Surface Superrotation

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ABSTRACT

Equatorial superrotation is commonly observed in simulations of Earth and planetary climates, but it is almost without exception found to occur only at upper levels, with zero or easterly winds at the surface. Surface superrotation—a state with climatological zonal-mean westerlies at the equatorial surface—would lead to a major reorganization of the tropical ocean circulation with important consequences for global climate. Here, we examine the mechanisms that give rise to surface superrotation. We identify four theoretical scenarios under which surface superrotation may be achieved. Using an axisymmetric model forced by prescribed zonal-mean torques, we provide concrete examples of surface superrotation under all four scenarios. We also find that we can induce surface superrotation in a full-complexity atmospheric general circulation model, albeit in an extreme parameter range (in particular, convective momentum transport is artificially increased by almost an order of magnitude). We conclude that a transition to surface superrotation is unlikely in Earthlike climates, including ancient or future warm climates, though this conclusion is subject to the currently large uncertainties in the parameterization of convective momentum transport.

1. Introduction

Atmospheric superrotation—a state in which zonal-mean winds have greater angular momentum than the equatorial value under solid-body rotation—is an observed feature of several planets and moons in the solar system, including Venus and Titan (Rossow et al. 1990; Bird et al. 2005). Superrotation readily arises in atmospheric models configured to simulate those planets (Del Genio and Zhou 1996; Yamamoto and Takahashi 2003; Hollingsworth et al. 2007; Schneider and Liu 2009; Lebonnois et al. 2010; Parish et al. 2011; Lebonnois et al. 2016; Tokano et al. 1999; Mitchell et al. 2011; Lebonnois et al. 2012a). Superrotation also occurs in simulations of tidally locked exoplanets (Showman and Polvani 2011; Kaspi and Showman 2015). Earth's troposphere does not superrotate (Lee 1999; Dima et al. 2005), but models show a spontaneous transition to superrotation in the upper troposphere when tropical surface temperatures become warm enough to match climate proxy reconstructions from warm periods in the deep past (Caballero and Huber 2010; Arnold et al. 2013).

It has long been understood that superrotation cannot arise in an axisymmetric atmosphere with purely diffusive viscosity (Hide 1969). Some mechanism for upgradient angular momentum transfer must exist to drive superrotation, the obvious candidate being zonal-mean momentum transport by nonaxisymmetric eddies (Gierasch 1975; Read 1986). In a superrotating atmosphere, we expect the climatological angular momentum maximum to occur at the equator, since an off-equatorial maximum would be inertially unstable and difficult to sustain (in fact, we provide an explicit realization of this expectation in section 3c below). Therefore, we generally expect a superrotating atmosphere to feature net climatological eddy momentum convergence onto the equator, balancing zonal-mean drag due to viscosity or momentum export by the mean meridional circulation.

An extensive literature has developed examining the eddy dynamics responsible for superrotation, spanning a hierarchy of model complexities including shallow-water models (Scott and Polvani 2008; Showman and Polvani 2010; Suhas et al. 2016), two-level models (Suarez and Duffy 1992; Saravanan 1993), dry

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three-dimensional models (Williams 2003; Kraucunas and Hartmann 2005; Schneider and Liu 2009; Liu and Schneider 2011; Mitchell and Vallis 2010; Potter et al. 2014; Wang and Mitchell 2014; Dias Pinto and Mitchell 2014; Laraia and Schneider 2015; Polichtchouk and Cho 2016; Dias Pinto and Mitchell 2016; Lutsko 2018), and full-complexity general circulation models (GCMs; Hoskins et al. 1999; Caballero and Huber 2010; Arnold et al. 2013; Carlson and Caballero 2016). The mechanisms responsible for generating the eddies driving superrotation fall into two broad categories. One is diabatic: equatorial heating anomalies—due to disorganized convection or to convection organized by a fixed surface temperature anomaly or by a traveling perturbation such as the Madden–Julian oscillation (MJO)—drive upper-level mass divergence, and the resulting Rossby wave source (Sardeshmukh and Hoskins 1988) generates off-equatorial waves that converge momentum onto the equator (Held 1999; Laraia and Schneider 2015). The other is adiabatic, relying on mechanisms such as baroclinic, barotropic, or Kelvin–Rossby instability to generate the waves (Suarez and Duffy 1992; Polichtchouk and Cho 2016; Wang and Mitchell 2014).¹ Horizontal momentum transport by off-equatorial Rossby waves plays a key role in both categories. Vertical momentum transport by equatorial Kelvin and gravity waves can also be important; this is true in the quasi-biennial oscillation on Earth, which is associated with transient stratospheric superrotation, and may also be relevant to the superrotation of Venus (Hou and Farrell 1987).

As noted above, some models predict that superrotation may have occurred on Earth during past warm periods and could therefore potentially recur in a future, warmer climate. The impact of superrotation on surface climate will remain muted, however, so long as the superrotation is confined to upper levels. Some authors (Pierrehumbert 2000; Tziperman and Farrell 2009) have pointed out that if superrotation were to extend to the surface, the results would be much more dramatic. Surface superrotation would lead to a major reorganization of the tropical ocean circulation and could account for the “permanent El Niño” state hypothesized to have occurred during the early Pliocene (Wara et al. 2005).

Curiously, however, surface superrotation does not arise in any of the three-dimensional simulations reported in the papers cited above, either in Earth or

planetary contexts. In papers where the eddy momentum convergence is shown, the equatorial maximum is always concentrated at upper levels—roughly coinciding with the Hadley cell outflow—and drives upper-level superrotation with zero or easterly equatorial surface wind. The one exception we are aware of (Schneider and Liu 2009; Liu and Schneider 2010, 2011) shows robust equatorial westerlies down to the surface; this work is also unusual in showing a reversed Hadley cell, with subsidence on the equator (Liu and Schneider 2011).

So why is it generally so difficult for superrotation to reach the surface? Under what conditions do we expect it to do so? In particular, is surface superrotation a possibility under warm Earthlike conditions and thus relevant for ancient or future climates? The aim of this paper is to address these questions. In section 2, we discuss the equatorial momentum balance in general terms, identifying four potential pathways to surface superrotation. In section 3, we test these hypothetical pathways in a simplified context—an axisymmetric model with an imposed zonal momentum forcing—and find that they do in fact all lead to surface superrotation under appropriate conditions. Section 4 considers the prospects for surface superrotation in a full-complexity atmospheric GCM configured to represent an Earthlike aquaplanet with warm surface temperatures. As in previous work (Caballero and Huber 2010; Arnold et al. 2013), these simulations feature an enhanced MJO-like mode that drives upper-tropospheric superrotation, and we study the momentum fluxes associated with this MJO-like mode in some detail. By manipulating the equator–pole temperature contrast and tropical viscosity, we find we can induce surface superrotation albeit in an extreme parameter regime. In section 5, we discuss our findings in a broader context and summarize our conclusions.

