Role of Continents in Driving the Hadley Cells

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ABSTRACT

GCM simulations are used to investigate how forcing that originates over land surfaces influences the Hadley circulation. The presence of continental surfaces is found to approximately double the intensity of the winter hemisphere Hadley cell and to halve the intensity of the summer cell. The strengthening of the winter cell occurs because the increase in surface friction associated with land enhances the angular momentum flux into the atmosphere. The development of strong monsoon circulations in the Northern Hemisphere summer and the convergence zones of the Southern Hemisphere (South Pacific, South Atlantic, and South Indian convergence zones) shifts mass out of the subtropics, lowers the zonal mean subtropical highs, and weakens the summer cells. The responses of the summer and winter cells are different signs and occur by different processes, because heating of the land surfaces in the summer is effectively communicated through the depth of the atmosphere, whereas cooling in the winter is not. Also, the higher surface wind speeds and shears of the winter hemisphere trade winds make the winter cell more sensitive to surface friction than the summer cell. These results suggest that the axisymmetric models that have provided a theoretical basis for the understanding of the Hadley circulation do not capture some of the important physical processes that determine the intensity of the mean meridional circulation.

1. Introduction

A fundamental understanding of the mean meridional circulation (MMC) is often founded on the results of axisymmetric models. In these models, the meridional flow is driven by zonally uniform heating provided by relaxing the atmospheric temperature to a specified zonally uniform thermal equilibrium temperature, with smaller-scale mixing parameterized by diffusing heat vertically.

A series of papers by Schneider and Lindzen (1977), Schneider (1977), Held and Hou (1980), Lindzen and Hou (1988), and Hou and Lindzen (1992) uses axisymmetric models to develop Hadley circulation theory, including a consideration of the seasonal behavior of the Hadley cells. Explanations of the intensity, latitudinal boundaries, and heat and momentum transport by the Hadley circulation are suggested by assuming that flow in the upper branch of the Hadley circulation conserves angular momentum, the zonal wind at the surface is nearly zero, and the extratropical atmosphere is in radiative equilibrium. These studies are aimed at understanding the portion of the large-scale flow that is driven by zonally uniform heating. The deliberate exclusion of longitudinal structure, including baroclinic and barotropic eddies, relates the circulation back to its most basic cause, namely, the latitude-dependent radiative forcing of the atmosphere.

Subsequent developments build on the understanding from this series of papers. Fang and Tung (1997) find that the strength of the Hadley circulation in an axisymmetric model is very sensitive to the value of the thermal relaxation time assumed. Other extensions of the basic theory address the influence of midlatitude baroclinic waves on a Hadley circulation driven by zonally uniform tropical heating (e.g., Becker et al. 1997; Kim and Lee 2001) or the influence of the steady-state assumption of the solutions (Fang and Tung 1999).

In this paper, a different approach is taken in adding to the first-order understanding of the Hadley circulation derived from axisymmetric models. Noting that the tropical heating has pronounced longitudinal structure closely tied to the land–sea distribution, this paper compares the physics of the Hadley circulation in the presence and absence of continents. A fully nonlinear general circulation model (GCM) is used, and so baroclinic instability and feedbacks between the atmospheric dynamics and diabatic heating fields are included and land surface temperature responds to the balance of radiative, sensible, and latent heating. The role of continents in determining the structure and intensity of the time-mean Hadley circulation is investigated; a companion paper (Cook et al. 2003, manuscript submitted to J. Atmos. Sci., hereafter CGB) will explore the seasonal behavior...
of the circulation and compare the GCM results more directly with the axisymmetric models.

The GCM simulations and climatologies are described in the following section. In section 3, a pair of simulations is analyzed to evaluate the most fundamental role of continents in the physics of the Hadley circulation by comparing an all-ocean simulation with one that includes idealized (featureless) land surfaces. Section 4 discusses the difference that adding land surface features, such as topography, makes in the simulation of the MMC. Section 5 contains conclusions and implications, including a discussion of how these results relate to the axisymmetric model results.

2. Model simulations

Three simulations with an R30, 14-level atmospheric GCM are discussed: the GCM is a version of the National Oceanic and Atmospheric Administration/Geophysical Fluid Dynamics Laboratory Climate Dynamics Group’s model. Physical parameterizations are simple, with the “bucket” hydrology at the surface and moist convective adjustment. The complete, nonlinear, and time-dependent atmospheric dynamics and thermodynamics are included, but various assumptions about the surface boundary conditions are made.

Because of its importance in the analysis of the Hadley circulation in the following section, the vertical transport of momentum from the surface and through the lowest levels of the model atmosphere is reviewed. The \( u \)- and \( v \)-momentum (zonal and meridional, respectively) fluxes from the ground to the atmosphere are modeled using the bulk aerodynamics formulation, with

\[
\tau_z = -\rho C_D Vu \quad \text{and} \quad \tau_y = -\rho C_D Vu
\]

respectively. In Eq. (1), \( \rho \) is air density and \( V, u, \) and \( v \) are the total wind speed, the zonal wind speed, and the meridional wind speed, respectively, all evaluated at the lowest model level (\( \sigma = 0.997 \)). The aerodynamic drag coefficient \( C_D \) is set to 0.001 over ocean and 0.003 over land to capture the more well-developed boundary layer that can develop over land.

