On the Discrepancies in Tropical Belt Expansion between Reanalyses and Climate Models and among Tropical Belt Width Metrics

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ABSTRACT

The arid subtropics are situated at the edges of the tropical belt, where subsidence in the Hadley cells suppresses precipitation. Any meridional shift in these edge latitudes could have significant impacts on surface climate. Recent studies have investigated past and future changes in the tropical belt width and have found discrepancies in the rates of expansion estimated with different metrics and between climate models and reanalyses. Here, CMIP5 simulations and four modern reanalyses are analyzed using an ensemble of objective tropical belt width metrics to reexamine if such inconsistencies exist. The authors do not find sufficient evidence to demonstrate this discrepancy between models and reanalyses, as reanalysis trends in the tropical belt width fall within the range of model trends for any given metric. Furthermore, only metrics based on the Hadley cells are found to exhibit robust historical and future expansion. Metrics based on the subtropical jet and the tropopause show no robust response. This differentiation may be due to the strong correlation, on all time scales, between the Hadley cell edge latitudes and the latitudes of the eddy-driven jets, which consistently shift poleward in response to radiative forcings. In contrast, the subtropical jet and tropopause metrics appear to be decoupled from the Hadley cells and the eddy-driven jets and essentially measure a different tropical belt. The tropical belt width metrics are inconsistently correlated with surface climate indices based on precipitation and surface evaporation. This may make assessing the surface impacts of observed and future tropical expansion challenging.

1. Introduction

Differential solar heating of Earth’s surface drives the Hadley cells—two tropical meridional overturning circulations that together cover half of the planet’s surface. The cells are the dominant mode of poleward (Keith 1995; Trenberth and Stepaniak 2003) and cross-equatorial (Heaviside and Czaja 2013) heat transport in the tropical atmosphere. Equatorward surface flow in the Hadley cells converges moisture into the tropics where air rises and the latent heat due to condensation increases its dry potential temperature. The upper branches of the circulation transport this warm air poleward until it subsides above Earth’s arid subtropics, suppressing precipitation by stabilizing and drying the atmosphere. The latitudes of these dry climates at the edges of the Hadley cells are referred to as the edges of Earth’s tropical belt.

These edges are not fixed but instead vary seasonally, interannually, and in response to climate forcings. They shift poleward in the summer and equatorward in the winter (Davis and Birner 2013), potentially driven by concurrent changes in insolation (Lindzen and Hou 1988) and transient (Bordoni and Schneider 2008) and stationary (Shaw 2014) eddy fluxes. Eddy fluxes also drive variability in the Hadley cell width, including a contraction of the Hadley cells in response to the warm phase of El Niño–Southern Oscillation (Lu et al. 2008; Stachnik and Schumacher 2011; Davis and Birner 2013; Nguyen et al. 2013). Many recent studies have reported an expansion of the tropical belt in observations, reanalyses, and model simulations, which could have significant impacts on surface climate (Seidel et al. 2008; Lucas et al. 2014; Birner et al. 2014).

Tropical expansion in reanalyses and models has been characterized using the edge latitudes of the Hadley cells (Hu and Fu 2007; Frierson et al. 2007; Lu et al. 2008; Johanson and Fu 2009; Stachnik and Schumacher 2011; Davis and Rosenlof 2012; Allen et al. 2012; Nguyen et al. 2013; Chen et al. 2014; Nguyen et al. 2015), with estimates of expansion in reanalyses ranging from 0.3° (Stachnik and Schumacher 2011) to 1.5° decade−1 (Davis and...
Rosenlof 2012). However, there may be spurious trends in the tropical belt width in reanalyses (Lucas et al. 2012; Quan et al. 2014) owing to inhomogeneities in the observing system (Bengtsson et al. 2004).

The Hadley cell is only one aspect of the tropical belt, and other characteristics can be used to estimate tropical expansion. The structure of the tropopause, which has an abrupt drop in altitude in the subtropics, can be used to characterize the edges of the tropical belt (Seidel and Randel 2006; Lu et al. 2009; Birner 2010; Wilcox et al. 2012; Davis and Rosenlof 2012; Lucas et al. 2012; Davis and Birner 2013; Ao and Hajj 2013). Tropopause metrics are advantageous because they can be estimated from radiosonde observations or from any remotely sensed temperature observations with sufficient vertical resolution, such as global navigation satellite system radio occultation profiles. Estimates of historical tropical expansion using tropopause metrics range from $-0.5^\circ$ (Davis and Birner 2013) to $3.1^\circ$ decade$^{-1}$ (Seidel and Randel 2006).

Another aspect used to characterize the edges of the tropical belt is the subtropical jet stream (Archer and Caldeira 2008; Davis and Rosenlof 2012; Allen et al. 2012). This can be inferred with some difficulty from observational data (Fu and Lin 2011; Davis and Birner 2013). Trends in the tropical belt width based on the subtropical jets have a smaller range than other metrics, at $0.1^\circ$ (Davis and Rosenlof 2012) to $0.6^\circ$ decade$^{-1}$ (Davis and Birner 2013).

The wide range of trends among metrics within the same dataset and between datasets poses a problem (Seidel et al. 2008; Staten et al. 2016; Davis et al. 2016). While they are in proximity to each other in the subtropics, it is not clear if or how the latitude of the subtropical jet, the tropopause break, the Hadley cell edge, and various surface metrics relate to one another climatologically, interannually, or in response to climate forcings. Given that the seasonality of the metrics and their response to El Niño–Southern Oscillation differ (Davis and Birner 2013), their temporal variability and basic response to radiative forcings may differ as well.

Another unresolved issue is that tropical expansion tends to be an order of magnitude weaker in climate models than in reanalyses over the historical period (Johanson and Fu 2009; Quan et al. 2014). Even trends assessed from climate model simulations with the most extreme greenhouse gas forcings are weaker than those assessed from reanalyses (Johanson and Fu 2009; Hu et al. 2013). This could be related to the relative insensitivity of the models’ Hadley cells to changes in global-mean surface temperature (Adam et al. 2014; Nguyen et al. 2015). Additionally, modes of natural variability such as the Pacific decadal oscillation (Grassi et al. 2012; Allen et al. 2014), the southern annular mode, and El Niño–Southern Oscillation (Lucas and Nguyen 2015) may have enhanced historical tropical expansion. Model simulations simply may not be able to reproduce the high rates of estimated historical expansion unless they are forced with the observed history of sea surface temperatures (Allen et al. 2014; Garfinkel et al. 2015).

When asking whether tropical expansion is different between reanalyses and models, it is important to define a null hypothesis that takes into account their fundamental differences. Reanalyses more or less reproduce observed natural variability by assimilating observations. Climate model historical experiments instead simulate an ensemble of possible histories of natural variability given the forcings on the system, such as insolation and greenhouse gas concentrations. If natural variability impacts tropical expansion trends then the models will produce a range of possible trends in the tropical belt width. If there are no relevant model biases that lead to biases in the trends then the real-world evolution of the tropical belt width should fall somewhere within this distribution (Quan et al. 2014; Garfinkel et al. 2015).

