The Response of the Extratropical Hydrological Cycle to Global Warming

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ABSTRACT

The change in the hydrological cycle in the extratropics under global warming is studied using the climate models participating in the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report. The changes in hydrological quantities are analyzed with respect to the increases expected from the Clausius–Clapeyron (C–C) equation, which describes the rate of increase of a hydrological quantity per temperature increase. The column-integrated water vapor increases at a rate close to the C–C rate, which is expected if relative humidity remains nearly constant. The poleward moisture transport and the precipitation increase with temperature at a rate less than the C–C rate, with the precipitation increasing the least. In addition, the intermodel variance of poleward moisture transport and precipitation is explained significantly better when the zonal-mean zonal wind change as well as the temperature change is taken into account. The percent increase in precipitation per temperature increase is smallest during the warm season when energy constraints on the hydrological cycle are more important. In contrast to other hydrological quantities, the changes in evaporation in the extratropics are not explained well by the temperature or zonal wind change. Instead, a significant portion of the intermodel spread of evaporation change is linked to the spread in the poleward ocean heat transport change.

1. Introduction

It is well known that the global-mean precipitation response to global warming predicted by the current generation of climate models is more uncertain than the temperature response (Wild et al. 1997; Watterson 1998; Roeckner et al. 1999; Boer et al. 2000; Allen and Ingram 2002; Yang et al. 2003; Meehl et al. 2005). Because the global-mean precipitation is dominated by the Tropics, however, the uncertainty in the response of extratropical precipitation to global warming has not received sufficient attention. Therefore, this study focuses on the changes in the hydrological cycle in the extratropics. Among our findings is that the change in extratropical precipitation in climate models is robustly linked to the temperature and zonal wind change in the extratropics.

In addition to the extratropical precipitation response, the changes in certain other aspects of the hydrological cycle are robust across climate models (Held and Soden 2006). One robust response to CO₂ increase in climate models is the water vapor content, which increases such that the relative humidity remains nearly constant (Manabe and Wetherald 1975; Cess et al. 1990; Held and Soden 2000; Ingram 2002). This model result is supported by observational evidence of constant relative humidity in interannual variability and trends (Wentz and Schabel 2000; Trenberth et al. 2005; Soden et al. 2005). Because most places in the atmosphere are far from saturation, the reasons for nearly constant relative humidity under global warming are not obvious. In the free troposphere, air motions act to keep the atmosphere unsaturated by precipitating out moisture in regions of upward motion. Thus the relative humidity of an air parcel in the free troposphere can be understood by considering the parcel’s trajectory backwards to the point the air was last saturated (Yang and Pierrehumbert 1994; Held and Soden 2000; Pierrehumbert 2007). By considering parcel trajectories and ne-
glecting mixing from neighboring air parcels, Held and Soden (2000) showed that the assumption of fixed relative humidity is basically equivalent to the assumption that the change in temperature of the air parcel at last saturation is on average the same as the change in the temperature of the air parcel itself. Thus, constant relative humidity is expected when the air trajectories are constant (no change in circulation) and the temperature change is spatially uniform. While circulation does change and the temperature change is not uniform, the fact that relative humidity is nearly constant suggests that these changes are of secondary importance when considering the water vapor content of the atmosphere.

Constant relative humidity means that the specific humidity \( q \) increases rapidly with increasing temperature \( T \):

\[
q = \frac{RH \cdot e_{\text{sat}}}{p} = \frac{RH}{p} \cdot ec_0 \cdot \exp \left( - \frac{L}{R_v} \frac{1}{T} \right),
\]

where \( RH \) is the relative humidity, \( e \) is the ratio of the molecular weight of water to the molecular weight of air, \( e_{\text{sat}} \) is the saturation vapor pressure, \( p \) is the pressure, \( L \) is the latent heat of vaporization, and \( R_v \) is the gas constant for water vapor. Assuming that the relative humidity is constant and that the temperature change due to global warming is small compared to the current temperature, we can write the ratio of the future specific humidity, \( q_1 \), to the current specific humidity, \( q_0 \), as

\[
\frac{q_1}{q_0} = \exp \left[ - \frac{L}{R_v} \left( \frac{1}{T_1} - \frac{1}{T_0} \right) \right] = \exp (\alpha(T_1 - T_0)),
\]

where \( \alpha = LR_v^{-1}T_0^{-2} \). Thus, constant relative humidity means that the specific humidity increases at the exponential “Clausius–Clapeyron \( (C-C) \) rate” (Boer 1993; Allen and Ingram 2002; Trenberth et al. 2003; Held and Soden 2006):

\[
q_1 = q_0 \exp(\alpha \Delta T).
\]

