Ground-Truth Model Evaluation of Subgrid Orographic Base-Flux Parameterization

STEPHEN T. GARNER

NOAA/GFDL, Princeton, New Jersey

(Manuscript received 30 November 2017, in final form 30 July 2018)

ABSTRACT

High-resolution simulation can be a powerful means of evaluating and tuning orographic drag schemes, but connecting the parameterized drag, which is a local forcing, with the model drag, which is fundamentally global, is not entirely straightforward. The simplest idea is to filter the velocity down to its divergent component and exploit Bernoulli's law to define a local form drag. Using regional simulations over the Rockies, the Andes, and Greenland, we investigate the validity of this approach, which assumes that both the included nonorographic divergence and the missing orographic deformation will not significantly alter the diagnostic. The local drag is checked for consistency with the nonlocal drag at scales containing most of the gravity wave drag and blocking drag. The agreement is found to be satisfactory unless the drag is weak and nonlinear. In that case, we find it necessary to remove a steady pattern from the nonlocal drag in order to uncover a correlation. We test a specific mountain drag scheme using the proposed diagnostic and describe procedures for tuning the scheme's drag coefficients and treatment of anisotropy.

1. Introduction

The purpose of this study is to develop a high-resolution model-based evaluation strategy for orographic drag parameterizations. This type of parameterization has been used in global climate models for most of their history. Every model in the CMIP5 database employs a mountain drag parameterization of some kind. The necessity of subgrid mountain drag was demonstrated by [Palmer et al. \(1986\)](#). Afterward, quite a number of parameterizations were developed. A summary and review are provided by [Kim and Arakawa \(1995\)](#) and [Kim et al. \(2003\)](#). The schemes estimate the total column drag, called the base flux, and the vertical distribution of the drag. We should acknowledge that a vertical column is an approximation of the three-dimensional wave-activity path (e.g., [Preusse et al. 2008](#); [Marks and Eckermann 1995](#)) that becomes increasingly unrealistic with increasing depth. In this study we focus entirely on the base flux and leave the question of the vertical distribution for later work.

Parameterizations in general follow from fundamental physical and fluid dynamical principles, but they carry an assumption of scale separation in time and space that inevitably adds error and uncertainty. Minimizing the error and uncertainty is the motivation for strategies of

various kinds to test and constrain the algorithms. The strategies can be categorized as 1) comparisons of model climatological fields with large-scale observations, 2) comparisons of parameterized drag with observations at the same scale, and 3) comparisons of parameterized drag with high-resolution model simulations, primarily at the same scale.

Evaluating the impact of drag schemes on critical diagnostics of the climate (the first category) has been carried out with both global (e.g., [Boer and Lazare 1988](#); [Klinker and Sardeshmukh 1992](#)) and regional (e.g., [Kim and Arakawa 1995](#); [Zadra et al. 2003](#)) models. Compared to process-scale simulation or observation, this approach, especially in global models, can be a blunt tool that requires a great deal of faith in the attribution of model errors to specific parameters of the scheme. Direct process-scale observations (second category) have been used to constrain parameterizations of moist convection, microphysics, and boundary layer turbulence, often in conjunction with single-column simulations (e.g., [Ghan et al. 2000](#); [Lock et al. 2000](#)). For mountain drag, this approach is less satisfactory because direct observations of vertical momentum fluxes in the free troposphere and of form drag at the surface are less reliable than observations of boundary layer heat and momentum fluxes and precipitation, while satellite-based observations of stratospheric momentum fluxes are compromised by indirect

Corresponding author: Steve Garner, steve.garner@noaa.gov

DOI: 10.1175/JAS-D-17-0368.1

For information regarding reuse of this content and general copyright information, consult the [AMS Copyright Policy \(www.ametsoc.org/PUBSReuseLicenses\)](#).

detection and undersampling (Alexander et al. 2009; Alexander et al. 2010; Geller et al. 2013). In the present approach using high-resolution simulation (third category), the challenge is to extract from the ground-truth simulations the same type of local forcing that is estimated by subgrid drag schemes.

The momentum budget is the starting point for both developing and evaluating mountain-drag schemes. The zonal momentum budget is a volume integral of

$$\frac{\partial \rho u}{\partial t} = -\nabla \cdot \rho \mathbf{V} u - \frac{\partial p}{\partial x} + F_x, \quad (1)$$

which is the zonal component of the equation of motion, written with standard notation. The Coriolis term is excluded because it is routinely neglected in drag schemes. The volume integral can be written

$$\frac{d}{dt} \iiint \rho u dV = \oint dA (\rho u V_n + p x_n), \quad (2)$$

where V_n and x_n are the outward normal components of the velocity and zonal unit vectors, respectively. We have ignored the viscosity F_x . The integral on the rhs, taken over the surface of the volume, is obtained with Gauss's theorem. In pursuit of a diagnostic for the local drag, we ignore the net flux across lateral boundaries. Then,

$$\begin{aligned} \frac{d}{dt} \iiint \rho u dV = \int \int_{\text{Bottom}}^{\text{Top}} dA \left[\rho u w - \rho u v \frac{\partial h}{\partial y} \right. \\ \left. - (\rho u^2 + p) \frac{\partial h}{\partial x} \right], \end{aligned} \quad (3)$$

where the double integral is over the horizontal or quasi-horizontal surfaces at the top and bottom of the volume and $h(x, y)$ is the height of these surfaces. Thus, the upward momentum flux is

$$M_x^\uparrow = \rho u w - \rho u v \frac{\partial h}{\partial y} - (\rho u^2 + p) \frac{\partial h}{\partial x}. \quad (4)$$

On a strictly horizontal surface, where $\partial h / \partial x = \partial h / \partial y = 0$, we have

$$M_x^\uparrow = \rho u w, \quad (5)$$

which we will call the simple momentum flux. On a material surface, where $w = u \partial h / \partial x + v \partial h / \partial y$, we have

$$M_x^\uparrow = -p(\partial h / \partial x), \quad (6)$$

which is the form drag.

In an inviscid steady state, the assumption of zero lateral flux convergence implies that the form drag at the

solid boundary is equal to the simple momentum flux through an overlying level surface.¹ Evaluating the simple momentum flux requires vertical interpolation to horizontal surfaces that are far enough above the ground to avoid downward extrapolation into the topography. In a terrain-following model, especially, the form drag is the more logical choice for diagnosing the resolved drag, but we will also briefly consider the simple momentum flux in our discussion.