2. Equatorial momentum balance

a. Pathways to surface superrotation

The zonal-mean zonal momentum balance in pressure coordinates can be written as

$$\partial_t \bar{u} = (f - \partial_y \bar{u})\bar{v} - \overline{\omega \partial_p \bar{u}} - \partial_y \overline{u'v'} - \partial_p \overline{u'\omega'} - g \partial_p \bar{\tau}, \quad (1)$$

where the first two terms on the rhs represent momentum redistribution by the mean meridional circulation (MMC), the second two terms represent momentum redistribution by eddies, and the last term represents vertical viscosity, expressed as the convergence of the zonal-mean stress $\bar{\tau}$ (Pa) due to vertical transport of zonal momentum by small-scale eddies and convection. Throughout, overbars represent the zonal mean,

¹ The two-level models studied by Suarez and Duffy (1992) and Saravanan (1993) in fact fall into both categories. These models show multiple equilibria; in one equilibrium, superrotation is maintained by a diabatically driven stationary wave, while in the other, it is maintained by transient eddies presumably of baroclinic origin.

primes a deviation therefrom, and other symbols have their standard meaning.

Integrating vertically and assuming steady state, we have

$$-\langle \bar{v} \partial_y \bar{u} \rangle - \langle \bar{\omega} \partial_p \bar{u} \rangle - \partial_y \langle \bar{u}' \bar{v}' \rangle - \bar{\tau}_s = 0, \quad (2)$$

where $\langle \cdot \rangle = g^{-1} \int_0^{p_s} (\cdot) dp$, p_s is surface pressure, and $\bar{\tau}_s$ is the zonal-mean zonal surface stress, which can be related to surface zonal wind $\bar{u}_s$ by the linearized bulk aerodynamic formula $\bar{\tau}_s = \gamma \bar{u}_s$, where γ is an exchange coefficient (defined positive; note that since we are using pressure coordinates, the momentum flux is positive downward). For simplicity, we have assumed the surface to be axisymmetric so that there is no form drag contribution.

The balance in (2) is most often invoked to explain the midlatitude surface westerlies: near the center of the Ferrel cell, where both relative vorticity and $\bar{\omega}$ are small, column-integrated horizontal momentum convergence by baroclinic eddies must be balanced predominantly by surface stress, requiring positive surface winds.

At the equator, the situation is different since $\bar{\omega}$ is generally not small. Let us disregard for the moment the relative vorticity term $-\langle \bar{v} \partial_y \bar{u} \rangle$. Given positive eddy momentum convergence confined to the equatorial upper troposphere, the most natural response is for the zonal-mean wind to develop westerly shear. Assuming a conventional Hadley cell with updraft on the equator, $-\langle \bar{\omega} \partial_p \bar{u} \rangle$ will be negative: mean upward advection can balance eddy momentum convergence (Shell and Held 2004; Kraucunas and Hartmann 2005; Showman and Polvani 2010) with zero or even negative surface winds. In essence, the MMC evacuates momentum from the equatorial strip at the same rate as the eddies converge it, with no need to involve surface stress in the momentum balance. This suggests a possible way to generate equatorial surface westerlies: if the internal viscous stress $\bar{\tau}$ is strong enough to vertically homogenize the zonal wind, then $-\langle \bar{\omega} \partial_p \bar{u} \rangle$ will be zero and surface westerlies will develop to balance eddy momentum convergence. In this picture, momentum is injected into the equatorial strip at upper levels, is rapidly transported down to the surface by viscosity, and is dissipated by surface stress.

A second possible pathway to surface superrotation is for eddy momentum convergence to be concentrated in the lower troposphere, in immediate contact with the surface. In this case, we expect easterly vertical shear and positive $-\langle \bar{\omega} \partial_p \bar{u} \rangle$, so the budget in (2) can only be balanced by positive surface stress (again disregarding the relative vorticity term). A third, more exotic possibility is to have a reversed Hadley cell, with subsidence on the equator. If eddy momentum convergence is

concentrated in the upper troposphere, $\bar{\omega}$ will advect westerly shear downward, $-\langle \bar{\omega} \partial_p \bar{u} \rangle$ will be positive, and (2) will again require positive surface stress. In both these cases, viscosity plays no essential role.

The role of the relative vorticity term $-\langle \bar{v} \partial_y \bar{u} \rangle$ is more delicate. Under equinoctial conditions, $\bar{v}$ is antisymmetric about the equator and must cross zero there and must do so smoothly since a discontinuity in v would result in infinite vertical velocity. At the same time, $\bar{u}$ is symmetric and therefore continuous at the equator; its first derivative (i.e., the relative vorticity) will be antisymmetric, but there is no strong a priori constraint on its smoothness. Given sufficient horizontal vorticity mixing, the relative vorticity will cross zero smoothly at the equator, and $-\langle \bar{v} \partial_y \bar{u} \rangle$ will be exactly zero at the equator and negligible in its vicinity. However, if vorticity mixing is very small, the relative vorticity can become discontinuous at the equator (in which case $\bar{u}$ takes the form of a sharp cusp-like jet centered at the equator). Under such conditions, relative vorticity remains finite all the way to the equator, and $-\bar{v} \partial_y \bar{u}$ can play a substantial role in the momentum budget in the immediate vicinity of the equator, affecting the equatorial value of $\bar{u}$ by continuity (by which we mean mathematical rather than mass continuity). We find that this situation does in fact arise in axisymmetric simulations presented in section 3.

The arguments in the preceding paragraph suggest that the $-\bar{v} \partial_y \bar{u}$ term can be important but say nothing about its sign. In thinking about this issue, it is helpful to take a Lagrangian perspective and consider the sources and sinks of westerly angular momentum encountered by a ring of air moving in the Hadley cell up from the equatorial surface, back down to the surface at a higher latitude, and back to the equator along the surface. In steady state, the ring must return to its starting point with the same angular momentum as when it left. If the ring experiences a net sink of momentum during the free-tropospheric portion of its trip—as will be the case in an Earthlike situation where extratropical eddies generate a net momentum export from the tropics—then it must accelerate as it moves along the surface back to the equator (which implies surface easterlies and an upward frictional momentum flux from the surface). In this situation, $\bar{u}$ will have a relative maximum at the equatorial surface, and $-\bar{v} \partial_y \bar{u}$ will be negative. On the other hand, if the ring experiences a net gain of momentum in the free troposphere, then it must decelerate in the surface branch, yielding an equatorial minimum in surface $\bar{u}$ and positive $-\bar{v} \partial_y \bar{u}$. In this case, it is conceivable that the ring could remain in superrotation at all stages of the cycle, including the surface branch near the equator. This provides a fourth potential pathway to surface superrotation, driven by downward advection of

superrotating air in the subtropical branch of the Hadley cell (analogously to equatorial downward advection in the reversed Hadley cell case), which would show up in Eulerian diagnostics as being forced by positive low-level $-\bar{v}\partial_y\bar{u}$.