The question that assumes the least about model and reanalysis deficiencies is whether trends based on reanalyses fall within the distribution of model trends. If the reanalyses’ trends fall outside of the range of model trends, then we may have a basis to claim that the reanalyses’ and models’ trends disagree. If they do not, then the level of evidence necessary to declare such a discrepancy will not have been met. This study will assess tropical expansion in a large sample of climate models and in only the most modern reanalyses. While we do not directly quantify internal variability as in Quan et al. (2014) or Garfinkel et al. (2015), we instead take a holistic view and ask whether the statistics alone demonstrate a discrepancy between models and reanalyses. Formally stated, our null hypothesis is that the reanalyses’ trends are not significantly different from all model trends using a given tropical belt width metric.

This intercomparison focuses on a representative subset of metrics. In addition to exploring the trends, it will also examine the intermodel and temporal relationships among the metrics in climate models and reanalyses. These results will be used to interpret any differences in tropical expansion between the different metrics. We will investigate the connection between these different tropical belt metrics and surface climate indices to determine which, if any, are most useful for climate impact studies.

2. Data

We use 25 models from phase 5 of the Coupled Model Intercomparison Project (CMIP5) historical and representative concentration pathway 8.5 (RCP8.5)

We also briefly use 17 models from the chemistry climate model validation activity for stratosphere–troposphere processes and their role in climate (SPARC-2) (CCMVal-2) reference B1 (REF-B1) experiment: CAM3.5, Centre for Climate System Research/National Institute of Environmental Studies (CCSRNIES), CMAM, CNRM–ARPEGE–Climat coupled MOCAGE (CNRM-ACM), ECHAM4-L39(DLR)/CHEM/ATTILA (E39CA), ECHAM/Messy Atmospheric Chemistry model (EMAC), EMAC/FUBRad model (EMAC-FUB), GEOS-CCM, Meteorological Research Institute (MRI), National Institute of Water and Atmospheric Research Solar–Climate–Ozone Links (Niwa-SOCOL), SOCOL, Università degli Studi dell’Aquila (ULAQ), Unified Model with Eulerian Transport and Atmospheric Chemistry (UMETRAC), Unified Model-SLIMCAT (USSLIMCAT), Unified Model/U. K. Chemistry Aerosol Community Model–Met Office (UMUKCA-METO), UMUKCA–University of Cambridge (UMUKCA-UCAM), and WACCM. The CCMVal-2 REF-B1 experiment is analogous to the CMIP5 historical experiment as it simulates Earth’s past climate using observed forcings (Eyring et al. 2010). CMAM is the only CCMVal-2 experiment that is fully interactive chemistry.

Monthly mean output from four modern reanalyses are used in this study: the European Centre for Medium-Range Weather Forecasts’ interim reanalysis (ERA-Interim) (Dee et al. 2011), the National Aeronautics and Space Administration’s Modern-Era Retrospective Analysis for Research and Applications 2 (MERRA2) (Bosilovich et al. 2015), the Japanese Meteorological Agency’s Japanese 55-year Reanalysis (JRA-55) (Kobayashi et al. 2015), and the National Centers for Environmental Prediction’s Climate Forecast System Reanalysis (CFSR) (Saha et al. 2010a).

Our analysis of the historical time period using the reanalyses, CMIP5 historical experiment, and CCMVal-2 REF-B1 experiment spans 1979–2005, while our analysis of the future climate projections in the CMIP5 RCP8.5 experiment spans 2006–2100.

3. Methods

We use five objective metrics that measure different aspects of the subtropical circulation and climate to characterize the edge latitudes of the tropical belt. These metrics are in many cases inspired by other metrics in the existing literature. However, none of the metrics examined here are evaluated with subjectively chosen numerical thresholds or on arbitrary vertical levels. Such objective metrics are especially critical when examining trends (Birner 2010; Davis and Rosenlof 2012). For example, changes in the mean structure of the atmosphere, such as an increase in the depth of the troposphere, could alias into expansion trends if metrics are evaluated on specific vertical levels or with numerical thresholds. The metrics examined here define the tropical belt edges based on the latitudes of the subtropical tropopause breaks $\Delta \theta$, the latitudes of the subtropical jets $U_{\text{max}}$, the latitudes of the edges of the Hadley cells $\int \Psi dp$, the latitudes of maximum subsidence in the Hadley cells $\delta_{\text{max}}$, the latitudes of zero zonal-mean surface zonal wind $U_{\text{500}}$. We also employ the commonly used Hadley cell metric based on the 500-hPa mean meridional streamfunction $\Psi_{500}$ to compare some results with previous work.

a. Hadley cell edge latitudes: $\int \Psi dp$

The Hadley cells are defined by the mean meridional streamfunction, which measures the overturning circulation in the zonal mean (Fig. 1, shading in top panel). The Hadley cell edge latitudes are often measured by the latitudes of the zero contour of the 500-hPa mean meridional streamfunction (Frierson et al. 2007; Lu et al. 2008; Davis and Rosenlof 2012; Allen et al. 2012; Hu et al. 2013; Tao et al. 2016). However, evaluating the streamfunction on arbitrary vertical levels could alias vertical shifts of the circulation into poleward shifts, as the zero contour varies in latitude throughout the troposphere (Fig. 1). Such metrics also implicitly assume that Hadley cell expansion is homogeneous in the vertical, whereas model simulations indicate it is not (see Fig. 2 of Tao et al. 2016). While the Hadley cell edge can similarly be estimated using the latitude of the zero contour of a midtropospheric layer average of the mean meridional streamfunction (Hu and Fu 2007; Johanson and Fu 2009; Stachnik and Schumacher 2011; Nguyen et al. 2013, 2015), it could still be subject to the same problems.

To avoid these issues, the $\int \Psi dp$ metric estimates the Hadley cell edge latitudes at the latitude where the vertically averaged mean meridional streamfunction is
The mean meridional streamfunction $C(p, f)$ is defined as follows:

$$C(p, f) = \frac{2\pi a \cos(f)}{g} \int_{p}^{0} [v] dp,$$

where $p$ is the pressure, $f$ is the latitude, $a = 6,371$ km is the mean radius of Earth, $g = 9.81$ m s$^{-1}$ is the acceleration due to gravity, $v$ is the meridional wind, and square brackets indicate the zonal mean. The mean meridional streamfunction is then averaged in pressure from 1000 hPa to the top of the atmosphere. This quantity measures the average meridional overturning circulation strength at a given latitude, and the latitude at which it is zero indicates the average latitude of the poleward edge of the Hadley cell. As the stratosphere represents approximately 10% of the mass of the tropical atmosphere and the streamfunction values in the stratosphere are orders of magnitude smaller than those in the troposphere, the contribution by the stratospheric circulation is negligible. Linear interpolation is used to estimate the edge latitude between grid points.