Mountain disturbances can, of course, be nonsteady and contain divergent lateral momentum flux (e.g., Pierrehumbert and Wyman 1985; Garner 1995). This will present a challenge for any diagnostic of the local drag that is designed to be directly comparable with drag parameterizations. Averaging in time can take care of most of the transience, but capturing all of the drag that is delivered to the vertical column is a more subtle problem.

The high-resolution model that we use for the ground-truth drag is a regional model to be described in the section 2. The proposed evaluation method is meant to be generally applicable, but we confine our testing to the scheme proposed by Garner (2005), which is also summarized in the next section. In section 3, we establish the diagnostic, and in section 4 we test it for internal consistency. The evaluation of the parameterization scheme is the focus of section 5.

All of the equations that we show here explicitly are for the proposed drag diagnostic. We hope to avoid confusion by leaving the analytical descriptions of the scheme and the ground-truth model to the references. These are specific choices from many that are available. One of the parameters of all drag schemes is the filter scale used in converting the raw topography to statistics at the resolution of the model using the scheme. We intend to smooth the drag diagnosed from the ground-truth model to the same scale. It can be chosen arbitrarily because it is that of a hypothetical model. We take it to be 50 km, which is about half the present state of the art in global climate models. In the scheme that we will be testing in section 5, the resolved flow is evaluated

¹ The equivalence is evident in the expressions for the velocity and pressure in linear waves. In the absence of rotation, stationary hydrostatic internal waves have (using primes for the disturbance and overbars for the mean state) $w' = \bar{u} \partial h / \partial x$, $u' = \bar{N} \hat{h}$, and $p' / \rho_0 = -\bar{u} \bar{N} \hat{h}$, where u is the wind component across the topography (assumed two-dimensional without loss of generality), N is the buoyancy frequency, and $\hat{h}$ is the Hilbert transform of h . After phase averaging, the form drag is the same as the simple momentum flux. Linearity is the limit $H \ll U/N$, where H is a scale for topographic height variations and U and N are scales for the undisturbed wind and buoyancy frequency, respectively.

at the top of the diagnosed mixed layer. We emulate this approach by taking data from the top of the ground-truth model's mixed layer to diagnose the drag.

2. Model and methodology

The resolved drag to be compared to the parameterized drag comes from a set of wintertime regional simulations over western North America, Greenland, and South America, downscaled from 2003 reanalysis. The model domains are defined in Table 1. The “Rockies” and “Andes” experiments run for 2 weeks and the Greenland experiment runs for 10 days. We also run a short simulation over western North America driven by an idealized westerly flow. The simulations are conducted on a C grid at a resolution of roughly 5 km, so that the smallest possible meaningful scale—at least for the divergent motions—is 10 km.² The vertical grid is terrain following with 120 levels up to an altitude of 20 km. We need this many levels to avoid extreme vertical grid stretching, which would distort complex vertical structures capable of changing the base flux. The vertical grid spacing ranges from about 90 m at the bottom of the model to about 170 m at the top.

We use a regional model with a compressible, non-hydrostatic dynamical core. Some of the details of the model will be given here and more information can be found in Pauluis and Garner (2006). The bulk microphysics is a five-species scheme (Krueger et al. 1995). The vertical diffusivity is a local one-dimensional TKE scheme (Mellor and Yamada 1982). There is no convective parameterization or subgrid orographic drag. The atmosphere is coupled to the land model described in Donner et al. (2011). It predicts soil temperature and moisture as well as albedo for use by the radiative transfer code, which is described in Anderson et al. (2004). The radiative forcing includes the diurnal cycle. Roughness lengths used by the Monin–Obukhov turbulence scheme are static over land and diagnosed over ocean as described in Anderson et al. (2004). At ocean points, the sea surface temperature field is linearly interpolated in time from 6-hourly mean Reynolds data. The atmosphere is nudged to the NCEP-1 (Kalnay et al. 1996) reanalysis at the lateral boundaries on a time scale of 2 h, while spectral nudging on a 12-h time scale is applied in the interior. The physical and computational

TABLE 1. Experiment domains.

	Latitude	Longitude
Rockies	22°–55°N	132°–92°W
Greenland	55°–85°N	75°–10°W
Andes	45°S–15°N	90°–45°W

setup is the same as in the tropical cyclone studies by Knutson et al. (2007), Garner et al. (2009), and Wu et al. (2014), except that we use many more levels.

We integrate through the first part of January 2003 for the northern regions and through the first 2 weeks of July 2003 for the Andes. High resolution is not a great computational burden in this study, because 2 weeks is long enough to sample an adequate range of synoptic patterns for the season. The parameterized drag is computed offline from the velocity, density, and static stability at the top of the mixed layer. These input fields are coarsened from the simulations up to a horizontal resolution of about 50 km using a spectral filter (the model is not actually run at the coarse resolution). The mixed layer is diagnosed from the dry static stability profile in the coarsened model data and varies in time and space from 0 to about 1200 m in depth. We evaluate the resolved drag, be it the form drag or the simple momentum flux, at the same level. However, for the form drag, the slope is evaluated from the resolved topography because the mixed-layer depth varies more smoothly.

The drag scheme is built on an analytical expression for the base flux over arbitrary topography and expanded to include a nonlinear contribution (Garner 2005). It uses a dimensional analysis to estimate the nonlinear drag, in a manner similar to the widely used scheme of Lott and Miller (1997), but uses an exact linear analysis of the drag to avoid their more ad hoc representation of anisotropy. The raw topographic data, a merger of Shuttle Radar Topography Mission 1 Arc Second (SRTM30m) and Global 30 Arc-Second Elevation (GTOPO30) datasets from the USGS, is gridded at 1/120°. The scheme requires a high-pass topographic filter length scale. For the largest passed scale, we use 50 km, the grid scale of our hypothetical global model. Except for the drag coefficients, the configuration is the same as in the CMIP6-member climate simulations reported in Zhao et al. (2018).

Base-flux schemes vary widely in their treatment of terrain height and anisotropy. A common feature among them, however, is the separation of the drag into additive linear and nonlinear contributions. Thus, the drag can generally be expressed as $D = a_0 D_0 + a_1 D_1$, where the subscripts 0 and 1 refer to linear and nonlinear contributions, respectively. When the true drag is known, the drag coefficients, a_0 and a_1 , can be chosen to maximize the agreement. Specifically, we can apply a least squares fit to

²The effective resolution is undoubtedly much larger for non-divergent circulations (Frehlich and Sharman 2008; Skamarock 2004). However, we often see 10-km-scale features over mountains that appear to be physical in that they match the drag scheme at that scale.