For earth-like temperatures, \( \alpha \) has a value of about 0.07 K\(^{-1}\). Thus, because specific humidity is tied to the temperature, the spread among climate models in the water vapor pressure is explained very well by the temperature increase. In addition, the exponential increase in specific humidity implies that the poleward moisture transport must also go up at the \( C-C \) rate provided that the circulation does not change (Held and Soden 2006). Since the spread of poleward moisture transport among climate models is also explained well by the temperature increase (Kutzbach et al. 2005; Held and Soden 2006), the changes in circulation appear to be secondary compared to the thermodynamics embodied in the Clausius–Clapeyron relation.

While moisture and moisture transport increase at approximately the \( C-C \) rate, the precipitation and evaporation (i.e., the strength of the hydrological cycle) increase at a rate less than the \( C-C \) rate. Unlike moisture content, the strength of the hydrological cycle is constrained by energetics, which limit the slope of the precipitation/temperature relationship (Boer 1993; Allen and Ingram 2002; Pierrehumbert 2002). Indeed, a few models show basically zero increase in the global-mean precipitation due to global warming (Houghton et al. 2001; Räisänen 2002; Yang et al. 2003; Meehl et al. 2005). Betts (1998) and Held and Soden (2006) point out that because the precipitation increases at a rate less than the water vapor content, the strength of the atmosphere circulation between the boundary layer and the free atmosphere must decrease. Recent observational evidence seems to support this suggestion (Vecchi et al. 2006; Zhang and Song 2006). The importance of circulation change in explaining the precipitation response is puzzling in light of the fact that fixed circulation is the easiest way to fix relative humidity. The decrease in strength of the circulation might also be implicated in the slope of the moisture transport versus temperature, which is actually slightly less than the \( C-C \) rate (Held and Soden 2006). The tendency for the circulation to try to decrease the moisture transport under global warming was also noted by Watterson (1998).

Many studies have focused on the \( CO_2 \)-induced changes to the global-mean hydrological cycle, which is dominated by the Tropics. Because of this tropical bias, previous studies have been pessimistic regarding the existence of robust constraints on the change in the hydrological cycle in response to global warming (Allen and Ingram 2002; Yang et al. 2003; Meehl et al. 2005). In this paper, we focus on the large-scale changes in moisture transport, precipitation, and evaporation in the extratropics. In contrast to the Tropics, we find that the precipitation change averaged over the extratropics is robustly linked to the temperature and zonal wind change. Our finding that circulation change can be used as a robust predictor of hydrological cycle change is novel, since previous studies generally portray circulation change as a complicating factor. In this study, we first confirm that the changes in specific humidity are to leading order consistent with constant relative humidity. Next, we look at the changes in moisture transport and precipitation in the extratropics and demonstrate the importance of the model spread in temperature and large-scale zonal wind in determining the model spread in moisture transport and precipitation. We also look at the latitudinal and seasonal dependence of the change in moisture transport and precipitation versus the change in temperature. Last, we look at the factors con-
Table 1. The IPCC models used in this study. The last column tells whether ocean heat transport data are available for that model.

<table>
<thead>
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<th>Model</th>
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<th>Ocean heat transport</th>
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<tr>
<td>UKMO-HadCM3</td>
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<td>20c3m, A2, B1</td>
</tr>
</tbody>
</table>

trolling the spread of model evaporation over the extratropical oceans and the factors controlling the spread of model relative humidity in the subtropics.