**b. Subtropical jet latitudes: $U_{\text{max}}$**

The subtropical jets are located in the subtropical troposphere where the vertically integrated meridional temperature gradient, or thermal wind shear, is largest. The $U_{\text{max}}$ metric estimates the subtropical jet latitudes as the latitudes of maximum upper-tropospheric/lower-stratospheric zonal-mean zonal wind in the subtropics between 1000 and 50 hPa above the tropopause in each hemisphere (Fig. 1) (Davis and Birner 2013). Limiting the search to 50 hPa above the tropopause prevents an erroneous identification of the stratospheric jets. The 1000-hPa wind speed is subtracted from each column so that the zonal wind at a given level represents the vertically integrated thermal wind shear (Davis and Birner 2016). This makes the subtropical jet distinct in all seasons from the eddy-driven jet, which is characterized by strong surface westerly winds. Linear interpolation of the meridional gradient of the zonal-mean zonal wind is used to identify the jet latitude between grid points.
c. Tropopause break latitudes: $\Delta \theta$

The tropopause abruptly drops from its tropical to extratropical altitude in the region of maximum baroclinicity near the subtropical jet. Here, the $\Delta \theta$ metric estimates the latitudes of the tropopause breaks as the latitudes of maximum zonal-mean tropospheric dry bulk stability in each hemisphere (Fig. 1) (Davis and Birner 2013). The tropospheric dry bulk stability is defined as the difference in potential temperature between the tropopause and the surface, which exhibits an unambiguous maximum value in the subtropics at the tropopause break. Because this field is discontinuous at the tropopause break, no interpolation method can be used to estimate the tropical belt edge latitude between grid points. This metric itself enters into scaling theories for the width of the Hadley cell (e.g., Held 2000) and is most similar to tropopause metrics based on the meridional gradient of tropopause height (Davis and Rosenlof 2012).

d. Latitudes of maximum downwelling: $\partial_y \int \Psi dp$

The Hadley cell edge is typically measured as the latitude where the meridional overturning circulation transitions from the Hadley cell to the Ferrel cell. However, the location of maximum subsidence within the Hadley cell may be more relevant for surface climate as it suppresses convection and dries the atmosphere, contributing to the formation of deserts. As shown in Fig. 1, the minimum of precipitation minus evaporation does not occur at the edge of but within the subsidence of the Hadley cells, where the meridional gradient of the streamfunction (the vertical velocity) is the largest. The $\partial_y \int \Psi dp$ metric estimates the latitudes of maximum vertically averaged subsidence in the Hadley cells in each hemisphere (Fig. 1). Linear interpolation of the second meridional gradient of the vertically averaged mean meridional streamfunction is used to estimate the edge latitude between grid points.

e. Latitudes of zero surface zonal wind: $U_{sfc}$

In the time and zonal mean and assuming linear surface drag, the vertically averaged zonal-mean quasi-geostrophic zonal momentum equation reduces to a balance between the vertically averaged eddy momentum flux convergence and surface drag on the zonal-mean surface zonal wind:

$$\frac{[u]}{\tau} = -\left\langle \frac{\partial}{\partial y} [u^* v^\theta] \right\rangle \tag{2}$$

where $u$ and $v$ are the zonal and meridional winds, $u_s$ is the surface zonal wind, $\tau$ is a time scale for drag, $[u^* v^\theta]$ is the eddy momentum flux where stars indicate deviations from the zonal mean, and angled brackets indicate the vertical average. Any latitude with nonzero zonal-mean zonal wind at the surface has a net convergence or divergence of momentum in the column by the eddies. Net eddy momentum flux divergence out of the tropics balances the drag on the surface easterlies while net eddy momentum flux convergence into the midlatitudes balances the drag on the surface westerlies. The latitude with zero zonal-mean surface zonal wind divides these two regimes. The $U_{sfc}$ metric estimates the latitudes of zero zonal-mean surface zonal wind as the first latitudes where the zonal-mean zonal wind is zero at 1000 hPa, poleward of the tropical easterlies in each hemisphere (Fig. 1). Linear interpolation is used to estimate the edge latitude between grid points. Modeling groups may extrapolate winds to 1000 hPa differently, but we do not have the information necessary to quantify or correct any errors this may introduce. The results are insensitive to using the 10-m wind field, at least in reanalyses.

This method is similar to the mean sea level pressure metric used by Choi et al. (2014), which measures the edge of the tropical belt as the location where the meridional derivative of sea level pressure within the subtropical ridge is zero. In the zonal mean this corresponds to the latitude where the zonal-mean sea level zonal wind changes sign from westerlies to easterlies.

f. Latitudes of the 500-hPa Hadley cell edge: $\Psi_{500}$

We also employ the commonly used $\Psi_{500}$ metric for continuity with past literature. The $\Psi_{500}$ metric estimates the tropical belt edge latitudes as the latitudes where the 500-hPa mean meridional streamfunction is zero, poleward of its tropical extrema in each hemisphere.

g. Eddy-driven jet metric

Eddy fluxes associated with Rossby waves connect the eddy-driven jets to the Hadley cells and significantly impact the mean meridional circulation (Kim and Lee 2001; Schneider 2006). Rossby waves propagating away from their source leave behind westerly zonal wind anomalies that reinforce the eddy-driven jet. When they propagate into and break within the Hadley cells they force easterly zonal wind anomalies, modifying the strength and position of the subtropical jets and the Hadley cell edges. The direct balance condition between the eddy momentum flux convergence and the mean meridional wind in the zonal-mean quasigeostrophic zonal momentum equation implies that the Hadley cell edge could be sensitive to shifts in the eddy-driven jet and its associated eddy momentum fluxes.