FIG. 1. Bernoulli-based (top) perturbation pressure (N m^{-2}) and (bottom) base flux (N m^{-2}) over western North America averaged over 6 h on (left) 6 and (right) 13 Jan 2003 from a 2-week-long regional downscaling.

the large sample of base-flux values, D_0 and D_1 , obtained by collecting all grid points with ground-truth drag above a certain threshold. The resulting coefficients are extremely variable in time, owing to the similarity in the spatial patterns of D_0 and D_1 . The vertical distribution would be needed to reduce the sensitivity. We therefore decide to fix the ratio of the coefficients and find the best fit to the

high-resolution data of a single overall drag coefficient. We will look at how this optimal drag coefficient varies in time.

3. A local form drag

The total form drag is obtained by integrating (6) and the corresponding meridional component along the

FIG. 2. (top) Perturbation pressure (N m^{-2}) and (bottom) base flux (N m^{-2}) over Greenland averaged for 24 h on (left) 2 and (right) 9 Jan 2003 from a 10-day-long regional downscaling.

lower boundary. In regional simulations, this has to be adjusted for any difference in vertically integrated pressure between lateral boundaries, which easily dominates the local drag if there is topography at these boundaries (in a model using pressure as a native variable, it is simpler and more exact to volume integrate the horizontal pressure gradient). For the local drag, we also rely on (6) but cannot use the full pressure. Surface pressure multiplied by slope

is dominated by high-amplitude dipoles due to the interaction of large-scale pressure with small-scale topography. Drag parameterizations implicitly deal with the rectified³ flux and ignore the dipoles. Therefore, we look

³ The term “rectify” is taken from the signal-processing or electrical-engineering vernacular.

FIG. 3. (top) Perturbation pressure (N m^{-2}) and (bottom) base flux (N m^{-2}) over South America averaged for 24 h on (left) 1 and (right) 5 Jul 2003 from a 2-week-long regional downscaling.

for a component of the pressure that is as strongly coupled to the high-resolution topography as possible.

We anticipate that spatial filters will not work well to isolate the mountain-scale pressure because the disturbance does not in general have a characteristic physical scale. A physically based filter is suggested by properties of nonrotating, hydrostatic linear internal waves, in

which the velocity perturbation is purely divergent (e.g., Garner 2005). These waves radiate drag up to breaking levels. The divergence field will also capture some drag-producing blocked flow in a general, nonlinear disturbance. The divergence will not detect drag due to deformational blocked flow around large-amplitude orography. We have not found a way to include this component.

FIG. 4. Meridional profiles from the Rockies simulation of zonally integrated zonal component of nonlocal form drag (black) and local form drag (red) for the two averaging periods analyzed in Fig. 1. The linear regression slopes are approximately 0.6, with correlation coefficients of about 0.8.

To obtain the local form drag, we first connect the low-level pressure and velocity fields using Bernoulli's law. We use the anelastic form so that we can write the pressure term as p/ρ_0 , where ρ_0 is a height-dependent large-scale density profile and p is the departure from the corresponding hydrostatic pressure. For the potential energy term, we estimate the buoyancy b by linearizing entropy conservation about a profile with constant buoyancy frequency, N^2 (e.g., Hoskins and James 2014, p. 121). Thus, $b = -N^2\delta z$, where δz is the vertical displacement. Then,

$$p = p_\infty - \rho_0[|\mathbf{u}|^2 - |\mathbf{u}_\infty|^2 + N^2(z - z_\infty)^2]/2, \quad (7)$$

where $\mathbf{u} = (u, v)$ is the horizontal velocity and the subscript ∞ refers to conditions far upstream. We now assume that this Lagrangian far field can be taken from a suitably defined “large scale,” from which the orographic gravity waves have been filtered. Thus, with LS

and MS denoting the large scale and mountain scale, respectively, the value of $M_x^\dagger$ at the lower boundary (the top of the mixed layer in our implementation) is

$$D_x = \rho_0 \frac{|\mathbf{u}_{LS} + \mathbf{u}_{MS}|^2 - |\mathbf{u}_{LS}|^2}{2} \frac{\partial h}{\partial x} + \text{Dipoles}. \quad (8)$$

We are defining as $\mathbf{u}_{LS} = \mathbf{u} - \mathbf{u}_{MS}$ and taking the mountain-induced velocity $\mathbf{u}_{MS}$ to be the divergent part of $\mathbf{u}$. The “dipoles” include the product of the terrain gradient with the large-scale pressure, $p_\infty \equiv p_{LS}$, and with the square of the vertical displacement, $\delta z = z - z_\infty$. There are also monopolar but non-gravity wave interactions between p_{LS} and large scales in the orography that we ignore.

The first part of the rhs of (8) should now be strongly rectified and the same should be true of the meridional component D_y obtained by replacing $\partial h/\partial x$ with $\partial h/\partial y$. We take (D_x, D_y) to define the base flux. For the density ρ_0 , we use the local value at the top of the mixed layer.

To isolate the divergent part of the flow, we use the Fourier transform that we also employ to coarsen the input fields for the drag scheme. For simplicity, we approximate the geometry as Cartesian with a zonal grid interval equal to the regional average. The divergence, $\nabla \cdot \mathbf{u}$, is calculated on quasi-horizontal surfaces defined by smoothing the model surfaces to a coarser resolution. Divergence measured on non-horizontal surfaces is not exactly coupled to gravity waves, but we wish to keep the vertical regridding as local as possible in the vertical near steep topography. We have used 50 km for the smoothing scale, the same scale that will be used to define the environment and high-pass-filtered topography for inputs into the drag scheme.