b. Equatorial Péclet number

As highlighted above, one possibility for generating surface superrotation is through vigorous downward transport of upper-level momentum by viscosity, balancing upward momentum advection by the MMC. This is reminiscent of an advection–diffusion problem, which can be made explicit by modeling $\bar{\tau}$ using a simple diffusive closure,

$$\bar{\tau} = -g\rho^2\nu\partial_p\bar{u} = -\frac{p^2}{gH^2}\nu\partial_p\bar{u}, \quad (3)$$

where ν is the kinematic viscosity coefficient (m^2s^{-1}), ρ is density, and $H = RT/g$ is the local-scale height, with T the temperature and R the gas constant. Advection–diffusion problems are characterized by the Péclet number (Pe), a nondimensional parameter that compares the advective and diffusive parts of the governing equation. If we scale the vertical advection term in (1) as $\bar{\omega}\partial_p\bar{u} \sim WU/p_s$, where W is a typical value of equatorial pressure velocity and U is the typical zonal-mean wind difference between the upper and lower equatorial troposphere, and scale the viscous term using (3) as $g\partial_p\bar{\tau} \sim \nu U/H^2$, we can define an equatorial Péclet number:

$$\text{Pe} = \frac{WH^2}{p_s\nu}. \quad (4)$$

Physically, Pe is the ratio of the time-scale p_s/W on which a parcel rises through a scale height to the time H^2/ν on which diffusion acts over that distance.² When $\text{Pe} \gg 1$, advection dominates and we expect zero or negative surface winds at the equator, while when $\text{Pe} \ll 1$, diffusion is dominant and surface superrotation is possible; we expect the transition to surface superrotation to occur at $\text{Pe} \sim 1$.

c. Summary

To summarize, we have identified four hypothetical scenarios that can give rise to surface superrotation. All scenarios require equatorial eddy momentum convergence, which may be either concentrated near the

surface or at upper levels, where it is transported down to the surface by viscosity or advection. More specifically, the four scenarios are as follows:

- 1) Conventional Hadley cell, eddy momentum convergence concentrated away from the surface, high viscosity (low Péclet number)
- 2) Conventional Hadley cell, eddy momentum convergence concentrated at the surface
- 3) Reversed Hadley cell
- 4) Conventional Hadley cell, eddy momentum convergence concentrated away from the surface with net positive convergence into the tropical zone embraced by the Hadley cell

3. Axisymmetric model

This section presents a series of axisymmetric model simulations providing concrete examples of surface superrotation achieved via the four pathways discussed in the previous section. Since existing Earth and planetary simulations most commonly show a conventional Hadley cell and eddy momentum convergence in the upper troposphere, we pay most attention to scenario 1 in the list above.

a. Model description

We employ an axisymmetric model forced by a prescribed distribution of zonal momentum sources mimicking the effect of eddy momentum convergences. The model solves the axisymmetric primitive equations in pressure coordinates, as described in Caballero et al. (2008). The equations are solved in flux form using a simple upwind scheme and discretized with 91 points in latitude and 45 vertical levels. Newtonian relaxation restores the temperature field to a background radiative–convective equilibrium state as specified in Held and Suarez (1994). In addition, an extra heating Q is imposed at the equator with compensating cooling in the subtropics so that the globally integrated heating is zero. The structure of Q is shown in Fig. 1a. Its purpose is to control Hadley cell strength (and thus the equatorial updraft) by simply varying the maximum value of the equatorial heating Q_{max} .

Eddy momentum convergence is represented by a time-invariant distribution S of momentum sources and sinks, as in previous work (Schneider 1984; Singh and Kuang 2016). The structure of S , shown in Fig. 1 (top row), is designed to capture the main features of eddy momentum convergence found in the full-complexity model simulations to be described in section 4. The structure varies sinusoidally in latitude and decreases exponentially with height from a maximum at 300 hPa,

² The Péclet number defined this way is the inverse of the parameter τ_D/τ_v , defined in Held and Hou (1980), evaluated at the equator.

FIG. 1. Results from axisymmetric simulations with varying Péclet number (Pe). (a)–(c) Imposed zonal momentum source S . Gray contours in (a) show the tropical heating Q normalized by $Q_{\max}$ (contours at 0.1, 0.3, 0.5, ...; negative dashed). Black dotted lines in (b) and (c) indicate the region of enhanced equatorial viscosity. (d)–(f) Zonal-mean zonal wind (shading; blues indicate easterlies) and meridional mass streamfunction (gray contours at 1, 5, 10, and $15 \times 10^{10} \text{ kg s}^{-1}$; negative dashed). Black lines outline regions of superrotation, where angular momentum is greater than the equatorial value in solid-body rotation Ωa^2 . (g)–(i) Zonal-mean surface zonal wind. (j)–(l) Terms in the vertically integrated zonal momentum budget: MMC term ($-\langle \bar{v} \partial_y \bar{u} \rangle - \langle \bar{\omega} \partial_p \bar{u} \rangle$; orange line), imposed zonal momentum source (blue), and surface stress (green).

transitioning smoothly to zero in the stratosphere, and features positive values near the equator and in the extratropics compensated by negative values in the subtropics. The amplitudes of the lobes are adjusted so that the global integral of S is exactly zero to ensure global conservation of angular momentum. For simplicity, we do not attempt to represent eddy energy transports, though it should be noted that these can affect Hadley cell strength and thus the equatorial momentum balance (Singh and Kuang 2016; Singh et al. 2017).

Vertical viscosity is represented using the diffusive closure in (3). The diffusivity has the structure $\nu = \max[\nu_b, \nu_e g(\varphi)]$, where $\nu_b = 2 \text{ m}^2 \text{ s}^{-1}$ is a uniform background diffusivity with a value in the range observed in free-tropospheric turbulence in the real atmosphere (Wilson 2004), ν_e is an equatorial diffusivity

that varies from run to run, and $g(\varphi)$ is a Gaussian meridional profile centered at the equator with half width of 3° latitude; the total diffusivity is vertically uniform in the troposphere but tapers rapidly to the background value above the tropopause at 200 hPa. The purpose of this structure is to arbitrarily enhance equatorial viscosity (and control the equatorial Péclet number) without excessively distorting the global general circulation. Physically, we take this equatorial enhancement to represent convective momentum transport (CMT) in the intertropical convergence zone. Roms (2012, 2014) shows that diffusion is in fact a reasonable first approximation for CMT, at least for wind profiles with long vertical wavelength, and given typical values of convective mass flux and entrainment rates yields a diffusivity of $10\text{--}30 \text{ m}^2 \text{ s}^{-1}$, an order of magnitude higher than in clear air. Note that this

enhanced diffusivity applies only to momentum; potential temperature diffusion is applied but with the background value ν_b everywhere.