Momentum diffuses upward into the atmosphere, communicating the presence of the surface up to the 777-hPa level in the model. The acceleration of the flow by this process is given by

\[
\frac{\partial u}{\partial t}_{\text{friction}} = F_u = \frac{g}{\rho*} \frac{\partial}{\partial \sigma} \left( -\frac{\rho^* g}{\rho*} K_v \frac{\partial u}{\partial \sigma} \right)
\]

(2)

and

\[
\frac{\partial v}{\partial t}_{\text{friction}} = F_v = \frac{g}{\rho*} \frac{\partial}{\partial \sigma} \left( -\frac{\rho^* g}{\rho*} K_u \frac{\partial v}{\partial \sigma} \right), \quad (3)
\]

with \( K_v = \frac{Fr}{\rho g/p^* \partial \ln \sigma} \). Here, \( l \) is a height-dependent mixing length that has a maximum value of 30 m at 75-m height, \( p^* \) is surface pressure, and \( \sigma = p/p^* \) is the vertical coordinate of the GCM. As a lower boundary condition, \( \tau_z \) and \( \tau_y \) [Eq. (1)] replace the quantity in parentheses in Eqs. (2) and (3).

For each simulation, the model is initialized from a dry isothermal atmosphere at rest; the spinup period is January through June of the first year of the integration. The simulations are of different length because models with more complex climates— for example, due to having more structure introduced at the lower boundary— require longer averaging times to capture the climatology.

The “OCEAN” simulation is an 8-yr-long all-ocean integration with prescribed zonally uniform SSTs from Shea et al. (1990). These SSTs are denoted by the solid lines in Figs. 1a,b for January and July, respectively. Note that the observed SSTs are not simple cosine functions of latitude, as they would be if they closely reflected the structure of the solar forcing. The SST distribution is nearly flat across the equator in January, with slight maxima near 8°N and 8°S. In July (Fig. 1b), the SST distribution is less symmetric about the equator, with a single maximum of about 301 K near 10°N. The structure of the observed zonal mean SST is influenced by ocean boundary currents and upwelling/downwelling processes. Thus, although continents are not explicitly included in this simulation, the imposed zonally averaged SSTs reflect some influence of continents and have a structure that is significantly different from the solar forcing at the top of the atmosphere.

“CONT” is a 12-yr simulation that has the same zonally uniform SSTs as OCEAN but also includes continents. These continents are idealized land surfaces, with no topography, uniform albedo of 0.1, and uniform soil moisture of 5 cm (1/3 of the maximum possible value). The purpose of this simulation is to isolate the most fundamental influences of continents on the MMC. Realistic asymmetry is introduced into the heating fields in the free atmosphere (i.e., condensation) and at the lower boundary and also into the momentum flux from the surface. Unlike the ocean surfaces with prescribed SSTs, land surface temperatures are free to change as a result of the surface heat balance in CONT. Surface friction is increased over land [see Eq. (1)] to account for the more vigorous boundary layers that can develop over the rougher land surface.

The dashed lines in Figs. 1a,b are January and July zonally averaged surface temperatures from CONT, respectively. As compared with OCEAN (solid line), January surface temperatures in CONT have a single maximum instead of a double maximum, are less flat in the deep Tropics, and are more symmetric about the sub-solar latitude. Surface temperatures at low latitudes in the summer hemisphere are significantly higher than in OCEAN. The meridional surface temperature gradient throughout the winter hemisphere and the summer hemisphere subtropics is steeper in CONT than in OCEAN. In July (Fig. 1b), the summer (Northern) hemisphere zonal SST is quite flat from about 5° to 30°N and is up to 3 K warmer than the zonal SSTs. The winter (South-
Fig. 1. Zonal-mean surface temperatures (K) for OCEAN (solid line) and CONT (dashed line) for (a) Jan and (b) Jul. Differences in surface temperature between CONT and OCEAN for (c) Jan and (d) Jul. Contour interval is 3 K, and negative values are indicated by dashed contour lines; the zero contour is not shown.
ern) hemisphere gradient is a little steeper and up to 2 K cooler.

Figures 1c and 1d show maps of the surface temperature differences between CONT and OCEAN for January and July, respectively. As expected, the land surface is warmer than the ocean in the summer hemispheres and cooler in the winter hemispheres. The magnitude of the winter surface cooling is much larger (about 2 times) that of the summer warming in subtropical and middle latitudes. One reason for this is the fourth-power dependence of the surface longwave emission on temperature. Another factor is the temperature dependence of the Clausius–Clapeyron equation, which makes evaporation more important in the surface heat budget in the summer than in the winter.

Figure 2 shows zonal mean precipitation and maps of precipitation differences for CONT and OCEAN in the same format as Fig. 1. The zonal mean tropical precipitation maximum in CONT (dashed lines in Figs. 2a,b) is larger than in OCEAN for both January and July (solid lines in Figs. 2a,b), and it is shifted a little deeper into the summer hemisphere. Precipitation rates are larger in the summer hemispheres in CONT, even in middle latitudes. In the winter hemispheres, precipitation rates are smaller in CONT than in OCEAN within the Hadley regime and are about the same in middle latitudes.

The difference maps (Figs. 2c,d) show the pronounced zonal asymmetry introduced by the presence of land. In the summer hemispheres especially, monsoons over Africa, Asia, and South America are associated with precipitation rates that are factors of 2–3 larger than the zonal average. Over parts of the oceans, rainfall is lower in CONT than in OCEAN and is one-half of the zonal mean rate in some regions. (Note that the Asian monsoon is very weak in the CONT simulation, which excludes topography. As discussed in section 4, the Asian monsoon is captured more realistically in a simulation with topography.)

A third GCM integration, with realistic, prescribed surface conditions, is discussed in section 4. The dotted lines in Figs. 1a,b and Figs. 2a,b show zonal mean surface temperature and precipitation from this simulation.

3. Role of continental surfaces

The mass transported within the MMC is measured by the Stokes streamfunction \( \psi \), defined as

\[
\psi(\phi, p) = \frac{2\pi a \cos \phi}{g} \int_p^{p_{top}} [\nu] \, dp,
\]

where \( \phi \) is latitude, \( a \) is the earth’s radius, \([\nu]\) is the time-averaged zonal-mean meridional wind, \( p_s \) is the surface air pressure, and \( p_{top} \) is the pressure at the top of the atmosphere.