The eddy-driven jet latitude (EDJ) is defined as the latitude of maximum wind at 850 hPa in each hemisphere (Fig. 1) (Kang and Polvani 2011). Using the 1000-hPa wind speed or a lower-tropospheric average produces...
indistinguishable results. A more objective metric might define the jet as the location of maximum vertically averaged eddy momentum flux convergence. However, this is difficult to obtain with the available data for the model simulations examined here. Linear interpolation of the gradient of the 850-hPa zonal-mean zonal wind is used to estimate the EDJ latitude between grid points.

h. Surface climate indices: \( \min(P - E), P - E = 0, \) and area of \( P - E < 0 \)

Three indices are used to understand the impact of variations in the tropical belt edge latitudes on surface climate: the latitudes of the minimum in precipitation minus evaporation \( \left[ \min(P - E) \right] \) (Zhou et al. 2011; Hu et al. 2011; Allen et al. 2012), the latitudes where precipitation minus evaporation is zero \( \left( P - E = 0 \right) \) (Davis and Rosenlof 2012; Allen et al. 2012), and the surface area of negative precipitation minus evaporation in the subtropics \( \left( \text{area of } P - E < 0 \right) \). The \( \min(P - E) \) latitudes are defined as the latitudes of the minimum of precipitation minus evaporation \( P - E \) in each hemisphere (Fig. 1). These latitudes reflect the location of the maximum in subtropical aridity and could be associated with the latitudes of maximum subsidence in the Hadley cells \( \delta_i \Psi dp \). The \( P - E = 0 \) latitudes are defined as the latitudes where \( P - E \) is zero poleward of the \( \min(P - E) \) latitudes (Fig. 1). As the transition between the arid subtropics and the rainy midlatitudes, this metric could be most correlated with the \( \left[ \Psi dp \right] \) metric edge latitudes, which divide the Hadley and Ferrel cells. A new metric, the area of \( P - E < 0 \) in the subtropics, is calculated as the area of negative \( P - E \) in the subtropics in each hemisphere weighted by the cosine of latitude (Fig. 1) and may be interpreted as the surface area prone to arid or desert climates.

i. Calculation details

The tropical belt width is defined as the difference in degrees latitude between the tropical belt edge latitudes. All correlations and linear least squares regressions are performed on monthly mean deseasonalized tropical belt edge latitudes and widths. We deseasonalize by removing the climatological-mean monthly value from each monthly mean (e.g., the mean of all Januaries is subtracted from each January), where the climatological mean is taken over the historical (1979–2005) or future projection (2006–2100) time periods as appropriate.

Trends are calculated based on linear least squares regressions. The significance of trends is assessed using each time series’ effective degrees of freedom based on each time series’ lag-1 autocorrelation (Santer et al. 2008), on average yielding 1 degree of freedom per 2 to 3 months. Trends are considered significantly different from zero if their 95% confidence intervals do not include zero. Using annual-mean data does not impact the trend values, but it is disadvantageous in that it reduces the number of degrees of freedom.

j. Reanalysis-mean time series

Reanalyses are atmospheric forecast models coupled to data assimilation systems that ingest satellite and in situ observations to produce a dynamically consistent estimate of the historical evolution of the atmosphere. Because they are in principle simulating the same evolution of Earth’s atmosphere, averaging their time series provides a potentially more robust estimate of the historical tropical belt width and edge latitudes. Because of the nonlinearity of linear least squares regression, the trends based on the reanalysis-mean time series are not guaranteed to be equal to the mean of the reanalyses’ trends. We will always examine the individual reanalyses’ results in conjunction with the reanalysis mean.

4. Results

We begin with a basic examination of the time series of tropical belt widths based on the five objective metrics (Fig. 2). The \( \Delta \theta \), the \( U_{\text{max}} \), and especially the \( U_{\text{sk}} \) metric widths are well constrained in reanalyses (Figs. 2a,c,j). On the other hand, the streamfunction-based metric widths disagree and are less constrained (Figs. 2e,g). The intensity of the overturning similarly varies across reanalyses (Stachnik and Schumacher 2011). While the \( U_{\text{max}} \) and \( \delta_i \Psi dp \) metric widths contract in response to the 1997/98 El Niño in the reanalyses, the \( \Delta \theta \) metric widths contract in response to the Mount Pinatubo eruption in both models and reanalyses. The inconsistency in the responses of the tropical belt width to these particular events is an indication that different metrics do not necessarily measure the “same” tropical belt.

A majority of the metrics have greater natural variability in the CMIP5 historical simulations than in the reanalyses. This could be caused by a model deficiency in the representation of the seasonal cycle of the tropical belt edge latitudes. CMIP5 models do not produce a consistent seasonal cycle, especially in the Northern Hemisphere where the tropical belt edge latitudes shift poleward by 10° to 15° latitude over one month in the early summer (Davis and Birner 2016). Minor year-to-year differences in the phasing of the seasonal cycle may act as a source of some of this enhanced variability. This could be why, for example, the \( \left[ \Psi dp \right] \) metric width (Fig. 2f) has near-annual periodicity in its tropical belt edge latitude anomalies.

a. Temporal and intermodel covariability

A direct assessment of the coupling between metrics is the temporal correlation, or covariability, between their
tropical belt edge latitudes within individual model simulations and reanalyses. The latitudes of the Hadley cell edge $\Delta \theta$, the latitudes of maximum downwelling $\partial_y \Psi dp$, and the latitudes of zero zonal-mean surface zonal wind $U_{sfc}$ are all correlated with each other in both hemispheres in the reanalyses (Figs. 3 and 4), with mean correlation coefficients ranging from 0.8 to 0.9. On the other hand, the subtropical jet $U_{max}$ and tropopause break $D_u$ latitudes are weakly correlated with only each other with correlation coefficients of 0.3. As a further point of distinction between the two sets of metrics, the latitude of the eddy-driven jet is correlated with the streamfunction and surface wind metric edge latitudes and uncorrelated with the subtropical jet and tropopause break metric edge latitudes in both hemispheres (Figs. 3 and 4).

The CMIP5 historical simulations have weaker correlations among the metrics than the reanalyses, but most of the same relationships emerge in the Southern Hemisphere. There are some models with correlations between the $\Psi dp$, $\partial_y \Psi dp$, and $U_{sfc}$ metric edge latitudes nearly as high as in the reanalyses. At the same time, some model correlations between these metric edge latitudes are nearly zero. There is even a spurious anticorrelation between the $U_{max}$ metric edge latitudes and the $\Psi dp$ and EDJ metric latitudes (Fig. 3). Correlations among all the metrics are poor in the Northern Hemisphere in the models, consistent with results from the CCMVal-2 REF-B1 simulations (not shown). Especially concerning is the lack of correlation between the Hadley cell edge and surface wind metrics and between the subtropical jet and tropopause break metrics. The cause of these deficiencies in simulated circulation variability over the Northern Hemisphere is unclear.

To provide a different perspective, we also examine the intermodel (or across model) correlations between the mean tropical belt edge latitudes measured by the different metrics (Fig. 5). If two metrics’ edge latitudes scale across models, they may be impacted by similar physical processes that not only set their mean edge latitudes but also contribute to any tropical expansion trends. We show the mean edge latitudes, rather than...
the correlation coefficients alone, to emphasize their large range among the models. The only correlation that is significant in both hemispheres is between the mean latitude of the Hadley cell edge $\Delta \phi$ and the mean latitude of maximum downwelling $\partial_y \Psi dp$ (Fig. 5h). Dynamical fields such as the distribution of eddy momentum fluxes consistently scale with these measures of the tropical belt width, as well (Davis and Birner 2016).

