Effects of Large-Scale Mountains on Surface Climate —A Coupled Ocean-Atmosphere General Circulation Model Study

Akio KITOH

Meteorological Research Institute, Tsukuba, Japan

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Abstract

Effect of mountain uplift on climate is investigated by a global coupled ocean-atmosphere general circulation model with an emphasis on surface temperature changes. Results of the no-mountain run (NM) are compared with those of the control run with the present-day orography (M). When the lapse-rate effect is eliminated, continent interior becomes warmer with mountain uplift because clouds become fewer and the surface drier due to decreased moisture transport. On the other hand, South Asia and East Asia become cooler because summer monsoon precipitation is stronger, which makes the land surface wetter and increases clouds. Over the ocean, the existence of orography has a role to reduce sea surface temperatures (SST), particularly over the subtropical eastern oceans. This occurs because evaporation is larger due to stronger trade winds and also less solar radiation reaches the surface due to more low-level clouds, both associated with stronger subtropical anticyclones in M. The subtropical gyre is stronger in M than in NM, and therefore, the Kuroshio Current is stronger in M. When the effect of the ocean general circulation is not included, the SST over the western north Pacific becomes much lower in M than in NM because of stronger cold air outbreak from Siberia in winter in M. Thus, the ocean circulation changes act to reduce the SST changes by heat transport.

1. Introduction

For the Earth's climate, tectonic boundary forcing is important, as well as the Milankovitch forcing (Earth's orbital parameters), and atmospheric CO₂ levels (Ruddiman 1997). The tectonic boundary forcing includes continental configuration, sea level, orography, ocean gateways and seafloor bathymetry (Crowley and Burke 1998). Large-scale mountains, such as the Tibetan plateau, play an important role in making a global climate system through dynamical and thermodynamical effects (Kutzbach et al. 1993). Climate model experiments have long indicated that orographic changes

have a significant effect on the development of the Asian monsoon (e.g., Manabe and Terpstra 1974; Kutzbach et al. 1989).

Most of the previous model studies are conducted by atmospheric general circulation models (e.g., Kasahara et al. 1973; Manabe and Terpstra 1974; Hahn and Manabe 1975; Tokioka and Noda 1986; Kutzbach et al. 1989; Broccoli and Manabe 1992), and focused on climate over land. The uplift of the Tibetan plateau is also a theme of studying late Cenozoic climate (Ruddiman et al. 1989; Kutzbach et al. 1989; Raymo and Ruddiman 1992). However, mountain uplifts not only change land surface temperatures but can also affect sea surface temperatures (SST). Experiments using an atmospheric GCM coupled with a 50 m slab ocean have been done by Kutzbach et al. (1993) and Kitoh (1997). Rind et al. (1997) used a global coupled atmosphere-ocean general circulation model (CGCM) to study the effects of the

Corresponding author: Akio Kitoh, Climate Research Department, Meteorological Research Institute, 1-1 Nagamine, Tsukuba, Ibaraki, 305-0052 Japan.

E-mail: kitoh@mri-jma.go.jp

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Tibetan uplift on ocean-atmosphere circulation.

Kitoh (1997) showed by using a coupled atmosphere-mixed layer ocean model that the global mean SST dropped 1.4 K with the mountain uplift, mainly due to increased low-level clouds over the subtropical eastern oceans. The increased frequency of low-level clouds was related to subtropical anticyclones, intensified by a strong temperature contrast between the continent and the ocean. The land surface temperature drops due to the lapse-rate effect. When this effect was eliminated, the continent interior became warmer with the mountain uplift, because clouds became fewer and the surface drier due to decreased moisture transport. Southern Asia has become cooler because monsoon-induced precipitation wets the ground and increases clouds. We revisit this problem using the Meteorological Research Institute (MRI) CGCM, and investigate the effects of orography in the context of the atmosphere-ocean-land system. Results are compared with those done by a coupled atmosphere-mixed layer ocean model to illustrate the effect of incorporating the ocean general circulation on the surface temperature changes.

2. Model and experiment

A global coupled atmosphere-ocean general circulation model (MRI CGCM1, Tokioka et al. 1995) is used for this simulation. The atmospheric part of the model has a horizontal resolution of 4° latitude by 5° longitude, and 15 vertical levels with the model top at 1 hPa. The oceanic part adopted a variable resolution of 0.5° – 2.0° for the latitudinal direction, with a finer grid in the tropics and a fixed 2.5° for the longitudinal direction. There are 21 vertical levels with the bottom at a depth of 5000 m. A realistic sea ice model is included. Seasonal and diurnal cycles are included in the model integration. The model uses flux adjustments for heat and fresh water.

