Quantification of the responses of equatorial Pacific surface wind to uncertain cloud-related parameters in GAMIL2

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Abstract

The response of surface winds over the equatorial Pacific to cloud-related parameters is quantified by using a uniform sampling method and conducting a large number of perturbed parameter simulations. The results show that the surface winds are highly sensitive to, and even linearly dependent on some parameters that include the precipitation efficiency for deep convection (C0_deep), the time scale for the consumption rate of shallow CAPE (CMFTAU), and the threshold value for RH for deep convection (RHCRT), indicating the potential to improve surface wind simulations by tuning these parameters. Parameter changes indirectly affect surface winds through changes in vertical velocity or deduced circulation linked to diabatic heating (DH) above the surface. The relative contribution of static stability changes induced by perturbed parameter to balancing DH and affecting vertical velocity exceeds 10% in most of the troposphere.

Keywords: uncertain parameter; surface wind; diabatic heating; static stability

1. Introduction

Surface winds exert a profound influence on sea surface temperature (SST) and large-scale ocean processes through air–sea heat exchange and driving ocean currents over the equatorial Pacific and vice versa. Nevertheless, it is challenging to accurately simulate surface winds in general circulation models (GCMs) (Chang et al., 2007; Guilyardi et al., 2009). There is also no virtually improvement in equatorial surface winds from Phase 3 to Phase 5 of the Coupled Model Intercomparison Project (CMIP3 to CMIP5) (Zhang et al., 2015). Strong and nonlinear interactions with El Niño-Southern Oscillation make it difficult to improve simulated surface winds (Guilyardi et al., 2009; Ohba and Ueda, 2009; Choi et al., 2015). There are two main sources for surface wind biases: SST (Lindzen and Nigam, 1987) and diabatic heating (DH) (Gill, 1980). SST affects surface winds by cross-frontal sea-level pressure gradient. SST could also induce atmospheric convection, thus creating low-level convergence (Wallace et al., 1989; Chung et al., 2002). DH from both shallow convection and long-wave radiation at low levels (850–700 hPa) and from cumulus convection at high levels could also contribute to surface wind biases through atmospheric response (Nigam and Chung, 2000; Chiang et al., 2001; Wu, 2003; Zhang and Hagos, 2009). In addition, the vertical DH profiles, precipitation amounts and boundary layer entrainment are thought to cause surface wind biases (Schumacher et al., 2004; Chang et al., 2008; Zermeño-Díaz and Zhang, 2013).

In GCMs, uncertain parameters in model physical schemes introduce large uncertainties into climate simulations and projections (Murphy et al., 2004; Stainforth et al., 2005). Uncertainty quantification and parameter tuning are generally applied to reduce model biases (Wang et al., 2007; Jackson et al., 2008; Mauritzen et al., 2012). Previous studies are mainly focused on model variables or physical processes including precipitation, cloud amount, and radiation, which are directly controlled by or closely associated with uncertain parameters. However, as a dynamic quantity that is not directly influenced by uncertain parameters, is there the potential to improve simulated surface wind by parameter tuning? Hence, the aims of this study are to: (1) explore the potential for improving simulated surface winds by parameter tuning; and (2) quantify the response of surface winds over the equatorial Pacific to cloud-related parameters.

In this study, we focus on surface zonal winds over the equatorial Pacific, because the zonal components of wind vectors are dominant. The zonal winds at 1000 hPa are used to characterize surface zonal winds. The remainder of this paper is organized as follows. The model and its experimental design are briefly described in Section 2, and a detailed analysis is provided in Section 3. Section 4 gives a summary and discussion.
Table 1. Default values, ranges, sampling intervals, number of samples, and descriptions of the nine parameters.