There is a useful interpretation of (8). After removing the dipoles, the local base flux can be rewritten as

$$D_x = -\rho_0 \frac{|\mathbf{u}_{MS}|^2}{2} \frac{\partial h}{\partial x} + \rho_0 v \mathbf{u}_{MS} \cdot \nabla_\perp h + \rho_0 u_{MS} w, \quad (9)$$

where we have identified $w = \mathbf{u} \cdot \nabla h$ and defined $\nabla_\perp h = (-\partial h/\partial y, \partial h/\partial x)$. The meridional component of the base flux has a similar form:

$$D_y = -\rho_0 \frac{|\mathbf{u}_{MS}|^2}{2} \frac{\partial h}{\partial y} - \rho_0 u \mathbf{u}_{MS} \cdot \nabla_\perp h + \rho_0 v_{MS} w. \quad (10)$$

The middle terms in (9) and (10) vanish for linear waves because the disturbance velocity is parallel to the topographic gradient (e.g., Garner 2005). If this is also approximately true for nonlinear disturbances, we can

FIG. 5. Meridional profiles from the Greenland simulation of zonally integrated zonal component of nonlocal form drag (black) and local form drag (red) for the two averaging periods analyzed in Fig. 2. The linear regression slopes are 0.39 and 0.64 for the first and second periods, respectively, with correlation coefficients of 0.67 and 0.82, respectively.

focus on the first and last terms in the two expressions. Let x be the direction perpendicular to a ridge and suppose that $u_{MS} = \alpha u_{LS}$ for different choices of α in the near field. Then, (9) gives

$$D_x = \left(1 + \frac{\alpha}{2}\right) \rho_0 u_{MS} u_{LS} \frac{\partial h}{\partial x}. \quad (11)$$

In the linear limit, $\alpha \ll 1$, we have $D_x = \rho_0 u_{MS} w_{LS}$, which is the last term in (9) but with w linearized as $w_{LS} = u_{LS} \partial h / \partial x$. For a nonlinear example, consider the full stagnation condition $\alpha = -1$, or $u_{MS} = -u_{LS}$. In this case, (11) gives $D_x = -\rho_0 (u_{LS}^2 / 2) \partial h / \partial x$, which is the first term in (9).

4. The base flux in ground-truth simulations

A drag parameterization like the one that we evaluate in section 5 is based on numerous approximations and cannot validate or invalidate the diagnosed base flux,

FIG. 6. Meridional profiles from the Andes simulation of zonally integrated zonal component of nonlocal form drag (black) and local form drag (red) for the two averaging periods analyzed in Fig. 3. The linear regression slopes are 0.15 and 0.47 in the (left) first and (right) second periods, with correlation coefficients of 0.40 and 0.65, respectively. The profile in the right panel is shifted south by 15°.

except perhaps in the limit of small-amplitude topography. A qualitative indication of success in diagnosing the base flux is rectification of the flux over small topographic features. Normally, the local drag should be single signed across a ridge or mountain. This condition will be checked in section 4a. Averaging the diagnosed and parameterized drag up to the 50-km coarse scale, similar to phase averaging, makes it easier to achieve. Quantitative validation should be based on consistency between the global form drag obtained without any filtering and global averages of the proposed base flux. In section 4b, we will compare the zonal integrals of the nonlocal form drag and local form drag after removing the large scales in the meridional direction. In a global climate model, most of the drag at these large scales would be resolved, but we will encounter an exception in the case of drag from ridgelike topography with large scales in only one direction.

FIG. 7. Time series of optimal drag coefficient for the (top) Rockies, (middle) Greenland, and (bottom) Andes experiments. Time averages are 1.35, 1.73, and 1.22, respectively. The top panel includes a shorter Rockies experiment driven by an idealized westerly flow, for which the time-average coefficient is 1.17.

One must be aware, when associating all of the horizontal divergence with the orography, of the other main sources of horizontal divergence, namely moist convection, fronts, and coastal circulations. We proceed on the assumption that these are relatively unimportant as sources of stationary gravity waves, especially after some Eulerian time averaging of the divergent velocity. Large-scale divergence can be relatively important in the tropics, but any significant interaction with topography should stand out as an unrectified component of the momentum flux.

a. Dynamically filtered pressure and form drag

In Fig. 1, we show the local Bernoulli-based pressure perturbation (top) and base flux (bottom) over western

North America, averaged over two 6-h periods during January 2003 that are characterized by strongly contrasting large-scale flows. The pressure variable is defined as $p_{MS} \equiv p - p_{LS}$ using (7), with p_{∞} and $\mathbf{u}_{\infty}$ replaced by p_{LS} and $\mathbf{u}_{LS}$, respectively. On 6 January, the flow is easterly or northeasterly across the southern two-thirds of the domain, whereas on 12 January, the northern two-thirds of the domain is in strong westerly flow. The simulated pressure perturbation exhibits dipoles throughout the region across a wide range of scales. The weak offshore disturbances are due to downstream effects of the coastal mountains. At places and times of weak large-scale flow, when the mountain disturbance is more nonlinear, the dipoles tend to be replaced by negative monopoles concentrated in the lees of mountains

FIG. 8. Base-flux vector and amplitude (shading, N m^{-2}) from the Rockies simulation averaged over the first period analyzed in Fig. 1. The (left) diagnosed base flux and (right) parameterized base flux. Vectors are plotted at every tenth model grid point.

and ridges. This type of asymmetry arises in (11) between, for example, $\alpha = -1$ upstream and $\alpha = 1$ downstream, but it might also be enhanced by the boundary layer viscosity. The time-averaged base flux for the two periods is shown in the bottom panels. The aforementioned smoothing up to 50 km has been applied. As expected, the terrain slope rectifies the pressure variations, so that the sign of the flux is dictated by the large-scale flow, rather than the topographic gradient.

Two periods with contrasting weather over Greenland are analyzed in Fig. 2. The averaging periods each span 24 h. The first begins at 0800 UTC 2 January 2003 and is shown in the left panels of Fig. 2. The second begins at 2200 UTC 8 January and is shown in the right panels. Between these periods, the easterly flow in the south strengthens considerably, while the westerly flow in the northeast weakens slightly. We again see dipoles in the time-averaged pressure disturbance, with scales ranging up to the width of the island (as in the Rockies experiment, there are downstream effects of the coastal mountains over water). On the other hand, the time-averaged local drag is mostly single signed over large areas, being positive

(eastward) in the south and along the west coast and negative (westward) along the northeast coast.