Figures 3a and 3b show the monthly mean Hadley circulations in OCEAN for January and July, respectively. Two Hadley cells are evident, a winter hemisphere cell and a weaker one in the summer hemisphere. This is, of course, the basic observed structure of the low-latitude MMC, but the circulation features in OCEAN have different magnitudes and are more symmetric about the equator than the observed. Figures 3c and 3d show the low-latitude Stokes streamfunction from the National Centers for Environmental Prediction (NCEP) 40-yr climatology (Kalnay et al. 1996), and Trenberth et al. (2000) show the MMC for both the NCEP and European Centre for Medium-Range Weather Forecasts (ECMWF) reanalyses (Figs. 10 and 11 in that paper). (The NCEP climatology shown in Trenberth et al. is a slightly different version of the climatology plotted in Figs. 3c,d.)

As in OCEAN, the Southern Hemisphere winter cell is stronger than the Northern Hemisphere winter cell in both the NCEP and ECMWF reanalyses. The winter Hadley cells simulated in OCEAN are less intense than in any of the reanalysis products, however. For example, the Northern Hemisphere winter cell in OCEAN has a maximum mass transport of about \( 10^{10} \) kg s\(^{-1} \), as compared with \( 12 \times 10^{10} \) kg s\(^{-1} \) in the 40-yr reanalysis (Fig. 3c) and \( 15 \times 10^{10} \) kg s\(^{-1} \) in the ECMWF and NCEP reanalysis products shown in Trenberth et al. (2000). At \( 12 \times 10^{10} \) kg s\(^{-1} \), the Southern Hemisphere winter cell in OCEAN is similar in magnitude to that in the NCEP reanalysis (Fig. 3d) but is significantly weaker than in the reanalyses shown in Trenberth et al. (2000). In contrast to the winter cells, the summer cells simulated in OCEAN are much stronger than in any of the reanalysis products.

Another structural difference between the Hadley cells in OCEAN and the reanalysis products is the location of the Hadley-cell division. In OCEAN, the zero mass flux line in essentially on the equator in January and near 5°N in July. In the reanalysis products, the cell divisions are at about 12°S in January and 18°–20°N in July.

Agreement between the model and reanalyses improves greatly when the simple land surfaces of the CONT simulation are included. The Northern Hemisphere winter cell in CONT (Fig. 3e) has a maximum of \( 16 \times 10^{10} \) kg s\(^{-1} \), very similar to the reanalysis products discussed above. The Southern Hemisphere winter cell (Fig. 3f) has the same magnitude as in the NCEP and ECMWF reanalyses shown by Trenberth et al. (2000), which are stronger than in the 40-yr reanalysis shown in Fig. 3d. The summer cells in CONT are appropriately weak, and the Hadley cell division lines are positioned away from the equator in CONT as in the reanalyses.

In summary, forcing over land surfaces strengthens the winter Hadley cells and weakens the summer cells (Greene 1998). These differences represent the influence of continental forcing in its most basic form, without topography and other continental features. Explaining why these differences occur is the main topic of this paper.
FIG. 2. Zonal-mean precipitation rates (mm day$^{-1}$) for OCEAN (solid line) and CONT (dashed line) for (a) Jan and (b) Jul. The dotted line indicates precipitation from the realistic GCM simulation, discussed in section 4. Differences in precipitation between CONT and OCEAN for (c) Jan and (d) Jul. Contour interval is 3 mm day$^{-1}$, and negative values are indicated by dashed contour lines.
Fig. 3. Monthly mean Stokes streamfunction from the OCEAN climatology for (a) Jan and (b) Jul, the NCEP reanalysis climatology for (c) Jan and (d) Jul, and the CONT simulation for (e) Jan and (f) Jul. Contour intervals are $2 \times 10^{10}$ kg s$^{-1}$. 
The zonal mean wind is examined to provide further model validation and to support the momentum budget analysis below. Figures 4a–c show the July zonal mean meridional wind from OCEAN, the 40-yr NCEP reanalysis, and CONT, respectively. Consistent with the Stokes streamfunction (Fig. 3), the low-level equatorward flow is stronger in the winter (Southern) hemisphere, as is the poleward flow in the upper tropical troposphere, in all three cases. When compared with the NCEP reanalysis (Fig. 4b), the meridional flow in OCEAN (Fig. 4a) is a little too strong at low levels in the summer hemisphere, too weak aloft, and too symmetric about the equator. When land surfaces are included in the CONT simulation, the upper-level wind speed increases realistically but the low-level flow is still too strong.

Since the Coriolis force couples the two horizontal directions of motion, the Hadley circulation is also reflected in the zonal-mean zonal wind field. Figures 4d–f show the zonal-mean zonal wind climatology for July from OCEAN, NCEP, and CONT, respectively. In OCEAN, the westerly jet in the winter hemisphere (Fig. 4d) is about 15 m s−1 weaker than in the reanalysis (Fig. 4e) and is located 10° farther south. The Southern Hemisphere summer jet (not shown) is about 5 m s−1 weaker than in the NCEP reanalysis, and the Northern Hemisphere summer jet in OCEAN (Fig. 4d) is about 5 m s−1 stronger than in NCEP (Fig. 4e). The tropical easterlies in OCEAN are stronger than in the reanalysis in both hemispheres. Adding land surfaces improves the simulation of the zonal wind. The winter hemisphere jet strengthens and shift equatorward, and the summer hemisphere trades weaken (Fig. 4f). The winter hemisphere trade winds, however, remain too strong.