| Parameter      | Description                               | Default | Minimum | Maximum | Interval | Number of Samples |
|----------------|-------------------------------------------|---------|---------|---------|----------|-------------------|
| C0_deep        | Precipitation efficiency for deep convection | 3.e−3   | 1.e−3   | 5.e−3   | 5.e−5    | 99                |
| KE             | Evaporation efficiency for deep convection | 7.5e−6  | 1.e−6   | 1.e−5   | 1.e−6    | 91                |
| CAPELMT        | Threshold value for CAPE for deep convection | 70      | 20      | 200     | 2        | 91                |
| ALFA           | Initial cloud downdraft mass flux          | 0.1     | 0.05    | 0.6     | 0.005    | 111               |
| CMFTAU         | Time scale for consumption rate of shallow CAPE | 7200    | 1800    | 14400   | 100      | 127               |
| RHCRIT         | Threshold value for RH for deep convection | 0.9     | 0.65    | 0.95    | 0.005    | 61                |
| C0_shc         | Precipitation efficiency for shallow convection | 5.e−5   | 3.e−5   | 2.e−4   | 2.5e−6   | 69                |
| RHMINL         | Threshold RH for low clouds                | 0.915   | 0.8     | 0.99    | 2.5e−3   | 77                |
| RHMINH         | Threshold RH for high clouds               | 0.78    | 0.65    | 0.85    | 2.5e−3   | 81                |

2. Models and experiments

This study uses the Grid-Point Atmospheric Model of IAP (Institute of Atmospheric Physics), LASG (National Key Laboratory of Numerical Modeling for Atmospheric Sciences and Geophysical Fluid Dynamics) Version 2 (GAMIL2), the version for Atmospheric Sciences and Geophysical Fluid Dynamics (Zhang et al., 2013). GAMIL2 uses a finite difference dynamical core that conserves mass and effective energy while solving the primitive hydrostatic equations of a baroclinic atmosphere (Wang et al., 2004), with the 128 × 60 horizontal and the 26-sigma vertical resolutions.

Moist physical processes in GAMIL2 are parameterized as three components: deep convection, shallow convection, and stratiform cloud processes. The Zhang and McFarlane, 1995 and Zhang and Mu, 2005 parameterizations are used for deep convection. GAMIL2 employs the shallow convective parameterization scheme proposed by Hack (1994). The cloud macrophysics (Zhang et al., 2003) provides a subgrid environment for microphysical processes. Cloud microphysics is represented by the two-moment bulk microphysics parameterization of Morrison and Gettelman (2008). The boundary layer scheme in GAMIL2 is a nonlocal K-profile scheme, which calculates diffusion coefficients based on the diagnosed boundary layer height and the related characteristic velocity scale (Holtslag and Boville, 1993; Vogelezang and Holtslag, 1996).

In this study, we focus on nine tunable cloud-related parameters. A description of these parameters and their perturbation is provided in Table 1. The perturbed ranges of parameters, based on the experience of model developers and previous studies (Jackson et al., 2008; Yang et al., 2015; Li, 2015, pers. comm.), could generally represent the uncertainty of parameters. Among the nine parameters, the following are from the deep convection scheme: C0_deep, KE, ALFA, CAPELMT, and RHCRIT. C0_shc and CMFTAU are shallow convective parameters. RHMINL and RHMINH are cloud fraction parameters that are closely related to the stratiform processes.

To explore the parameter sensitivities, the uniform sampling (US) method is used to divide the range of a given parameter into $n$ equal bins, and then the lower and upper limits of the parameter range and boundaries of $n$ bins are sampled in the parameter space.

The purpose of the US method is to sample a small quantity of points to represent the entire parameter space. In this study, the US sampling for the parameters was conducted as follows: when one of the nine parameters was sampled in its space, the other eight parameters remained unchanged at their default values. All sampled parameter sets are applied to conduct the experiments from July 2002 to December 2003. Each experiment is conducted following the standard setting for phase II of the Atmospheric Model Intercomparison Project (AMIP-II) with the same forcing recommended by the CMIP5 project, including the time-varying solar constant, greenhouse gases, aerosols (including sulfate, black and organic carbon, dust, and sea salt), ozone, and Hadley Centre Sea Ice and Sea Surface Temperature data (HadISST). More details related to the initial and boundary conditions can be found in Li et al. (2013). The model outputs for 2003 are used in the analyses and comparisons.