Finally, in Fig. 3, we show the pressure and base flux over South America during two 24-h periods starting at 1600 UTC 1 July and 1100 UTC 5 July 2003. The earlier period, shown in the left-hand panels of Fig. 3, is characterized by relatively strong easterlies in the tropics and subtropics. The other period, analyzed in the right-hand panels, is notable for strong westerlies in the subtropics and extratropics. The strongest easterlies are always weaker than the strongest westerlies. This is the main reason for the systematically weaker drag in low latitudes. The perturbation pressure field generally conforms to the terrain and the strength of the low-level flow. However, the base flux is not completely rectified, even at the coarse scale. We will find in section 5 that our parameterization, while agreeing with the dominant sign of the diagnosed drag at different latitudes, does not reproduce the weak sidelobes of negative (positive) drag in the north (south). The northern part of the domain develops daily convection with a strong diurnal signal in

FIG. 9. As in Fig. 8, but for the second period analyzed in Fig. 1.

the horizontal divergence, but the time averaging eliminates any significant impact on the pressure.

b. Integral constraint on the local drag

To validate the local base flux quantitatively, we propose a consistency check that refers to the nonlocal drag, that is, the form drag obtained without dynamical filtering. The horizontal integral of the local drag need not match or even approximate the nonlocal drag, as the latter includes contributions from the large-scale circulation. However, we can isolate the small scales that contribute to the nonlocal drag by analyzing zonal integrals of drag as a function of latitude and removing large scales in the meridional direction. There is still ambiguity in the choice of scales to retain in the meridional profiles. Our approach is simply to optimize the correlation between the zonal integrals. We interpret the optimized high-pass filter as that which retains the bulk of the drag from stationary gravity waves that are coupled to the topography without including significant drag from synoptic scales. By using moving-mean filters, we have found that removing scales larger than about 100 km produces the best matches. Across regions and over time, this scale varies between about 50 and 200 km.

There will be parts of the local drag that the comparison of meridional profiles after filtering cannot validate. Over zones containing sufficiently narrow north–south ridges, profiles of local zonal drag contain large contributions that drag schemes are designed to capture but which are removed by the filter because they appear at the meridional scale of the ridge. The corresponding global form drag profiles contain this drag plus non-gravity wave drag that may be an order of magnitude larger. There being no clear way to isolate the gravity wave drag on these scales, we must be satisfied with validating the along-ridge variations and trust that the match would extend to the mean amplitude of the ridge. Ridges are the main reason why the meridional profiles decorrelate quickly as the filter scale is increased beyond 100 km.

At each latitude, the integral of the zonal base flux between the western and eastern model boundaries, $x = x_W$ and $x = x_E$, is

$$\overline{D_{lx}} = \int_{x_W}^{x_E} D_x dx, \quad (12)$$

where D_x is given by (8). The same component of the integrated nonlocal drag is

FIG. 10. Base-flux vector and amplitude (shading, N m^{-2}) from the Greenland simulation averaged over the first period analyzed in Fig. 2. The (left) diagnosed base flux and (right) parameterized base flux. Vectors are plotted at every tenth model grid point.

$$\overline{D_{gx}} = - \int_{x_W}^{x_E} p(\partial h / \partial x) dx - \int_h^H [p(x_E) - p(x_W)] dz, \quad (13)$$

where h and H are the heights of the topography and model top, respectively. The pressure in the first integral in (13) and the density and velocity used for D_x in (12) will be interpolations at the top of the mixed layer. The functions $\overline{D_{gx}}(y)$ and $\overline{D_{fx}}(y)$ are the meridional profiles before filtering.

We start with the Rockies simulation. In Fig. 4, the meridional profile of zonally integrated nonlocal form drag $\overline{D_{gx}}$ is graphed in black and the integral of the base flux $\overline{D_{fx}}$ is graphed in red. Both profiles have been subjected to a 100-km high-pass spatial filter. The panels on the left and right sides of Fig. 4 correspond to the periods analyzed in the corresponding panels of Fig. 1. Linear regression over the whole range of latitudes yields correlations of approximately 0.8 in both regimes and slopes of approximately 0.6. A regression slope of much less than unity indicates that the local drag is underestimating the nonlocal drag at the relevant scales. The best amplitude matches are in the zones of strongest drag.

The same type of analysis for the two periods previously analyzed from the Greenland simulation is shown in Fig. 5. The correlations for both averaging periods are comparable to the Rockies simulation. The regression slope for the strong easterly regime is also similar. However, the slope for the weak easterly regime is only 0.39. As in the previous experiment, the match is best when the drag is strongest.

The consistency check for the Andes is shown in Fig. 6 for the same averaging periods as in Fig. 3. These are the smallest correlations and smallest regression slopes of the three simulations. There is reasonable agreement wherever there is strong large-scale flow and strong drag, but the correlation disappears when the zonal flow and drag are weak, especially in the deep tropics. The correlation between the two profiles in the left panel of Fig. 6 (strong tropical easterlies) is 0.40 but the regression slope is only 0.15. North of the equator (not shown), there is no correlation. In the right panel (strong extratropical westerlies), we find a correlation of 0.65 and a slope of 0.47. During this period, there is no correlation north of 15°S.

FIG. 11. As in Fig. 10, but for the second period analyzed in Fig. 2.

5. Comparisons with parameterized drag

The overall drag coefficient that produces the best match between the magnitude of the parameterized base flux and the magnitude of the diagnosed local drag is obtained by using least squares regression on each hourly map. We choose to set the individual drag coefficients a_0 and a_1 equal based on time averages from a bivariate regression. For the base flux (as opposed to the vertical drag profiles, which we are not considering), there is little sensitivity to the ratio. The results of the regression on the single overall coefficient $a = a_0 = a_1$ are plotted in Fig. 7 as a function of time for the three regions. The curves are smoothed using a 10-point running mean. The analysis ignores drag values less than 0.2 N m^{-2} . The parameterized drag is evaluated on the high-resolution grid from smoothed input fields and then averaged back up to the coarse scale (about 50 km). The time-averaged drag coefficients in each region based on the regression fall between 1 and 2. These averages will be used later to scale the parameterized base flux.

The time series for the downscaled Rockies experiment is shown in the top panel of Fig. 7. The time mean is 1.35, with a standard deviation of about 0.2. We also

show in the same panel the experiment with idealized westerly flow, which runs for 4 days. The average for this experiment is 1.17. A perfect drag scheme coupled with a perfect diagnostic would yield a constant optimal drag coefficient independent of time and geographic region. The coefficients for Greenland stand out as significantly larger than for the other two experiments and are also more variable in time. The average is 1.73, with a standard deviation of about 0.6 and a period of much higher coefficients around hour 210 (9 January). This period appears to coincide with an easterly tip-jet episode (e.g., Moore and Renfrew 2005). Since the local drag for Greenland shows good internal consistency (Fig. 5, right), most of the blame must lie with the scheme, which is not capturing the extremely high drag during this event.