Surface stress is computed using the linearized formulation $\bar{\tau}_s = \gamma \bar{u}_s$, where $\bar{u}_s$ is the zonal wind in the lowest model layer and $\gamma = 0.065 \text{ Pa s m}^{-1}$. A diffusive horizontal viscosity term of the form $-a^{-2} \cos^{-1} \varphi \partial_\varphi (\nu \cos^{-1} \varphi \partial_\varphi \bar{u})$ is also included, with the same ν as for vertical viscosity, but because of the small aspect ratio ($a \gg H$), it plays a much smaller role than vertical diffusion.

b. Transition to surface superrotation: Scenario 1

Figure 1 presents steady-state results from three simulations demonstrating the transition to surface superrotation at low Péclet number. All simulations employ the same momentum source distribution S , shown along the top row of the figure. One simulation (left column) uses $Q_{\max} = 0.5 \text{ K day}^{-1}$, chosen to give a Hadley cell with roughly the same strength as in the full-complexity model (section 4) and spatially uniform diffusivity $\nu_e = \nu_b = 2 \text{ m}^2 \text{ s}^{-1}$. The Péclet number for this simulation, computed using (4), taking W as the 200–1000-hPa vertically averaged equatorial ω , $H = 8000 \text{ km}$, and $\nu = \nu_e$, turns out to be about 30. In the second simulation (middle column), equatorial viscosity is enhanced by increasing ν_e to $16 \text{ m}^2 \text{ s}^{-1}$ while $Q_{\max}$ is reduced to 0.1 K day^{-1} ; the weaker thermal driving reduces equatorial upwelling, which combined with the moderate enhancement of diffusivity brings the Péclet number down to about 1. The third simulation (right column) again uses $Q_{\max} = 0.5 \text{ K day}^{-1}$ but ν_e is increased to $2000 \text{ m}^2 \text{ s}^{-1}$, yielding $\text{Pe} = 0.02$.

In the $\text{Pe} = 30$ case, the positive equatorial torque imposed by S leads to superrotation in the equatorial upper troposphere (Fig. 1d), but winds remain easterly throughout the lower tropical troposphere down to the surface (Fig. 1g). Unrealistic upper-level easterlies appear in the subtropics; this happens because the poleward branch of the Hadley cell is unrealistically weak in this model, so the negative momentum source in the subtropics has ample time to act and induce easterly winds. Strong jets appear at the poleward edges of the Hadley cell, as well as midlatitude Ferrel cells and surface westerlies in response to the positive momentum source there. The vertically integrated momentum balance (Fig. 1j) shows that the equatorial momentum forcing is balanced almost entirely by the MMC, with surface stress providing a small additional source of westerly momentum. In the $\text{Pe} = 1$ case, the situation is much the same, but the equatorial vertical shear is weaker, and winds are westerly throughout the column; the region of superrotation now stretches downward and just touches the surface (Fig. 1e). In the $\text{Pe} = 0.02$ case,

the very strong equatorial viscosity eliminates vertical shear almost entirely, and zonal winds on the equator are nearly uniform and westerly at all levels, including the surface. In this case, vertically integrated momentum forcing is balanced partly by surface drag, though the MMC continues to provide a drag of comparable magnitude (Fig. 1l).

Figure 2 presents vertical profiles of the zonal wind and momentum budgets averaged within 2° latitude of the equator. The zonal wind profile in the $\text{Pe} = 30$ case shows a layer of unsheared easterlies in the lower troposphere, transitioning to strong superrotation in the upper troposphere. The sheared transition region is centered around 300 hPa, where S peaks, and the momentum balance there is mostly between acceleration by S and drag due to vertical advection by the MMC (thin orange line in Fig. 2d), with viscosity playing a smaller role redistributing momentum downshear. Near the surface, the MMC also exerts significant drag on the zonal wind, because of the horizontal vorticity flux term (dotted orange line): as can be seen in Fig. 1g, the near-surface wind has a cusp-like profile giving large values of relative vorticity in the immediate vicinity of the equator. The resulting drag is balanced by viscosity and by mean vertical advection, which both transport positive momentum upward from the surface (note the reversed vertical shear) and converge it within the near-surface layer. Note that since there is zero vertical shear between 600 and 850 hPa, there is no momentum transport across this layer by either advection or viscosity; the upper and lower troposphere are effectively decoupled, each achieving a local momentum balance.

In the $\text{Pe} = 1$ case, on the other hand, vertical shear is westerly at all levels below 200 hPa, and viscous momentum transport is downward throughout the column, coupling the upper troposphere to the surface. Viscous transport converges in the lower troposphere to balance drag by vertical and horizontal advection but with enough left over to drive modest westerlies at the surface. The high-viscosity case ($\text{Pe} = 0.02$) is a more extreme version of the previous case: vertical shear is almost absent (though still slightly westerly), and S is almost entirely balanced by viscosity. Viscous stress converges in the lower troposphere to balance drag by horizontal advection—which is stronger and fills a deeper layer than in previous cases—but still leaving enough to drive substantial surface superrotation.

To provide a broader overview of the transition, Fig. 3 shows results for a large number of simulations with Pe spanning five orders of magnitude. The Péclet number is varied in these simulations by changing the equatorial diffusivity ν_e and by using three different values of

FIG. 2. Equatorial profiles (averaged within 2° latitude of the equator) for the same simulations as in Fig. 1. (a)–(c) Zonal-mean zonal wind. (d)–(f) Terms in the zonal-mean zonal momentum budget: vertical momentum advection (dashed orange line), horizontal momentum advection (dotted orange line), their sum (solid orange line), imposed momentum source (blue), and viscosity ($-g\partial_p\bar{u}$; green).

tropical heating, as indicated in the figure. All simulations use the same momentum source S . As expected, surface winds are easterly at high Pe and westerly at low Pe , with the transition to surface superrotation occurring at $Pe \sim 1$ (Fig. 3a). The results do not collapse onto a single curve, however, because Pe only captures the role of vertical advection by the MMC and neglects horizontal advection, which as we saw above is important near the surface.

To better appreciate the role of horizontal advection, we can rewrite the vertically integrated momentum budget in (2), setting $-\partial_y\langle u'v' \rangle = \langle S \rangle$, as

$$-\frac{\bar{\tau}_s}{\langle S \rangle} - \frac{\langle \bar{w}\partial_p\bar{u} \rangle}{\langle S \rangle} - \frac{\langle \bar{v}\partial_y\bar{u} \rangle}{\langle S \rangle} + 1 = 0. \quad (5)$$

Figures 3b–d show the first three terms in this expression. For $Pe \gg 1$, the upper troposphere is decoupled from the surface layer, as discussed above; vertical advection balances momentum forcing in the upper

troposphere (so $-\langle \bar{w}\partial_p\bar{u} \rangle / \langle S \rangle + 1 \approx 0$; Fig. 3c) while horizontal advection balances surface stress in the surface layer (cf. Figs. 3b and 3d). Increasing $Q_{\max}$ strengthens the Hadley cell, causing an increase in $\bar{v}$ in the lower branch of the Hadley cell; this makes low-level drag stronger (Fig. 3d) and leads to stronger surface easterlies at the equator (Fig. 3a). As viscosity increases and Pe gets close to 1, the troposphere becomes increasingly coupled; horizontal advection in the surface layer can be partially balanced by westerly momentum transported from the upper troposphere by viscous stress. The transition to surface superrotation occurs when this balance is complete and some of the momentum transported from the upper troposphere can be deposited directly to the surface. However, a stronger Hadley cell requires stronger viscosity (smaller Pe) for the transition to occur. At $Pe \ll 1$, vertical shear tends to vanish, vertical advection no longer plays a role, and all the momentum injected at upper levels that is

FIG. 3. (a) Equatorial surface wind and (b)–(d) terms in the vertically integrated momentum budget as a function of equatorial Péclet number (Pe) for simulations with varying equatorial diffusivity ν_e (colors) and tropical heating $Q_{\max}$ (symbols; see legend).

not consumed by horizontal advection is deposited to the surface, driving strong surface superrotation; for a given Pe , surface superrotation strengthens as the Hadley cell weakens.