2. Models and data

We use output from climate change scenario integrations prepared for the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (IPCC 2007). Future climate data used here come from the A2 scenario, a “business-as-usual” scenario in which carbon dioxide (CO₂) concentrations rise unchecked and reach 856 ppm (ppm) by the year 2100, and from the B1 scenario, in which CO₂ concentrations are controlled and stabilized at 549 ppm by 2100. Simulations of present-day climate from the same models were obtained from the Coupled Model Intercomparison Project’s “20th Century Climate in Coupled Models” (20C3M) data archive [Data available from the Program for Climate Model Diagnosis and Intercomparison (PCMDI) Web site; http://www-pcmdi.llnl.gov]. Output from the models listed in Table 1 was available for the 20C3M, A2, and B1 scenario archives for the variables of interest here. (More information available online at http://www-pcmdi.llnl.gov/ipcc/model_documentation/ipcc_model_documentation.php.) All data used here are archived as monthly averages. Complete details on the forcing used in the A2 and B1 scenarios are given in appendix 2 of the 2001 IPCC report (Houghton et al. 2001). In the results presented here, climate change is defined as the difference between the climatologies of years 2080 to 2099 in either the A2 or B1 scenario and years 1980 to 1999 in the 20C3M simulations. The “twenty-first-century climate” is defined as the average over years 2080–99 and the “twenty-first-century climate” is defined as the average over years 1980–99. For most variables, there are 15 models for the A2 scenario and 16 models for the B1 scenario.

3. Results

a. Column-integrated water vapor

If relative humidity remains constant under global warming and the temperature increase is uniform in height, then (3) describes the dependence of column-integrated water vapor, \( Q \), on temperature. Dividing the twenty-first-century water vapor, \( Q_{21} \), by the twentieth-century water vapor, \( Q_{20} \), and taking the logarithm, we get

\[
\log \left( \frac{Q_{21}}{Q_{20}} \right) = \alpha (T_{21} - T_{20}).
\]  

(4)

The changes in hydrological cycle are sometimes given in terms of the percent increase of a particular variable per temperature increase (Boer 1993; Allen and Ingram 2002; Trenberth et al. 2003; Held and Soden 2006). The above (4) is approximately the same as the formulation in terms of percent increase because

\[
\log \left( \frac{Q_{21}}{Q_{20}} \right) \approx \frac{Q_{21} - Q_{20}}{Q_{20}}.
\]  

(5)

We choose to use the formula with the logarithm because 1) the scatter among models is smaller and 2) the column-integrated water vapor results are more consistent with the relative humidity. In Fig. 1, we scatter 100log(\( Q_{21}/Q_{20} \)) on the y axis and \( T_{21} - T_{20} \) on the x axis for each month and model. The \( Q \) and \( T \) are zonal averages at a particular latitude, where \( T \) is the temperature at 850 mb.¹ We use \( T \) at 850 mb because low-level water vapor dominates the column-integrated water vapor. There are 12 × 15 points for the 15 models in the A2 scenario and 12 × 16 points for the 16 models in the B1 scenario (≈ 372 points). The linear least squares fit is shown as a solid line with a slope, \( a \), given in the top-right corner of the plot. The expected slope assuming that the specific humidity increases at the C–C rate is given by the dashed line. The value of the C–C slope (= \( 100L_{cr}^{-1}T^{-2} \), where \( T \) is the 850-mb temperature) is given above each plot. The factor of 100 in front of the

¹ The zonal averaging places an additional assumption in order for (4) to hold: the temperature increase must be uniform in longitude.
logarithm means that the slope is approximately the percent increase in column-integrated water vapor per temperature increase.

As expected from previous studies, the observed slope from the climate models is quite close to the C–C slope consistent with the idea that the relative humidity changes are small compared to the large change in water vapor (Fig. 2). The same tight relationship between water vapor and temperature also holds in the Tropics (not shown) and for the interannual variability and trends in observations (Wentz and Schabel 2000; Trenberth et al. 2005; Soden et al. 2005). Closer inspection of Fig. 1 shows that the observed slope tends to be less than the C–C rate at 40° latitude and that the slope increases relative to the C–C slope at 55° latitude. In the Southern Hemisphere the observed slope even becomes larger than the C–C slope at 55°S. This behavior can also be seen in the ensemble mean change in relative humidity for the A2 scenario (Fig. 2). Focusing on the lower levels, which dominate the column-integrated water vapor, we see that the relative humidity tends to decrease in the subtropics and into the midlatitudes and that relative humidity then increases as one moves poleward past 50°.