The European Centre for medium-range weather forecasts (ECMWFs) Interim Reanalysis (Simmons et al., 2006) and the National Centers for Environmental Prediction (NCEP) Reanalysis II (Kanamitsu et al., 2010) are used to evaluate the model results.

The thermodynamical equation in the tropics is simplified to a balance between DH and adiabatic vertical cooling (Mitias and Clement, 2006):

$$ DH = -S_p \omega $$

(1)

where, the static stability parameter is $$ S_p \equiv -(T/\theta)(\partial \theta/\partial p) $$, and the pressure velocity is $$ \omega \equiv d\theta/dt $$. Yu and Zwiller (2010) estimated the changes of DH by differentiating Equation (1) into two terms. Inspired by the study, the response of DH to parameters can be separated into two contributions as follows:

$$ \frac{\partial DH}{\partial \alpha} = S_p \frac{\partial (-\omega)}{\partial \alpha} + (-\omega) \frac{\partial S_p}{\partial \alpha} \equiv DH_D + DH_T. $$

(2)

where, $\alpha$ is the parameter. The term $DH_D$ indicates the response of DH related to variations in dynamics, while $DH_T$ is associated with variations in thermodynamics. Note that the contribution ratio (R) for $DH_T$ is estimated as,

$$ R = \frac{|DH_T|}{|DH_D| + |DH_T|} $$

(3)
3. Results

The observed and simulated annual mean zonal winds at 1000 hPa over the tropical Pacific are given in Figures 1(a)–(c). Both of the reanalysis shows similar spatial distributions with easterly wind cores over the regions of 150°E–120°W, 10–20°N, and 150–90°W, 0–20°S (Figures 1(a) and (b)). Easterly cores in NCEP2 are about 4 m s\(^{-1}\) stronger than in ERA-interim. The westerly wind over the Maritime Continent and west of Panama in NCEP2 is slightly weaker than that in ERA-interim. GAMIL2 captures the main features of the observed zonal wind, including the tropical easterlies with intensities closer to those of NCEP2 values (Figure 1(c)).

Model responses are represented by the regression coefficients between output variables and parameters (Figures 1(d)–(l)). Six of the nine parameters (C\(_0\) deep, KE, CMFTAU, RHCRIT, RHMINL, and RHMINH) can significantly affect surface winds. The time-scale for consumption rate of shallow CAPE (CMFTAU) and threshold value for RH for deep convection (RHCRIT) are the most important parameters, having strong but opposite influences over the whole tropical Pacific. Larger CMFTAU (smaller RHCRIT) causes weaker easterlies over the central–eastern Pacific, which can significantly reduce the simulated biases of strong easterlies when compared with those of ERA-interim (Figures 1(g) and (i)). However, larger CMFTAU (smaller RHCRIT) induces weaker westerlies over the Maritime Continent, which enhances the biases of weak westerlies to some degree, indicating that the optimal parameters may not be transferable over regions. In the physical parameterization, enhanced CMFTAU decreases the intensity of shallow convection by prolonging the adjustment time, which indirectly strengthens deep convection through the interaction among moist processes. However, increased RHCRIT directly makes the trigger condition difficult and suppresses deep convection. The opposite impacts of CMFTAU and RHCRIT on deep convection are closely related to the Walker Circulation, which can cause opposite changes in surface zonal wind. Although C\(_0\) deep and KE are also important, there is little overlap over the affected areas of them (Figures 1(d) and (e)). It is considered that the parameter impacts are highly dependent on the climate regimes (Yang et al., 2015). Larger RHMINL and RHMINH induce stronger easterlies over the central–eastern Pacific (Figures 1(k) and (l)).