In summary, horizontal advection affects the precise value of Pe required for transition to superrotation but does not qualitatively modify the nature of the transition. As discussed in section 2, horizontal advection plays a significant role here because of the cusp-like structure of low-level $\bar{u}$ in these simulations. This seems to be a generic feature of axisymmetric models (e.g., Held and Hou 1980, their Fig. 8).

c. Transition to surface superrotation: Scenarios 2, 3, and 4

We briefly discuss three further simulations demonstrating surface superrotation via alternative scenarios. Scenario 2 is exemplified by a simulation with weak, uniform viscosity $\nu = \nu_b$ (as in the $Pe = 30$ run above) and $Q_{\max} = 0$ (so the only thermal driving is by Held–Suarez relaxation). The momentum forcing S is modified to have a positive maximum near the surface at the equator (Fig. 4a). In this configuration, the momentum source directly forces surface westerlies at the equator; in addition, the Hadley cell advects momentum upward from the low-level source, producing equatorial westerlies at all levels (Fig. 4b). Near the surface, the momentum source is balanced by drag due to the MMC and viscosity (Fig. 4c). Surface winds (Fig. 4d) are easterly through most of the tropics, as required to balance the net deceleration provided by tropical-mean S but westerly around the equator.

To generate scenario 3, we use the same settings as in the previous case but configure S to have a broad upper-level positive lobe straddling the equator and negative lobes in the extratropics (Fig. 4e). This configuration yields extremely strong superrotation, approaching 200 m s^{-1} at its maximum (Fig. 4f). To accommodate such strong westerlies while maintaining thermal wind balance, temperatures at the equator need to be very

warm—warmer in fact than the Held–Suarez radiative–convective equilibrium. To generate the required equatorial warm anomaly, the Hadley cell reverses so as to converge energy onto the equator (this can be seen as a consequence of the Hadley cell being entirely eddy driven and thermally indirect; see Liu and Schneider 2011). The Hadley cell now transports momentum downward from upper levels, driving low-level westerlies that are balanced by frictional drag (Fig. 4g).

Finally, we generate scenario 4 by repeating the previous simulation but adding strong diabatic heating ($Q_{\max} = 1 \text{ K day}^{-1}$; Fig. 4i), strong enough to yield a thermally direct Hadley cell (Fig. 4j). In this case, air moving poleward in the upper branch of the Hadley cell is continuously accelerated by the imposed momentum source, forcing an off-equatorial angular momentum maximum. This leads to symmetric instability at upper and middle levels, and the simulation never settles into a time-invariant state (the figure shows averages over 100 days of the equilibrated statistically steady state). Transient motion associated with symmetric instability acts to homogenize angular momentum along isentropes, which are essentially horizontal. As a result, the entire tropical upper troposphere superrotates with horizontally uniform angular momentum. The subsiding branch of the Hadley cell descends along a sloping surface where strong vertical shear develops, as also seen in the nonsuperrotating simulations of Held and Hou (1980). As it subsides along this surface, the air is gradually decelerated by viscosity but is still superrotating when it reaches the surface at around 7° latitude and in fact remains in superrotation all the way to the equator. As discussed in section 2a, $-\bar{v}\partial_y \bar{u}$ is positive at low levels near the equator in this case (Fig. 4k) and appears as the main driver of surface superrotation.

4. Full-complexity model

We now examine the equatorial momentum balance of a full-complexity GCM in detail and search for a

FIG. 4. Axisymmetric simulations showing surface superrotation via (a)–(d) scenario 2, (e)–(h) scenario 3, and (i)–(l) scenario 4. (a),(e),(i) Imposed momentum source S (shading) and tropical heating Q (gray contours; conventions as in Fig. 1). (b),(f),(j) Zonal wind and mass streamfunction with plotting conventions as in Fig. 1. Black lines outline regions of superrotation, where angular momentum is greater than the equatorial value in solid-body rotation Ωa^2 . (c),(g),(k) Equatorial momentum balance as in Figs. 2d–f. (d),(h),(l) Zonal-mean surface zonal wind (m s^{-1}).

parameter regime in which the model exhibits surface superrotation.

a. Model and simulations

We use the Community Atmospheric Model, version 4 (CAM4), developed by the National Center for Atmospheric Research (Neale et al. 2010). The dynamical equations are solved using the spectral-element dynamical core High-Order Method Modeling Environment (HOMME) with 1° resolution. Unlike the default CAM4 finite-volume dynamical core, HOMME conserves angular momentum to high accuracy (Lauritzen et al. 2014), providing confidence that the superrotation observed here is not the result of nonphysical momentum sources (Lebonnois et al. 2012b). Subgrid-scale processes are modeled using the full set of CAM4 parameterizations, including radiation, moist convection, turbulence, and clouds [see Neale et al. (2010) for details]. The moist convective parameterization includes a description of convective momentum transport following the method of Gregory et al. (1997). The model is configured as an aquaplanet with fixed sea surface

temperature (SST). The SST is zonally uniform and equatorially symmetric, with a cosine-squared latitudinal dependence dropping from a maximum T_{max} at the equator to a minimum T_{min} at 80° latitude, remaining constant poleward of that latitude. Orbital parameters are set to yield permanent equinox insolation (zero obliquity and eccentricity), though a diurnal cycle is retained. Atmospheric CO_2 is set at 367 ppm. Simulations are run for 6 years, and results presented below are statistics over the last 5 years.

In the reference (control) simulation, we set $T_{\text{max}} = 36^\circ\text{C}$ and $T_{\text{min}} = 7^\circ\text{C}$. These values are chosen to roughly match the best available proxy surface temperature reconstructions for the early Eocene (Huber and Caballero 2011). We perform three sets of sensitivity simulations. In one set, which aims to explore the effect of Hadley cell strength, the equator–pole temperature gradient is reduced by raising T_{min} in several steps up to a maximum of 30°C . Note that this has the additional effect of reducing extratropical baroclinicity, leading to a reduction in momentum export from the tropics by baroclinic eddies; this can

be expected to further favor surface superrotation by decreasing the drag on the subsiding branch of the Hadley cell, as discussed in [section 2](#). In a second set, $T_{\min}$ is held fixed at $T_{\min} = 7^{\circ}\text{C}$ while the strength of CMT is artificially increased. This is accomplished by simply multiplying the instantaneous u and v tendencies produced by the CMT parameterization by a fixed constant $c > 1$ at each time step in the simulation. We find that the model becomes numerically unstable when $c > 8$ but produces reasonable results below that value. In the third set, the SST gradient is reduced, and CMT is simultaneously increased. We present results for three end-member simulations: the control run, a low-gradient run ($T_{\min} = 30^{\circ}\text{C}$, $c = 1$), and a low-gradient CMT $\times 8$ run ($T_{\min} = 30^{\circ}\text{C}$, $c = 8$). Our aim here is not to present a full exploration of the GCM's sensitivity to these parameters but rather—as with the axisymmetric model—simply show under which conditions a transition to surface superrotation occurs in the GCM.