The control run (M run) was integrated for 150 years with a realistic land-sea distribution and orography (Kitoh et al. 1999). Elevation of the highest point over Tibet is 4,400 m in M (Fig. 1). Figure 2 shows the annual mean distributions of the SST, precipitation and evaporation in M. As we used the flux adjustments, overall features of the SST pattern are reproduced reasonably well. The simulated

Fig. 1. Orography used in the control run (M). Contours are 500, 1000, 2000, 3000 and 4000 m.

Fig. 2. Annually averaged climatology of the CGCM control run (M) in (a) SST, (b) precipitation, and (c) evaporation.

large-scale precipitation distribution generally agrees with the observed climatology (Xie and Arkin 1996). Greater precipitation is simulated over warm waters, e.g., a warm pool area in the western Pacific, the Intertropical Convergence Zone (ITCZ) and the South Pacific Convergence Zone (SPCZ), and also over the tropical land. Due to a stronger cold tongue, the simulated precipitation shows a double ITCZ-like structure with a dry region over the equator, extending far westward than the observations. The evaporation is large over the subtropical oceans and the eastern coasts of the continents. It is dry over the central Eurasian continent.

The same GCM was integrated for 30 years with a flat surface everywhere by keeping the same land-sea distribution (NM run). The initial condition was taken from 1 January of year 30 of the M run. The same flux adjustment values were used for the M run and the NM run. Figure 3 shows the time series of some indices of the NM run, chosen to represent

the Kuroshio Current intensity (ocean stream function at 140°E , 20°N), the SST at the central equatorial Pacific and near Hawaii (SST at 180°E , 0°N , and at 150°W , 20°N), and the Equatorial Current (ocean zonal current at 1.7 m at 150°E , 0°N), respectively. They show that the upper oceans have adjusted within 5 years to the imposed change of orography. Deep oceans are not in equilibrium, as the time scale of the deep oceans is much longer. In this report we focus on land and ocean surface temperature changes and associated atmospheric circulation changes. We calculate the averages from the last 10 years of integration. This set of experiments is called the CGCM experiment.

In the second set of experiments, we used the atmosphere/mixed layer ocean-coupled model (SGCM), where the ocean model was replaced by a constant 50-m depth slab. This model was integrated for 12 model years. The last 5 years of data were used. Results of this SGCM run were described by Kitoh (1997). By comparing the CGCM experiment and the SGCM experiment, the role of the ocean general circulation, at least in the upper oceans, may be separated.

3. Atmospheric circulation changes

Figure 4a shows the stationary eddy part of the geopotential height at 500 hPa of the CGCM control run (M) in January. The largest amplitude is located around 50°N , with troughs around Japan and northeastern Canada, and ridges over the west coasts of North America and Europe. When compared with the observed climatology of the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP/NCAR) reanalysis (Kalnay et al. 1996), the amplitude of the simulated stationary waves is smaller. Although a trough over Japan is rather well reproduced, a ridge over the northeastern Pacific is displaced westward than the observations.

The stationary eddy is created by the effect of land-sea distribution and mountains. Figure 4b shows the eddy geopotential height in NM, and Fig. 4c shows the difference between M and NM. They can be ascribed as the thermally forced component (NM) and orographically forced component (M–NM), respectively (Tokioka and Noda 1986). To estimate relative roles of the two effects, a variance of the 500 hPa stationary eddy field is calculated

Fig. 3. Time series of the indices of the NM run. (a) ocean stream function at 140°E , 20°N , (b) SST at 180°E , 0°N , (c) SST at 150°W , 20°N , and (d) ocean zonal current at 1.7 m at 150°E , 0°N .

Fig. 4. Simulated atmospheric stationary waves of the CGCM at 500 hPa in January. (a) M, (b) NM, and (c) M-NM. The contour interval is 40 m.

Fig. 5. Simulated 200 hPa winds of the CGCM in (a) M in January, (b) NM in January, (c) M in July, and (d) NM in July. Contours are for zonal wind with the contour interval 10 m s⁻¹.

over the region northward of 20°N. They are 3,800 m² in M, 1,600 m² in NM and 2,200 m² in M-NM, respectively. The land-sea distribution effect explains about 40%, and the mountain effect explains about 60% of the total variance. Therefore, the mountain effect is slightly larger than the land-sea distribution effect, but

they are of the same order of magnitude. This is in accord with the previous studies (e.g., Held 1983; Tokioka and Noda 1986). For a creation of the trough around Japan, both effects are of comparable magnitude.

