Climate Model Biases in the Width of the Tropical Belt

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ABSTRACT

Earth’s arid subtropics are situated at the edges of the tropical belt, which encircles the planet along the equator and covers half of its surface area. The climate of the tropical belt is strongly influenced by the Hadley cells, with their subsidence and easterly trade winds both sustaining the aridity at the belt’s edges. The understanding of Earth’s past, present, and future climates is contingent on understanding the dynamics influencing this region. An important but unanswered question is how realistically climate models reproduce the mean state of the tropical belt. This study augments the existing literature by examining the mean width and seasonality of the tropical belt in climate models from phase 5 of CMIP (CMIP5) and experiments from the second phase of the Chemistry–Climate Model Validation (CCMVal-2) activity of the Stratospheric Processes and Their Role in Climate (SPARC) project. While the models overall reproduce the structure of the tropical belt width’s seasonal cycle, they underestimate its amplitude and cannot consistently reproduce the seasonal cycle lag between the Northern Hemisphere Hadley cell edge and subtropical jet latitudes found in observations. Additionally, up to 50% of the intermodel variation in mean tropical belt width can be attributed to model horizontal resolution, with finer resolution leading to a narrower tropical belt. Finer resolution is associated with an equatorward shift and intensification of subtropical eddy momentum flux convergence, which via the Coriolis torque explains essentially all of the grid-size bias and a large fraction of the total intermodel variation in Hadley cell width.

1. Introduction

Earth’s subtropical arid regions are situated under the downward branches of the Hadley cells, zonal-mean meridional overturning cells where subsidence stabilizes the atmosphere and suppresses precipitation. Midlatitude eddies and the lower branch of the Hadley cells flux moisture out of the subtropics and contribute to their dryness. The region straddling the equator and bounded by these arid regions is commonly referred to as the tropical belt. Because of the strong meridional temperature and precipitation gradients across the tropical belt edge latitudes, any meridional shift of the circulation will project as major changes to surface climate (Birner et al. 2014). In fact, precipitation decreases and increased drought frequency have been observed in subtropical arid regions throughout the world (Lucas et al. 2014), and the area covered by arid land has increased (Feng and Fu 2013).

The meridional movement of the edge of regional overturning circulations modulates regional precipitation in the Northern Hemisphere subtropics and subtropical South America (Chen et al. 2014), supporting the notion that changes to the width of the Hadley cells could be responsible for these broad changes in precipitation and aridity. Future precipitation shifts in phase 5 of the Coupled Model Intercomparison Project (CMIP5) projections of global warming have been linked to changes in large-scale dynamics, rather than thermodynamic changes (Scheff and Frierson 2012). Indeed, the poleward shift of the boundaries of the Hadley cells in model projections of global warming is concurrent with a poleward displacement of subtropical dry zones (Lu et al. 2007).

In general, there are three broad classes of metrics typically employed to measure the tropical belt width: circulation-based metrics (e.g., based on the subtropical jets and the Hadley cells), temperature-based metrics (e.g., based on tropopause structure), and surface climate metrics (e.g., based on surface winds and precipitation). Some of these are difficult to assess from observations, such as the commonly used mean...
meridional overturning circulation comprising the Hadley cells (Waliser et al. 1999), adding to the uncertainty in the tropical widening calculated from reanalyses (Lucas et al. 2014). Additionally, different metrics and models exhibit a wide range of historical and projected tropical expansion rates (Davis and Rosenlof 2012; Davis and Birner 2013; Lucas et al. 2014), which could be due to a physical decoupling of the metrics (Davis and Birner 2013). Examining the mean climate of the tropical belt may illuminate why the metrics differ.

Davis and Birner (2013) explored the seasonality and interannual variability of the edges and width of the tropical belt in reanalyses and observations using three objective metrics for the edges of the Hadley cells and the latitudes of the subtropical jet and tropopause break. The key findings were that 1) the seasonal cycle amplitude in reanalyses is weaker than that measured by observations and 2) the Northern Hemisphere edge of the Hadley cell lags the latitudes of the subtropical jet and tropopause break by one month on monthly time scales. This final result, found to be robust in reanalyses, was argued to be a critical measure of the dynamics of the Northern Hemisphere zonal-mean circulation.

While care needs to be taken in interpreting trends assessed from reanalyses (Trenberth et al. 2001; Bengtsson et al. 2004), model projections indicate that the expected widening resulting from anthropogenic forcings is roughly an order of magnitude weaker than that measured in reanalyses (Johanson and Fu 2009; Hu et al. 2013; Quan et al. 2014). There is clearly a discrepancy between models, reanalyses, and observations concerning the magnitude of observed and projected widening, a now long-outstanding issue (Seidel et al. 2008).

It may be worthwhile, then, to augment the existing analyses of projected tropical widening and assess the climate models’ ability to reproduce the climatological mean tropical belt width. The models’ fitness in reproducing the observed climate of the Hadley cells, jet streams, and tropopause structure should be weighed when considering model projections, which may help to resolve the discrepancy in their trends.

In this study, we document the seasonality of the tropical belt width in a host of climate models from both the CMIP5 and the second phase of the Chemistry–Climate Model Validation (CCMVal-2) activity of the Stratospheric Processes and Their Role in Climate (SPARC) project experiments using three objective metrics for the tropical belt width in order to eliminate uncertainties and biases that may arise from the use of arbitrary or threshold-based metrics. Models generally underestimate the seasonal cycle amplitude of the tropical belt width compared to observations. Only a subset of models reproduce the observed seasonal cycle lag between the edge of the Hadley cell and the latitude of the subtropical jet core and tropopause break in the Northern Hemisphere, as well as their synchronicity in the Southern Hemisphere. Additionally, we find evidence that the grid size of the models biases the tropical belt width, with finer horizontal resolution leading to a narrower tropical belt. An analysis of the eddy momentum flux convergence and wave fluxes reveals that it is changes to wave propagation and the balance between the momentum flux convergence and the Coriolis torque by the mean meridional circulation that is associated with this bias.

## 2. Data

We use historical runs from the CCMVal-2 and CMIP5 experiments to have a sample of models spanning an array of model configurations. We also employ two modern reanalyses, one legacy reanalysis, and Constellation Observing System for Meteorology, Ionosphere, and Climate (COSMIC) global positioning system radio occultation (GPS-RO) measurements as benchmarks.

Monthly mean output from the first ensemble member, initialization, and physics package (r1i1p1) from 25 model simulations of the CMIP5 historical scenario was used, with details of each model listed in Table 1. The

<table>
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<th>Model</th>
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<tr>
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<td>GISS-E2-H</td>
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<td>GISS-E2-R</td>
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<td>HadGEM2-ES</td>
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<td>MIROC-ESM-CHEM</td>
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<td>MIROC5</td>
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<td>MRI-CGCM3</td>
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<td>NorESM1-M</td>
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historical simulation runs from 1850–2005 and imposes observed forcings on the climate system (Taylor et al. 2012).

The CCMVal-2 transient reference simulation forced by observations (REF-B1) scenario runs from 1960–2006 and imposes observed forcings on the climate system, as well as sea surface temperatures and sea ice concentrations (Eyring et al. 2008), which is unlike the CMIP5 historical scenario. Monthly mean output from 15 model simulations from the REF-B1 scenario are used in this study and are detailed in Table 2. The CMAM model is the only CCMVal-2 model with a coupled ocean (Eyring et al. 2008).