b. Péclét number estimate

For each simulation, we diagnose the net zonal-mean viscous tendency term $-g\partial_p\bar{\tau}$ —which comprises the time and zonal average of all parameterized subgrid drag terms in the model—as a residual in the momentum budget (1) assuming $\partial_t\bar{u} = 0$. We then integrate vertically to obtain $\bar{\tau}$. For a simple order-of-magnitude estimate of the equatorial Péclét number, we take the vertical average of $\bar{\tau}$ in the 100–400-hPa layer (where wind shear is strong) and estimate ν_e from (3) using the vertical-mean shear in the same layer. This yields $\nu_e = 1.6$, 1.5, and $4.8\text{ m}^2\text{ s}^{-1}$ and $\text{Pe} = 16$, 4, and 3 for the control, low-gradient, and low-gradient CMT $\times 8$ cases, respectively.

c. Control case

We now discuss the control case—which has the most realistic parameter setting among our simulations—in some detail and compare it with the other two cases in the next subsection. Results for the control case are presented in the left column of [Figs. 5 and 6](#). The eddy kinetic energy [$\text{EKE} = (u'^2 + v'^2)/2$] distribution for this case ([Fig. 5a](#)) shows maxima in the midlatitudes and another peak at the equator. Eddy momentum convergence produces a westerly equatorial acceleration of up to $4\text{ m s}^{-1}\text{ day}^{-1}$ in the upper troposphere but weak easterly acceleration in the equatorial boundary layer. [Figure 6g](#) shows a breakdown of the eddy momentum flux convergence into horizontal and vertical components, revealing that very strong horizontal convergence concentrated near the tropopause is almost exactly balanced by vertical eddy fluxes, which transport

momentum downward and deposit it in the 300–500-hPa layer as also seen in previous work ([Norton 2006](#); [Lutsko 2018](#)). The net westerly acceleration in the upper troposphere drives strong equatorial superrotation in the layer between 100 and 600 hPa ([Fig. 6a](#)), but winds are easterly in the lower troposphere and at the surface with reversed shear in the near-surface layer, qualitatively resembling the high-Pe axisymmetric simulation ([Fig. 2a](#)).

The momentum balance ([Fig. 6d](#)) also resembles its axisymmetric counterpart ([Fig. 2d](#)) in the upper troposphere, where eddy acceleration is mostly balanced by mean upward advection [consistent with [Kraucunas and Hartmann \(2005\)](#)], with a small role for viscosity. The boundary layer balance is very different from the axisymmetric case, however. The horizontal structure of $\bar{u}$ near the equator is much smoother in CAM4 than in the axisymmetric model (cf. [Figs. 1g and 5g](#)); relative vorticity is therefore small near the equator, making the drag due to the $-\bar{v}\partial_y\bar{u}$ negligible at all levels. Instead, there is drag because of horizontal eddy momentum divergence in the equatorial boundary layer ([Fig. 6g](#)), playing a role somewhat analogous to mean horizontal advection in the axisymmetric model and balancing acceleration by vertical advection and surface stress.

Why does the eddy momentum convergence profile have the particular structure shown in [Fig. 6d](#), with westerly acceleration at upper levels and drag in the boundary layer? To answer this question, we take a closer look at the nature and structure of tropical eddies in the simulation. As in previous work ([Caballero and Huber 2010](#); [Carlson and Caballero 2016](#)), we identify the leading mode of tropical variability using empirical orthogonal function (EOF) analysis. Specifically, we apply EOF decomposition to 5 years of 6-hourly snapshots of the zonal wind field at 100 hPa (where equatorial EKE peaks) in the 60°S – 60°N band. This yields two leading modes with equal variance, which in combination explain 42% of the total variance. The two modes have almost identical spatial structure—which is almost purely wavenumber 1 in the zonal direction (see below)—but are in quadrature with each other. The associated principal components (PCs) have maximum lag correlation at around 4 days. Together, the two modes describe an eastward-propagating wavenumber-1 wave with a period of 16 days and a phase speed of around 30 m s^{-1} .

To identify the spatial structure of the global anomalies associated with this wave, we regress the two leading PCs onto other fields; to minimize sampling error, we present an average of the two regressions after zonal shifting by 90° to bring the two modes into phase. Results are presented in [Fig. 7](#). An equatorial cross

FIG. 5. Results from CAM4 simulations using (a),(d),(g),(j) reference parameters, (b),(e),(h),(k) reduced meridional SST gradient, and (c),(f),(i),(l) reduced SST gradient plus enhanced convective momentum transport. (a)–(c) Total eddy momentum convergence ($-\partial_y u'v' - \partial_p u'w'$; shading) and $\sqrt{\text{EKE}}$ (gray contours; dotted, solid, and thick solid lines at 5, 10 and 15 m s^{-1} , respectively). (d)–(f) Zonal-mean zonal wind and mass streamfunction, with plotting conventions as in Fig. 1. Black lines outline regions of superrotation, where angular momentum is greater than the equatorial value in solid-body rotation Ωa^2 . (g)–(i) Zonal-mean surface zonal wind. (j)–(l) Terms in the vertically integrated zonal momentum budget: MMC term ($-\langle \bar{v} \partial_y \bar{u} \rangle - \langle \bar{w} \partial_p \bar{u} \rangle$; orange line), eddy momentum convergence ($-\langle \partial_y u'v' \rangle$; blue line), and surface stress (green line).

section through the leading mode (Fig. 7a) shows a remarkable resemblance to the observed MJO (Kiladis et al. 2005; Adames and Wallace 2014a,b). The wave has a first-baroclinic-mode vertical structure, with the upper-level height anomaly shifted roughly 90° east with respect to the low-level anomaly. Air rises through a layer of cooler air in the lower troposphere and then into a strong warm anomaly in phase with the upper-level height anomaly. The amplitudes of these anomalies are around 4 times greater than in observations and similar to those found in previous warm-climate simulations (Carlson and Caballero 2016).

A horizontal cross section at upper levels shows the familiar combination of an equatorial Kelvin wave with off-equatorial, geostrophically balanced Rossby gyres (Fig. 7b). As in observations (Adames and Wallace

2014a), the gyres are tilted so as to yield a chevron pattern that converges westerly momentum onto the equator. Figure 8 shows that the Rossby gyres are confined within the subtropical jets both horizontally and vertically. This confinement can be interpreted as Rossby wave trapping by the ambient shear (Barlow 2012; Monteiro et al. 2014)—note in fact that the anomalies become evanescent in the region where their phase speed exceeds the local value of the zonal-mean wind (delineated by the thick contour in Fig. 8). Midlatitude features coherent with the tropical waves are also apparent, again in agreement with observations (cf. Adames and Wallace 2014a, their Fig. 13).