The $U_{sfc}$ metric edge latitudes also scale with the $\Psi dp$ and $\partial_y \Psi dp$ metric edge latitudes but only in the Southern Hemisphere (Figs. 5i,j), perhaps reflecting the models’ poor correlations among metrics in the Northern Hemisphere (Fig. 4).

There is no relationship between the climatological $\Delta \theta$ and $U_{max}$ metric edge latitudes in either hemisphere, suggesting a lack of covariability between these two edge latitude metrics in both hemispheres in the models. Surprisingly, in the Northern Hemisphere the $U_{max}$ metric edge latitudes scale with the $\partial_y \Psi dp$ metric edge latitudes. While they may be situated in the subtropics and their mean latitudes may scale relative to each other across models, their month-to-month variability is uncorrelated (Fig. 4).

The results so far illustrate distinct properties of two sets of tropical belt metrics. Metrics based on the streamfunction $\Psi dp$ and $\partial_y \Psi dp$ and the latitudes of zero zonal-mean surface zonal wind $U_{sfc}$ are temporally coupled with each other and the latitudes of the eddy-driven jet. On the other hand, metrics based on the subtropical jet $U_{max}$ and tropopause break $D_u$ latitudes are weakly correlated with each other and not at all with the eddy-driven jet. There are no correlations across these classes of metrics within a given realization of Earth’s atmosphere, and only the mean tropical belt edge latitudes based on the Hadley cell itself scale robustly across models in both hemispheres.

b. Historical trends in the tropical belt width

Having assessed the covariability and intermodel scaling among the metrics, in this section we seek to answer 1) whether the metrics respond differently to observed and projected forcings on the climate system and 2) whether reanalysis trends in the width of the tropical belt disagree with model trends. There is a well-documented seasonality to tropical expansion in both hemispheres, with expansion generally enhanced in each hemisphere’s
respective summer and autumn seasons (Hu and Fu 2007; Hu et al. 2013; Allen et al. 2012; Tao et al. 2016). Here we focus primarily on the total change in the tropical belt width and do not examine its seasonality. The trends are more robust and the results that follow are similar when individual seasons and hemispheres are examined.

Trends in the tropical belt width in reanalyses assessed from the streamfunction and surface wind metrics are generally between 0.08 and 0.68 decade\(^{-1}\) (Fig. 6), similar to those found by Stachnik and Schumacher (2011), Allen et al. (2012), Nguyen et al. (2013), and Davis and Birner (2013), though there are some outliers. Only ERA-Interim’s \(\Delta \Psi\) metric expansion trend is as large as those found in Hu and Fu (2007) and Johanson and Fu (2009). The small range of reanalysis \(U_{\text{max}}\) metric width trends, from \(-0.1^\circ\) to 0.7\(^\circ\) decade\(^{-1}\), agree well with the jet-based tropical belt width trends found by Archer and Caldeira (2008), Davis and Rosenlof (2012), Allen et al. (2012), and Davis and Birner (2013). For the \(\Delta \Psi\) metric width the trends range from \(-0.4^\circ\) to 1.2\(^\circ\) decade\(^{-1}\) with a reanalysis-mean trend of 0.6\(^\circ\) decade\(^{-1}\), generally lower than tropopause-based metrics that use numerical thresholds (Seidel and Randel 2006; Davis and Rosenlof 2012; Lucas et al. 2014) and more similar to objective tropopause-based metrics (Birner 2010; Davis and Rosenlof 2012). These objective metrics measure different aspects of the structure of the tropopause but arrive at similar historical trends.

The large spread in tropical expansion for the \(\Delta \theta\) and \(\partial_y \Psi\) metrics among the reanalyses could be interpreted as a consequence of assimilation system changes, though it may also include some genuine uncertainty regarding the historical changes in the tropical belt width. At least in the case of the \(\partial_y \Psi\) metric, a substantial fraction of the disagreement among the reanalyses may come from the inability to constrain the mean meridional streamfunction (Fig. 2g). In spite of this, the \(\Psi\) metric expansion trends have the smallest range of any metric in the reanalyses. None of the reanalysis trends for each of the \(U_{\text{max}}, \Psi_{500}, \partial_y \Psi,\) and \(U_{\text{sfnc}}\) metric tropical belt widths are significantly different from each other at the 95% confidence level. Further, none of the reanalysis-mean tropical belt width trends are significantly different from each other at the 95% confidence level. However, the reanalysis-mean trends should not be interpreted as representative of the observed trends. With only four reanalyses, it is unlikely that all individual reanalysis biases have been averaged out. The reanalysis mean may instead be more appropriate for determining which individual reanalysis trends are significant outliers. For example, MERRA2’s and ERA-Interim’s \(\partial_y \Psi\) trends could be considered outliers because they are significantly different than the reanalysis-mean trend. Conversely, it may be possible to say that the reanalyses’ \(\Psi\) metric tropical expansion trends could be assigned some additional confidence because none are significantly different from the reanalysis-mean \(\Psi\) metric expansion trend. However, without an estimate of historical tropical
expansion derived from observations, it may not be possible to further conclude whether the reanalysis trends actually reflect the observed trends.

The multimodel-mean trends in the tropical belt width average out much of the natural variability in the trends, which manifests differently in different model simulations. The actual historical evolution of the tropical belt width contains natural variability and is therefore hard to discern from the multimodel mean. The multimodel-mean trends are small, ranging between 0.08 decade$^{-1}$ for the $U_{\text{max}}$ metric width and 0.28 decade$^{-1}$ for the $U_{\text{sfc}}$ metric width. For the subtropical jet and tropopause break metrics, the multimodel-mean trends are not statistically significant, while for the streamfunction and surface wind metrics, the multimodel-mean trends are significant and range from 0.18 to 0.28 decade$^{-1}$. This is similar to the 0.1° to 0.3° decade$^{-1}$ Hadley-cell-based tropical widening trends found by Hu et al. (2013) for the same data and time period.

It is worth considering the range of model trends as well, as multimodel-mean trends describe only the forced response. There is a 1.0°–1.5° decade$^{-1}$ spread in the individual model trends for any given metric, and for each metric there are CMIP5 simulations with positive and negative trends. However, all of the significant trends based on the streamfunction and surface wind metrics indicate expansion, while there are both significant expansion and contraction trends based on the subtropical jet and tropopause break metrics. The $\Psi_{500}$ metric expansion trends are significant at 44%. This drops to 12% for the objective version of this metric ($\partial \Psi dp$), and no trends based on the latitudes of maximum downwelling ($\partial_y \Psi dp$) indicate tropical expansion. The multimodel-mean trends and the ranges of trends for the streamfunction and surface wind metrics are similar to those that Johanson and Fu (2009) found using the $\Psi_{500}$ metric in CMIP3 historical simulations, despite modeling advances between CMIP3 and CMIP5.

