NOTES AND CORRESPONDENCE

Impact of Surface Drag of Islands in the Maritime Continent
on the Atmospheric General Circulation

By Akio Kitoh and Koji Yamazaki

*Meteorological Research Institute, Tsukuba, Ibaraki 305, Japan**(Manuscript received 10 January 1990, in revised form 13 February 1991)*

Abstract

In most general circulation models, the “maritime continent” is poorly resolved because there are many islands too small to be represented in the model. We investigate a possible effect of these small islands on the atmospheric general circulation by introducing surface drag over islands in model’s ocean grid points.

It is shown that the increased drag coefficient in the maritime continent is very effective in increasing the precipitation there by inducing stronger moisture flux convergence. Walker-type circulation anomaly is distinct over the entire equatorial Pacific along with the westerly anomaly in the lower troposphere over the Indian Ocean. There are also large impacts on both the Asian and the Australian monsoon not only for their strength but also for the onset/duration. These results imply the need of carefully handling surface conditions around the maritime continent where moisture flux convergence/heating feedback is strong, by considering the surface heterogeneity.

1. Introduction

Tropical islands exert dynamical as well as thermal effects on the atmospheric circulation. They are orographic and diurnal effects. In the “maritime continent” (Ramage, 1968, 1971), there are many islands which enhance convection in a continental way under an oceanic condition in terms of heat and moisture supply. In most general circulation models (GCMs), a land grid point is usually defined when the actual land area is larger than the ocean area in a given latitude-longitude grid box. In the maritime continent and the surroundings, however, there are many small islands which are not resolvable in GCMs. For climate modeling purposes, a model’s horizontal resolution must be limited despite the appearance of faster computers in order to make decades to hundreds of integrations feasible. Thus the parameterization of sub-grid scale processes becomes a big problem.

Recent development of the *land* surface parameterization in GCMs includes the specification of the soil and vegetation variables using the $1^\circ \times 1^\circ$ soil type and vegetation datasets (Dickinson, 1984; Sellers *et al.*, 1986; Warrilow and Buckley, 1989).

As the importance of including the heterogeneity of *land* surfaces into GCMs has been recognized (Avisar and Verstraete, 1990), the land surface in future GCMs will incorporate sub-grid heterogeneity (Henderson-Sellers, 1990). There are already some attempts to include heterogeneous land surfaces into numerical models (Sud and Smith, 1984; Mahrt, 1987; Wetzel and Chang, 1988; Avissar and Pielke, 1989) and also one attempt to consider localized precipitation in a model grid box (Pitman *et al.*, 1990). However, such models consider only the land surface and neglect the possible effect of small islands and simply prescribe an ocean grid point.

In this paper, we suggest that such an attempt should also be done over a usual *ocean* grid box, showing the possible effect of tropical small islands on the atmosphere by increasing surface drag coefficients even at ocean grid points. Results show that the impact on the atmosphere of introducing a surface drag over islands is enormous in the tropics, increasing dynamical convergence in the lower troposphere and divergence in the upper troposphere. This leads to moisture flux convergence and, through condensation of latent heat, affects atmospheric circulations, thus implying the need for careful handling of surface processes in this region.

2. The model and the experiment

The model used for this study (MRI-GCM) is a 5-level tropospheric atmospheric GCM with a horizontal resolution of 4° latitude and 5° longitude with a top at 100 mb. A parameterization of the planetary boundary layer (PBL) is based on Randall (1976) where the turbulent fluxes at the surface are determined by the bulk method of Deardorff (1972). See Tokioka *et al.* (1984) for details. In the current MRI-GCM, the effect of sub-grid scale topography is incorporated by increasing surface drag coefficients *over the continents* as a function of the standard deviation of sub-grid scale topography after Yagai and Tokioka (1987). The increased drag coefficient, C'_{DN} , is given by $C'_{DN} = C_{DN}(1 + \delta_m)$, where C_{DN} is the neutral value of the Deardorff's (1972) drag coefficient and δ_m is the standard deviation of sub-grid scale topography normalized by 100 m and calculated in the $4^\circ \times 5^\circ$ region by using $0.2^\circ \times 0.25^\circ$ data of topography.

In this study, δ_m is defined in a different way to include anisotropy of sub-grid scale topography. For instance, δ_m for the u-component is defined as "mean silhouette height seen in the east-west direction" after subtracting "the area-averaged height." It is normalized by 100 m. δ_m for the v-component is defined in a same way. Figure 1a and 1b shows δ_m for the u- and the v-component, respectively.