Figure 5 shows the wind vector and zonal wind velocity at 200 hPa of the CGCM in Jan-

Fig. 6. Simulated surface winds of the CGCM in (a) M in January, (b) NM in January, (c) M in July, and (d) NM in July. Contours are for zonal wind with the contour interval 5 m s^{-1} .

uary and July. In January, the control run (M) simulates the strong jets over Japan and eastern North America. In the tropics, easterlies are located over Africa, Indonesia and South America. In July, the westerly becomes weak and displaced northward, but its location over Siberia is too far northward than the observations (Kitoh et al. 1995). The region with easterlies over Africa and the Indian Ocean becomes wider and the easterly jet is stronger in July than in January. The Tibetan anticyclone can be seen with an axis around 30°N .

The NM run shows weaker westerly jet maximum in January. The wind flows more zonal than in M. The Asian subtropical jet in M is more than 15 m s^{-1} stronger than that in NM, while wind around Hawaii is weaker in M than in NM. As indicated by a line plot in Fig. 5, the magnitude of the zonal mean zonal wind is about the same between M and NM. In the tropics, the region with easterlies is broadly extended in M than in NM. In July the westerly jet in NM does not shift northward, and is

found at 30°N , a latitudinal position close to that in winter. The tropical easterly jet in NM over the equatorial Africa and Indian Ocean is weak.

Figures 6a and 6b compare the surface wind in January between M and NM. It is noted that the surface wind in M is stronger than in NM over the ocean. In the North Pacific the westerly wind greater than 10 m s^{-1} is found in M, but not in NM. The northwesterly winds from Eurasia to the Far East (winter cold surge) are weaker in NM. The easterly trade winds in the subtropics are also stronger in M than in NM. The strengthened trade winds are more clearly seen in July, particularly over the subtropical North Pacific (Figs. 6c and 6d). Over the Asian monsoon region, the control run shows a maximum wind along the Somali coast, while the NM run has a more elongated westerly jet with its maximum over India. Over the Arabian Sea, the wind blows from southwest to northeast along the coast, which induces ocean upwelling. On the other hand,

Fig. 7. Simulated precipitation of the CGCM in (a) M in January, (b) NM in January, (c) M in July, and (d) NM in July.

wind is weak in NM and there is a northerly component over the eastern Arabian Sea. Therefore, the ocean upwelling is weaker in NM. This change in wind direction between M and NM could make an SST difference, which will be shown later. These changes in the wind distribution are associated with monsoon precipitation changes.

Figure 7 shows the precipitation of the CGCM in January and July for M and NM. An overall precipitation pattern is similar between the M and the NM run. This indicates that the land-sea configuration and the SST distribution are the main factors for the precipitation distribution. However there are some systematic differences. For example, differences are found in the Asian summer precipitation. The NM run shows a rain belt along 10°N over the Bay of Bengal and Indochina peninsula, while the M run has more rainfall inland along the southern and eastern peripheries of the Tibetan plateau. The Indian monsoon rainfall in July is larger in M than in NM. The role of mountains for this relocation of monsoon rain-

fall is pointed out by previous studies (e.g., Hahn and Manabe 1975). Broccoli and Manabe (1992) demonstrated the existence of dry inland areas due to mountains. The present model also shows less precipitation over the continent in M than in NM. Winter precipitation off the coast of Japan is larger in M than in NM. On the other hand precipitation over the eastern North Pacific is larger in NM. This feature is also seen in the North Atlantic. Another difference between the M and the NM run is found in the shape of the ITCZ. While the M run shows a sharp ITCZ north of the equator over the Pacific, the NM run shows a more broad distribution with large rainfall extending to Hawaii. The SPCZ in the NM run extends eastward, consisting a double ITCZ-like structure over the eastern tropical Pacific. These regions are where there are warmer SSTs in NM than in M (see Fig. 9). This is associated with a weaker subtropical anticyclone in NM, as shown immediately below.