We now examine whether the reanalyses’ trends in the tropical belt width fall outside the range of model trends. A reanalysis trend could be said to fall outside of the range of model trends if it is significantly different from every model trend for a given metric. An alternative and perhaps simpler formulation may be to ask whether the reanalysis trends were drawn from a distribution with

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**Fig. 5.** The intermodel correlation between the mean tropical belt edge latitudes as measured by each metric in the CMIP5 historical simulations, for the Northern (red) and Southern (blue) Hemispheres. The percentage of variance explained $R^2$ in each hemisphere is shown above each plot in red and blue, respectively, for correlations significant at the 95% confidence level. Correlations that are not significant are shown in gray, with the corresponding edge latitude colors lightly faded. Note the change in axes for the $\Delta \theta$ and $\partial_y \Psi dp$ panels.
the mean given by the multimodel mean and the standard deviation given by the standard deviation of model trend estimates. However, since the distribution of trends from which the models draw is unknown, we take into account the uncertainties in the individual model trend estimates.

We assess the significance of the differences by performing two-sided Student’s t tests for the difference of means for each reanalysis and model trend combination for a given tropical belt width metric. Note that one cannot typically visually compare 95% confidence intervals to determine whether two trends are statistically significantly different. For two trend estimates, their difference is not statistically significant at the 95% confidence level if either of the trend estimates themselves fall within the other’s 95% confidence intervals [cf. Eq. (2) of Lanzante (2005)]. However, in any other case, the difference may or may not be statistically significantly different.

In some cases it appears that the reanalysis trends are larger than the model trends. However, for every reanalysis trend using a given metric, at least one model trend exists that is not statistically significantly different from the reanalysis trend. That is, for any given metric none of the reanalysis trends are statistically significantly different from every CMIP5 historical simulation’s trend.

As an illustrative example, consider CFSR’s Ψ500 trend, the highest reanalysis trend estimate using the Ψ500 metric. Two climate model trend estimates fall within its 95% confidence interval, and nearly half of the model trend estimates are not significantly different from its trend estimate at the 95% confidence level. In this specific case, we conclude that CFSR’s trend does not fall outside of the range of model trends because it is not statistically significantly different from every model trend.

We find that none of the reanalysis trends fall outside the range of model trends and thus cannot reject our null hypothesis. As a result, we do not find sufficient evidence to claim there is a discrepancy in tropical expansion estimated in climate models and reanalyses.

FIG. 6. Trends in the tropical belt width for the 1979–2005 time period. Bars indicate the mean trend from the CMIP5 historical simulation (light red) and the reanalysis-mean time series (white). For the CMIP5 historical simulation, whiskers indicate the maximum and minimum 95% confidence interval bounds among CMIP5 models and the 95% confidence intervals for the multimodel-mean trend. Individual model estimates are indicated by black dots. Individual reanalysis estimates are shown for ERA-Interim (purple), MERRA2 (blue), JRA-55 (green), and CFSR (yellow). For the reanalysis-mean time series and individual reanalyses, whiskers indicate 95% confidence intervals. The percentage of CMIP5 model trends that are significantly (top) positive and (bottom) negative is shown at the bottom of each panel.
over the historical period. One could argue that there is still a distinction between tropical expansion in the reanalyses and the models: few models have multiple significant trends estimated by different metrics (only five models have significant trends in at least two metrics), while the majority of trends within the reanalyses are significant. However, this could be due to the enhanced variability in the models’ tropical belt widths (Fig. 2), which increases the significance thresholds for a majority of the metrics.

c. Projected trends in the tropical belt width

The multimodel-mean trends in the CMIP5 RCP8.5 experiment (Fig. 7) are virtually the same as they are in the historical experiment. As in the historical experiment, all of the streamfunction and surface wind metric trends that are significant indicate tropical expansion. For the $\Psi_{500}$ and $\int \Psi \, dp$ metric widths, all models predict statistically significant tropical expansion, ranging between 0.1° and 0.5° decade$^{-1}$. The $\Delta \Psi$ metric indicates significant expansion for only 68% of the models, with rates of expansion ranging between $-0.1°$ and 0.4° decade$^{-1}$. Based on the $U_{slc}$ metric 76% of the models have significant tropical expansion, with rates of expansion ranging between 0.0° and 0.3° decade$^{-1}$.

The subtropical jet and tropopause metrics have no robust forced response, with approximately equal numbers of models exhibiting tropical belt expansion and contraction. The trends in the tropical belt width range between $-0.7°$ and 1.0° decade$^{-1}$ for the $\Delta \theta$ metric and between $-0.2°$ and 0.2° decade$^{-1}$ for the $U_{max}$ metric, with no significant multimodel-mean trends.

While the range of CMIP5 model trends for each metric generally overlap, this does not mean that models with stronger expansion in one metric necessarily have stronger expansion in another metric. There is a robust scaling in tropical expansion trends among the streamfunction and surface wind metrics in the RCP8.5 simulations (Figs. 8h–j) that is stronger in the Southern Hemisphere than in the Northern Hemisphere. In the Northern Hemisphere the $U_{slc}$ metric edge latitude trends unexpectedly scale with all other tropical belt edge latitude metric trends. There is also an unexpected correlation between $\Delta \theta$ and $\int \Psi \, dp$ metric expansion in the Northern Hemisphere only. These metrics have no covariability (Figs. 3 and 4) and do not scale in the mean (Fig. 5). In the models there are spurious correlations between some metrics and a lack of correlation between metrics that are correlated in the reanalyses (Figs. 3 and 4). Whether this potential misrepresentation of the Northern Hemisphere zonal-mean circulation results in these unexpected scalings is unclear, and we hesitate to comment further without a deeper analysis.

d. Relation to the eddy-driven jet

Why do the streamfunction and surface wind metrics exhibit robust tropical expansion in response to radiative forcings while the tropopause and subtropical jet metrics do not? There is a well-documented relationship between interannual variations in the eddy-driven jet latitude and the edge latitude of the Hadley cell in the Southern Hemisphere (Kang and Polvani 2011; Staten and Reichler 2014). While several mechanisms could govern a poleward shift of the eddy-driven jet in response to radiative forcings (Chen and Held 2007; Kidston et al. 2011), they all result in a poleward shift of the region of wave breaking and eddy momentum flux divergence in the subtropics, which would tend to drag the Hadley cell edge poleward (neglecting any changes to its internal dynamics).