Three runs are performed: 1) the control run (C) with $\delta_m = 0$ over the ocean grid points; 2) the anomaly run (A1) with $\delta_m \neq 0$ even over the ocean grid points in the maritime continent (20°S – 20°N , 100°E – 180°E) where actual land area occupies more than 10 % of the grid point shown in Fig. 1c; 3) another anomaly run (A2) with increased drag over all grid points. All runs have non-zero δ_m over the land grid points. All runs are integrated for one year from September to August.

3. Results

a. Mean fields

Figure 2a shows the annual mean precipitation in C. Large precipitation areas simulated in C are southern India, Indonesia, New Guinea, the central Pacific off the equator in both hemispheres corresponding to the Intertropical Convergence Zone (ITCZ) and the South Pacific Convergence Zone (SPCZ), equatorial South America and central Africa. There is also much precipitation over the east coast of Australia. In the tropics, a precipitation peak is at the equator over the continents but it is off the equator over the oceans, particularly in the Pacific. In the maritime continent, precipitation has its maxima over the islands of Sumatra, Borneo and New Guinea.

There are large positive precipitation anomalies, A1–C, in the maritime continent (Fig. 2b). At the

periphery of the maritime continent, there is below-normal precipitation. Another decrease in precipitation is found in Central America. Anomalous evaporation (Fig. 2c) is positive in the maritime continent but its magnitude is no more than 1 mm d^{-1} and much less than the anomalous precipitation. The moisture flux convergence accounts for the rest (Fig. 2d). There is a large anomalous westerly moisture flux in the equatorial Indian Ocean and an anomalous easterly moisture flux over the entire Pacific. These anomalous moisture fluxes converge in the maritime continent, explaining the difference between precipitation and local evaporation. Notable decrease in precipitation around 160°E , 10°N and 70°W , 10°N is accounted for by anomalous moisture flux divergence. A northerly cross-equatorial flux is responsible for the former and the basin scale easterly flux over the equatorial Pacific is responsible for the latter. This larger contribution of moisture transport than *in situ* evaporation is similar to that found in previous sea surface temperature anomaly experiments (*e.g.*, Shukla and Wallace, 1983; Mechoso *et al.*, 1987). Anomalies in precipitation and moisture flux convergence in the region 20°S – 20°N , 100°E – 180°E , are statistically significant at the 99 % level and that in evaporation is significant at the 95 % level, based on the standard deviations of twelve annual means of the 12-year simulation of the MRI-GCM (Tokioka *et al.*, 1986).

At the grid points where the increased drag coefficient is used (Fig. 1c), on the average we get weaker PBL mean wind speeds in A1 except for the December–January–February season. However, the increased drag coefficient effect exceeds the reduced wind speeds, resulting in a larger (latent) heat flux from the ocean to the atmosphere. When local evaporation is increased and anomalous precipitation occurs in the potentially unstable and low-level convergent regions, there could be a positive feedback among precipitation, heating and moisture flux convergence as discussed in many sea surface temperature anomaly studies. This positive feedback is very effective in the December–January–February season, when convergence into the maritime continent is intensified with increased surface easterlies (trade winds) over the western Pacific and increased surface westerlies over Indonesia just south of the equator. The PBL mean wind speeds thus increase considerably in this season and the feedback becomes exceptional because the mean flow is convergent into the region. This leads to the difference in the intensification due to the current parameterization between the East Asian summer monsoon and the Australian summer monsoon.

Figure 3 shows the longitude-vertical cross section of the zonal wind, vertical p-velocity and total diabatic heating rate averaged for 12°S – 12°N . It is seen that mid-tropospheric heating has increased

Fig. 1. (a) x-component of δ_m . (b) y-component of δ_m . Contours of 1, 2, 4, 8 and 16 are drawn. Values greater than 4 are shaded. (c) Grid points where increased δ_m was used in run A1.

between 100°E and 160°E with some compensation outside of this domain, *i.e.*, the Indian Ocean and the central equatorial Pacific (Fig. 2b). Anomalous east-west circulations are seen there. Another large-scale anomalous Walker-type circulation is found in the Pacific with anomalous westerlies in the upper troposphere and anomalous easterlies in the lower troposphere, intensifying the existing basin-scale Walker circulation. A decrease in precipitation in Central America (Fig. 2b) is related to this large-scale anomalous circulation and thus to the precipitation increase in the maritime continent. This teleconnection can be seen in the velocity potential field.

b. Seasonal change

Figure 4 shows the longitude-time section of the velocity at 200 mb averaged for 12°S – 12°N using the 10-day mean data. In C, there are two stages of minima, indicating vigorous convective activity, one in December–January and the other in the June–July–August season, corresponding to the Australian summer monsoon and the East Asian summer monsoon, respectively. The latter is stronger and located more westward than the former. During a two month's transition period of March and April, tropical convection activity is rather weak. There are also minima around 60°W in October–November

Fig. 2. Annual mean precipitation for C (a) and annual mean differences between run A1 and run C in (b) precipitation, (c) evaporation and (d) moisture flux. Contour interval is 1 mm/day for (a) and (b), and 0.5 mm/day for (c). Negative values are dotted for (b) and (c). Reference vector for (d) is 30 kg m s^{-1} .