Figure 8 shows the mean sea-level pressure of the CGCM in January and July for M and

Fig. 8. Simulated mean sea-level pressure of the CGCM in (a) M in January, (b) NM in January, (c) M in July, and (d) NM in July. The contour interval is 4 hPa. For the purpose of comparison, a constant value is added for NM so that global mean values become the same.

NM. In the experiments, the atmospheric mass was kept the same between M and NM. However, a reduction from the surface pressure to the mean sea-level pressure results in different global mean sea-level pressure values. Therefore, in order to facilitate an easy comparison, a constant value (about 7 hPa) is added for NM so that the global mean values become the same in the figure (NB: this is done only for the plotting purpose).

The control run (M) reproduces the Siberian High and the Aleutian Low in January, a strong heat low in India and a strong subtropical anticyclone in July. When compared to the observations, the Aleutian Low is displaced westward, and the Pacific subtropical anticyclone is stronger. This may be related to a somewhat distorted upper troposphere stationary wave structure (Fig. 4a). The stationary wave at the surface in the Northern Hemisphere becomes weaker when mountains are removed. The winter anticyclone no longer exists over Siberia, but is located over China in

NM. Apparently, the existence of the Tibetan plateau has a role to keep cold air over Siberia. A strong cold surge from Eurasia to the Far East is weak in NM.

In July, although low pressures cover land area, no deep cyclone is found in India in NM. A contrast in the mean sea-level pressure between the land and the oceans is small in NM. In particular, the strong Pacific subtropical anticyclone found in M becomes substantially weaker in NM.

4. Surface temperature changes

Figure 9 shows the surface temperature changes (M–NM) of the CGCM. Over the ocean, this corresponds to the SST changes. The globally averaged surface temperature change is -2.56°C for January, -1.81°C for July, and -2.24°C for the annual mean (Table 1). Land surface temperature change is large negative over the mountainous regions because of higher elevations in M. A large contrast is found between January and July over the mid-

Fig. 9. The surface temperature differences (M–NM) of CGCM in (a) annual mean, (b) January, and (c) July. The contour interval is 4°C with additional contours of -2°C and 2°C .

latitude land of Eurasia and North America, where negative temperature anomalies in January are replaced by positive anomalies in July. The SST changes are nearly the same between January and July.

As the temperature decreases with elevation, most of the temperature changes of Fig. 9 over

the mountainous regions are due to this “lapse-rate effect.” As in Kitoh (1997), we adjust the surface temperature changes by assuming a typical environmental lapse rate of 6.5 K km^{-1} in comparing the M run with the NM run. Figure 10a shows the difference in the adjusted annual mean surface temperature (M–NM) of

Table 1. Surface temperature differences (M–NM) in the CGCM. Global* and land* denote lapse rate adjusted surface temperature differences over land.

	global	ocean	land	global*	land*
Annual	–2.24	–0.84	–4.57	–0.64	–0.32
January	–2.56	–0.78	–5.51	–0.96	–1.26
July	–1.81	–0.89	–3.34	–0.21	+0.92

the CGCM. For comparison, the result by Kitoh (1997) is reproduced in Fig. 10b that is derived by the atmosphere/mixed layer ocean-coupled model (SGCM). They are qualitatively and quantitatively similar to each other except for some regions discussed below. Both in the CGCM and SGCM, most interior regions of Eurasia and North America are covered with positive temperature anomalies, while negative temperature anomalies are found over South Asia and East Asia on the southern and eastern peripheries of the Tibetan plateau, and also over tropical lands.

As there are three distinct characteristics in the surface temperature changes, we consider them separately; that is, (1) tropical lands and southern and eastern sides of the continents, (2) continental interiors, and (3) the SST.

4.1 South Asia and Eastern Asia

Figure 10a shows that there is a surface temperature cooling over South Asia and Eastern Asia even when the lapse-rate effect is removed. Figure 11 shows the annual mean differences (M–NM) of the CGCM in precipitation, evaporation, soil moisture and total cloudiness. As was shown in Section 3, the Asian summer monsoon in M is stronger than in NM. Mountains create a heat source in the mid-troposphere and extend monsoon circulation and precipitation farther north onto the Asian continent (Hahn and Manabe 1975). Negative temperature anomalies over South Asia and East Asia are associated with positive precipitation anomalies there. Increased precipitation moistens the ground (Fig. 11c), and wet ground favors local evaporation (Fig. 11b). Wet ground also decreases the sensible heat flux and lowers the surface temperature. The model simulates increased cloudiness associated with increased precipitation (Fig.