The time-averaged optimal coefficient for the Andes experiment is 1.22. This time series shows a strong diurnal signal. This time series shows a strong diurnal signal, indicating a disagreement between the amplitudes of the diurnal signals in the parameterized drag and the diagnosed drag. The latter exhibits the weaker

FIG. 12. Base-flux vector and amplitude (shading, N m^{-2}) from the Andes simulation averaged over the first period analyzed in Fig. 3. The (left) diagnosed base flux and (right) parameterized base flux. Vectors are plotted at every tenth model grid point.

variation (not shown). It is influenced by diurnal moist convection in the northern Andes but not enough to explain the disagreement. The two local drag calculations evidently make inconsistent assumptions about the impact of the diurnally varying resolved variables on the drag. Moist convection also occurs in the other experiments, but only in association with fronts.

In Figs. 8 and 9, we show the geographical distribution of diagnosed (left panel) and parameterized (right) drag in the Rockies simulation for the first and second periods highlighted in the previous section, respectively. The arrows show the base-flux vector and the color shading the amplitude of the vector; that is, $D = \sqrt{D_x^2 + D_y^2}$. The fields are smoothed to the 50-km scale and vectors are plotted at every tenth model grid point. Recall that the drag coefficient is different for the three regions but fixed in time for a given region. The agreement seems acceptable upon inspection in both synoptic regimes. We will quantify the overall similarity in the next section.

Maps of the vector drag for the weak easterly period in the Greenland experiment are presented in Fig. 10 and for the strong easterly period in Fig. 11. The main

problem is that the parameterization underestimates the drag at the southern tip of the big island as well as over Baffin Island to the west. This problem is worse in the strong easterly regime.

The diagnosed and parameterized drag fields for the Andes experiment are shown in Figs. 12 and 13. There is good agreement during both averaging periods. The main discrepancy between the diagnosed and parameterized drag is that the latter is slightly stronger in the strong westerly regime. The parameterized drag is like the diagnosed drag in that it does not correlate very highly with the nonlocal form drag (Fig. 6). We will return to this problem in the next section.

With the drag coefficient optimized for the magnitude of the drag vector, we expect high correlations between the diagnosed and parameterized drag amplitudes. Indeed, the correlations easily exceed 0.8 in all of the periods and regions that we are focusing on. For a measure of the success in matching the drag direction, we also consider correlations between the meridional components of the drag vectors. The scatterplot and regression line for the meridional component can be seen in Fig. 14.

FIG. 13. As in Fig. 12, but for the second period analyzed in Fig. 3.

Between diagnosed and parameterized values of D_y , the easterly period in the Rockies experiment shows a correlation of 0.82 and a slope of 0.81. This slope is calculated for a zero intercept. The results are virtually the same for the zonal component (not shown).

The same type of analysis for the strong easterly regime in the Greenland experiment is shown in Fig. 15. The correlation is 0.63, while the slope is 0.43. The zonal component gives a correlation of 0.82 and nearly the same slope (not shown). For the data north of 69°N, which excludes the problematic southern tip of Greenland and Baffin Island, the regression slopes increase to 0.78 and 0.96, with correlations near 0.8 (not shown).

Finally, the scatter for the strong westerly regime in the Andes experiment is shown in Fig. 16. The correlation between the meridional components is 0.71 and the slope is 0.51. The zonal components yield a correlation of 0.80 and slope of 0.97 (not shown). The slope for the zonal components is artificially “improved” by the suspicious sidelobes discussed in Fig. 3, which, taken separately, show a negative correlation and negative slope between diagnosed and parameterized D_x .

An alternative strategy for tuning the drag coefficient is to give a unit slope to the best-fit regression line. This always increases the mean squared error as well as the coefficient. For the Rockies, Greenland, and Andes experiments, the time-averaged optimal coefficients increase from 1.35, 1.73, and 1.22, respectively, to 1.80, 3.14, and 1.61. The correlations are unaffected, but the root-mean-squared errors increase from 0.47, 0.59, and 0.45 on average to 0.67, 0.96, and 1.00, as normalized by the (time varying) standard deviation of the diagnosed drag.

For another measure of how well the scheme estimates drag direction, we consider the correlation between the meridional drag components after rotating the parameterized drag vector to the direction opposite the low-level wind. Orienting the drag opposite the low-level wind was the common approach before schemes began using information about the anisotropy of the terrain. Since the optimization is based on the vector magnitudes, the rotation does not affect the drag coefficients. For the easterly flow period in the Rockies experiment, the correlation between meridional components drops from 0.82 to 0.73 and, for the westerly regime, from 0.80 to 0.28. All other

FIG. 14. Meridional component of the local base flux vs the same component of the parameterized base flux (N m^{-2}) from the Rockies experiment for the averaging period analyzed in the left-hand panels of Fig. 1. The best-fit line with zero intercept is plotted over the scatter.

FIG. 15. Meridional component of the local base flux vs the same component of the parameterized base flux (N m^{-2}) from the Greenland experiment for the averaging period analyzed in the right-hand panels of Fig. 2. The best-fit line with zero intercept is plotted over the scatter.

periods and regions show slight to strong degradations. Thus, the scheme's treatment of terrain anisotropy significantly improves the agreement with ground truth.

6. Discussion and conclusions

We have developed and tested an algorithm that extracts orographic drag from high-resolution simulations in a form that is consistent with the output from drag parameterizations. Such parameterizations implicitly estimate the drag that is rectified across sign changes in the terrain slope and remains local in the horizontal. Drag schemes are based on statistical or analytical extensions of the phase-averaged momentum flux, which is rectified implicitly.

The challenge in isolating the local drag from ground-truth simulations is in removing the parts of the form drag that arise from interactions between the terrain and nonorographic circulations and interactions that do not take the form of gravity waves or blocking. This requires a pressure perturbation that is dynamically coupled to the high-resolution (submesoscale) terrain. Our way of estimating this pressure is probably the simplest and may be the best practical approach overall. It consists of using only the divergent part of the velocity together with a simplified Bernoulli law. Divergence from nonorographic circulations is a potential source of contamination, but we

find that time-averaging removes the bulk of the spurious signal.

We validate the local base flux by requiring consistency with the form drag defined by the full pressure, which we have called the nonlocal drag. The local drag is usually well correlated with the small meridional scales in the zonally integrated nonlocal drag, but generally falls short of the full amplitude, especially when the drag is weak. Some of this shortfall is the inevitable result of subjecting the meridional profiles to a high-pass filter, but another possible problem is that the Bernoulli approximation can be too energy conserving to correctly identify the pressure perturbation. A further dynamical issue is the neglect of deformational flow in calculating the Bernoulli pressure. However, after some attempts to redefine the pressure to include nondivergent flow at small scales, we failed to improve the match with the nonlocal drag.