The three reanalyses used in this study include the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim, hereafter ERA-I; Dee et al. 2011), the Modern-Era Retrospective Analysis (MERRA; Rienecker et al. 2011), and the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (hereafter NCEP; Kalnay et al. 1996). We use each reanalysis’ monthly mean, pressure-gridded product and additionally employ ERA-I’s 6-hourly model level product to subsample ERA-I to COSMIC observations. MERRA, ERA-I, and NCEP are supplied on a \( \frac{1}{6}^\circ \times \frac{1}{6}^\circ \), \( \frac{1}{4}^\circ \times \frac{1}{4}^\circ \), and \( 2.5^\circ \times 2.5^\circ \) latitude–longitude resolution grid, respectively. The three reanalyses’ pressure-gridded outputs have the same vertical resolution about the tropical tropopause. MERRA and NCEP institute a three-dimensional variational data assimilation (3D-Var) scheme (Rienecker et al. 2011; Kalnay et al. 1996), whereas ERA-I assimilates COSMIC GPS-RO and uses a four-dimensional variational data assimilation (4D-Var) scheme (Dee et al. 2011). Only the 1979–2013 period is used in this analysis.

Finally, we employ GPS-RO retrievals from COSMIC supplied by the University Corporation for Atmospheric Research (UCAR). The bending angle of GPS radio waves measured by the COSMIC satellite constellation is proportional to the vertical gradient of the refractive index, which below the ionosphere is itself a function of the air pressure, the air temperature, and the water vapor partial pressure (Anthes et al. 2008). At temperatures below about 250 K, the contribution from water vapor is negligible. That is, in the upper troposphere and stratosphere, temperature profiles can be constructed as a function of pressure, neglecting the water vapor contribution; these are referred to as dry profiles. The COSMIC data supplied by UCAR also provide so-called wet profiles, which are constructed by running a one-dimensional variational data assimilation (1D-Var) scheme with ECMWF operational weather analyses to separate the water vapor and temperature contributions to the refractive index [see Anthes et al. (2008) for further details].

We average daily profiles of COSMIC temperature and pressure into 181 equally spaced latitude bins between 90°S and 90°N and 200 equally spaced altitude bins between 0 and 40 km to produce monthly zonal means. These zonal means on altitude are then interpolated onto a pressure grid to create a zonal mean of temperature and geopotential height on pressure. COSMIC retrievals typically do not penetrate to the surface, so we use the ERA-I climatology for zonal-mean surface potential temperature as the surface field for COSMIC for the tropopause metric described later. We use retrievals from 2007–13, excluding 2006 because

<table>
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<th>Model</th>
<th>Expansion</th>
<th>Resolution (lat × lon)</th>
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<td>CCSRNIRES</td>
<td>Center for Climate System Research/National Institute for Environmental Study GCM</td>
<td>2.8° × 2.8°</td>
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<tr>
<td>CMAM</td>
<td>Canadian Middle Atmosphere Model</td>
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</tr>
<tr>
<td>CNRM-ACM</td>
<td>CNRM–ARPEGE Coupled with Modèle de Chimie Atmosphérique de Grande Echelle (MOCAGE)</td>
<td>2.8° × 2.8°</td>
</tr>
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<td>E39CA</td>
<td>ECHAM4.L39 (Deutsches Zentrum für Luft- und Raumfahrt)–Chemistry–Atmospheric Tracer Transport in a Lagrangian Model (ATTILA)</td>
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<td>EMAC</td>
<td>ECHAM5 Middle Atmosphere with Chemistry</td>
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<td>GEOS Chemistry–Climate Model</td>
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<td>Meteorological Research Institute</td>
<td>2.8° × 2.8°</td>
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<td>NIWA-SOCOL</td>
<td>National Institute of Water and Atmospheric Research Solar–Climate–Ozone Links</td>
<td>4° × 4°</td>
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<td>Met Office Unified Model–U.K. Chemistry Aerosol Community Model</td>
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<td>WACCM</td>
<td>Whole Atmosphere Community Climate Model</td>
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TABLE 2. CCMVal-2 horizontal model resolution in degrees.
of the low number of profiles during this spinup period. In our analysis of tropical belt widths, COSMIC is downsampled to the ERA-I pressure grid to ensure any differences are not due to the vastly different vertical resolutions.

We employ the dry profiles rather than the wet profiles as there is evidence, presented in appendix A, that the wet profiles exhibit a warm bias in the subtropical stratosphere that may influence our tropical belt calculations. Some values may differ with respect to COSMIC between this paper and a previous paper, Davis and Birner (2013), which used the wet profiles to calculate the tropical belt width.

3. Methods and metrics

Tropical belt width metrics that rely on subjectively determined numerical thresholds or that are computed on arbitrary pressure levels may be subject to individual models’ biases, and mean-state changes associated with greenhouse gases or other forcings may erroneously project as tropical widening trends. The metrics used here instead objectively measure distinct phenomena and are not subject to such limitations.

We evaluate the tropical belt edge latitudes using the latitudes based on the subtropical jet cores (the $U_{\text{max}}$ metric), the latitudes based on the location where the vertically averaged mean meridional streamfunction (MMS) vanishes (the $\int \Psi \, dp$ metric), and the latitudes of the tropopause break based on the maximum tropospheric dry bulk stability (the $\Delta \theta$ metric) (Davis and Birner 2013). These straightforward metrics are computed from zonal-mean, monthly mean fields, and the tropical belt width is defined as the difference (in degrees latitude) between the Northern and Southern Hemisphere edge latitudes. The $\int \Psi \, dp$ metric is qualitatively similar to other streamfunction-based metrics evaluated at a particular pressure level, such as 500 hPa (Lu et al. 2007; Frierson et al. 2007), or over a mid-tropospheric layer, such as 600–400 hPa (Hu and Fu 2007; Johanson and Fu 2009) or 700–400 hPa (Stachnik and Schumacher 2011). The $\Delta \theta$ metric is similar to tropopause height-based metrics that attempt to measure the latitude of the subtropical tropopause break (Davis and Rosenlof 2012); however, it specifically examines the dry bulk static stability, a critical quantity in the two-layer baroclinic instability Hadley cell scaling theory (Held 2000). The subtropical jet is theoretically connected to the poleward-flowing upper branch of the Hadley cell through angular momentum conservation (Held 2000), although the poleward flow in the real atmosphere is not angular momentum conserving due to eddy fluxes (Schneider 2006). Through the thermal wind relation, the subtropical jet core also corresponds to a region of high baroclinicity, characteristic of the tropopause break in the subtropics.

a. Subtropical jet cores $U_{\text{max}}$

The phenomenological difference between the subtropical and eddy-driven jets is their surface component. While the eddy-driven jet is characterized by westerly surface winds sustained against frictional dissipation by a net eddy momentum flux convergence, the subtropical jet is characterized by near-zero surface wind. The strong vertical shear in the subtropical jet is owing to the strong meridional temperature gradient in the tropics, whereas the weak vertical shear of the eddy-driven jet is due to eddy heat fluxes in baroclinic zones constantly damping the temperature gradient. The location of the maximum vertical shear from the surface in the zonal-mean zonal wind column, then, is a natural method for identifying the subtropical jet core.