The fact that the Hadley circulation becomes much more realistic when the idealized continental surfaces of the CONT simulation are included reinforces the idea that forcing over land is of central importance for establishing the MMC. Differences in the wind components are generally consistent with differences in the Hadley cells, especially for the meridional component by definition [Eq. (4)], but there are discrepancies. For example, while the July winter cell increases in intensity by 75% when land surfaces are added in the model (Figs. 3b,d), the Southern Hemisphere tropical easterlies at low levels do not change by very much (Figs. 4d,f).

The u- and v-momentum equations provide insight into the processes that cause the large-scale circulation to be different in the two simulations. The goal is to relate the simulated differences in the circulation summarized in Figs. 3 and 4 to the surface boundary condition differences in OCEAN and CONT.

For the zonal mean, denoted by square brackets, and long-time mean, denoted by overbars, the u-momentum equation is

\[ \frac{\partial [\mathbf{u}]}{\partial t} = 0 = -f[\mathbf{u}] - \frac{\partial [\mathbf{v} \cdot \mathbf{u}]}{\partial y} + [\mathbf{F}_u]. \]  

Here, \( \mathbf{v} \) is wind velocity, \( f \) is the Coriolis parameter, and \( \phi \) is now geopotential height. Accelerations \( \mathbf{F}_u \) and \( \mathbf{F}_v \) are defined in Eqs. (2) and (3).

Figures 5a and 5b provide an overview of the zonally averaged \( u \)-momentum balance [Eq. (5)] within the Hadley regime in the OCEAN climatology for January and July at 935 hPa, which is representative of the lower troposphere. The nonlinear advection term \( -[\mathbf{v} \cdot \nabla \mathbf{u}] \) is calculated as a residual. The east–west zonal-mean momentum balance is primarily between the Coriolis and friction terms. Between ±30°N latitude, Coriolis forces associated with the equatorward return flow of the Hadley circulation accelerate the low-level air to the west to conserve absolute angular momentum. This easterly acceleration is balanced by westerly acceleration from friction, as the rotating earth injects positive (eastward) momentum into the low-level atmosphere. Very close to the equator and near ±30°, where Coriolis accelerations vanish, advection of zonal momentum balances small dissipation. The momentum balance is less symmetric about the equator in July than in January, consistent with the circulation and SSTs being less symmetric (Figs. 1, 3, and 4).

Figures 5c and 5d show the lower-troposphere u-momentum balances in CONT for January and July, respectively. In the summer hemispheres, both Coriolis and dissipation terms are 30%–50% smaller than in OCEAN (Figs. 5a,b). In the Northern Hemisphere summer (Fig. 5d), both Coriolis and frictional accelerations become very small in the northern half of the Hadley regime (~15°–30°N) in CONT, consistent with the very weak equatorward flow seen in Fig. 4c. Between about 5° and 15°N, the sign change in the meridional flow (cf. Figs. 4a with 4c) changes the sign of the Coriolis term. Advection plays an important role in this region, where \( u \) (and, therefore, \( F_u \)) is small and \( v \) (and, therefore, the Coriolis force) is relatively large.

Differences in the low-level u-momentum balance between OCEAN and CONT are smaller in the winter hemispheres than in the summer ones. Both Coriolis and frictional forces are only about 15%–25% larger in the Tropics and sub-tropics in CONT when compared with OCEAN, despite the large differences in the Hadley cells (Fig. 3).

Away from the surface, friction is small and the westerly (positive) Coriolis acceleration of the poleward flow in the upper branch of the Hadley circulation is balanced by the advection of easterly (negative) u momentum. Figures 6a and 6b show these dominant components of the upper-level angular momentum balance for January and July, respectively, at 257 hPa for the all-ocean simulation (OCEAN). The advection term \( -[\mathbf{v} \cdot \nabla \mathbf{u}] \), indicated by the thick dashed line, is essentially equal and opposite to the Coriolis term (solid line) at all latitudes. It produces easterly acceleration in the Hadley regime.
Fig. 4. Zonal-mean meridional wind in Jul from (a) OCEAN, (b) the NCEP reanalysis, and (c) CONT. Contour intervals are 1 m s$^{-1}$. Zonal-mean zonal wind in Jul from (d) OCEAN, (e) the NCEP reanalysis, and (f) CONT. Contour intervals are 5 m s$^{-1}$. 
and westerly acceleration in middle latitudes, balancing the westerly Coriolis acceleration of the low-latitude poleward flow and the easterly Coriolis acceleration of the equatorward (Ferrel) circulation. Many authors (e.g., Edmon et al. 1980; Pfeffer 1981; Becker et al. 1997) discuss the important role of eddy momentum transport in determining the MMC.

The velocity components can be expressed as the sum of zonal-mean and stationary-eddy components, and time-mean and transient-eddy components. For example,

\[ u = [\mathbf{\pi}] + [u'] + \pi^* + u^*, \]

where the square brackets indicate the zonal mean, the asterisk indicates the deviation from the zonal mean, the overbar indicates the time mean, and the prime indicates the deviation from the time mean. Then, the advection term in Eq. (5) expands to

\[ -[\mathbf{\nabla} \cdot \mathbf{u}] = -[\mathbf{\nabla}] \frac{\partial [\mathbf{\pi}]}{\partial y} - [\mathbf{\omega}] \frac{\partial [\mathbf{\pi}]}{\partial \rho} + E, \]

where

\[ E = -[\mathbf{\nabla} \cdot \mathbf{\nabla} \pi^*] - [\mathbf{\nabla} \cdot \mathbf{\nabla} u^*] \]

\[ -[\nu'] \frac{\partial [u']}{\partial y} - [\omega'] \frac{\partial [u']}{\partial \rho} \]

(9)

expresses the zonal acceleration of the flow by stationary- and transient-eddy activity.