11d), thus reducing solar insolation. These processes, i.e., the precipitation–soil moisture–evaporation positive feedback and the precipitation–cloudiness–insolation relationship, maintain an increased precipitation and cooler surface temperatures in M on the southern and eastern peripheries of the Tibetan plateau in summer.

Over Eastern Asia, a cold surge from Siberia is stronger in M (Figs. 6 and 8). The surface temperature in this region is decreased by the northerly cold winds due to orography (Tokioka and Noda 1986; Kutzbach et al. 1993).

4.2 Continental interior

Surface temperatures in the continental interior became warmer due to mountains (Fig. 10a). In the continental interior, precipitation generally decreases with the mountain uplift (Fig. 11a). Figure 12 shows the annual mean moisture flux and its convergence. In NM, there is a large westerly moisture flux over northern Eurasia from the Atlantic Ocean, while in M it is much reduced. First of all, atmospheric moisture content (precipitable water) is larger in NM than in M, mainly due to surface and lower tropospheric temperature differences. The annual mean precipitable water difference between M and NM (not shown) has a very similar spatial pattern to that in the surface temperature. In addition, the surface wind is weaker in M than in NM over the continent except for high mountain regions (Fig. 6). Due to these effects, the moisture flux convergence in M is less over the continental interior than in NM (Fig. 12). This is in accord with the findings by Broccoli and Manabe (1992). Ground is drier, and cloudiness is less in M than in NM. Drier conditions in M result in the lapse rate adjusted surface temperature warmer than in NM, because of less evaporative cooling.

4.3 SST

The SST decreased over large parts of the ocean with the mountain uplift. The globally averaged annual mean SST in M is -0.84°C colder than that in NM (Table 1). A large SST drop is found over the eastern Pacific, over the tropical Atlantic, and over the ocean near Japan (Fig. 10a). Figure 13 shows the differences in the annual mean surface heat fluxes between M and NM. It can be found that two components maintain the warm SST anomalies

Fig. 10. Annually averaged difference (M-NM) in the surface temperature of (a) CGCM and (b) SGCM. Lapse rate effect due to orography is adjusted by $6.5^{\circ}\text{C km}^{-1}$, the environmental lapse rate. The contour interval is 2°C . (c) Difference between (a) and (b).

over the subtropical eastern Pacific. They are the net surface short wave radiation and the latent heat flux. Net short wave flux is about 40 W m^{-2} lower, and the latent heat flux is about 20 W m^{-2} lower in M than in NM to the southeast of Hawaii. Differences in the long-

wave radiation and sensible heat flux are small.

As indicated in Kitoh (1997), an intensity of the subtropical anticyclone is important for these SST changes. Figure 14 compares the annual mean sea-level pressure and surface

Fig. 11. Annually averaged difference (M–NM) of the CGCM in (a) precipitation, (b) evaporation, (c) soil moisture in the first level, and (d) total cloudiness. The contour interval is 1 mm d^{-1} in (a), 1 mm d^{-1} in (b), 10% in (c), and 10% in (d).

wind stress. Although the subtropical anticyclone still exists over the North Pacific in NM, it is much stronger in M. An intensification of the northern summer subtropical highs due to orography has been noted by Ruddiman and Kutzbach (1989) in their mountain, half-mountain and no-mountain experiments. This should be related to intensified monsoon heating in M over the Tibetan and Colorado (North America) plateaus. Hoskins (1996) proposed a mechanism of the summer subtropical anticyclones, where continental monsoon precipitation and associated latent heat release creates Rossby wave responses to its west, making atmospheric downward motions and anticyclones. However, the effect of the Tibetan plateau on the North Pacific subtropical high may not be ruled out yet. Using an idealized model, similar to Hoskins and Rodwell (1995), Ose (1998) examined northern summer circulation responses to the Asian summer monsoon heat sources.

Ose (1998) showed some hint of an influence of the Asian monsoon heating, not only on the Indian monsoon trough, but also on the subtropical anticyclones. Recent studies by Rodwell and Hoskins (2001) also showed that the Asian monsoon forcing alone induces realistic North Pacific subtropical anticyclones. Rodwell and Hoskins (2001) interpreted the equatorward part of subtropical anticyclones as the Kelvin wave response to the monsoon heating over the continent to the west, and concluded that the North American forcing is required to induce northerly flows off the west coast of North America to form a closed-off anticyclone. In our present experiment strategy, relative roles of the Eurasian and North American orography cannot be separated out. Therefore, we are continuing other experiments where either the Eurasian or North American orography is reduced, whose results will be reported elsewhere.