b. Moisture transport

Held and Soden (2006) note that the poleward moisture transport should also increase close to the C–C
rate like the column-integrated water vapor, assuming that changes in circulation are negligible. To see this consider the poleward moisture transport across a latitude circle:

\[
\int_0^{2\pi} \int_0^{\rho_2} q_{21} v_{21} \frac{\partial p}{g} R \partial \lambda = \int_0^{2\pi} \int_0^{\rho_2} q_{20} v_{20} e^{\alpha(T_{21} - T_{20})} \frac{\partial p}{g} R \partial \lambda = e^{\alpha(T_{21} - T_{20})} \int_0^{2\pi} \int_0^{\rho_2} q_{20} v_{20} \frac{\partial p}{g} R \partial \lambda,
\]

where \( q \) is the specific humidity, \( v \) is the meridional wind, \( g \) is the acceleration of gravity, \( R \) is the radius of the earth times the cosine of the latitude, \( \lambda \) is the longitude, the overbar is a time average, and the other notation is as in (4). In the first step in (6) we wrote \( q_{21} \) in terms of \( q_{20} \) and the average temperature change assuming that relative humidity is constant for each individual air parcel and that the temperature for each individual air parcel increases by \( T_{21} - T_{20} \) compared to the twentieth-century climate. In addition, we substituted \( v_{21} \) with \( v_{20} \) assuming that the statistics of the circulation do not change (i.e., both the magnitude of the circulation and the correlation between the circulation and various other fields is unchanged). For the last step we assume that the average temperature increase is uniform in pressure and longitude so that the temperature term can be placed outside the integral. With these assumptions, the integral term on the far right is now the twentieth-century moisture transport so that (6) can be written in the form of (4) with the column-integrated water replaced by the poleward moisture transport.

The zonal-mean poleward moisture transport across a particular latitude circle is calculated by integrating precipitation minus evaporation \((P - E)\) over the area poleward of that latitude circle. We then scatter 100log\((qv_{21}/qv_{20})\) versus the zonal-mean 850-mb temperature, where \( qv \) stands for the vertically integrated
poleward moisture transport (Fig. 3). The model spread is much larger when the climatology for each individual month is plotted as in Fig. 3 than when the annual mean quantities are plotted as in Kuzbach et al. (2005) and Held and Soden (2006). At 40° latitude, the moisture transport tends to increase more in the winter than in the summer in both hemispheres. At 55° latitude, the seasonal difference between summer and winter disappears in the Northern Hemisphere, but the scatter among models remains just as large. In the Southern Hemisphere, the seasonality actually reverses, with summer having a larger increase in moisture transport than winter.

The large scatter in the moisture transport suggests that the approximations used in (6) are not accurate. Since the water vapor scatter is rather small, the source of the moisture transport scatter is likely due to changes in the circulation. Indeed, Yin (2005), for example, find significant changes in the midlatitude circulation, which can be seen in the zonal wind changes due to global warming in the A2 scenario (Fig. 4). In all seasons, the ensemble-mean 850-mb zonal wind change consists of a combination of a poleward shift and a strengthening of the midlatitude jets. Since the zonal jets coincide with the transients and the mean-meridional circulation that transport water poleward, the change in zonal-mean zonal wind might explain the scatter in moisture transport. (The daily mean fields required to calculate the transients explicitly are only available for a small subset of models.) Figure 5 shows the moisture transport versus temperature change scatter with the points color coded by zonal-mean zonal wind change at 850 mb. There is a clear tendency for models with increasing (decreasing) zonal wind at a particular latitude to have moisture transport increase more (less) than the C–C curve. Another way to see this is through a two-parameter fit of the moisture transport to both the temperature change and the zonal wind ($u$) change:

$$100\log \left( \frac{q_{21}}{q_{20}} \right) = a(T_{21} - T_{20}) + b(u_{21} - u_{20}).$$