By construction, the diagnostic does well when the mountain waves are linear. The biggest concern, then, is our inability to validate the weak, nonlinear drag in the tropical Andes. We notice that most of the high-pass signal in the meridional profiles across this region is constant in time. We cannot yet definitively attribute this steady component to a physical or numerical cause. However, if we remove it by forming differences between strong and weak drag episodes, we do uncover a correlation, especially south of the equator. Thus, in

FIG. 16. Meridional component of the local base flux vs the same component of the parameterized base flux (N m^{-2}) from the Andes experiment for the averaging period analyzed in the right-hand panels of Fig. 3. The best-fit line with zero intercept is plotted over the scatter.

Fig. 17, we compare the integrals of local and nonlocal drag after differencing the maps between simulation hour 193 (weak eastward drag) and hour 277 (strong eastward drag). These times occur on 9 and 12 July, respectively. The correlation is 0.55 and the slope is 0.49. This appears to validate the diagnostic, but we have set aside the possible issue with the uncorrelated stationary component that dominates the nonlocal drag.

An alternative to the local form drag is the simple momentum flux. If we use the divergent velocity only, the components are

$$D_x = \rho u_{\text{MS}} w, \quad D_y = \rho v_{\text{MS}} w, \quad (14)$$

with w measured at the top of the mixed layer. This singles out the last terms in the form-drag vector in (9)

and (10). The components in (14) are often quite similar to those in (9) and (10), but we have not found a period or region in which they are better matched to the nonlocal form drag. Resemblance to the local form drag is to be expected in a nearly linear disturbance. However, in the case of the tropical Andes, the drag is far from linear. We find that the linearized third term in (9) has the opposite sign to D_x in (14) as well as to the full zonal form drag and parameterized drag, which are all positive.

The comparison between the local diagnostic and the parameterized drag is meant to reveal problems with the parameterization. We find that the scheme that we tested generally captures the orientation and amplitude of the drag vector accurately. However, the optimized drag coefficient has too much low-frequency variability in one region (Greenland) and too much diurnal variability in another (South America). It is also not sufficiently uniform between these regions. We have not investigated the reasons for the strong variations between regions. However, we suspect that the period of large optimal drag coefficients in the Greenland simulation is due to a high-drag tip jet, which the scheme is apparently not handling well. In the case of the diurnal variability in the Andes, the problem boils down to disagreement about the dynamics of the mixed layer as it affects the drag.

Much of the uncertainty in orographic drag schemes attaches to the vertical distribution of forcing due to breaking, which is something that we have not looked at. Using high-resolution simulation to evaluate the vertical forcing profile is a formidable challenge, partly because vertical momentum flux is more difficult to diagnose away from the boundary, but mostly because wave-breaking and wave-activity conservation (or nonconservation) are difficult to do realistically in a model. Still, validating the vertical profile is necessary for independently tuning the linear and nonlinear parts of the parameterized forcing by means of ground-truth simulation. We intend to tackle this problem in future work.

FIG. 17. High-pass meridional profiles from the Andes simulation of zonally integrated zonal component of nonlocal form drag (black) and local form drag (red) differenced between hours 193 and 277. The linear regression slope is 0.49 and the correlation is 0.55.