In OCEAN, the stationary-eddy terms are negligible \((u^* = v^* = \omega^* = 0)\), so the advection term only includes the advection of zonal-mean \(u\) momentum by the zonal-mean wind [first two terms on the rhs of Eq. (8)] and advection by transient eddies [second two terms on the rhs of Eq. (9)]. Since the time-mean vertical-advection term is small,

\[ -[\mathbf{\nabla} \cdot \mathbf{\nabla} u] = -[\mathbf{\nabla}] \frac{\partial [\mathbf{\pi}]}{\partial y} - E', \]

(10)

where

\[ E' = -[\nu'] \frac{\partial [u']}{\partial y} - [\omega'] \frac{\partial [u']}{\partial \rho}. \]

(11)

In Figs. 6a,b, each of the two terms on the rhs of Eq.
Fig. 6. Terms ($10^{-5} \text{ s}^{-1}$) in the zonally averaged $u$-momentum balance [Eq. (5)] at 257 hPa for (a) OCEAN in Jan, (b) OCEAN in Jul, (c) CONT in Jan, and (d) CONT in Jul. Solid lines indicate the Coriolis term, and dashed lines indicate the advection term. The zonal-mean and transient-eddy components of the advection term are indicated by the thinner dotted and dotted–dashed lines, respectively.

(10) is shown for the OCEAN climatology (dotted and dot–dashed thin lines, respectively). In midlatitudes, nearly all of the momentum advection is accomplished by the transient eddies in both winter and summer. In the Hadley regime, transient transports tend to dominate but advection by the time-mean meridional wind is not negligible. In Southern Hemisphere winter (Fig. 6b), this term is as important as the transients term.

In CONT, the upper-level $u$-momentum balance is still between the Coriolis and advection terms, both of which are larger (smaller) than in OCEAN in the winter (summer) hemisphere (Figs. 6a,b). The differences in the advection term in both summer and winter are primarily due to differences in angular-momentum advection by the time-mean flow (dotted lines in Fig. 6). This term changes in proportion to the strength of the Hadley circulation itself, since the poleward flow in the upper branch of the MMC is not only subject to Coriolis acceleration but also advects easterly time-mean momentum out of the Tropics. The eddy-advection term, which includes stationary-eddies-momentum advection in addition to transient-eddy advection in CONT, is only slightly stronger than in OCEAN.

The zonally averaged $v$-momentum balance [Eq. (6)] includes a large meridional pressure gradient force in addition to friction, advection, and Coriolis forces, and so zonal wind speeds that are more than an order of magnitude greater than meridional velocities can be supported. At all levels, during every month, the primary balance is geostrophic. In OCEAN at 935 hPa, as seen in Figs. 7a,b, poleward Coriolis forces associated with easterly flow in the lower tropical troposphere balance equatorward height gradient forces that are directed into the equatorial trough from the subtropical highs. The nonlinear term is small and, unlike the low-level $u$-momentum balance (Figs. 5a,b), friction does not play a notable role. Friction is not as important in the $v$-momentum balance partly because meridional velocities are smaller than zonal velocities [Eqs. (2) and (3)] but mainly because the other meridional forces (Coriolis and height gradient) are so large.

The low-level $v$-momentum balance in CONT is
shown in Figs. 7c,d. In the summer hemispheres, both Coriolis and height gradient forces are weaker in CONT than in OCEAN, consistent with the weakening of the circulation. However, despite the near doubling of the winter hemisphere Hadley cell intensity because of the presence of land (Fig. 3), the low-level $v$-momentum balance in the winter hemispheres is not very different from the balance in OCEAN (Figs. 7a,b). Also, there is very little hemispheric asymmetry in the winter cells' $v$-momentum balance; that is, the January Northern Hemisphere components are essentially identical to the July Southern Hemisphere components, except for the expected sign change.

In summary, $u$-momentum dissipation is smaller in the summer hemispheres when land is present. This is consistent with having lower surface friction because of a weaker circulation, but not with an increase in friction caused by an increase in $C_D$ over land surfaces [Eq. (1)]. In the $v$-momentum balance, the magnitude of the meridional height gradient is smaller in CONT and, since the balance is essentially geostrophic, the zonal wind is smaller. These facts, taken together, indicate that the cause of the weakening of the summer hemisphere cells when land is present is the weakening of the meridional height gradient. The response in the $u$-momentum dissipation term is an effect of the reduced strength of the circulation, and not the cause, although there may be a feedback from the reduction in friction that weakens the circulation more than the height-gradient differences alone would.

In contrast, the doubling of the winter hemisphere circulation intensity is not associated with a comparable increase in the meridional height gradient, and the $v$-momentum balance is not very different in CONT and OCEAN. Instead, the influence of land on the winter Hadley cells comes through the stronger $u$-momentum dissipation term at low levels and the stronger angular-momentum advection by eddies in the upper troposphere. Again, the possibility of a feedback exists if increases in $C_D$ over land enhance $F_x$ and, in turn, the low-level wind speed.