For Fig. 6, we plot $100\log(q_{21}/q_{20})$ versus $(T_{21} - T_{20}) + (b/a)(u_{21} - u_{20})$ along with a line of slope $a$. We divide the right side of (7) by $a$ so that the axis units remain the same as the other figures. As expected, the scatter is significantly reduced such that it is comparable to the column-integrated water vapor (Fig. 1). The data in Fig. 6 are limited to the climatological monthly mean data during the three-month winter season (i.e., three data points per model and scenario). The

![Figure 4](image-url)
reduction in scatter about the linear fit is shown in Table 2, which shows the ratio of the unexplained variance to the total variance as a function of the number of predictors. The unexplained variance drops by a substantial amount when the zonal wind is included, especially at 55°S where the unexplained variance decreases from 56.3% to 17.7%. The slope of the moisture transport increase tends to be close to but less than the C–C slope at most latitudes, with larger deviations from the C–C slope in the Northern Hemisphere. In addition, the Northern Hemisphere results exhibit more scatter than the Southern Hemisphere.

The reasons for the discrepancy between the C–C slope and the moisture transport slope even after taking into account some changes in the circulation might be a result of using zonal wind rather than the transients or some other measure of the meridional wind change. We have tried the two-parameter fit with low-level transient kinetic energy instead of zonal wind using the limited number of models with daily mean output. The slope does become more like the C–C slope, but the scatter becomes larger.

We have also looked at the relationship between 850-mb zonal wind and 850-mb transient meridional wind to see if the magnitude of the effect of \( u \) on \( qv \) [i.e., the parameter \( b \) in (7)] is consistent with the expected relationship between meridional wind and \( qv \). For example, if \( qv \) is dominated by the transients and the correlation between \( q \) and \( v \) does not change in the future climate, then

\[
100\log \left( \frac{qv_{21}}{qv_{20}} \right) = 100\log \left( \frac{\sqrt{q_{21}^2 + v_{21}^2}}{\sqrt{q_{20}^2 + v_{20}^2}} \right),
\]

where the prime stands for transient. Using (3) for the change in \( q' \) and simplifying, we get

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Fig. 5. Same as in Fig. 3, but the points are color coded by the zonal-mean zonal wind change at 850 mb at each particular latitude, month, and model. The color key is given at the top right of the plot and the units in the key are m s\(^{-1}\).
Using the linear least squares fit between $\Delta u$ and $\Delta \sqrt{v^2}$ from the IPCC models, \(^{(9)}\) becomes

\[
100\log \left( \frac{q_{v21}}{q_{v20}} \right) = 100 \left[ \alpha \Delta T + \log \left( \frac{\sqrt{v^2}}{\sqrt{v_2^2}} \right) \right] 
\approx 100 \left( \alpha \Delta T + \frac{\beta}{\sqrt{v_2^2}} \right). \tag{9}
\]

where $\beta$ is the slope of the linear fit. Substituting the values for the constants in front of $\Delta u$, we get a $b$ value of 1.0 for (7), which is significantly less than the $b$ values of around 4 and 5 given in Fig. 6. We think the discrepancy between the slopes is due to the fact that the Ferrel cell (and possibly the stationary waves in the Northern Hemisphere) is an important contributor to the poleward moisture transport in the midlatitudes. In fact, the fractional change of the meridional wind in the lower branch of the Ferrel cell is much larger than the

\[
\text{TABLE 2. The ratio of unexplained variance to total variance (in percent) for the linear least squares fit with either 1) 850-mb temperature as a predictor, or 2) both 850-mb temperature and 850-mb zonal wind as predictors. The predictand is 100log}(H_{21}/H_{20}), \text{ where } H \text{ is either poleward moisture transport (}q\text{) or precipitation integrated poleward of a latitude circle (}P\text{cap). Climatological monthly mean data during the three-month winter season are used (i.e., three data points per model and scenario). There is no intercept in the least squares fit.}
\]

<table>
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<th>No. of predictors</th>
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<td>17.7</td>
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<td>2</td>
<td>11.4</td>
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</table>
fractional change in transient root-mean-square meridional wind. Thus, when the Ferrel cell meridional wind is used in (10) the slope $b$ is 9.6. Evidently, the observed dependence of $100\log(\frac{q_{21}}{q_{20}} - \frac{q_{21}}{q_{20}})$ on $\Delta u$ is a weighted average of a small contribution from the transients and a large contribution from the Ferrel cell.