Fig. 12. Annually averaged moisture flux (vector) and its convergence (contour) in the CGCM of (a) M, (b) NM, and (c) M–NM. Moisture flux is vertically integrated through the whole atmosphere. Shaded (negative) region denotes divergence.

The subtropical anticyclones and SST are closely related through evaporative as well as radiative heat flux (Klein et al. 1995). Intensified subtropical anticyclones and resultant stronger northeasterly trade winds in M enhance evaporative heat loss and reduce the

SST over the eastern subtropical Pacific. Cloud-radiation feedbacks also contribute to reduce the SST. In M, large-scale atmospheric downward motions, associated with strong anticyclones, keep moisture within the atmospheric boundary layer and make stratus clouds off

Fig. 13. Annually averaged difference (M–NM) of the CGCM in heat flux at the surface. (a) net heat flux, (b) latent heat flux plus sensible heat flux, (c) net shortwave flux, and (d) net longwave flux. The contour interval is 20 W m^{-2} . Positive value indicates heat flux from atmosphere into the ocean.

California. The increased stratus clouds in turn reduce solar radiation at the surface, again contributing to lower the SST. The net surface heat flux difference is negative, which is balanced by an anomalous oceanic downwelling in M compared to NM by anticyclonic circulation anomalies.

In Fig. 10b, the surface temperature change obtained by the atmosphere/mixed layer ocean-coupled model (SGCM) is shown. On the first approximation, the response of the model to mountains is similar to each other between the CGCM and the SGCM. However, there is a difference between them, for example, in the temperature changes to the east of Japan (Fig. 10c).

At sea level, the Siberian high and the Aleutian low characterize the Eurasia–North Pacific region in the present-day winter climate. Both are more intense in M than in NM (Fig. 8). A

cold surge from Siberia is stronger in M than in NM (Fig. 6). The SST drop near Japan found in the SGCM (Fig. 10b) is related to changes in the cold air advection from the colder continent toward the warmer ocean in winter (Kutzbach et al. 1993). However, the SST decrease to the east of Japan in the CGCM (Fig. 10a) is much smaller. This difference may be explained by the changes in the ocean general circulation, which are present in the CGCM but is absent in the SGCM. Figure 15 shows the ocean stream function. As implied by a stronger subtropical anticyclone (Fig. 14) in M, the oceanic subtropical gyre is stronger in M than in NM. The Kuroshio Current and its heat transport are stronger in M. The lowered surface temperature by orography in the SGCM to the east of Japan is compensated by the larger heat transport, resulting in a smaller temperature response in the CGCM.

Fig. 14. Annually averaged sea level pressure (contour) and surface wind stress over the ocean (vector) in the CGCM of (a) M, (b) NM, and (c) M–NM.

Table 2 compares surface temperature changes (M–NM) of the three GCM experiments with different ocean configurations, i.e., the atmospheric GCM (AGCM), the SGCM and the CGCM. Here the SST is prescribed in the AGCM, but ocean surface temperature changes are not null because “ocean” includes sea-ice

grid points. In the AGCM, the global and annual mean temperature change with mountain uplift is only -1.34°C . By allowing ocean temperature changes, including a 50 m slab ocean, surface temperatures decrease more (-1.34°C in the AGCM versus -2.86°C in the SGCM). Larger sensitivity of the coupled models than

Fig. 15. Annually averaged ocean stream function in the CGCM of (a) M, (b) NM, and (c) M–NM.

Table 2. Annual mean surface temperature differences (M–NM) in the AGCM (atmospheric GCM), the SGCM (atmosphere-slab ocean coupled model), and the CGCM (coupled atmosphere-ocean GCM). Global* and land* denote lapse rate adjusted surface temperature differences over land. “Ocean” includes sea ice points.

	global	ocean	land	global*	land*
AGCM	–1.34	–0.05	–3.47	+0.26	+0.79
SGCM	–2.86	–1.40	–5.28	–1.26	–1.02
CGSM	–2.24	–0.84	–4.57	–0.64	–0.32

that of the AGCM is due to SST changes, where the SST is not allowed to change in the AGCM but the SST decreased in the SGCM and CGCM. It is noted that even the land surface temperature in SGCM decreases more than that in the AGCM. This is a remote and indirect effect of the SST change (Kitoh 1997). When ocean circulations are explicitly included in the CGCM, the magnitude of global temperature changes in the CGCM are smaller than that in the SGCM because of the cancellation as denoted earlier. Thus how much the climate system responds to forcing (mountain uplift in this case) greatly depends on what feedback processes are allowed.