Fig. 3. Annual mean zonal wind and vertical p-velocity and total diabatic heating rate averaged between 12°S and 12°N for (a) C and (b) A1-C. Reference vectors: 5 (1) m s^{-1} and 5 (1) mb hour^{-1} for a (b). Contour interval: $0.5^{\circ}\text{C d}^{-1}$ for a and b. Negative values are shaded.

and in February–March. They correspond to diabatic heatings over Central America and northern South America. It is noted that the upper-level divergence there is weaker in December and January when divergence in the Australian monsoon region is strong. Such teleconnection is also seen in northern summer season when Central America is covered with strong convergence while there is strong divergence in the maritime continent.

For the northern summer, the seasonal mean divergent center in C at 200 mb, at 95°E , 16°N with the value of $-12.5 \times 10^6 \text{ m}^2 \text{ s}^{-1}$, is located systematically far westward of the observed mean position ($\sim 135^{\circ}\text{E}$, 18°N , $-15.8 \times 10^6 \text{ m}^2 \text{ s}^{-1}$, 1979–1987 mean, not shown). It is shifted to 115°E , 18°N in A1 and it is intensified to $-13.8 \times 10^6 \text{ m}^2 \text{ s}^{-1}$. As far as the velocity potential minimum at 200 mb is concerned, A1 is closer to the observation than C in both its location and intensity.

Velocity potential anomalies shown in Fig. 4c reveal upper-level divergence in the maritime continent and convergence in the Central America to African longitudes throughout the year. The Australian summer monsoon, peaked in its strength in late December in C, becomes more vigorous in A1. At the same time, convergence in South America also becomes strong. For the East Asian summer monsoon season, low-frequency anomalies with a pe-

Fig. 4. Longitude-time sections of the velocity potential at 200 mb averaged between 12°S and 12°N for (a) C, (b) A1 and (c) A1-C based on the 10-day means. Contour interval: $1.0 \times 10^6 \text{ m}^2 \text{ s}^{-1}$. Positive values are shaded.

Fig. 5. Power spectral distributions of the wavenumber 1 component of the velocity potential at 200 mb over the equator for the one-year period. (top) C, (bottom) A1.

riod of about two months are distinct with anomalous divergence in June and August and anomalous convergence in July at the longitude of the maritime continent. This is because the phase of the low-frequency oscillations is different between C and A1 (Fig. 4a and 4b).

The frequency of the low-frequency oscillation changes as the intensity and/or the vertical distribution of diabatic heating changes (Tokioka *et al.*, 1988). Activity of low-frequency oscillation is also different between C and A1. Figure 5 shows the power spectrum of the eastward- and westward-moving wavenumber 1 component of the velocity potential at 200 mb over the equator. The spectrum is calculated from daily time series after removing seasonal cycle (yearly and half-yearly) components. A five-day running mean is applied to the original data to correct a sampling problem, as the 5-day period appears due to sampling interval of 30 hours (four data are stored in every 5 days).

Both runs have large power of eastward-moving wavenumber 1 in the period between 10 and 60 days. The power of low-frequency oscillations in A1 is larger than that in C. Moreover, a spectral peak in C is found at 24 to 30 days while in A1 a 30–60 day spectral peak is more clearly defined. Inspection of daily velocity potential charts of A1 (not shown, but partly seen in Fig. 4b) reveals that the divergent center in the East Asian summer monsoon strengthens with a period of about 30 to 40 days. It reaches its peaks in late April, early June, early July and mid August.

Influence of the increased heating in the maritime continent can also be seen in a different development of the monsoon circulation. Figure 6 shows the time-series of indices relevant to the Asian sum-

Fig. 6. Daily time-series (5-day running mean) of indices for C (solid line) and A1 (dashed line). (a) Geopotential height at 200 mb averaged for 16°N–36°N, 40°E–100°E. (b) Zonal wind at 850 mb averaged for 4°N–12°N, 55°E–80°E. (c) Precipitation averaged for 0°N–24°N, 60°E–100°E. (d) Latitude of the East Asian subtropical jet axis defined as the location of the maximum zonal wind at 200 mb between 120°E and 180°E.

mer monsoon. The indices for the Tibetan high, the Arabian Sea lower troposphere westerlies, and the rainfall in the monsoon region (indices used by Kitoh and Tokioka, 1987) all show an earlier development of the Asian summer monsoon in A1 than in C. This difference comes not from the different phases of the 30–60 day oscillation between the two runs, but from