c. Precipitation

Unlike the increase in moisture content and transport, which is expected to scale like (4) and (6) under the assumptions outlined above, the increase in precipitation is not expected to scale in the same way a priori. Nevertheless, we also scale the precipitation increase like (4) because other hydrological parameters scale in this way and because previous studies have looked at the precipitation change in this way (Boer 1993; Allen and Ingram 2002). As discussed in the introduction, the precipitation is not expected to increase at the C–C rate because the energy budget constrains changes in the strength of the hydrological cycle (Boer 1993; Trenberth 1998; Pierrehumbert 2002; Allen and Ingram 2002). We first consider the precipitation integrated over the polar cap enclosed by a latitude circle ($P_{cap}$) and use a two-parameter least squares fit as before:

$$100\log\left(\frac{P_{cap_{21}}}{P_{cap_{20}}}\right) = a(T_{21} - T_{20}) + b(u_{21} - u_{20}).$$

The scatter of $100\log(P_{cap_{21}}/P_{cap_{20}})$ versus $(T_{21} - T_{20}) + (b/a)(u_{21} - u_{20})$ about the linear fit (Fig. 7) is surprisingly less than that for moisture transport, which perhaps one would expect to be more constrained than precipitation by the arguments given in the previous section (see Table 2 for the percent unexplained variance for $q_v$ and $P_{cap}$). The scatter is larger if one does not take into account the zonal wind changes, especially at 55°S (Table 2). As expected, the slope of the precipitation averaged over the polar cap increases at a rate less than the C–C rate. In addition, the rate of $P_{cap}$ change tends to increase as one moves poleward. Like the moisture transport, the scatter in the Northern Hemisphere is greater than in the Southern Hemisphere. The $P_{cap}$ scatter with the global-mean temperature increase rather than the local temperature increase is similar although the spread about the linear fit is larger.

The zonal-mean annual-mean precipitation change tends to show increases poleward of about 40° latitude and decreases or zero change equatorward of 40° latitude (not shown). The zonal-mean precipitation
changes scatter much more than the precipitation averaged over the polar cap, except at latitudes poleward of about 50° where the scatter is only slightly more than for Pcap (not shown).

d. Slope versus latitude and month

The dependence of the slope of different hydrological parameters versus temperature for different latitudes is shown in Fig. 8. The column-integrated water vapor (Q) slope is computed with a one-parameter fit to 850-mb temperature. The moisture transport (qv) and the precipitation over the polar cap (Pcap) slopes are computed with a two-parameter fit to 850-mb zonal-mean temperature and the 850-mb zonal-mean zonal wind (just as in the previous section). The slope is the coefficient for the temperature dependence [a in (7) and (11)]. The C–C slope is \( LR_{cc}^{-1} T^{-2} \), where \( T \) is the 850-mb ensemble-mean zonal-mean temperature for the twentieth century. The C–C curve increases gradually toward higher latitudes because of its weak \( T^{-2} \) dependence. In the Southern Hemisphere, the water vapor slope is smaller than the C–C slope equatorward of 40°–45°S and is larger than the C–C slope at higher latitudes. Thus, the water vapor slope depends more strongly with latitude than the C–C slope. In the Northern Hemisphere, the water vapor increase is smaller than the C–C predicted increase at all latitudes, but the amount of variation with latitude is about the same for both. The increase in poleward moisture transport (qv) is significantly less than either the water vapor or the C–C slope at almost all latitudes and seasons. Thus, changes in circulation, other than those encapsulated in the zonal-mean zonal wind change, must be involved. The fact that the moisture transport decreases less than the water vapor means that the horizontal as well as the vertical circulation must slow down (Held and Soden 2006). The slope of the precipitation averaged over the polar cap (Pcap) increases as one moves from the sub-tropics toward the pole.