5. Concluding remarks

In this paper we revisited the classic study on the mountain's role on climate. Effects of large-scale mountains on the climate are investigated by a global coupled ocean-atmosphere general circulation model. Results of the no-mountain run (NM) are compared with those of the control run with the present-day orography (M) with an emphasis on surface temperature changes. Responses in the atmospheric circulation are qualitatively similar to earlier studies. In northern winter, the subtropical jet in NM is more zonal and weaker than in M, while zonal mean wind intensity is close to each other. The summertime subtropical jet in NM does not show a seasonal northward migration. Monsoon circulations do exist in NM, but their structure and magnitude are different from those in M. Associated with a stronger monsoon rainfall, the subtropical anticyclone over the ocean is stronger in M.

Land surface temperature changes by orography can be summarized as a warmer continental interior, and colder coastal area over land, in M than in NM. Land surface temperatures drop due to the lapse-rate effect. When this effect is eliminated, the continent interior becomes warmer with the mountain uplift because clouds become fewer and the surface drier due to a decreased moisture transport. On the other hand, South Asia and East Asia become cooler because the summer monsoon is stronger, and heavier precipitation makes the land surface wetter, and increases the cloud formation.

The SST can also be modified by the existence of orography. Both the CGCM and SGCM show qualitatively similar results in a cooler SST in the model with orography, particularly over the subtropical eastern oceans. This occurs because less solar radiation reaches the surface due to more low-level clouds that are induced by a strong subtropical anticyclone. The subtropical gyre is stronger in M than in NM. The Kuroshio Current is stronger in M. This difference in ocean circulation may explain the quantitatively different response in SST to the east of Japan between the CGCM and SGCM.

In our experiment, surface temperatures cooled both over land, and the ocean, by the mountain uplift. Rind et al. (1997) reported the

result of the Tibetan plateau uplift experiment with the GISS coupled atmosphere-ocean GCM. In their model, the global mean SST has warmed slightly by uplift. The two models give qualitatively opposite conclusions on temperature changes, thus warranting further investigation on physical mechanism of climate change. One possibility is that this difference comes from different experimental design (all mountain uplift versus only the Tibetan uplift). We are now investigating the effects of the Asian and the American mountains separately. We are also performing the same experiment with a new version of the MRI coupled GCM whose physical package is quite different to the MRI CGCM1. Even minor changes in elevation may result in different types of climate (Crowley and Baum 1985; Hay and Wold 1998). Specification of the correct elevation will affect model results. Climate could respond nonlinearly to mountain heights (Liu, personal communication). We are attacking this problem by conducting a series of experiments with different mountain heights. Analysis of these results will help to understand the relationship between tectonic uplift and climate change.

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References

- Broccoli, A.J. and S. Manabe, 1992: The effects of orography on middle latitude northern hemisphere dry climates. *J. Climate*, **5**, 1181–1201.
- Crowley, T.J. and S.K. Baum, 1995: Reconciling Late Ordovician (440 Ma) glaciation with very high (14X) CO₂ levels. *J. Geophys. Res.*, **100**, 1093–1101.
- and K.C. Burke, (ed.), 1998: *Tectonic Boundary Conditions for Climate Reconstructions*, Oxford University Press, New York, 285 pp.
- Hahn, D.G. and S. Manabe, 1975: The role of mountains in the south Asian monsoon circulation. *J. Atmos. Sci.*, **32**, 1515–1541.
- Hay, W.W. and C.N. Wold, 1998: The role of mountains and plateaus in a Triassic climate model. In *Tectonic Boundary Conditions for Climate Reconstructions*, T.J. Crowley and S.K. Baum (ed.), Oxford University Press, New York, pp. 116–143.