Accordingly, we define the latitude of the subtropical jet core as the latitude of the most equatorward maximum in the maximum column wind speed (Strong and Davis 2006), with the surface component of the wind subtracted, in each hemisphere. Davis and Birner (2013) noted difficulty in locating the subtropical jet core in the zonally mean in the Southern Hemisphere in austral summer, resorting to a gradient-based approach that was not conclusive. Here, by subtracting the zonal-mean surface wind from each zonal-mean zonal wind column we find a well-defined subtropical jet core distinct from the eddy-driven jet core in all cases. The maximum column wind speed is simply the maximum zonal-mean zonal wind speed in each surface-wind-subtracted column. We limit the vertical extent of our search for the jet core to 50 hPa above the tropopause, in order to avoid mistakenly identifying the stratospheric jet, and only search poleward of 10° latitude in order to exclude small tropical wind maxima. While this domain is necessarily subjective, the actual metric is not. Linear interpolation is used to find the maximum between grid points. The resulting tropical belt width and maximum column wind speed are here referred to as the $U_{\text{max}}$ metric and $U_{\text{max}}$ diagnostic field.

An observational wind product can be obtained from COSMIC by solving the balanced zonal-mean meridional momentum equation in height coordinates on the sphere (Davis and Birner 2013):

$$f[u] + \frac{\tan(\phi)}{a}[u]^2 = -\rho \frac{1}{a} \frac{\partial p}{\partial \phi}$$  \hspace{1cm} (1)

for $[u]$, where $f = 2 \Omega \sin(\phi)$ is the Coriolis parameter, $\Omega = 7.292 \times 10^{-5} \text{ s}^{-1}$ is Earth’s angular velocity, $a = 6371 \text{ km}$...
is Earth’s mean radius, φ is the latitude, ρ is the dry air density, p is the pressure, [v] is the zonal-mean zonal wind, and square brackets denote the zonal mean. The geostrophic wind field is then used in lieu of the zonal-mean zonal wind field to calculate the \( U_{\text{max}} \) metric and is referred to instead as the \( U_{g,\text{max}} \) metric.

\section*{b. Hadley cell edge \( \int \Psi \, dp \)\)

We define the edge latitude of the Hadley cell as the latitude where the vertically averaged MMS vanishes, poleward of its tropical maximum (minimum) in the Northern (Southern) Hemisphere. The MMS is the vertically integrated meridional mass flux between a given pressure level and the top of the atmosphere (Holton 1994):

\begin{equation}
\Psi(p, \phi) = \frac{2\pi a \cos(\phi)}{g} \int_0^\phi \left[ \frac{\partial T}{\partial z} \right] dp,
\end{equation}

where \( g = 9.81 \text{ m s}^{-1} \) is the acceleration due to gravity, \( p \) is the pressure, \( \phi \) is the latitude, \([v]\) is the zonal-mean meridional wind, and \( \Psi(p, \phi) \) is the MMS at the given pressure and latitude. We then vertically average the mass-weighted MMS over the depth of the atmosphere, measuring the average meridional overturning mass flux at a given latitude. Because the stratosphere only represents approximately 10\% of the mass of the tropical atmosphere, and the MMS is orders of magnitude weaker in the stratosphere than in the troposphere, the stratospheric contribution to the vertically averaged MMS is negligible (less than 1\%). In other words, averaging over the entire atmosphere produces results indistinguishable from averaging up to some arbitrary level but comes with the advantage that no specific upper-limit pressure level must to be specified. The interpretation of the zero contour of the vertically averaged MMS is straightforward: it is the boundary between the Hadley and Ferrel cells. Unlike typical streamfunction-based metrics, this metric does not evaluate the zero contour on a fixed pressure level or over an arbitrary layer and thus should not be subject to individual model biases in the structure of the circulation. Linear interpolation is used to find the edge latitude between grid points. The resulting width and vertically averaged MMS are here referred to as the \( \int \Psi \, dp \) metric and the \( \int \Psi \, dp \) diagnostic field.

\section*{c. Tropopause break \( \Delta \theta \)\)

The latitude of the tropopause break is defined as the latitude of maximum tropospheric bulk stability in each hemisphere, where the tropospheric bulk stability is defined as the difference in potential temperature between the tropopause and the surface. The bulk stability field exhibits an unambiguous peak in the subtropics at the subtropical tropopause break (Frierson and Davis 2011; Davis and Birner 2013). We define the tropopause as the level of maximum curvature in the temperature profile (Birner 2010), characterizing the sharp transition between the troposphere and the highly stable stratosphere. This objective method captures the standard WMO thermal tropopause (WMO 1957) in the extratropics and the cold-point tropopause in the tropics and hence does not require a priori knowledge of the tropical belt edge latitudes to determine which particular definition to use. As in Davis and Birner (2013), we calculate the curvature of the lapse rate (\( \partial_{zzz} T \)) at every vertical level and interpolate in geopotential height space to the level where \( \partial_{zzz} T = 0 \). Once the tropopause height is known, temperature and pressure can be uniquely defined using the lapse rate and the hypsometric equation, respectively. To calculate the surface potential temperature from reanalyses and models, 2-m temperature and surface pressure are used. For COSMIC, no direct observations of surface variables are available, so the climatological surface potential temperature from ERA-I is used for COSMIC. This substitution is reasonable because the nearly 40-K variation in the \( \Delta \theta \) diagnostic field from the equator to the tropopause break is dominated by the temperature structure of the tropical tropopause. Interpolation is not used to identify the edge latitude between grid points, as the \( \Delta \theta \) diagnostic field is discontinuous at the tropopause break. The resulting tropical belt width and tropospheric bulk stability field are here referred to as the \( \Delta \theta \) metric and \( \Delta \theta \) diagnostic field.

\section*{4. Tropical belt seasonality\)

The \( \int \Psi \, dp, \Delta \theta, \text{ and } U_{\text{max}} \) diagnostic fields (not the tropical belt width metrics) as a function of latitude are shown for an example January and July in Fig. 1. These fields distill the zonal-mean climate into a single dimension that is more easily compared across models.

For the reanalyses, the CCMVal-2 REF-B1 simulations, and COSMIC, we use the year 2007, as COSMIC was undergoing a spinup period in 2006. The CMIP5 historical simulation ends in 2005, so we use the year 2005 for this model ensemble only. The diagnostic fields for these particular years represent typical monthly mean conditions.

The \( \int \Psi \, dp \) diagnostic field shows a distinct maximum (minimum) in January (July) associated with the strong cross-equatorial winter Hadley cell, as well as a weak summer cell with an overturning strength similar to the Ferrel cells. The CMIP5 historical scenario ensemble-mean Hadley cell in both January and July is stronger than the CCMVal-2 REF-B1 ensemble mean and also
shows a greater model spread. This behavior does not appreciably differ in other years and may warrant further investigation.

The ensemble-mean $U_{\text{max}}$ diagnostic fields are nearly coincident, although there is one outlier model from the CCMVal-2 experiments with an equatorward-shifted $U_{\text{max}}$ diagnostic field: the CMAM, which is the only CCMVal-2 model with a coupled ocean. CMAM’s $U_{\text{max}}$ diagnostic field is curious because it does not resemble those of the CMIP5 models, which similarly have coupled oceans. There are only small differences between COSMIC’s $U_{g,\text{max}}$ and ERA-I’s $U_{\text{max}}$ diagnostic fields outside of the tropics, indicating that the balanced wind is likely a good approximation.