Accordingly, there is a statistically significant correlation between tropical expansion measured by the latitudes of the Hadley cell edge $\int \Psi \, dp$, the maximum downwelling $\partial_z \int \Psi \, dp$, and zero zonal-mean surface zonal wind $U_{slc}$ as well as the shift in the latitude of the
eddy-driven jet in the Southern Hemisphere in the RCP 8.5 experiment (Figs. 9c–f). Note that these correlations and regressions use only models with statistically significant trends in both the eddy-driven jet latitude and tropical belt edge latitude, though using all trends does not substantially change these relationships. There is no significant correlation between either the subtropical jet or tropopause break edge latitude trends and the trend in the latitude of the eddy-driven jet. The trends in the eddy-driven jet latitude are also not significantly correlated with any of the meridional circulation metric trends in the Northern Hemisphere (not shown), owing to both the low number of significant trends and the poor temporal coupling between the two in the Northern Hemisphere in the models (Fig. 4). Additionally, the eddy-driven jets in the Northern Hemisphere tend to have basin-specific variability and change (Barnes and Polvani 2013). The resulting impacts on the zonal-mean eddy momentum flux divergence may not be as linear as they are in the Southern Hemisphere.

The scaling between the trend in the latitude of the eddy-driven jet and tropical expansion in the Southern Hemisphere is strongest in December–January–February (Fig. 10). Between 50% and 87% of the total intermodel variation in tropical expansion in the streamfunction and surface wind metrics can be explained by the poleward shift of the latitude of the eddy-driven jet. There remain no significant correlations between the shift in the latitude of the eddy-driven jet and tropical expansion as measured by the subtropical jet or tropopause metrics (Figs. 10a,b).

The regression coefficients based on trends calculated for the full time series indicate a ratio of a 0.2° to 0.4° decade⁻¹ trend in the Hadley cell edge for every 1.0° decade⁻¹ trend in the eddy-driven jet latitude, slightly less than the interannual ratios reported in Kang and Polvani (2011) and Staten and Reichler (2014) and similar to the trend ratios in the Southern Hemisphere in austral winter reported by Staten and Reichler (2014). In austral summer, the regression coefficients double, confirming that the Hadley cell is more strongly connected to shifts in the latitude of the eddy-driven jet when the baroclinicity is low (Staten and Reichler 2014).

e. Connection to surface climate

To conclude our exploration of the different metrics we perform a simple analysis of how they relate to three zonal-mean indices of surface climate in the reanalyses: the latitudes of the minimum of precipitation minus evaporation \([\min(P - E)]\), the latitudes where precipitation minus evaporation is zero \((P - E = 0)\), and the total surface area of the arid subtropics \((\text{area of } P - E < 0)\).

On monthly time scales, there is no substantial correlation between any of the surface climate indices and

![Figure 8](https://example.com/fig8.png)
the tropical belt edge latitudes in either hemisphere. Soil moisture, vegetation, and other factors that communicate circulation and precipitation variability to evaporation rates may operate on seasonal to annual time scales, so we extend these correlations to non-deseasonalized, annual-mean data (Fig. 11).

Each surface climate metric is correlated with slightly different sets of tropical belt edge latitude metrics. In the Southern Hemisphere, the latitude where \( P - E = 0 \) is not significantly correlated with any metric edge latitudes, while in the Northern Hemisphere it is correlated with only the EDJ latitude. In both hemispheres, the streamfunction metrics are significantly correlated with the area of \( P - E < 0 \) and the \( \text{min}(P - E) \) latitudes. The strongest correlations are in the Northern Hemisphere, where the \( \Psi dp \) metric edge latitudes and the \( \text{min}(P - E) \) latitudes and area of \( P - E < 0 \) are correlated with mean \( R^2 \) values of 0.3 and peak values of 0.5. This highlights the potential impact of variations in the Hadley cell width on surface climate on time scales as short as a year. The \( \Delta \theta \) metric latitudes are correlated with the area of \( P - E < 0 \) in both hemispheres and the \( \text{min}(P - E) \) latitudes in the Southern Hemisphere only, and the \( U_{\text{max}} \) metric has no significant correlation with any surface climate metric.

While there are some broad differences in the surface climate connections between the two sets of metrics, there is a substantial range in the values of the correlation coefficients among reanalyses. No particular reanalysis stands out with consistently weaker or larger correlation coefficients.

5. Summary and discussion

Numerous metrics have been used to study tropical expansion. There are two open questions: 1) Why do different metrics measure different rates of historical and future tropical expansion? 2) Why do rates of tropical expansion in reanalyses appear to be greater than those measured in climate models (Seidel et al. 2008; Staten et al. 2016; Davis et al. 2016)?
We explored these questions with five objective tropical belt width metrics. Metrics that measure the latitudes of the Hadley cell edges, the latitudes of maximum downwelling in the Hadley cells, and the latitudes of zero zonal-mean surface zonal wind are temporally correlated within simulations, scale across simulations in the mean and in their rates of tropical expansion, and exhibit robust historical and future tropical expansion. On the other hand, metrics that measure the latitudes of the tropopause breaks and subtropical jets are weakly coupled and have no detectable forced response to radiative forcings, even in the most extreme radiative forcing scenario in CMIP5. There are few correlations in any sense across these two sets of metrics.

To understand these differences, consider the zonal-mean thermal wind equation:

\[
\frac{\partial u}{\partial p} = \frac{1}{f} \frac{\partial}{\partial y} \left( \frac{-T}{R} \right),
\]

which holds approximately—away from the surface and the equator—and follows from geostrophic balance in the zonal-mean meridional momentum equation. The term \(T\) is the temperature and \(R\) is the gas constant for dry air. The subtropical jet metric quantifies an aspect of the zonal-mean zonal circulation, which is related to zonal-mean temperature. The tropopause metric also quantifies an aspect of zonal-mean temperature, which by Eq. (3) is related to the zonal circulation as well. In this sense both metrics are “zonal circulation” metrics as their connection occurs through momentum balance.

On the other hand, the zonal-mean meridional circulation is by definition ageostrophic and is connected to the eddy momentum fluxes and the zonal-mean zonal wind at the surface through the zonal-mean zonal momentum equation:

\[
-f[v] = -\frac{\partial}{\partial y} [u^*\nu^*] - \begin{cases} 0 & \text{if } p < p_s \\ \frac{[U_{sfc}]}{\tau} & \text{if } p = p_s \end{cases},
\]

where \(p_s\) is the surface pressure. This equation provides a relation between the streamfunction metrics, which are based on \([u]\), and the zonal-mean surface
zonal wind metric. Both are constrained by the eddy momentum fluxes. Taking the vertical integral as in Eq. (2) shows that the drag on the zonal-mean surface zonal wind balances the vertically integrated eddy momentum flux convergence. In the free atmosphere the eddy momentum flux convergence is balanced by the Coriolis torque exerted by the mean meridional wind. The zonal-mean surface zonal wind at the surface is unlike the zonal-mean zonal wind everywhere else in the atmosphere because it is not proportional to the meridional gradient of temperature. It follows that metrics that measure the tropical belt edge latitudes based on the mean meridional streamfunction or the zonal-mean surface zonal wind are physically linked to the distribution of eddy momentum fluxes. The latitude of zero zonal-mean surface zonal wind shares characteristics of and is conceptually linked to the meridional circulation and the Hadley cell edges. Zonal circulation metrics based on the tropopause and subtropical jet are instead physically linked to the distribution of temperature. While it is true that eddy momentum fluxes also modify the zonal circulation and could physically couple the zonal and meridional circulation metrics, this coupling does not emerge in our analysis.