The next step in understanding how the changed surface conditions modify the Hadley circulation is to relate the results from the momentum budget analysis to the
lower boundary conditions associated with land. Understanding the connection between the modified surface boundary conditions and the reduced meridional height gradient in the lower troposphere will explain why the summer cells are weaker. For the winter cells, the processes that cause frictional dissipation to be modified need to be understood. In addition, the differences between the responses in the winter and summer hemispheres must be addressed. Why is the physics of the Hadley cell responses different in the winter and the summer? Why is it that the winter hemisphere meridional height gradient is not changed very much by the presence of land, especially since the surface temperature differences are larger over land in summer than in winter (Fig. 1)? Why is the increase in surface drag associated with land effective for enhancing the winter cell but not the summer cell?

a. Summer hemisphere Hadley cell response

Figures 8a and 8b show zonal-mean 935-hPa heights for OCEAN (solid line) and CONT (dashed line). Differences in the height fields for CONT minus OCEAN for January and July are plotted in Figs. 8c and 8d, respectively. The weakening of the zonal-mean meridional height gradients that the $\tau$-momentum budget analysis associates with the weakening of the summer hemisphere cells is seen here to be due to a lowering of zonal-mean heights in the subtropics (Figs. 8a,b). During Southern Hemisphere summer (Fig. 8c), the lowering of the subtropical highs is fairly zonally uniform. Heights are lowered by up to 80 gpm from 80°W (South America) to 160°E (east of Australia). The only east–west compensation for the lower heights occurs over the eastern Pacific, but it is minimal. The subtropical height decrease is less zonally uniform during Northern Hemisphere summer (Fig. 8d). Here, more distinct lows develop over North America, northern Africa, and Southeast Asia, and heights increase over most of the Atlantic and Pacific Oceans.

Differences in the geopotential height field can be related to the surface boundary conditions by examining temperature and surface pressure since

$$\phi(p) = -R \int_{p_E}^{p} T \, d \ln p,$$

(12)

where $R$ is the gas constant. Differences in 935-hPa heights (Fig. 8) are virtually identical in structure to differences in the surface pressure field (not shown). In particular, the lowering of zonal-mean geopotential heights in the subtropics during summer and the lack of a similar response in winter are reflected in the surface pressure fields. By inspection, it is clear that the differences in heights and surface pressure are related to the surface temperature differences over land seen in Figs. 1c,d, but changes in surface pressure and heights over the oceans (where the surface temperature is prescribed) are also part of the response. This means that the low-level dynamical response to the thermal lows is an important determinant of the Hadley circulation intensity in summer. This response is somewhat different in the Northern and Southern Hemispheres because of differences in the land–sea distributions.

In the Northern Hemisphere, the lowering of zonal-mean surface pressures and heights in the subtropics in summer is due to the development of lows that are mainly confined to land (Fig. 8d). These lows, of course, mark the North American, African, and Asian monsoon systems. The degree to which the zonal-mean heights are lowered, and the Hadley circulation is weakened, depends on the many factors that determine the strength of each monsoon. Thus, the monsoon dynamics are coupled to the summer Hadley circulation dynamics through controls on the magnitude of the subtropical highs in the Northern Hemisphere.

In Southern Hemisphere summer, the thermal lows are not closely confined to land surfaces, and they extend off the continents poleward and to the east (Fig. 8c). The diagonal extension of the lows over the ocean is strong in the Southern Hemisphere because of the continental configuration, and the result is the well-formed South Pacific, South Atlantic, and South Indian convergence zones (SPCZ, SACZ, and SICZ, respectively; Kodama 1993; Cook 2000). Even though there is less land in the Southern Hemisphere, the zonal mean surface pressure and height differences are just as strong in the Southern Hemisphere as in the Northern Hemisphere. Because of the diagonal extension of the land-based convergence zones in the Southern Hemisphere, the anomalous low pressure extends deeper into the mid-latitudes in the Southern Hemisphere.

High pressures and heights over the oceans are reinforced when the thermal lows form over the continents in summer, especially in the Northern Hemisphere, but by too little to offset the lowering of pressure over the continents in the zonal average (Fig. 8). In other words, the shifting of mass that occurs when continents are introduced is not confined to the east–west plane, and mass flows meridionally out of the subtropics. The resulting weakening of the zonal-mean subtropical highs, with the associated reduction in the meridional height gradient, is the cause of the weakening of the summer Hadley cells.

According to the momentum budget analysis, the lowering of the summer hemisphere subtropical highs and the associated reduction in the meridional height gradient are communicated into the Hadley circulation through the zonal wind field. Differences in the zonal wind, and in the full-wind vectors, between CONT and OCEAN are plotted at 935 hPa in Figs. 9a,b for January and July, respectively. In the Southern Hemisphere summer (Fig. 9a), the monsoon flow into the thermal lows over South America, Africa, and, to a lesser extent, Australia accelerates the easterlies close to the equator to the northeast of the land. A stronger, westerly per-
Fig. 8. Zonal-mean heights (gpm) for OCEAN (solid line) and CONT (dashed line) for (a) Jan and (b) Jul. Differences in the 935-hPa geopotential height field between CONT and OCEAN for (c) Jan and (d) Jul. Contour interval is 20 gpm.
Fig. 9. Contours of differences in the 935-hPa zonal wind speed between CONT and OCEAN for (a) Jan and (b) Jul. Contour interval is 2 m s\(^{-1}\). Vectors show the full wind differences (CONT - OCEAN) at 935 hPa, and the vector scale is indicated for 10 m s\(^{-1}\).

b. Winter hemisphere Hadley cell response

Despite the fact that the cooling over land surfaces in winter is generally larger in magnitude than the warming over land surfaces in summer (Fig. 1), zonal-mean lower troposphere heights (Figs. 8a,b) and surface pressures do not change much in the winter hemisphere Tropics and subtropics. This occurs because the cooling at the surface in winter is not communicated into the lower troposphere very effectively. Lower-tropospheric temperature profiles for OCEAN (solid lines) and CONT (dashed lines) at 15\(^\circ\)N for January and July are plotted in Figs. 10a and 10b, respectively; they are representative of the response throughout the Tropics and subtropics. The cooler temperatures in CONT during January are confined below about 750 hPa, and there is warming above this level (Fig. 10a). In contrast, the warmer temperatures in CONT during July (Fig. 10b) are communicated through the depth of the troposphere. Thus, the vertically integrated temperature of the atmospheric column changes very little in the winter hemisphere, and there is a correspondingly small change in the subtropical surface pressure and height fields (Fig. 8). In the summer hemisphere, the surface warming is smaller in magnitude than the winter cooling because the warming is spread through the entire atmospheric column, primarily by latent heat releases. The result is a warmer, less dense atmospheric column in the summer, and strong responses in the surface pressure and geopotential height fields.
Poleward of 30° in the winter hemispheres, the atmosphere’s response to the cooling of land surfaces is wavy. This is a Rossby wave response that is expected in the presence of strong zonal flow, and it redistributes mass mostly in the east–west direction. Thus, the strong land surface cooling in midlatitudes does not redistribute mass meridionally.