The $\Delta \theta$ diagnostic field shows a large spread between models, with differences in the $\Delta \theta$ diagnostic field spanning nearly 30 K (standard deviation of 8 K), which is greater than the model spread in tropopause temperature (Kim et al. 2013). However surface temperature and tropopause pressure differences can certainly “conspire” to amplify model differences in the bulk stability. For the most part, the CCMVal-2 and CMIP5 models reproduce the broad-scale structure of the tropospheric bulk stability, but the variability in the latitude of the tropopause break makes this difficult to assess.

Figure 2 aligns the annual-mean $\Delta \theta$ diagnostic field from each model about the tropopause break or the location of the maximum $\Delta \theta$. COSMIC exhibits a relatively sharp tropopause break compared to the models and reanalyses and also tends to have a higher tropical and lower extratropical $\Delta \theta$. NCEP is a clear outlier with a very gently sloping transition between the tropical and extratropical tropopauses. Overall, the models capture the structure of the tropopause break in the $\Delta \theta$ diagnostic field, in particular its alignment with the maximum $\Delta \theta$ in each hemisphere, but differ substantially in the actual values of $\Delta \theta$. The $\Delta \theta$ metric then, unlike a threshold-based tropopause metric, will objectively capture the latitudes of the tropopause break in climate models without being subject to an individual model’s biases in tropopause height, temperature, or pressure.

The seasonal cycles of the ensemble-mean tropical belt widths based on the $\int \Psi \, dp$, the $U_{\text{max}}$, and the $\Delta \theta$ metric are shown in Fig. 3. In an annual-mean sense, the CMIP5 models have a much narrower $\Delta \theta$ metric.
width than the CCMVal-2 models (Fig. 3b), with the CMIP5 models on average narrower than COSMIC and ERA-I. However, there is tremendous spread about the ensemble-mean \( \Delta U \) metric widths. The CMIP5 and CCMVal-2 models have a slightly narrower \( \Psi dp \) metric width than ERA-I (Fig. 3a), and it appears that the models, ERA-I, and COSMIC tend to be in closer agreement on the mean \( U_{max} \) metric width (Fig. 3c).

With regard to amplitude, here defined as one-half the difference between the maximum and minimum climatological monthly mean tropical belt widths, Davis and Birner (2013) found that reanalyses slightly underestimate the seasonal cycle amplitudes for the \( \Delta U \) metric and \( U_{max} \) metric widths compared with COSMIC. The models also underestimate the \( \Delta U \) (Fig. 3b) and \( U_{max} \) (Fig. 3c) seasonal cycle amplitudes. The ensemble-mean \( \Delta U \) seasonal cycle amplitudes from CCMVal-2 and CMIP5 are 4.9° and 6.7° latitude, respectively, compared to 8.7° latitude from COSMIC. However, the models’ seasonal cycle amplitudes are overall comparable to those in ERA-I.

The phases of the CCMVal-2 \( \Psi dp \) metric widths are slightly different from those in CMIP5 and ERA-I, with a peak in tropical belt width in August rather than September. This phase difference in CCMVal-2 is not apparent in the other metrics. If the tropical belt width metrics were strongly coupled, the other metrics should exhibit a similar shift in seasonal cycle phase in CCMVal-2, but there may also be a difference in their dynamical relationships in the CCMVal-2 models than in either CMIP5 or ERA-I.

Davis and Birner (2013) found that the Northern Hemisphere edge latitude of the \( \Psi dp \) metric lags the edge latitudes of the \( U_{max} \) and \( \Delta U \) metrics by one month in the seasonal cycle, whereas the three metrics are synchronized in the Southern Hemisphere. This was hypothesized to be as a result of enhanced zonally asymmetric circulations in the Northern Hemisphere, such as the monsoons causing the Hadley cell edge to lag the seasonal cycle of the subtropical jet latitude. Similar to Davis and Birner (2013), for each model, we lag correlate 12 months of \( \Psi dp \) metric edge latitudes with 12 months of \( \Delta U \) and \( U_{max} \) metric edge latitudes and find the monthly lag with the highest correlation coefficient. This is done for every year in the time series, and the average of these monthly lags is defined as the best-fit lag for each model. Histograms of the best-fit lags for the CMIP5 and CCMVal-2 models are shown in Fig. 4 with bins centered on integer lags. Typical values of the correlation coefficients for the seasonal cycle range between 0.5 and 0.9, and no best-fit lags outside of...
±2 months were found. A positive lag indicates the \( f\Psi dp \) metric edge lagging the other metrics.

In the mean, the CMIP5 historical models tend toward a 1-month lag in both hemispheres, while the CCMVal-2 models tend toward a 0-month lag in both hemispheres, in both cases resulting in an error in one hemisphere. The 1-month lag between the \( f\Psi dp \) and \( U_{\text{max}} \) metric edge latitudes in the Northern Hemisphere is captured by 80% of the CMIP5 models (Fig. 4a), but fewer than half produce a 0-month lag in the Southern Hemisphere (Fig. 4b). Only half of the CCMVal-2 models reproduce the 1-month Northern Hemisphere lag, while only 40% reproduce the 0-month Southern Hemisphere lag. The outlier models with ±2-month lags are overwhelmingly from CCMVal-2. The results for the \( f\Psi dp \) metric edge latitudes lagging the \( \Delta\theta \) metric edge latitudes are similar, although the lag distributions are broader. In total, some CCMVal-2 models do not accurately reproduce the dynamics of the Northern Hemisphere seasonal cycle, while some CMIP5 and CCMVal-2 models do not accurately reproduce some of the dynamics of the seasonal cycle in the Southern Hemisphere.

Some models have particularly poor performance with respect to the seasonal lag. The ULAQ, CAM3.5, and UMUKCA-METO models from CCMVal-2 show the \( f\Psi dp \) metric edge leading both the \( U_{\text{max}} \) and \( \Delta\theta \) metric edge latitudes in the Northern Hemisphere by 2 months, while the E39CA and EMAC models from CCMVal-2 and the FIO-ESM from CMIP5 show a 2-month lag for the same. The inability of some of the CCMVal-2 models to capture the lag may be related to their more irregular seasonal cycle (Fig. 3c).

However, the ACCESS1.1, ACCESS1.3, MRI-CGCM3, and NorESM1-M models from CMIP5 are exceptional in capturing the Northern Hemisphere 1-month lag and Southern Hemisphere 0-month lag between the \( f\Psi dp \) and the \( \Delta\theta \) and \( U_{\text{max}} \) metric edge latitudes. The CMAM is the only CCMVal-2 model that accurately reproduces these seasonal lags. Since CMAM is also the only CCMVal-2 model with a coupled ocean, this might indicate that reproducing the observed lag is aided by coupled ocean–atmosphere dynamics. Many of the CMIP5 models have a coupled ocean and show overall better performance in the Northern Hemisphere.