The responses of surface zonal wind root-mean-square errors (RMSE), based on ERA-interim data, to parameters over the tropical Pacific are given in Figure 2. Five of the nine parameters (C\(_0\) deep, KE, CMFTAU, RHCRIT, and RHMIN) have significant effects on the model skill at the 95\% confidence level. Especially for CMFTAU, RHCRIT, and RHMINL, the responses of RMSEs are almost linear, in agreement with Figures 1(h), (i), and (k). The strong linear relationships between surface wind and parameters provide a good basis for model calibration or tuning. Better model results can be generally derived with larger (smaller) values for C\(_0\) deep, KE, and CMFTAU (RHCRIT and RHMINL), indicating the potential for
improving simulations of surface wind by parameter tuning. Given that the parameters CAPELMT, ALFA, and C0_shc have little or no effect on the zonal surface wind, they are excluded from further analysis.

To investigate the association of DH with surface wind, we examine the response of vertical DH profiles and vertical velocity to parameters over the equatorial Pacific. For the six selected parameters, the response of the vertical DH profile is well correlated with that of vertical velocity (Figure 3). Increased (decreased) DH is accompanied by stronger (weaker) upward motion, which substantiates that DH (cooling) and adiabatic upward cooling (downward heating) are in balance in the tropics (Holton, 1992). Larger precipitation efficiency for deep convection (C0_deep) induces more DH with stronger upward motion in the western part of the Maritime Continent, and less DH with weaker upward motion in the convection center between 150°E and the dateline (Figures 3(a) and (g)). Because of the constraint of continuity, the changes in vertical velocity along longitudes would give rise to anomalous easterlies (westerlies) at 1000 hPa in the west (east) of the dateline (Figure 1(d)). The dominant region of evaporation efficiency for deep convection (KE) is quite different from that of C0_deep. Larger KE leads to less DH with weaker upward motion over 120°–150°E, and less DH with stronger downward motion in the east Pacific (Figures 3(b) and (h)), which causes the response of surface zonal wind shown in Figures 1(e). The possible reason for different affected areas is that, the impact of C0_deep could be more significant over regions with strong convective condensation, while KE could be more important over regions of subsidence with relatively dry conditions that favor the evaporation of rainwater, such as the eastern Pacific (Yang et al., 2015). The response of surface wind to other parameters could also be attributed to changes in velocity and DH. That is, the impact of parameter changes on surface winds is through changes in vertical velocity or deduced circulation in response to DH above the surface. The results of the strong El Nino year (1998) and the La Nina year (2000) are also examined (not shown). The responses during the El Nino year (1998) are weaker than those during the normal year (2003). The largest DH response to C0_deep is located at 150°E, 15°W of the largest DH response in the normal year. Additionally, larger C0_deep induces weaker DH over the eastern Pacific at 600–200 hPa. During the La Nina year (2000), the response locations are slightly west of those during the normal year. In both the strong El Nino year and the La Nina year, the responses of the vertical DH profile are well correlated to those of vertical velocity.

Besides influenced by DH, vertical velocity would weaken (strengthen) as static stability $S_p$ increases (decreases) (Equation (1)). Yet the effect of $S_p$ on vertical velocity has been neglected in previous
studies. In addition, $S_p$ is not independent of DH. A larger value of RHCRIT significantly destabilizes the lower–middle troposphere below 300 hPa, while stabilizing the upper troposphere, especially west of the dateline (Figure 4(d)). The dominant areas of the other five parameters are almost in the upper level, where increased (decreased) $S_p$ is caused by larger CO\_deep and RHMINH (KE, CMF\_TAU, and RHMINL). For all six parameters, the ratio $R$, representing the relative contribution of $S_p$ to the DH response, is more than

Figure 3. Vertical cross-sections of the responses of (a)–(f) diabatic heating (units: K/day) and (g)–(l) vertical velocity ($-\omega$) (units: $10^{-2}$ Pa s$^{-1}$) to different parameters over the equatorial (5°S–5°N) Pacific. Model responses are calculated as in Figure 1.