One could alternatively segregate the metrics in terms of their height in the atmosphere. The subtropical jet and tropopause break are both located at the top of the troposphere. On the other hand, the zonal-mean surface zonal wind could be cast as a “surface” metric. The streamfunction is by definition a vertically integrated measure, so it is not easily classified by a single level in the atmosphere, though the maximum occurs near 500 hPa and could be classified as “midtropical.” However, the subtropical jet is a product of the vertically integrated meridional temperature gradient. Similarly, the zonal-mean surface zonal wind, while technically located at the surface, is related to the vertical integral of the eddy momentum flux convergence. The tropopause height is linked to surface temperature, average tropospheric lapse rates, and the stratospheric circulation (Thuburn and Craig 1997; Juckes 2000), even on regional scales (Wu and Shaw 2016). Segregating metrics by their height may neglect the physical processes important for their covariability and response to forcings.

The poleward shift of the Southern Hemisphere eddy-driven jet is a consistent model response to historical climate forcings (Chen and Held 2007), to idealized greenhouse gas forcings (Kushner et al. 2001), and to the suite of forcings in simulations of future climate models.
climate (Yin 2005; Miller et al. 2006; Kidston and Gerber 2010; Swart and Fyfe 2012; Barnes and Polvani 2013; and confirmed here in Figs. 9 and 10). We hypothesize that its coupling on all time scales with the meridional circulation metric edge latitudes through subtropical eddy momentum flux divergence may result in their robust tropical expansion in historical and future climate simulations.

Conversely, we hypothesize that the zonal circulation metrics may not have robust tropical expansion because they do not have such direct theoretical and statistical relationships with the eddy momentum flux divergence or the eddy-driven jet. The subtropical jets shift upward through a thermal wind response to lower-stratospheric cooling and upper-tropospheric warming produced by increasing greenhouse gas concentrations (Shepherd and McLandress 2011). Whether they shift poleward or equatorward may be sensitive to the structure of those temperature anomalies and not a change in eddy-driven jet dynamics.

The CMIP5 historical experiments simulate an envelope of possible tropical expansion trends given the external constraints on the climate system. We find there is insufficient evidence to conclude that there is a discrepancy between tropical expansion trends in reanalyses and models, as all of the reanalysis trends fall within the range of model trends for any given tropical belt width metric.

The model spread is not only due to natural variability but also differences in parameterizations. Reanalyses’ trends are impacted by data assimilation methods. Our analysis did not take such effects into account and only demonstrates that the level of evidence necessary to declare a discrepancy has not yet been met. Differences in model and reanalysis configurations and natural variability need to be disentangled to reach a more conclusive result.

While Johanson and Fu (2009) concluded that there was a discrepancy between the tropical expansion trends in models and in reanalyses and observations, they did not consider the uncertainties in the trend estimates. In the case of model trend distributions, the trend estimates do not form a “known” distribution but instead have their own uncertainties that should be considered. Johanson and Fu (2009) also used an older set of reanalyses, which may further contribute to the apparent contradiction between their results and ours.

While there is a large range of tropical expansion rates in the reanalyses, for a majority of tropical belt width metrics they are not statistically significantly different. In other words, there is scant evidence to demonstrate a discrepancy among the reanalyses.

The multimodel-mean trends, while small, illustrate the forced component of tropical expansion, while the reanalysis trends indicate one particular realization impacted by natural variability. Given that the reanalysis trends tend toward being greater than the multimodel mean, the observed widening of the tropical belt over the past 30 years may be the result of a forced signal amplified by natural variability. However, simulations that are forced with observed sea surface temperatures do not have constrained tropical belt width climatologies (e.g., Figs. 5 and 6 of Davis and Birner 2016). Further, the range of tropical belt width trends in the CCMVal-2 REF-B1 experiment, which is forced by observed sea surface temperatures, is as large as the range of tropical belt width trends in the CMIP5 historical experiment (Fig. 12). A naïve expectation may be that forcing models with observed sea surface temperatures would lead to a convergence in their tropical expansion trends. Differences in climate feedbacks among models (Feldl and Bordoni 2016) and both internal atmospheric (Garfinkel et al. 2015) and coupled ocean–atmosphere variability (Kang et al. 2013) can produce significant intermodel variations in tropical expansion. The physical processes driving such internal variability in the tropical belt width are unclear.

Regarding surface climate, the Hadley cells’ strongest impact appears to be on the most arid regions where \( P - E \) is at a minimum and on the areal extent of the arid regions themselves, with little impact on regions situated near or within the Ferrel cell where \( P - E = 0 \). Precipitation declines at the poleward flanks of the subtropics, where precipitation minus evaporation is nearly zero, may be associated with the poleward shift of the midlatitude storm tracks (Scheff and Frierson 2012). There is a weak association between the eddy-driven jet latitudes and the latitudes of \( P - E = 0 \) on annual time scales in the Northern Hemisphere (Fig. 11b) but not in the Southern Hemisphere (Fig. 11a). However, the ratio of precipitation to potential evapotranspiration has recently been used to study global changes in arid climates (Scheff and Frierson 2012; Feng and Fu 2013). This quantity may be better for assessing the connections between circulation variability and surface climate as it measures the precipitation deficit against the evaporative demand of the atmosphere (Feng and Fu 2013).

The future of Earth’s subtropical circulations are unclear, even if the forced signal is known. The lack of intermodel scaling in tropical expansion between the meridional and zonal circulation metrics suggests that in any given individual realization of Earth’s future climate the Hadley cells, subtropical jets, and surface circulations may shift in latitude at different rates and potentially in different directions. Distinct realizations may even exhibit
different structural changes to the circulation. This presents a challenge for assessing historical tropical expansion from observations. While the metrics based on the Hadley cells are consistent with each other, in situ radiosonde measurements have insufficient sampling to accurately construct an estimate of the zonal-mean meridional winds comprising the Hadley cells (Waliser et al. 1999). Observations can be used to assess tropical expansion based on the tropopause and the subtropical jets, but these metrics have little connection in any sense to metrics based on the Hadley cells. However, the metric based on the zonal-mean surface zonal wind is well constrained in reanalyses, is generally representative of the Hadley cell edge latitude, and can be obtained from observations, similar to the sea level pressure metric used by Choi et al. (2014).

At the most basic level, tropical expansion can only be understood by examining more than just the mean meridional streamfunction. Each metric holds unique information about variability and change in the subtropical circulation, and assessing their relationships can yield a more complete story about the circulation response to radiative forcings.

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FIG. 12. As in Fig. 6, but for the CMIP5 historical and CCMVal-2 REF-B1 experiment tropical belt width trends.
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