The dependence of the slope of different hydrological parameters versus temperature for different months is shown in Fig. 9. The values for each month are smoothed by using the 3-month period centered on the month in question for the linear least squares calculation. The C–C curve depends weakly on temperature with the larger values in winter when the temperature is colder. As noted before, the water vapor increase tends to be greater than the C–C rate at higher latitudes,
especially in the Southern Hemisphere, and lower than the C–C rate toward the equator. At almost all seasons and latitudes, the moisture transport increase is significantly less than the water vapor increase. This is most pronounced in the Northern Hemisphere, particularly in summer. The slope of precipitation integrated over the polar cap (Pcap) has a clear seasonal cycle with larger increases in the winter than in the summer. This is consistent with the idea that energy constraints on the hydrologic cycle are less significant at lower temperatures (Pierrehumbert 2002). Thus the wintertime precipitation increase is closer to the rate expected from the Clausius–Clapeyron relation.

Figure 10 shows the ratio of the unexplained variance to the total predictand variance as a function of month. The values for each month are smoothed by using the 3-month period centered on the month in question for the linear least squares calculation. For all seasons and latitudes, the scatter about the linear fit is very small for the column-integrated water vapor (Q). For 55°S, the error about the linear fit for poleward moisture transport (qv) and precipitation integrated poleward of a latitude circle (Pcap) is also very small. At other latitudes, there is a distinct seasonal cycle to the percent unexplained variance for qv and Pcap with large values in the summer and small values in the winter. In winter the fit is better for Pcap than for qv, while in summer the fit is better for qv than for Pcap. Thus, the better fit for Pcap seen in the scatter diagrams shown previously is not robust. At 40°N in summer, the percent unexplained variance for Pcap is actually greater than 100%—this occurs because we do not include a y intercept in our linear fit. Including a y intercept hardly changes our results when the fit is good (<40%) but it does constrain the percent unexplained variance to be less than 100% when the fit is poor. We also show the fit of evaporation averaged poleward of a latitude circle (Ecap) to demonstrate the poor fit to the temperature and zonal wind change regardless of season. We will hypothesize on some other factors controlling the extratropical evaporation in the next section.

e. Evaporation

The model spread of the evaporation is about the same as the precipitation (not shown), however, as mentioned previously, the model changes in temperature (at 850 mb) and zonal wind do not explain the model spread (Fig. 10). Moreover, the scatter does not improve when surface air temperature is used instead of 850-mb temperature. Looking at the annual-mean ensemble-mean change in evaporation for the A2 sce-
nario (Fig. 11), we see that evaporation actually decreases in some regions under global warming. The largest decreases in evaporation occur in the North Atlantic Ocean, which also corresponds to the region where the sea surface temperature (SST) increases the least (Fig. 11b). In the Southern Hemisphere, evaporation decreases in a ring around Antarctica with the largest decreases occurring in the Pacific southeast of New Zealand. This region southeast of New Zealand also corresponds to the region with the smallest SST increase.

We hypothesize that the collocation of regions of small evaporation change and small SST change are not a coincidence, but that the small evaporation change may be a result of the small SST change. The evaporation ($E$) can decrease even if the SST increases because there are two temperatures that enter in the calculation for $E$: the surface temperature and the air temperature. If the air temperature increases significantly more than the surface temperature, this will act to reduce surface evaporation so that the net increase in evaporation will be less than the $C$–$C$ rate. The difference between the surface temperature increase and the air temperature increase does not have to be large for this to occur. For example, the evaporation in the current climate ($E_0$) is

$$E_0 = C[ q_{sat}(T_{a0}) - R q_{sat}(T_{a0})],$$  

(12)

where $C$ is a bulk formula parameter that depends on wind speed, $q_{sat}$ is the saturation specific humidity, $T_s$ is the surface temperature, $T_a$ is the air temperature, and $R$ is the relative humidity. In the future climate, the new evaporation ($E_1$) is

$$E_1 = C[ q_{sat}(T_{a1}) - R q_{sat}(T_{a1})].$$  

(13)

We assume that the wind speed does not change in the future climate so that $C$ is constant and we assume that $R$ is constant also. The difference between $E_1$ and $E_0$ can be written as

$$E_1 - E_0 = C \left[ \delta_s \frac{\partial q_{sat}(T_{a0})}{\partial T} - R \delta_a \frac{\partial q_{sat}(T_{a0})}{\partial T} \right],$$  

(14)

where we have used the fact that the temperature change due to global warming, $\delta$, is small. If $T_{a0}$ is close to $T_{a0}$, then the evaporation decreases if $\delta_s < R \delta_a$. In the region of the North Atlantic with a decrease in
evaporation, for example, the model-average 1000-mb relative humidity is about 80% and the ratio $\delta_e/\delta_a$ is less than 0.8 (not shown). Therefore, the condition $\delta_e < R\delta_a$ for evaporation decreases holds. The regions of small SST change correspond well to the regions of small $\delta_e/\delta_a$, suggesting that variations in the SSTs are involved in the evaporation decrease.