- Held, I.M., 1983: Stationary and quasi-stationary eddies in the extratropical troposphere: Theory. In *Large-scale Dynamical Processes in the Atmosphere*, B.J. Hoskins and R.P. Pearce, Eds., Academic Press, New York, pp. 127–168.
- Hoskins, B.J., 1996: On the existence and strength of the summer subtropical anticyclones. *Bull. Amer. Meteor. Soc.*, **77**, 1287–1292.
- and M.J. Rodwell, 1995: A model of the Asian summer monsoon. Part I. The global scale. *J. Atmos. Sci.*, **52**, 1329–1340.
- Kalnay, E., M. Kanamitsu, R. Kistler, W. Collins, D. Deaven, L. Gandin, M. Iredell, S. Saha, G. White, J. Woollen, Y. Zhu, M. Chelliah, W. Ebisuzaki, W. Higgins, J. Janowiak, K.C. Mo, C. Ropelewski, J. Wang, A. Leetmaa, R. Reynolds, R. Jenne and D. Joseph, 1996: The NCEP/NCAR 40-year reanalysis project. *Bull. Amer. Meteor. Soc.*, **77**, 437–471.
- Kasahara, A., T. Sasamori and W.M. Washington, 1973: Simulation experiments with a 12-layer stratospheric global circulation model. I. Dynamical effect of the earth's orography and thermal influence of continentality. *J. Atmos. Sci.*, **30**, 1229–1251.
- Kitoh, A., 1997: Mountain uplift and surface temperature changes. *Geophys. Res. Lett.*, **24**, 185–188.
- , A. Noda, Y. Nikaidou, T. Ose and T. Tokioka, 1995: AMIP simulations of the MRI GCM. *Pap. Meteor. Geophys.*, **45**, 121–148.
- , S. Yukimoto and A. Noda, 1999: ENSO-monsoon relationship in the MRI coupled GCM. *J. Meteor. Soc. Japan*, **77**, 1221–1245.
- Klein, S.A., D.L. Hartmann and J.R. Norris, 1995: On the relationships among low-cloud structure, sea surface temperature, and atmospheric circulation in the summertime north-east Pacific. *J. Climate*, **8**, 1140–1155.
- Kutzbach, J.E., P.J. Guetter, W.F. Ruddiman and W.L. Prell, 1989: Sensitivity of climate to Late Cenozoic uplift in southern Asia and the American West: numerical experiment. *J. Geophys. Res.*, **94**, 18393–18407.
- , W.L. Prell and W.F. Ruddiman, 1993: Sensitivity of Eurasian climate to surface uplift of the Tibetan plateau. *J. Geology*, **101**, 177–190.
- Manabe, S. and T.B. Terpstra, 1974: The effects of mountains on the general circulation of the atmosphere as identified by numerical experiments. *J. Atmos. Sci.*, **31**, 3–42.
- Ose, T., 1998: Seasonal change of Asian summer monsoon circulation and its heat source. *J. Meteor. Soc. Japan*, **76**, 1045–1063.
- Raymo, M.E. and W.F. Ruddiman, 1992: Tectonic forcing of late Cenozoic climate. *Nature*, **359**, 117–122.
- Rind, D., G. Russell and W.F. Ruddiman, 1997: The effects of uplift on ocean-atmosphere circulation. In *Tectonic Uplift and Climate Change*, W.F. Ruddiman (ed.), Plenum Press, New York, pp. 123–147.
- Rodwell, M.J. and B.J. Hoskins, 2001: Subtropical anticyclones and summer monsoons. *J. Climate*, **14**, 3192–3211.
- Ruddiman, W.F. (ed.), 1997: *Tectonic Uplift and Climate Change*. Plenum Press, New York, 535 pp.
- and J.E. Kutzbach, 1989: Forcing of late Cenozoic northern hemisphere climate by plateau uplift in southern Asia and the American west. *J. Geophys. Res.*, **94**, 18409–18427.
- , W.L. Prell and M.E. Raymo, 1989: Late Cenozoic uplift in southern Asia and the American west: Rationale for general circulation modeling experiments. *J. Geophys. Res.*, **94**, 18379–18391.
- Tokioka, T. and A. Noda, 1986: Effects of large-scale orography on January atmospheric circulation: A numerical experiment. *J. Meteor. Soc. Japan*, **64**, 819–840.
- , ———, A. Kitoh, Y. Nikaidou, S. Nakagawa, T. Motoi, S. Yukimoto and K. Takata, 1995: A transient CO₂ experiment with the MRI CGCM—Quick report—. *J. Meteor. Soc. Japan*, **73**, 817–826.
- Xie, P. and P. Arkin, 1996: Analysis of global monthly precipitation using gauge observations, satellite estimates and numerical model predictions. *J. Climate*, **9**, 840–858.