In summary, the dominant mode of tropical variability in our simulations is remarkably MJO-like in its spatial structure but with considerably faster eastward

FIG. 6. Equatorial profiles (averaged within 2° latitude of the equator) for the same simulations as in Fig. 5. (a)–(c) Zonal-mean zonal wind (solid line) and square root of EKE (dotted). (d)–(f) Terms in the zonal-mean zonal momentum budget: vertical momentum advection (dashed orange line), horizontal momentum advection (dotted orange line), their sum (solid orange line), total eddy momentum convergence ($-\partial_y \overline{u'v'} - \partial_p \overline{u'\omega'}$; blue), and viscosity estimated as a residual of the other terms (green). (g)–(i) Partitioning of the eddy momentum convergence term into vertical (solid line) and horizontal (dotted) components.

propagation than the observed MJO phase speed of $5\text{--}10\text{ m s}^{-1}$. This difference in phase speed may simply be due to Doppler shifting by the superrotating equatorial winds—with westerlies reaching 30 m s^{-1}

near the tropopause (Fig. 6a)—but we cannot exclude the possibility that the simulated mode has a fundamentally different dynamical origin from the real-world MJO.

FIG. 7. Regression of leading PC of tropical upper-tropospheric variability onto (a) geopotential height (shading), temperature (contours at intervals of 0.3 K; negative dashed; zero thick), and zonal–vertical wind (arrows; u' and w' have been scaled so arrows point in their true direction within the axes) averaged between 5°S and 5°N; (b) geopotential height (shading), ω (single contour at 2 Pa s⁻¹; negative dashed) and horizontal wind (arrows) averaged between 100 and 200 hPa; (c) as in (b), but averaged between 600 and 900 hPa; and (d) surface pressure (shading), ω at 900 hPa (thick and thin contours at 2 and 1 Pa s⁻¹ respectively; negative dashed) and horizontal wind averaged between 900 and 1000 hPa (arrows). All fields are scaled to show anomalies associated with one standard deviation of the PC.

FIG. 8. Zonal root-mean-square amplitude of height anomalies associated with the leading mode of variability (shading) and climatological $\bar{u}/\cos\phi$ (contours at intervals of 10 m s^{-1} ; zero dotted; 30 m s^{-1} thick).

The fact that the Rossby wave components are vertically trapped within the upper troposphere is also apparent in a horizontal cross section within the lower free troposphere (Fig. 7c), which shows that the off-equatorial height anomalies are much weaker at these levels though still tilted so as to provide a modest equatorward momentum transport. Things change markedly in the boundary layer (Fig. 7d). Here, wind anomalies are not geostrophically balanced, showing strong frictionally driven cross-isobaric flow. This Ekman transport causes horizontal mass divergence in regions of anomalous westerlies and convergence in regions of easterlies, as also seen in observations (Adames and Wallace 2014b). The role of frictionally driven moisture convergence in the MJO has been highlighted in many previous works (e.g., Salby et al. 1994; Maloney and Hartmann 1998; Adames and Wallace 2015); here, we also see that it induces wind anomalies oriented so as to diverge zonal momentum from the equator (note that wind vectors near the equator in Fig. 7d are oriented at right angles to those at other levels).

Finally, Fig. 9 compares the climatological momentum fluxes diagnosed from the simulation with those associated with the MJO-like mode, computed by zonally averaging the product of the regressed u and v (or u and ω) fields. The horizontal eddy momentum flux diagnosed from the model (shading in Fig. 9a) is very tightly peaked near the tropopause, as expected given the upper-level trapping of the Rossby wave components. Near the equator, the momentum flux has the same sign throughout the free troposphere but changes sign in the boundary layer; this yields the upper-level momentum convergence and boundary layer divergence shown in Fig. 6g. The momentum fluxes attributable to the MJO-like mode (contours in Fig. 9a) reproduce this

structure accurately and capture over half the magnitude of the full eddy fluxes, implying that this mode is the chief driver of eddy momentum transport in the equatorial zone. The vertical flux (Fig. 9b) is downward at all levels and is also concentrated in the upper troposphere though with a broader peak extending down to around 400 hPa. At low levels, the downward transport has two off-equatorial maxima, causing off-equatorial maxima of vertical momentum convergence into the boundary layer (see Fig. 5a). The fluxes associated with the MJO-like mode again reproduce this structure well except in the midtroposphere, where they become slightly negative. This is due to the first baroclinic mode structure of the MJO, with u' changing sign around 400 hPa (Fig. 7a).

d. Sensitivity to reduced SST gradient and increased CMT

In the low-gradient case with default CMT (middle columns in Figs. 5 and 6), superrotation is weaker than in the control case but still substantial in the upper troposphere, with weaker easterlies at low levels. Tropical eddy amplitudes (as measured by $\sqrt{\text{EKE}}$) are about 30% smaller than in the control case. The associated momentum convergences retain the same general structure as in the control case but are much weaker, dropping by about 90%. This combination of modestly weaker amplitudes and dramatically weaker momentum transports can only be explained by changes in phasing of u' , v' , and ω' , implying changes in the structure and phase-line tilting of the waves, likely related to the much weaker horizontal and vertical-mean shear in which the waves are embedded. The equatorial momentum balance has a qualitatively similar structure to that in the control case, but viscosity now plays a larger role (it has

FIG. 9. Eddy momentum fluxes in the control simulation. (a) Horizontal eddy momentum flux (shading; $\text{m}^2 \text{s}^{-2}$). Gray contours show the same quantity (with the same units and contour intervals) computed for the leading mode of tropical variability. (b) As in (a), but for the vertical momentum flux (m Pa s^{-2}). Note the nonlinear contour spacing.

the same size as vertical advection at around 200 hPa), consistently with the smaller estimated Pe for this simulation. Equatorial surface winds remain easterly, however.

Finally, the low-gradient $\text{CMT} \times 8$ case (right columns in Figs. 5 and 6) presents several surprises. One is that eddy momentum convergence is very much larger than in the previous case despite comparable eddy kinetic energies in the two simulations. The reason for this increase is not clear; it is possible that increased CMT affects the phase structure of tropical waves so as to increase their momentum transport and convergence; alternatively, it may be that increased CMT increases damping of the MJO's Kelvin wave component, which recent work suggests to be a key factor in permitting equatorial eddy momentum convergence (Zurita-Gotor and Held 2018). Another surprising aspect is that the Hadley cell is stronger than in the previous case, though there is no change in the imposed surface temperature gradient. In any event, the increase in CMT is reflected in a substantial role for viscosity in the momentum balance (Fig. 6f), and Pe is low enough to induce a transition to weak surface superrotation (Fig. 5i), as in the $Pe = 1$ axisymmetric simulation presented above. In fact, the low-gradient $\text{CMT} \times 8$ case is the only simulation in our set of sensitivity experiments to demonstrate surface superrotation; cases with $\text{CMT} \times 8$ but greater SST gradients fail to do so.

5. Discussion and conclusions

We have identified four potential scenarios that can lead to surface superrotation. Scenario 1 envisages a conventional, thermally direct Hadley cell with updraft

on the equator and equatorial eddy momentum flux convergence localized away from the surface. In this scenario, the only way to induce surface superrotation is via downward viscous momentum transfer. The Péclet number (Pe)—which compares the roles of viscosity and upward momentum advection by the Hadley cell—proves useful in predicting the transition to surface superrotation in this scenario, which occurs when $Pe \lesssim 1$. In scenario 2, positive eddy momentum convergence at the surface directly drives surface superrotation. In scenario 3, eddy momentum convergence is localized away from the surface, but a reversed, thermally indirect Hadley cell advects momentum down to the surface. In scenario 4, eddy momentum convergence provides positive acceleration throughout the upper branch of the Hadley cell, and upper-level superrotating air is advected by the Hadley cell all the way down to the surface at the equator.

