Effects on Asian Monsoon of Gigantic Qinghai–Xizang Plateau and Western Pacific Warm Pool

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ABSTRACT

A GCM study is performed of the effects on Asian summer monsoon initiation of the Qinghai–Xizang Plateau and western Pacific warm pool. Results show that the Plateau, being a prominent sensible heat source, acts as a basic factor for the formation of the monsoon circulation; the northward transported low-latitude and low-level warm, moist flow in relation to the sensible heating experiences dynamic lifting on the south and east sides of the highland, releasing vast quantities of latent heat through condensation, whereby the monsoon circulation pattern is further modulated; the temperature contrast between the Pacific warm pool and the Australian/marine continents serves as another basic factor for the northern SW summer monsoon genesis over the South–China Sea–the western Pacific, which, however, falls into a category of winter monsoon on a physical basis.

Key words: Qinghai–Xizang plateau, Western Pacific warm pool, Sensible heating, Formation of Asian summer monsoon

I. INTRODUCTION

South and East Asia is the most active and extensive of the monsoon regions on a global basis. When SW monsoon prevails, abundant rainfall takes place over this area each summer. But the local wetness/dryness (associated with anomaly in the wind vigor and stricken area) represent severe threat to socioeconomy and human activities. It is generally accepted that the reason that Asian summer monsoon becomes so strong lies in the vast thermal contrast between the landmass and the Pacific/Indian Oceans, the Qinghai–Xizang Plateau as a huge barrier having its part to play in preventing the summer monsoon from northward progress. Yeh et al. (1979) and Flohn (1968) showed that the Plateau in summer, as a lifted sensible heat source, is responsible for the genesis of an upper–troposphere warm anticyclonic circulation (i.e., South Asian high) core and favorable for lower–troposphere SW summer establishment in this region as well. In contrast, Chen et al. (1985), and Luo and Yanai (1984) argued that the summer monsoon heat source center is not situated over the Plateau but in the Bay of Bengal, where enormous amount of latent heat released through vigorously convective condensation is responsible. In their numerical experiments He et al. (1984) indicated that a summer monsoon circulation close to reality is obtained with the heating source centered around 20°N on the south side of the highland and not when the center is located just over the plateau for simulation.

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Among the monsoon-formation effects, which is dominant, the dynamic or thermal? And for the latter, which is prevalent, the highland sensible heating or the latent heating through convective condensation around the Plateau? Our focus is on the two issues.

Adopted for numerical study is the GCM developed by the U.S. State University of Oregon (1982), including in itself full dynamics and fuller physics, with the drag coefficient evaluated as

\[
C_D = \begin{cases} 
0.002(1 + 3Z_s / 5000) & \text{for non-water surface} \\
\min(0.001(1 + 0.07|\bar{V_s}|), 0.0025) & \text{for water surface}
\end{cases}
\]

where \(Z_s\) denotes the terrain height in units of meters and \(V_s\) is expressed in m/s. \(C_D\) at 4000 m altitude in the Plateau is found to be 0.007, according to the above expression in contrast to 0.004–0.005 calculated from measurements (Zhang et al., 1988). Clearly both vary considerably. Hence, the revised value of \(C_D\) is given as follows:

\[
C_D = \begin{cases} 
0.002(1 + 3Z_s / 5000) & \text{with } Z_s \leq 1600 \text{ for non-water surface} \\
0.004(1 + Z_s / 4400) & \text{with } Z_s > 1600 \text{ for non-water surface} \\
\min(0.001(1 + 0.07|\bar{V_s}|), 0.0025) & \text{for water surface}
\end{cases}
\]

With this revision, \(C_D\) is 0.004–0.005 for the Plateau itself, with little or no influence on the value in other regions.

Five experiments conducted for this study are described as follows:

Exp. 1 is the control run with the maximum height of 4400 m assumed for the Plateau.

Exp. 2 is the no-plateau run with the terrain altitude of 200 m (if the actual elevation exceeds 200 m) over 60°–110°E, 18–58°.

Exp. 3 is the run in which the Plateau’s height \(H\) is assumed to be increasing. The increment \(\Delta H\) is given by

\[
\Delta H(I, J) = 2000 \times \sin(\pi I / 70) \times \sin(\pi J / 48)
\]

where \(I(= 60, 65, 70, \cdots, 110)\) denotes the degrees of longitude and \(J(= 18, 22, 26, \cdots, 58)\) of latitude, with the maximum increment set to be 2000 m.

Exp. 4 is the run of no sensible heating of the Plateau.

Exp. 5 is the experiment with SST warm anomaly of the western Pacific warm pool.

These experiments have the same initial conditions and the atmosphere is isothermal (310 K), calm (\(|\bar{V_s}| = 0\)), with initial surface temperature equal to zonally averaged SST and initial ground humidity varying versus the underlying surface’s conditions. The integrating period covers 92 model days from May 1 to July 31, with the mean of the last 31 days as the summer (July) circulation.

II. ANALYSIS OF CONTROL EXPERIMENT’S RESULTS

Fig. 1a is a plot of the simulated summer 850 hPa flowfield, wherefrom one sees that, for the Northern Hemisphere (NH), South Asia, East America and North Africa (western North Pacific and North Atlantic) are under the control of a cyclonic (a strong anticyclonic) circulation with the ridge line around 30°N while for the Southern Hemisphere (SH), the subtropical anticyclone ridge line is about 15°S with its vigor greatly reduced as compared to its NH