The momentum budget analysis revealed that the more intense winter hemisphere circulations in CONT are due to increases in the vertical diffusion of $u$ momentum. Such an increase could have two sources. First, there could be an increase in the $u$-momentum flux $\tau$, from the surface into the lowest level of the atmosphere, defined in Eq. (1). Second, the wind shear in the lower troposphere could be modified, which would affect the vertical diffusion of angular momentum according to Eq. (2).

The zonal mean of $\tau$ is plotted in Fig. 11a for January and shows an approximate doubling in the winter hemisphere Tropics and subtropics. The same pattern occurs in July (not shown). Figure 11b shows the zonally averaged $C_D$ from CONT; the value is 0.001 at all latitudes in OCEAN. The low-wind speed $V$ and the zonal wind $u$ are very similar in OCEAN and CONT, as seen in Fig. 11c (and also Fig. 4), but the product of the wind and drag coefficient (as represented by the product $C_D u$ in Fig. 11d) is larger in the winter hemisphere subtropics, and this makes $\tau$ larger. Thus, the enhancement of $\tau$ is due to the increase in the surface drag coefficient over land surfaces.

Note in Fig. 11a that $\tau$ is smaller in the summer hemisphere in CONT. This is a result of the lower wind speeds (Fig. 11c) despite larger $C_D$ values (Fig. 11b). This supports the conclusion from the momentum budget analysis that the changes in $\tau$ in the summer hemisphere are an effect of the Hadley circulation modification and not a cause as in the winter hemisphere.

The second possible cause of the enhanced $u$-momentum fluxes is more vigorous vertical mixing of $u$ momentum in association with stronger wind shear [Eq. (2)]. Figure 12 shows vertical profiles of the zonal-mean zonal wind in OCEAN and CONT at 15°N in January and July; these regions are representative of other latitudes and times of year. In the winter hemisphere subtropics (Fig. 12a), adding continents increases the zonal wind shear below the 935-hPa level by about 20%, and this enhances the vertical transport of $u$ momentum. In the summer hemisphere (Fig. 12b), the wind shear is greatly reduced when land surfaces are present. These differences in shear are consistent with having a larger (smaller) $u$-momentum flux from the surface in the winter (summer).
4. Role of continental features

An additional GCM integration is used to verify that the idealized surface in the CONT simulation provides an accurate assessment of the role of the continents in forcing the Hadley circulation. In this simulation, realistic two-dimensional SSTs (Shea et al. 1990), soil moisture (Mintz and Walker 1993), and land surface albedo (Matthews 1984) are all prescribed from observations. Topography is included and is filtered to reduce Gibbs oscillations. The integration was run for 20 years.

Zonal-mean surface temperatures for the realistic simulation are indicated by the dotted lines in Figs. 1a,b. Adding land surface features and longitudinal structure in the SSTs changes the zonal-mean temperature distribution in significant ways, comparable in magnitude to the differences between the CONT and OCEAN climatologies (the dashed and solid lines in Figs. 1a,b, respectively) at most latitudes. In January (Fig. 1a), zonal-mean surface temperatures in the Tropics and Southern Hemisphere subtropics are up to 2.5 K cooler than in CONT and they are more similar to the surface temperatures in OCEAN than in CONT. In July (Fig. 1b), the realistic simulation is cooler than either OCEAN or CONT in the Tropics and has a well-defined surface temperature maximum in the Northern Hemisphere near 25°N that is mainly caused by high temperatures over northern Africa. In the Northern Hemisphere during both January and July, cooler temperatures at 30°–40°N in the realistic simulation are due to the elevation of the Tibetan Plateau.

The addition of land surface features leads to a pronounced decrease in the tropical precipitation maximum in January when compared with CONT (dotted line in Fig. 2a). In July, despite a large increase in Asian precipitation when the Tibetan Plateau is present, the zonal-mean precipitation in the realistic simulation is very similar to the CONT precipitation (Fig. 2b). Even though precipitation rates in the Asian monsoon are high in the model (up to 24 mm day\(^{-1}\)), they are localized and offset by precipitation decreases at other longitudes (e.g., northern Africa) and so do not affect the zonal mean very much.

January and July Stokes streamfunctions from the realistic simulation are shown in Figs. 13a,b, respectively. Despite the relatively large differences in the zonal mean surface temperature and precipitation that accompany the added realism of the continental surfaces, there is no evidence that the addition of orographically induced stationary waves into the model climate has weakened the MMC, and the jet is actually a little stronger in the realistic simulation than in CONT (not shown). Several studies, including those of Held and Phillips (1990) and Becker and Schmitz (2001), suggest that potential vorticity mixing associated with orographically induced Rossby waves decelerates the upper-level flow and weakens the Hadley circulation. An examination of the upper-level \(u\)-momentum balance in the realistic simulation (not shown) shows that the total eddy advection \(E\), defined in Eq. (9)\), is about 25\% stronger than in CONT in Northern Hemisphere winter. However, the meridional gradient of the zonal-mean velocity decreases at the same time, as the westerly jet becomes narrower, and this decreases the time-mean zonal-mean advection [first term on the rhs of Eq. (10)] so that the full-advection term [Eq. (10)] does not change significantly. During winter in the Southern Hemisphere, differences in \(E\) associated with adding continental features are very small.