The dynamics governing the seasonality of the tropical belt edge latitudes are not yet fully understood. Zonal asymmetries such as those that arise from the monsoons were suggested as a possible reason for the 1-month lag in the Northern Hemisphere (Davis and Birner 2013), with some indication that the stationary waves associated with the monsoons play a pivotal role in the seasonal transitions in the Northern Hemisphere zonal-mean circulation (Shaw 2014). This suggests, for example, that many of the CMIP5 models may be producing the dynamics critical for the Northern Hemisphere tropical belt edge seasonality while simultaneously incorrectly producing similar dynamics in the Southern Hemisphere, and vice versa for the CCMVal-2 models.

5. Tropical belt grid-size bias

In an attempt to understand the difference in mean tropical belt width and seasonality between the climate models, reanalyses, and observations, we performed an exhaustive analysis of plausible model biases in the tropical belt width. For example, increases in tropical surface air temperature can drive Hadley cell expansion (Adam et al. 2014; Frierson et al. 2007) by modifying tropical static stability via moist adiabatic adjustment. It is reasonable to hypothesize, then, that tropical surface air temperature may predict intermodel variations in the mean tropical belt width. For example, increases in tropical surface air temperature can drive Hadley cell expansion (Adam et al. 2014; Frierson et al. 2007) by modifying tropical static stability via moist adiabatic adjustment. It is reasonable to hypothesize, then, that tropical surface air temperature may predict intermodel variations in the mean tropical belt width. However, we find no significant correlations between the tropical surface air temperature and the mean tropical belt width, with insignificant correlation coefficients ranging from −0.01 to 0.17 (Fig. 5). Here, tropical surface air temperature is defined as the mean 2-m air temperature between 20°S
and 20°N (Adam et al. 2014) weighted by the cosine of latitude, and the mean tropical belt width is the mean of every monthly mean tropical belt width over the entire historical run. Changes in the gradient in surface air temperature might also be associated with changes in the Hadley cell width (Adam et al. 2014; Frierson et al. 2007), although we also find no such intermodel relationship with the mean tropical belt width. This indicates the quantities associated with changes in the tropical belt width within models are not necessarily the same quantities defining the intermodel variation in the mean tropical belt width.

The sole quantity dictating any significant intermodel variation in the mean tropical belt width that we were able to identify was horizontal grid size, with finer horizontal grid size leading to a narrower tropical belt with correlation coefficients ranging between 0.4 and 0.62 (Fig. 6). These correlations extend to the tropical belt edge latitudes as well, with similar values (not shown). The highest correlation for the CMIP5 historical and CCMVal-2 REF-B1 models separately is for CCMVal-2 REF-B1’s $\psi dp$ metric width ($R = 0.71$), indicating that 50% of the total intermodel variation in the $\psi dp$ metric width in the CCMVal-2 REF-B1 scenario is attributable to grid size. We have excluded the ULAQ model from this analysis as the correlations are sensitive to its large grid size of 11.5°; however, it obeys the relationship between grid size and tropical belt width (e.g., for the $\Delta \theta$ metric, its width is 79.2°, the widest tropical belt width for this metric).

The IPSL-CM5A model from CMIP5 is run at both low horizontal resolution (IPSL-CM5A-LR) and moderate horizontal resolution (IPSL-CM5A-MR). This grid-size bias manifests within this model, as the low-resolution $\psi dp$ metric width of 62.9° and $U_{\text{max}}$ metric width of 66.9° are larger than the moderate resolution $\psi dp$ metric width of 61.1° and $U_{\text{max}}$ metric width of 66.2°. However, the model’s $\Delta \theta$ metric width is marginally larger in the low-resolution run (46.9° versus 47.4°). This is evidence that the grid-size bias in the tropical belt width, at least in the circulation-based metrics, is consistent at the level of an individual model.

Most models have a $U_{\text{max}}$ metric tropical belt width that is wider than observed in COSMIC (Fig. 6c), but with finer resolution the modeled width approaches the observed value. However, increasing model resolution does not lead to a more accurate tropical belt width for the $\Delta \theta$ metric width, where all models (including reanalyses) with a resolution finer than 2° latitude exhibit a narrower tropical belt than COSMIC (Fig. 6b). Most of the models have a narrower $\psi dp$ width than the reanalyses, the opposite of which is true for the $U_{\text{max}}$ width.

Model performance in simulating the mean tropical belt width does depend upon the particular metric, but there is a significant, positive correlation between model grid size and mean tropical belt width regardless of the metric used. The source of this model bias is investigated next, as it may help to resolve the model differences in the mean tropical belt width and illuminate the processes governing the width of the tropical belt.

6. Eddy grid-size bias

Hadley cell scaling theories that assume angular momentum conservation (Held 2000) effectively assume an absolute vorticity of zero within the Hadley cell (the
The meridional derivative of zonal-mean angular momentum is the zonal-mean absolute vorticity. In these theories, the zonal-mean zonal momentum balance is between the meridional advection of zonal momentum $\nabla_y^2 [y z f]$ and the Coriolis torque $2 f [y]$, so that the sum $\zeta + f$, or the absolute vorticity, remains zero within the poleward flow of the upper branch of the cell. However, in reality, the magnitude of $f$ substantially increases in the subtropics, and the relative vorticity contribution becomes small. A more accurate balance outside the tropics is between the Coriolis torque and the eddy momentum flux convergence:

$$-f [u v] \approx - \frac{\partial [u' v']}{\partial y}, \quad (3)$$

where $[u' v']$ is the horizontal eddy momentum flux, where primes denote departures from the zonal mean, and we have assumed a steady state away from the surface and that $|\zeta| \ll f$. Under the quasigeostrophic scaling typically employed to study dynamics away from the equator, $-f [u z f]$ vanishes from the momentum equation—advection in the quasigeostrophic framework is done by the geostrophic wind, and the zonal mean of the geostrophic meridional wind is zero.

The horizontal resolution of a model has been shown to influence the eddy momentum fluxes (Boville 1991; Held and Phillipps 1993) as well as Rossby wave breaking frequency (Béguin et al. 2013). Given that $[u]$ vanishes where the eddy momentum flux convergence is zero, the edges of the Hadley cells, where $[u] = 0$, will be located near the latitude of the maximum eddy momentum flux and thus be subject to model differences in its strength and structure. The zonal wind and tropopause structure have no simple balance condition with the eddy fluxes, so we now focus exclusively on the Hadley cells and the $\Psi dp$ metric.

**a. Eddy momentum flux grid-size bias**

We wish to answer: how much of the grid-size bias in Hadley cell width, as measured by the $\Psi dp$ metric, can be explained by a grid-size bias in the eddy momentum flux convergence that projects onto changes in the meridional wind via the Coriolis torque? Perhaps more generally, how much of the total intermodel variation in Hadley cell width can be explained not only by a grid-size bias but by a grid-size bias in this eddy forcing?

CMIP5 historical output at daily time resolution that is necessary to directly calculate eddy fluxes is limited to a subset of vertical levels and models insufficient to study the vertical structure of any eddy grid-size bias across a large enough sample space of model grid sizes. However, the CCMVal-2 REF-B1 models span a large sample of grid sizes and allow the estimation of eddy fluxes. A method of calculating and validating the eddy fluxes from time-mean model output from the CCMVal-2 REF-B1 models is detailed in appendix B. We will use these derived eddy fluxes to study the grid-size bias.