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Figure 4. Vertical cross-sections of the responses of (a)–(f) static stability to different parameters (units: $10^{-5}$ K Pa$^{-1}$) over the equatorial ($5^\circ$S–$5^\circ$N) Pacific. Model responses are calculated as in Figure 1.

10% in most of the troposphere and 5–10% at some lower levels over both the equatorial western Pacific ($5^\circ$S–$5^\circ$E, 90–180$^\circ$E) and the equatorial eastern Pacific ($5^\circ$S–$5^\circ$E, 180$^\circ$E–90$^\circ$W; Figure 5). R over the eastern Pacific is slightly larger than R over the western Pacific. Despite the sensitive responses of $S_p$ in the upper troposphere, R reaches its maximum or submaximum in the lower level, exceeding 20%. These results reveal that the impact of $S_p$ on DH and vertical velocity cannot be ignored.

4. Summary and discussion

The impacts of uncertain parameters on simulated surface winds over the equatorial Pacific are explored by applying the uniform sampling method and conducting a large number of perturbed parameter simulations. Results show that the linear relationship between surface winds and parameters offers the potential to improve simulations of surface wind by parameter tuning. Six of nine selected parameters exhibit...
significant impacts on surface wind. RHCRIT is the most sensitive parameter, with a smaller value which induces weaker easterlies at 1000 hPa over the central–eastern Pacific and stronger westerlies over the Maritime Continent. Such responses reduce the surface wind biases compared with those from ERA-interim. CMFTAU has an opposite effect to RHCRIT on surface zonal wind. Model biases can be reduced in simulations with a larger CMFTAU.

The simulated DH is closely associated with these cloud-related parameters. A further look into the response of vertical DH profiles and vertical velocity explains the response of surface winds. For example, with larger C0_deep, the simulated DH and vertical velocity both strengthen simultaneously in the western Maritime Continent, and weaken in the western convection centers, favoring that DH (cooling) and vertical adiabatic cooling (heating) are in balance over the tropics. The changes in vertical velocity along longitudes lead to deduced easterlies at 1000 hPa in the west of the dateline. In short, parameter changes indirectly affect surface winds through changes in vertical velocity or deduced circulation linked to DH above surface.

The influence of static stability on vertical velocity is also emphasized. Static stability is most sensitive to the parameter RHCRIT. Larger RHCRIT destabilizes the lower–middle troposphere below 300 hPa and leads to greater stability above 300 hPa. Except for RHCRIT, the affected areas of the other parameters are mainly located in the upper level where the static stability is increased (decreased) by larger C0_deep or RHMINH (KE, CMFTAU, or RHMINL). The contribution ratio (R) for DHr is more than 10% in most of the atmosphere, and reaches a maximum or submaximum (>20%) near the surface, clearly indicating that the impact of Sp on DH and vertical velocity cannot be neglected.

Several limitations should be considered and deserve future investigation. First, parameter responses are represented by the regression coefficients between model variables and parameters here. The measure of responses could further be explained by the linear, high-order terms of parameters. Second, the US method used in this study perturbs one parameter, while the other parameters remain constant. Thus, the US method excludes interference from other parameters due to their possible interaction. Other efficient sampling approaches, such as the Latin hypercube sampling (LHS) method, can be used to evaluate the interaction among parameters and its effect on simulated surface winds. We further investigate the responses of DH and vertical velocity to the parameters using the LHS method. RHCRIT remains the most sensitive parameter. Some parameters (such as C0_deep and CMFTAU) induce weaker responses using the LHS method, compared with those from the US method, while the effects of KE are stronger. The affected areas of most parameters are changed with more or less shrink (not shown). These changes in the LHS method result from the interaction among parameters.

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