This apparent link between the SST and the evaporation, as well as the North Atlantic location of the Northern Hemisphere evaporation decrease, suggests that the ocean thermohaline circulation might be involved. The hypothesis is that a slowdown in the ocean thermohaline circulation causes the SSTs in the polar oceans to increase less than the SSTs in other regions. A scatterplot of the change in ocean heat transport in the midlatitudes versus the change in evaporation over polar oceans indeed suggests that model changes in ocean heat transport might be responsible for the model spread in evaporation (Fig. 12).

4. Conclusions

In this paper, we have looked at changes in the hydrological cycle using Clausius–Clapeyron (C–C) relationship scaling. In this perspective the changes in the hydrological cycle are fit to

$$100\log \left( \frac{H_{21}}{H_{20}} \right) = a(T_{21} - T_{20}).$$

(15)
where $H$ is a climatological hydrological quantity in either the twenty-first or twentieth century (subscripts), $T$ is the temperature, and $a$ is the linear least squares fit parameter. Because relative humidity is nearly constant, the column-integrated water vapor follows (15) with a slope given approximately by $100L R_e^{-1} T^{-2}$, where $L$ is the latent heat of vaporization and $R_e$ is the gas constant for water vapor. For poleward moisture transport and precipitation, a significantly greater fraction of the intermodel variance can be explained by including both the temperature change and the zonal wind change:

$$100 \log \left( \frac{H_{21}}{H_{20}} \right) = a(T_{21} - T_{20}) + b(u_{21} - u_{20}).$$

The change in zonal wind is a proxy for the change in the transients, which are directly responsible for the poleward moisture transport and precipitation in the extratropics. The dependence on temperature, $a$, is slightly less than the C–C rate for the moisture transport, implying that the circulation must slow down in some sense in response to global warming (Held and Soden 2006). For the precipitation, the dependence on temperature, $a$, is even smaller than that for the moisture transport. Precipitation does not increase at the C–C rate because energy constraints control the strength of the hydrological cycle (Boer 1993; Allen and Ingram 2002; Pierrehumbert 2002). The slope of precipitation integrated over the polar cap ($a$) has a clear seasonal cycle with larger increases in the winter than in the summer. This is consistent with the idea that energy constraints on the hydrologic cycle are less significant at lower temperatures (Pierrehumbert 2002). Thus the wintertime precipitation increase is closer to the rate expected from the Clausius–Clapeyron relation.

A surprising finding is that the temperature and zonal wind changes due to global warming do a very good job explaining the model spread in the precipitation change averaged over the extratropics in the cool season. Evidently, the uncertainties in the precipitation increase due to global warming (e.g., Wild et al. 1997; Watterson 1998; Roeckner et al. 1999; Boer et al. 2000; Allen and Ingram 2002; Yang et al. 2003; Meehl et al. 2005) are dominated by uncertainties in tropical precipitation. For the evaporation in the extratropics, the above scaling in terms of the C–C relation does not hold very well. The evaporation appears to be tied to the sea surface temperature (SST) increase and the change in poleward ocean heat transport.

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![Fig. 12. Scatterplot of annual-mean change in evaporation over the ocean vs the annual-mean change in poleward ocean heat transport (PW) across a latitude circle. (a) Ocean heat transport across 40°N and evaporation over oceans between 45° and 70°N; (b) ocean heat transport across 35°S and evaporation over oceans between 45° and 60°S. The units of evaporation are mm day$^{-1}$.](Unauthenticated | Downloaded 08/12/22 10:07 PM UTC)
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REFERENCES


Boer, G. J., 1993: Climate change and the regulation of the surface moisture and energy budgets. *Climate Dyn.*, 8, 225–239.