We first regress the CCMVal-2 model-mean estimated Eliassen–Palm (EP) flux and its divergence and eddy momentum flux convergence on CCMVal-2 model grid size and model-mean $\Psi dp$ edge latitudes, shown in Fig. 7. The Eliassen–Palm flux is parallel to the group velocity of linear Rossby waves, while its divergence measures the wave forcing on the mean flow (Edmon et al. 1980). Vectors of the Eliassen–Palm flux thus indicate Rossby wave propagation in the latitude–pressure plane.

An important point to remember for this and later regressions on grid size is that the patterns are not necessarily capturing a mode of physical variability, because the independent variable is not a physical index within the atmosphere of each model. For the edge latitude regressions, regressions are taken on each
respective hemisphere’s $\Psi dp$ edge latitude (e.g., the Northern Hemisphere regression coefficients correspond to the regression on the Northern Hemisphere edge latitude). The regression coefficients are all multiplied by either a one standard deviation reduction in model grid size or a one standard deviation equatorward shift in $\Psi dp$ edge latitude. The regression (shading) can thus be interpreted as the representative changes to the eddy momentum flux convergence associated with finer model resolution and a narrower Hadley cell.

There are strikingly similar changes to the patterns of wave breaking and propagation between the regression on grid size and $\Psi dp$ edge latitudes. With finer resolution and an equatorward-shifted $\Psi dp$ edge latitude, there is enhanced wave breaking ($\nabla \cdot F < 0$) and enhanced equatorward wave propagation (arrows) in the subtropical lower and upper troposphere around 700 and 150 hPa (Figs. 7a,b), with a reduction in wave breaking ($\nabla \cdot F > 0$) and anomalous poleward wave propagation in the subtropical middle troposphere. Both finer-resolution (Fig. 7c) and equatorward-shifted $\Psi dp$ edge latitudes (Fig. 7d) are associated with enhanced, equatorward-shifted eddy momentum flux divergence in the tropical upper troposphere and enhanced, equatorward-shifted eddy momentum flux convergence in the subtropical upper troposphere. Taken together, both indicate an equatorward shift of wave activity and enhanced wave breaking into the poleward flank of the Hadley cell with finer resolution and a narrower Hadley cell.

b. Impact of the bias on the Hadley cell width

To directly relate the impact of the grid-size-biased horizontal eddy momentum flux convergence on the Hadley cells, we diagnose the MMS in balance with the anomalous eddy momentum flux convergence by substituting Eq. (3) into Eq. (2) to obtain the balanced MMS associated with the anomalous horizontal eddy momentum flux convergence:

$$\Psi_{\text{eddy}}(p,\phi) = \frac{2\pi a \cos(\phi)}{g} \int_0^p \frac{1}{\partial y} \left( (u'v') \right) dp, \quad (4)$$

where $\Psi_{\text{eddy}}$ is the MMS in balance with the anomalous eddy momentum flux convergence. The grid used in these calculations avoids the equator so that there is no singularity in the balanced MMS.

The MMS in balance with the eddy momentum flux convergence associated with finer resolution exhibits thermally indirect midtropospheric cells that maximize along the edge of the Hadley cells in the subtropics (Fig. 8a), indicating a contraction of the Hadley cells. The MMS in balance with the eddy momentum flux convergence associated with equatorward-shifted $\Psi dp$ edge latitudes exhibits a more complex structure (Fig. 8b). It does include a similar subtropical midtropospheric Hadley cell contraction, but it also indicates a strengthening of the Hadley cells as they narrow. This suggests that part of the intermodel variation in $\Psi dp$ width is characterized by changes in the tropical eddy momentum flux convergence that are not associated with grid size.

To begin answering how much of the grid-size bias in the $\Psi dp$ metric width is connected to the grid-size-biased eddy momentum flux convergence, we regress the CCMVal-2 model-mean MMSs on grid size (Fig. 8c). The grid-size bias in the MMS looks almost identical to the MMS in balance with the grid-size-biased eddy momentum flux convergence, suggesting that the structural changes to the MMS with finer resolution are associated almost exclusively with changes to the eddy momentum flux convergence. Given the lack of any
significant equatorial structures in the regression, we conclude that the bias is not primarily associated with a grid-size effect in the convection scheme, although intensity changes are not necessarily the only effect convective scheme differences could have.

Finally, we regress the CCMVal-2 model-mean MMSs on the model-mean $\Psi dp$ edge latitudes to see the structure of the total intermodel variation in the MMS (Fig. 8d). This also appears similar to the MMS in balance with the eddy momentum flux convergence associated with a narrower Hadley cell (Fig. 8b), suggesting that the bulk of the intermodel variation in the MMS between models can be tied to the MMS in balance with the differences in the eddy momentum flux convergence, or at least wherever Eq. (3) is the dominant balance.

For the regressions in Figs. 8a, 8c, and 8d, we add the anomalous MMS to the CCMVal-2 multimodel-mean MMS and vertically average it (Fig. 9) so that we can apply the $\Psi dp$ metric and quantify the total effect of the eddy momentum flux grid-size bias on the $\Psi dp$ edge latitudes. The percentage of the total intermodel variation explained by the grid-size bias and the grid-size-biased eddies is found by taking the difference between their $\Psi dp$ edge latitudes (dashed and dotted curves) and the multimodel-mean edge latitudes (red curve) and dividing this by the difference between the total intermodel variation in the $\Psi dp$ edge latitudes (solid curve) and the multimodel-mean edge latitudes (red curve). Grid size alone explains 46% of the intermodel variation in $\Psi dp$ edge latitudes in the Northern Hemisphere.
and 59% in the Southern Hemispheres. The MMS in balance with the grid-size-biased eddy momentum flux convergence explains 51% of the intermodel variation in $\Psi dp$ edge latitude in each hemisphere, and MMS regression on (c) a one std dev reduction in model grid size and (d) a one std dev equatorward shift in model-mean $\Psi dp$ edge latitude in each hemisphere. Shading indicates the MMS regression coefficients multiplied by a one std dev (c) reduction in grid size and (d) equatorward shift in $\Psi dp$ edge latitude, while shading in (a) and (b) indicates the MMS in balance with the respective eddy momentum flux convergence fields from Fig. 7. Contours indicate the CCMVal-2 mean MMS (10^9 kg s^{-1}). Stippling indicates (c),(d) regression coefficients significant at the 95% confidence level or (a),(b) with an eddy-momentum-flux-convergence-weighted (Figs. 7c,d) $p = 0.05$ (95% confidence) vertically averaged from each level to the top of the atmosphere (i.e., similar to the calculation for the MMS itself).