REFERENCES

- Alexander, M. J., S. D. Eckermann, D. Broutman, and J. Ma, 2009: Momentum flux estimates for South Georgia Island mountain waves in the stratosphere observed via satellite. *Geophys. Res. Lett.*, **36**, L12816, <https://doi.org/10.1029/2009GL038587>.
- , and Coauthors, 2010: Recent developments in gravity-wave effects in climate models and the global distribution of gravity-wave momentum flux from observations and models. *Quart. J. Roy. Meteor. Soc.*, **136**, 1103–1124, <https://doi.org/10.1002/qj.637>.
- Anderson, J. L., and Coauthors, 2004: The new GFDL global atmosphere and land model AM2–LM2: Evaluation with prescribed SST simulations. *J. Climate*, **17**, 4641–4673, <https://doi.org/10.1175/JCLI-3223.1>.
- Boer, G. J., and M. Lazare, 1988: Some results concerning the effect of horizontal resolution and gravity-wave drag on simulated climate. *J. Climate*, **1**, 789–806, [https://doi.org/10.1175/1520-0442\(1988\)001<0789:SRCTEO>2.0.CO;2](https://doi.org/10.1175/1520-0442(1988)001<0789:SRCTEO>2.0.CO;2).
- Donner, L. J., and Coauthors, 2011: The dynamical core, physical parameterizations, and basic simulation characteristics of the atmospheric component AM3 of the GFDL Global Coupled Model CM3. *J. Climate*, **24**, 3484–3519, <https://doi.org/10.1175/2011JCLI3955.1>.
- Frehlich, R., and R. Sharman, 2008: The use of structure functions and spectra from numerical model output to determine effective model resolution. *Mon. Wea. Rev.*, **136**, 1537–1553, <https://doi.org/10.1175/2007MWR2250.1>.
- Garner, S. T., 1995: Permanent and transient upstream effects in nonlinear stratified flow over a ridge. *J. Atmos. Sci.*, **52**, 227–246, [https://doi.org/10.1175/1520-0469\(1995\)052<0227:PATUEI>2.0.CO;2](https://doi.org/10.1175/1520-0469(1995)052<0227:PATUEI>2.0.CO;2).
- , 2005: A topographic drag closure built on an analytical base flux. *J. Atmos. Sci.*, **62**, 2302–2315, <https://doi.org/10.1175/JAS3496.1>.
- , I. M. Held, T. Knutson, and J. Sirutis, 2009: The roles of wind shear and thermal stratification in past and projected changes of Atlantic tropical cyclone activity. *J. Climate*, **22**, 4723–4734, <https://doi.org/10.1175/2009JCLI2930.1>.
- Geller, M. A., and Coauthors, 2013: A comparison between gravity wave momentum fluxes in observations and climate models. *J. Climate*, **26**, 6383–6405, <https://doi.org/10.1175/JCLI-D-12-00545.1>.
- Ghan, S., and Coauthors, 2000: A comparison of single column model simulations of summertime midlatitude continental convection. *J. Geophys. Res.*, **105**, 2091–2124, <https://doi.org/10.1029/1999JD900971>.
- Hoskins, B. J., and I. N. James, 2014: *Fluid Dynamics of the Midlatitude Atmosphere*. John Wiley and Sons, 432 pp.
- Kalnay, E., and Coauthors, 1996: The NCEP/NCAR 40-Year Reanalysis Project. *Bull. Amer. Meteor. Soc.*, **77**, 437–471, [https://doi.org/10.1175/1520-0477\(1996\)077<0437:TNYRP>2.0.CO;2](https://doi.org/10.1175/1520-0477(1996)077<0437:TNYRP>2.0.CO;2).
- Kim, Y.-J., and A. Arakawa, 1995: Improvement of orographic gravity wave parameterization using a mesoscale gravity wave model. *J. Atmos. Sci.*, **52**, 1875–1902, [https://doi.org/10.1175/1520-0469\(1995\)052<1875:IOGWP>2.0.CO;2](https://doi.org/10.1175/1520-0469(1995)052<1875:IOGWP>2.0.CO;2).
- , S. D. Eckermann, and H.-Y. Chun, 2003: An overview of the past, present and future of gravity-wave drag parameterization for numerical climate and weather prediction models. *Atmos.–Ocean*, **41**, 65–98, <https://doi.org/10.3137/ao.410105>.
- Klinker, E., and P. D. Sardeshmukh, 1992: The diagnosis of mechanical dissipation in the atmosphere from large-scale balance requirements. *J. Atmos. Sci.*, **49**, 608–627, [https://doi.org/10.1175/1520-0469\(1992\)049<0608:TDOMDI>2.0.CO;2](https://doi.org/10.1175/1520-0469(1992)049<0608:TDOMDI>2.0.CO;2).
- Knutson, T. R., J. J. Sirutis, S. T. Garner, I. M. Held, and R. E. Tuleya, 2007: Simulation of the recent multidecadal increase of Atlantic hurricane activity using a 18-km grid regional model. *Bull. Amer. Meteor. Soc.*, **88**, 1549–1565, <https://doi.org/10.1175/BAMS-88-10-1549>.
- Krueger, S. K., Q. Fu, K. N. Liou, and H.-N. S. Chin, 1995: Improvements of an ice-phase microphysics parameterization for use in numerical simulations of tropical convection. *J. Appl. Meteor.*, **34**, 281–287, <https://doi.org/10.1175/1520-0450-34.1.281>.
- Lock, A. P., A. R. Brown, M. R. Bush, G. M. Martin, and R. N. B. Smith, 2000: A new boundary layer mixing scheme. Part I: Scheme description and single-column model tests. *Mon. Wea. Rev.*, **128**, 3187–3199, [https://doi.org/10.1175/1520-0493\(2000\)128<3187:ANBLMS>2.0.CO;2](https://doi.org/10.1175/1520-0493(2000)128<3187:ANBLMS>2.0.CO;2).
- Lott, F., and M. J. Miller, 1997: A new subgrid-scale orographic drag parametrization: Its formulation and testing. *Quart. J. Roy. Meteor. Soc.*, **123**, 101–127, <https://doi.org/10.1002/qj.49712353704>.
- Marks, C. J., and S. D. Eckermann, 1995: A three-dimensional non-hydrostatic ray-tracing model for gravity waves: Formulation and preliminary results for the middle atmosphere. *J. Atmos. Sci.*, **52**, 1959–1984, [https://doi.org/10.1175/1520-0469\(1995\)052<1959:ATDNRT>2.0.CO;2](https://doi.org/10.1175/1520-0469(1995)052<1959:ATDNRT>2.0.CO;2).
- Mellor, G. L., and T. Yamada, 1982: Development of a turbulence closure model for geophysical fluid problems. *Rev. Geophys. Space Phys.*, **20**, 851–875, <https://doi.org/10.1029/RG020i004p00851>.
- Moore, G. W. K., and I. A. Renfrew, 2005: Tip jets and barrier winds: A QuikSCAT climatology of high wind speed events around Greenland. *J. Climate*, **18**, 3713–3725, <https://doi.org/10.1175/JCLI3455.1>.
- Palmer, T. N., G. J. Shutts, and R. Swinbank, 1986: Alleviation of a systematic westerly bias in general circulation and numerical weather prediction models through an orographic gravity-wave drag parameterization. *Quart. J. Roy. Meteor. Soc.*, **112**, 1001–1039, <https://doi.org/10.1002/qj.49711247406>.
- Pauluis, O., and S. T. Garner, 2006: Sensitivity of radiative–convective equilibrium simulations to horizontal resolution. *J. Atmos. Sci.*, **63**, 1910–1923, <https://doi.org/10.1175/JAS3705.1>.
- Pierrehumbert, R. T., and B. Wyman, 1985: Upstream effects of mesoscale mountains. *J. Atmos. Sci.*, **42**, 977–1003, [https://doi.org/10.1175/1520-0469\(1985\)042<0977:UEOMM>2.0.CO;2](https://doi.org/10.1175/1520-0469(1985)042<0977:UEOMM>2.0.CO;2).
- Preusse, P., S. D. Eckermann, and M. Ern, 2008: Transparency of the atmosphere to short horizontal wavelength gravity waves. *J. Geophys. Res.*, **113**, D24104, <https://doi.org/10.1029/2007JD009682>.
- Skamarock, W. C., 2004: Evaluating mesoscale NWP models using kinetic energy spectra. *Mon. Wea. Rev.*, **132**, 3019–3032, <https://doi.org/10.1175/MWR2830.1>.
- Wu, L., and Coauthors, 2014: Simulations of the present and late-twenty-first-century western North Pacific tropical cyclone activity using a regional model. *J. Climate*, **27**, 3405–3424, <https://doi.org/10.1175/JCLI-D-12-00830.1>.
- Zadra, A., M. Roch, S. Laroche, and M. Charron, 2003: The sub-grid-scale orographic blocking parameterization of the GEM Model. *Atmos.–Ocean*, **41**, 155–170, <https://doi.org/10.3137/ao.410204>.
- Zhao, M., and Coauthors, 2018: The GFDL global atmosphere and land model AM4. 0/LM4.0: 2. Model description, sensitivity studies, and tuning strategies. *J. Adv. Model. Earth Syst.*, **10**, 735–769, <https://doi.org/10.1002/2017MS001209>.