracted by an equatorward anomaly along South America. Transport across the Himalayas is negligible for both scenarios.

c. Ocean throughflows and transports

Bering Strait (BS) transport is understood to be a function of the sea surface height (SSH) gradient from Pacific to Atlantic, which some have suggested is maintained by atmospheric moisture export over Central and South America and associated $E-P$ fluxes (Latif 2001; Schmitt 2008). Despite projections for increasing disparity between Pacific and Atlantic $E-P$, the six models analyzed here show a decreasing BS flow, from 0.80 to 0.68 Sv in the MMM, compared to observational estimates of 0.79 Sv (Schanze et al. 2010). While the Atlantic becomes more evaporative and would be expected to decrease in mean height from mass balance and halosteric effects, the Arctic experiences significant freshwater gain, which increases SSH there. Wind stress has been used to explain seasonal variability in the Bering Strait transport (Coachman and Aagaard 1988), as the mean wind field is southward and opposes the pressure-driven ocean transport. Wind stress does not appear to explain the projections for Bering Strait transport here, as the change in mean wind fields is actually for weaker southerly circulation through the strait. A probable explanation for decreased BS flow is the thermosteric rise associated with amplified warming in the Arctic, reducing the more local Pacific to Arctic pressure gradient.

Basin meridional transports for the historical simulations closely resemble those of observations (Baumgartner and Reichel 1975; Wijffels et al. 1992; Dai and Trenberth 2003, Schanz et al. 2010). The PIT transport has been neglected in Fig. 7 for simplicity, which induces a counterclockwise circulation around Australia [see Wijffels et al. (1992) and Dai and Trenberth (2003) for further discussion] and would significantly displace the Indian Ocean curve toward southward transport and the Pacific curve toward northward transport south of the PIT. The observational estimate presented here shows stronger northward transport in the Pacific and weaker southward transport in the Atlantic from approximately 10°N and all points southward relative to CMIP5. Because of the slight decrease in BS flow and increased precipitation in the tropical Pacific, the projected meridional flow in the Pacific becomes southward around the equator, unlike in the historical simulation, which produces northward flow throughout the Pacific.

5. Summary

This study has examined the estimated changes in the water cycle of relevance to ocean circulation from a multimodel ensemble of CMIP5 models for a high CO$_2$ emissions scenario for the twenty-first century. Drainage basins are defined and net water fluxes calculated between them. There are large changes in interbasin transports and strengthening of the patterns of midlatitude evaporation and high- and low-latitude precipitation. Substantial changes in ocean salinity are projected, with salty areas getting saltier and fresh areas fresher overall. Such changes have consequences for regional sea level rise and the strength of the meridional overturning circulation, and indicate the importance of sustained global salinity observations over the next century. However, the spatial response of SSS and $E-P$ to warming in CMIP5 is far from identical. These results suggest that if the salinity field is to be used as a global “rain gauge” for Earth’s water cycle, the effects of ocean mixing and advection and large-scale interbasin patterns of moisture convergence must be considered in addition to the predicted amplification of the local surface forcing.

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