Fig. 7. Difference between run A2 and run C in precipitation. Contour interval: 1 mm d^{-1} . Negative values are dotted.

the anomalous heating over the maritime continent in A1. Anomalous heating over the maritime continent accompanied by an anomalous upper-level anticyclonic circulation to the northwest of the heating region and an anomalous low-level westerly over the Indian Ocean flowing into Indonesia might be favorable to the earlier development of the Asian summer monsoon. Responses in A2 (not shown) are similar to those in A1. Other evidence of the early establishment of the northern summer climate in A1 can be seen in the plot of the latitude of the East Asian subtropical jet (Fig. 6d). The East Asian subtropical jet shows a jump from south of 50°N to north of it in early July in C. This northward shift is concurrent with the decay of the jet strength. In A1, this northward jump or weakening of the subtropical jet has occurred in mid-June, about 20 days earlier than in C. Therefore the early establishment of the summer climate has consistently been observed in the model in the tropics and in the northern extratropics.

c. The A2 run

The A2 run uses the increased drag coefficient, C'_D , everywhere. This is in contrast to A1 where δ_m over the ocean grid points is zero outside the maritime continent. Figure 7 shows the annual mean difference, A2–C, in precipitation. Precipitation in A2 increases in the maritime continent as well as in the coastal regions in the tropics. Again, the contribution by the anomalous *in situ* evaporation is smaller and the anomalous moisture flux convergence mainly accounts for the anomalous precipitation.

A2 has produced less precipitation than A1 in the maritime continent and more precipitation in Central America. In A1, the maritime continent is the only region with the increased drag coefficient and the Pacific Ocean-scale Walker circulation anomaly can be created. Thus the anomalous moisture flux is easterly throughout the equatorial Pacific, inducing less precipitation in Central America and larger pre-

cipitation in the maritime continent. In A2, Central America becomes a moisture flux convergence region due to the increased drag coefficient there. In this case, the moisture divergence area is located in the Central Pacific and the precipitation increase in the maritime continent is less intense than in A1. The difference, A2–A1, of the velocity potential at 200 mb (not shown) is characterized by divergence in A2 over Brazil and convergence in the equatorial Central Pacific. The important point here is that the impact of introducing islands' surface drag on the atmospheric circulation through moisture flux convergence/heating feedback is most effective in the maritime continent.

4. Concluding remarks

The impact on the atmosphere of introducing surface drag over islands in the model's ocean grid points is investigated. It is shown that the increased drag coefficient in the maritime continent is very effective in increasing the precipitation there by inducing stronger moisture flux convergence. Walker-type circulation anomalies are distinct over the entire equatorial Pacific along with the westerly anomaly in the lower troposphere in the Indian Ocean. Changes are significant for the strength of both the East Asian summer monsoon and the Australian summer monsoon, although the latter dominates the former in the present configuration of the experiment because the confluent PBL winds into the maritime continent become strong due to positive feedback. Monsoon fluctuations and the monsoon onset/withdrawal are also affected.

The heterogeneity of land surfaces will be included in future GCMs, but we stress that such an attempt should also be done even over a usual ocean grid point. The parameterization we employed in this paper for introducing heterogeneity in the model grid box where ocean and land coexist by increasing surface drag coefficients over the entire grid scale by keeping other oceanic surface conditions intact is rather primitive and may have overestimated its ef-

fect. A more sophisticated parameterization should be developed before being incorporated into GCMs. But it can be said that the heterogeneity of land and ocean should be important in the tropics where moisture flux convergence plays a pivotal role by positive feedback.

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海洋大陸の島嶼による地表摩擦の大気への影響について
鬼頭昭雄・山崎孝治
(気象研究所気候研究部)

インドネシア周辺の「海洋大陸」には多くの大気大循環モデルでは表現できない小さい島が多数存在する。この論文では従来無視されてきたこれらの小さい島の効果を導入する最初の試みとして、海の格子点であっても島がある場合には抵抗係数を増加させてその影響を評価してみる。

海洋大陸において島による抵抗係数増加の影響は水蒸気フラックスによる収束を強めて降水量を増加させる上で効果的であることが分かった。赤道太平洋では元来ある東西循環が強まりインド洋でも下部対流圏での西風偏差ができる。アジアの夏のモンスーン及びオーストラリアの夏のモンスーン双方にとってもその平均場のみでなくモンスーン循環の季節内変動やモンスーンオンセットにも影響する。これらの結果は水蒸気フラックス収束と加熱とのフィードバックの大きい海洋大陸において、グリッド内での海と陸の混在を考慮した地表面過程のパラメタ化が特に重要であることを示唆している。