The total intermodel variation indicates that narrower Hadley cells tend to have stronger overturning, whereas the regression on grid-size-biased eddy momentum fluxes indicates little change to the intensity of the circulation with decreasing width. However, there is a change in the intensity of the circulation with finer grid size alone that is not associated with the grid-size-biased eddy fluxes, which could be due to a grid-size effect in the convection schemes in the deep tropics (Lorant and Royer 2001; Frierson 2007; Rauscher et al. 2013). As for the intensity changes not associated with grid size, differences in convection schemes, such as the abruptness of the convection trigger (Frierson 2007) and surface evaporation (Numaguti 1993), can impact the strength of the Hadley cells. At least in boreal winter, intermodel variability in the strength of Northern Hemisphere stationary eddy momentum flux divergence can also impact the strength of the overturning (Caballero 2008). Grid size and its effect on the eddy momentum flux convergence only explain a portion of the structural differences in the Hadley cells between models, although they certainly explain a substantial fraction of the variation in the Hadley cells’ width.
7. Summary and discussion

We examined the basic climatology of the width of the tropical belt in the CMIP5 and CCMVal-2 historical scenarios. The metrics employed in this study measured the tropical belt width based on the latitudes of the edges of the Hadley cells, the latitudes of the subtropical jet cores, and the latitudes of the tropopause breaks. These objective metrics are by construction not susceptible to individual model biases and do not alias any structural changes into widening trends.

We found that the models generally overestimated the mean width of the tropical belt as measured by the latitudes of the subtropical jet cores when compared to COSMIC GPS-RO observations. Additionally, the models on average underestimated the seasonal cycle amplitude of the tropical belt width based on the latitudes of the subtropical jet cores and the tropopause breaks compared to reanalyses and observations. We also found evidence that the models have difficulty reproducing the observed 1-month lag in the Northern Hemisphere between the Hadley cell edge latitude and the latitudes of the jet cores and tropopause breaks (Davis and Birner 2013), with only a small subset of models accurately reproducing the phasing of the seasonal cycle of the different metric edge latitudes.

Tropical surface air temperature was found to have little correlation with the modeled Hadley cell width, despite the fact that changes in tropical surface air temperature are expected to be associated with changes in the Hadley cell width (Frierson et al. 2007; Adam et al. 2014). Instead, we found that climate model grid size statistically explains 15%–50% of the total inter-model variation in the mean tropical belt width, with finer horizontal resolution leading to a narrower tropical belt. Using an approximation that was found to accurately estimate the modeled eddy fluxes, we examined how the eddy momentum and Eliassen–Palm fluxes change with CCMVal-2 model grid size and model-mean Hadley cell width. Both finer-resolution and narrower model-mean Hadley cells were associated with enhanced equatorward wave propagation in the lower and upper troposphere and enhanced, equatorward-shifted eddy momentum flux convergence and divergence in the subtropical and tropical upper troposphere, respectively. By considering the balance between the Coriolis torque and the eddy momentum flux convergence, we have shown that these grid-size-biased eddy momentum fluxes are associated with an equatorward contraction of the MMS in the subtropics. The resulting effect on the Hadley cell edge latitudes was found to explain about half of the total intermodel variation in the edge latitudes of the Northern and Southern Hemisphere Hadley cells and explained essentially all of the grid-size bias in the Hadley cell width.

The ultimate source of the grid-size bias in the eddies is not fully clear. It seems plausible that grid size may impact the ability of a model to resolve the small-scale structures associated with wave generation and breaking. However, no causality was examined here, as a balance condition was invoked to study the interaction between the eddies and the Hadley cells. While a grid-size bias in the eddies is a plausible mechanism for producing the grid size bias in the tropical belt width, this does not preclude a mechanism by which the eddies are responding to a grid-size effect in, for example, the zonal wind.

The impact of grid size on the zonal momentum itself is more complicated than its impact on the eddies or the meridional circulation (Fig. 10a). With finer horizontal resolution, the zonal winds become more easterly in the deep tropics and more westerly in the midlatitudes, with a robust easterly response in the tropical surface zonal wind. This is the signature of an enhanced transfer of momentum out of the tropics and into the midlatitudes by eddies, consistent with our analysis of the eddy momentum flux convergence (Fig. 7c). In contrast, the changes to the zonal wind seen with narrower Hadley cells (Fig. 10b) indicate a strengthening of westerly zonal winds in the subtropics and a weakening of westerly zonal winds in the midlatitudes. The strengthening of winds is likely due to enhanced advection of momentum by the mean meridional wind, which intensifies as the Hadley cell narrows (Fig. 8d).
The surface wind response is similar to the grid-size response with tropical easterlies and midlatitude westerlies, although the pattern is shifted equatorward and is less robust. Despite the fact that the zonal wind response in the troposphere is of opposite sign, both finer-resolution and narrower Hadley cells are associated with enhanced vertical shear and subtropical relative vorticity throughout the subtropical troposphere. These are consistent with a narrower Hadley cell based on the Hadley cell scaling theory for a variable Rossby number (Kang and Lu 2012).

The large intermodel variation in the tropical belt width has important implications for intermodel comparisons. Multimodel means of these fields will not be representative of the actual models’ circulation structures as the subtropical jet core and Hadley cell edge vary by more than 5° latitude in each hemisphere across the models examined here. Further, given the impacts of the Hadley cells on precipitation and surface aridity, it is clear that attempts to quantify future climate impacts on a regional scale may be obscured by the intermodel variation in the tropical belt edge latitudes. Multimodel analyses that make use of relative coordinates (Scheff and Frierson 2012) will avoid such statistical problems.

While we explored intermodel variations in the mean state of the tropical belt width, we did not ignore the question of how a model’s mean state may impact its tropical widening trend (Fig. 11). None of the correlation coefficients between the trend and the mean tropical belt width are significant, neither for the CMIP5 and CCMVal-2 models together nor for the two model ensembles separately. In simulations of climate change, the midlatitude jets tend to shift more poleward the more equatorward they are in the twentieth century climatology (Kidston and Gerber 2010), but it appears that the eddy-mean flow processes biasing the poleward shifts of the midlatitude circulation do not similarly bias the trends in the tropical belt width, despite their interannual coupling (Kang and Polvani 2011).

We found that stronger eddy momentum flux divergence is associated with a climatologically narrower Hadley circulation. While the presence of eddies may widen the Hadley cells in comparison to axisymmetric circulations (Kim and Lee 2001), this does not necessarily dictate whether stronger or weaker eddy momentum flux divergence out of the cells will widen or narrow them. Hadley cell scaling theories based on a variable Rossby number would suggest that stronger eddy momentum flux divergence should permit wider Hadley cells by shifting poleward the latitude of the onset of baroclinic instability (Kang and Lu 2012). However, theories for the width of the Hadley cells based on angular-momentum-conservation thinking are by nature unable to account for the role of eddies in setting the width of the Hadley cells. Our results indicate that, on climatological time scales, the geography of the eddy fluxes may be more important than their strength.

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APPENDIX A

COSMIC Wet Profile Bias

We use COSMIC dry rather than wet retrievals because of an undocumented bias in lower-stratospheric temperatures, detailed below.

Figures A1a and A1b display the wet and dry COSMIC temperature retrievals, respectively, for January 2012. The troposphere and especially the tropics are unrealistically colder in the dry profile data compared to the wet profile data, reflecting the need to account for water vapor in the lower atmosphere. However, in the extratropical upper troposphere and the global stratosphere, the temperatures are qualitatively similar, with a deep tropical cold point at 100 hPa and a polar vortex poleward of 70°N.

For direct comparison with COSMIC temperatures, we subsample 6-hourly model-level ERA-I output to COSMIC observations. For each COSMIC observation in latitude–longitude–height space that falls within ±3 h of an ERA-I time step, the eight nearest ERA-I grid points in latitude–longitude–height space forming a box around the COSMIC observation are used to linearly interpolate temperature and pressure to the exact location of the COSMIC observation. Essentially, COSMIC-like temperature and pressure profiles are constructed from ERA-I model-level output. These COSMIC-like profiles are then gridded in the same procedure as COSMIC.