In an axisymmetric model where zonal torques representing the eddy momentum convergence can be arbitrarily specified, we are able to induce surface superrotation via all four pathways. Our GCM simulations—like essentially all superrotating three-dimensional simulations in the literature—robustly present a conventional Hadley cell and eddy momentum convergence peaked at upper levels, precluding scenarios 2 and 3, and net momentum export by extratropical eddies, precluding scenario 4. We found that by greatly reducing the meridional SST gradient and artificially enhancing convective momentum transport, we were able to reduce Pe to the point where weak surface superrotation appears via scenario 1. The only previous simulations we are aware of showing surface superrotation also featured a reversed Hadley cell as well as low-level momentum convergence

on the equator (Liu and Schneider 2011, their Fig. 4) and therefore follow a combination of scenarios 2 and 3.

On the basis of this analysis, the question of why surface superrotation is largely absent in the literature therefore breaks down into four parts: 1) the question of why eddy momentum convergences are always concentrated away from the surface, precluding scenario 2; 2) why Pe is too high to yield surface superrotation via scenario 1; 3) why a thermally indirect Hadley cell is not more commonly observed, precluding scenario 3; and 4) why the width of the tropical region with positive eddy momentum convergence is narrower than the width of the Hadley cell, precluding scenario 4.

The analysis of the MJO-like structure in our GCM results (section 4) provides some insight into question 1. The Rossby wave components of this mode are excited by mass divergence at the convective outflows near the tropopause, and Fig. 8 shows that they remain vertically trapped within their critical lines, becoming evanescent at low levels. This trapping should generalize to Rossby waves generated by stationary heating (in which case the upper-level stationary waves will be unable to propagate vertically across the zero-wind line into low-level easterlies) and by unorganized stochastic convection. As a result, horizontal eddy momentum convergence associated with Rossby waves excited by upper-level mass divergence will remain confined to the upper troposphere. Vigorous shallow convection could create a low-level Rossby wave source, but these waves would be excited into easterlies and would be horizontally evanescent and likely unable to drive strong momentum convergence. In addition, equatorial Kelvin waves driven by heating anomalies have a baroclinic structure and cannot efficiently transport momentum downward beyond their midtropospheric node (where $u' = 0$ and thus $\overline{u'\omega'} = 0$). Moreover, Ekman transport associated with the equatorial Kelvin wave will robustly drive horizontal momentum divergence in the boundary layer, as discussed in section 4c. In the dry adiabatic instability mechanism of Wang and Mitchell (2014), a Rossby wave trapped within the subtropical jets interacts with an equatorial Kelvin wave, so the same constraints on momentum transport will apply. In summary, so long as a conventional, thermally direct Hadley cell exists and generates upper-level westerlies and low-level easterlies, we expect eddy momentum convergence to be strongly peaked in the upper troposphere, with weak convergence or divergence at low levels.

Regarding question 2, on the value of Pe , we note that observed clear-air turbulent diffusivities on Earth are $\sim 1 \text{ m s}^{-2}$ (Wilson 2004) and Hadley cell ascent $\sim 0.01 \text{ Pa s}^{-1}$, giving $Pe \sim 10$. Convective momentum transport can be grossly approximated as a diffusive

transport with diffusivity $\nu_{\text{CMT}} = M/\rho\varepsilon$, where M is the convective mass transport, ρ is density, and ε is the fractional entrainment rate (Roms 2014). Typical observed values of M and ε give $\nu_{\text{CMT}} \sim 10 \text{ m s}^{-2}$. However, CMT is strong only in regions of zonally anomalous ascent and weak or suppressed in regions of anomalous descent. Further, in the case of the MJO, ascent occurs in a region of upper-level easterly anomalies (see Fig. 7) and therefore acts on a weaker shear than the zonal-mean value. As a result, CMT only moderately increases the zonal-mean viscosity (for the GCM considered here, which includes a CMT parameterization, we estimated an effective zonal-mean viscosity of $\sim 1 \text{ m s}^{-2}$, increasing to $\sim 5 \text{ m s}^{-2}$ when CMT was increased by a factor of 8; see section 4). As the climate warms, convective mass transport is expected to decrease (Held and Soden 2006), further diminishing the role of CMT. Increasing entrainment rates could compensate for the reduction in convective mass flux, but the response of entrainment to warming remains highly uncertain. In addition to diffusion, CMT also provides an effective downward advection of the vertical shear (Mapes and Wu 2001; Roms 2014). This effect could be important for surface superrotation and would merit further study. In the broader planetary context, atmospheric viscosities are largely unconstrained, and planet-specific GCMs generally inherit parameter values from the Earth-oriented GCMs from which they are derived; the high Pe values on Earth thereby carry over to planetary simulations too.

Concerning question 3, we surmise based on our results in section 3c that a thermally indirect Hadley cell arises when upper-tropospheric superrotation assumes the form of a very strong jet centered on the equator, requiring lateral energy convergence onto the equator to produce the warm temperature anomaly demanded by thermal wind balance. This in turn requires very strong equatorial eddy momentum convergence and a weak role for mechanisms that mitigate upper-tropospheric superrotation, such as seasonality (Mitchell et al. 2014)—which is absent in our simulations and in those of Schneider and Liu (Schneider and Liu 2009; Liu and Schneider 2010, 2011). It appears that this combination is hard to achieve in simulations configured to have realistic annual cycles and meridional temperature gradients.

Coming finally to question 4, we note that Rossby waves generated in the extratropics—that is, outside of the Hadley cell domain—and propagating toward the tropics will necessarily transport momentum poleward, precluding scenario 4, so long as the meridional gradient of zonal-mean potential vorticity is positive (e.g., Vallis 2006, chapter 12). It may be possible to construct an atmosphere with a reversed potential vorticity gradient, in

which case extratropical eddies would converge momentum into the tropics, but the possibility seems exotic.

An important motivation for the present work was the suggestion (Pierrehumbert 2000; Tziperman and Farrell 2009) that surface superrotation could arise on Earth in a warmer climate. In this regard, we can say that the presence of even a weak Hadley cell will generate upper-level westerlies and low-level easterlies that will robustly confine eddy momentum convergence to the upper troposphere. With observed Earthlike values of clear-air viscosity and Hadley cell strength, the Péclet number is an order of magnitude too large to yield surface superrotation, and convective momentum transport—as modeled in the GCM studied here—does not raise the viscosity above background clear-air values sufficiently to close the gap. On the basis of these results, a transition to surface superrotation appears unlikely in past or future warm climates. This conclusion should be tempered with a recognition of the large uncertainties in CMT parameterizations; surprises may be in store as understanding of this process develops and global-scale convection-resolving models become more established.

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