The most noticeable difference between the Hadley cells in the realistic-continent and idealized-continent simulations is the shape of the winter cells. In the realistic simulation, the January up branch slants to the
Fig. 12. Zonal-mean wind profiles at 15°N for (a) Jan and (b) Jul. Solid lines represent the OCEAN GCM climatology, and dashed lines show the CONT climatology.

Fig. 13. Monthly mean Stokes streamfunction from CONT2 for (a) Jan and (b) Jul. Contour intervals are $2 \times 10^{30}$ kg s$^{-1}$. 
south with decreasing pressure, being confined north of the equator in the lowest 200 hPa of the troposphere (Fig. 13a). The southward flow of the trade winds is weakened significantly over West Africa and the eastern Atlantic when realistic land surface conditions are specified, and this is largely responsible for the weakening of the low-level flow and Hadley circulation. With realistically dry soil moisture specified over northern Africa in January, the surface is 6–8 K warmer over the Guinean coast and the winter monsoon flow southward off the coast is weaker. This structure is also seen in the reanalyses (e.g., Fig. 3c).

5. Conclusions and implications

Much of our theoretical understanding of the Hadley circulation derives from axisymmetric model simulations forced by zonally uniform heating functions. Land–sea distributions modify the incoming zonally uniform radiative heating of the earth system, however, and the actual heating of the atmosphere is highly asymmetric. All-ocean, idealized-continent, and realistic-continent GCM simulations are compared to investigate how the Hadley cell dynamics are modified by the presence of continents.

Two GCM simulations are discussed in detail. One has an all-ocean surface, and the prescribed SSTs are observed zonal means. The other has the same SSTs, but flat, featureless continental surfaces are included. The winter hemisphere Hadley cells in the simulation with featureless continents are nearly 2 times as strong as in the all-ocean simulation, and the intensity of the summer cells is approximately halved.

Without further analysis, and just thinking about the lessons from axisymmetric theory, one might suppose that land surface warming in the summer hemisphere weakens the meridional temperature gradient and, thereby, the Hadley cell, and that land surface cooling in the winter hemisphere strengthens the meridional temperature gradient and the Hadley cell.

This is indeed what happens in the GCM to weaken the summer hemisphere cells when land surfaces are included. Land surface warming off the equator is spread through the mid- and lower troposphere by latent heating (precipitation) and is communicated into the meridional height field. The low-level dynamical response to continental thermal lows—that is, the development of the monsoon systems and the Southern Hemisphere convergence zones—determines the magnitude of the height-field response and, therefore, the summer Hadley cell response. Since the primary north–south momentum balance within most of the Hadley regime is geostrophic, the zonal wind weakness in response to the weakened meridional height gradient in the lower troposphere. The reduction in the zonal wind speed weakens the meridional flow when the vertical diffusion of $u$ momentum is reduced. This process is summarized schematically in Fig. 14a.

In contrast, the strengthening of the winter cells is unrelated to changes in meridional height gradients. The intensification of the winter cells is due to the increase in surface friction that accompanies the introduction of land surfaces in the model. In the climatological zonally averaged $u$-momentum equation, frictional dissipation balances the Coriolis acceleration. The meridional wind and the Hadley circulation are therefore strengthened when friction is increased. Figure 14b provides an outline of this mechanism.

The responses of the summer and winter cells are different for two main reasons. One is that cooling of the land surfaces in the winter hemispheres is not effectively communicated into the lower troposphere, whereas warming in the summer hemispheres is. Winter cooling increases vertical stability, confining the surface temperature perturbation to the near-surface atmosphere. Summer warming decreases stability, and enhanced precipitation causes the atmospheric column to warm. Thus, the height gradients in the 700–900-hPa layer are modified in the summer hemispheres but not in the winter hemispheres. The other reason is that the higher low-level wind speeds and vertical wind shears of the winter hemisphere trade winds make surface friction and dissipation relatively more important processes in supporting the winter cells. So, when the surface momentum flux changes, it has a large effect on the winter-cell intensity.

An additional simulation was conducted to ensure that the simulation with featureless land surfaces captures the essence of the influence of land on the Hadley circulation. This simulation has realistic conditions, including topography and soil moisture, prescribed on land and has realistic SSTs. Resulting differences in the Hadley circulation are minor.

This study suggests that axisymmetric models do not capture some of the important physical processes that determine Hadley circulation intensity. These models have been used to explain the seasonal behavior of the MMC and to evaluate the physical processes that determine the intensity and position of the Hadley cells. A companion paper (CGB) will describe findings that the seasonal behavior of the MMC in axisymmetric models is reproduced by the GCM when the GCM is forced with a zonally uniform heating distribution that mimics the heating function in those models. That is, in such a simulation, the Hadley circulation strength is correlated with the overall strength of the diabatic heating, with the concentration of the heating, and with the degree to which the heating is located off the equator, similar to the axisymmetric model results. However, it will be shown that the observed (reanalyzed) MMC does not behave in this way and that neither does the GCM when forced by observed zonally uniform SSTs (i.e., in OCEAN) or in model simulations with continentality included (e.g., in CONT). The physics of that response will be investigated using the results from this paper as a starting point.
Fig. 14. Schematic that summarizes the results of the momentum budget analysis concerning the processes associated with (a) the weakening of the summer cells and (b) the strengthening of the winter cells by the presence of land.

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