Figures A1c and A1d display the difference in temperature between the subsampled ERA-I output for January 2012 and the wet and dry gridded temperatures, respectively (i.e., $T_{ERA-I} - T_{COSMIC}$), while Figs. A1e and A1f display the difference between the COSMIC wet and dry temperatures and the full ERA-I gridded product, respectively. As expected, the dry COSMIC temperatures are much colder than both the subsampled and full ERA-I temperatures in the lower and middle troposphere; however the differences in the upper troposphere and lower stratosphere are generally less than 1 K.

In contrast, the wet COSMIC temperatures exhibit a very large warm bias of over 8 K in the Northern Hemisphere subtropics and midlatitudes at and above the tropopause. This bias aligns closely with the edge of the polar vortex where there is a strong meridional temperature gradient and temperatures remain constant with height. Further, the bias is consistent between years and also appears, albeit weaker, in the Southern
Hemisphere stratosphere at the edge of the polar vortex in austral winter (not shown).

As the bias is present when compared to both the full and subsampled ERA-I, we conclude that it does not arise from sampling or gridding error. Further, as the bias is present in a region where there is very little water vapor, we believe that the bias originates in the 1D-Var scheme used to generate the moisture-corrected profiles. To avoid this bias, we employ the dry profiles for calculating tropopause variables and estimating upper-tropospheric winds.

APPENDIX B

Estimation of Eddy Fluxes

As noted in section 6, CMIP5 daily output is limited to select vertical levels, but a subset of 9 CCMVal-2 models provide eddy fluxes as monthly mean model output. This section details the calculation of eddy fluxes for 15 CCMVal-2 models and their validation using the subset of 9 models providing eddy fluxes as model output. Models not included in this analysis that participated in the REF-B1 simulation did not supply the fields necessary to estimate the Eliassen–Palm flux and its divergence.

The zonal-mean zonal momentum equation in Cartesian coordinates is given by

$$\frac{\partial [u]}{\partial t} + [\omega] \frac{\partial [u]}{\partial p} - [v]([\zeta] + f) = -\frac{\partial [u'u']}{\partial y} - \frac{\partial [u'\omega']}{\partial p} + [F_x],$$

(B1)

where $[u]$, $[v]$, and $[\omega]$ are the zonal-mean zonal, meridional, and vertical components of the wind, $[\zeta] = -\partial_x [u]$ is the zonal-mean relative vorticity, $[u'u']$ and $[u'\omega']$ are the meridional and vertical eddy momentum fluxes, and $[F_x]$ is the zonal mean of
nonconservative forces, such as friction. Brackets indicate zonal means, and primed quantities denote departures from the zonal mean.

In the time mean and away from the surface, the horizontal eddy momentum flux convergence is proportional to the Coriolis torque and the advection of zonal momentum:

$$2 \frac{\partial}{\partial y} \left[ u_0 y_0 \right] \frac{\partial y'}{\partial y} \left[ y (f_1 [z]) \right] \frac{\partial}{\partial p}, \quad \text{(B2)}$$

where we have retained the vertical advection of zonal momentum to ensure an accurate calculation. It should be noted that we cannot calculate the vertical eddy momentum flux [the second-to-last term in Eq. (B1)] from CCMVal-2 model output, and it may be non-negligible in the subtropical midtroposphere, where there are strongly sloped isentropes and strong, zonally asymmetric subsidence. The eddy momentum flux convergence can be estimated by calculating the terms on the right-hand side of Eq. (B2) from the model-mean, time-mean fields from the CCMVal-2 models.

Similarly, in the time-mean and away from the surface, the Eliassen–Palm flux divergence is proportional to the Coriolis torque of the transformed Eulerian-mean meridional wind and the advection of zonal momentum by the transformed Eulerian-mean wind:

$$\nabla \cdot F = -\frac{\partial F_y}{\partial y} + \frac{\partial F_p}{\partial p} = -\frac{\partial [u']}{\partial y} + \frac{\partial (f \psi)}{\partial p}, \quad \text{(B3)}$$

where $[u']$ and $[\omega^*]$ are the transformed Eulerian-mean meridional and vertical winds (Andrews and McIntyre 1976). All 15 CCMVal-2 REF-B1 models examined here output the transformed Eulerian-mean winds. Thus, the Eliassen–Palm flux divergence can be estimated by calculating the terms on the right-hand side of Eq. (B3) from the model-mean, time-mean fields from the CCMVal-2 models.

Under quasigeostrophic scaling, the Eliassen–Palm flux divergence is given by

$$\nabla \cdot F = \frac{\partial F_y}{\partial y} + \frac{\partial F_p}{\partial p} = -\frac{\partial [u']}{\partial y} + \frac{\partial (f \psi)}{\partial p}.$$

where $F_y$ and $F_p$ are the meridional and vertical components of the Eliassen–Palm flux, $\psi$ is the potential temperature, and $[\theta_p]$ is the static stability. The eddy heat flux convergence, or the vertical component of the Eliassen–Palm flux divergence, can simply be estimated as the difference between the Eliassen–Palm flux divergence and the eddy momentum flux convergence.

The meridional component of the Eliassen–Palm flux is calculated by integrating the eddy momentum flux convergence meridionally with the boundary condition that the estimated horizontal component equals the multimodel-mean horizontal component at the equator from the subset of nine models. Similarly, the vertical component of the Eliassen–Palm flux is calculated by integrating the eddy heat flux convergence with the boundary condition that the estimated vertical
component equals the multimodel-mean vertical component at 500 hPa. The components of the Eliassen–Palm flux are used only for illustrative purposes in Fig. 7 and do not factor into our calculations of the impact of the eddy grid-size bias on the Hadley cell width.

A comparison between the estimated and actual Eliassen–Palm flux and its divergence and the eddy momentum flux convergence for the nine CCMVal-2 models is shown in Fig. B1. The Eliassen–Palm flux vectors are scaled as in Edmon et al. (1980), where the meridional and vertical components of the Eliassen–Palm flux are multiplied by the distance in each plot occupied by 1 rad of latitude and 1 Pa of pressure, respectively. The vector magnitudes are equivalent between Figs. B1a, B1b, and B1c but, as in Edmon et al. (1980), are scaled to an arbitrary maximum magnitude. The approximations produce reasonable results compared with the modeled values, although the magnitude of the Eliassen–Palm flux divergence in midlatitudes is underestimated. Midlatitude wave generation at the surface is poorly captured by the estimation method.

The estimated and actual eddy momentum flux convergence do differ lower in the troposphere, which likely reflects the need to account for other terms in the zonal momentum equation that we were not able to estimate from CCMVal-2 output, such as the vertical eddy momentum flux convergence. Nevertheless, the method is reliable over much of the troposphere, with a pattern correlation of 0.86 ($R^2 = 0.74$) between the actual and estimated eddy momentum flux convergences.

These approximations are extended to all REF-B1 models to calculate their model-mean eddy fluxes and Eliassen–Palm flux divergences, which are used to explore the tropical belt grid-size bias in section 6.

REFERENCES


