Maintenance of the Midtropospheric North African Summer Circulation: Saharan High and African Easterly Jet

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ABSTRACT

The conspicuous feature of the midtropospheric North African summer circulation is the Saharan high surrounded on its southern rim by the African easterly jet (AEJ). Like a major monsoon circulation, the Saharan high is juxtaposed with the North African divergent center to the east and the eastern Atlantic convergent center to the west. Different from a major monsoon circulation, these pronounced midtropospheric circulation components are overlaid by the western part of the Tibetan high. Because of the unique roles played by the Saharan high and the African easterly jet in the North African summer circulation, an effort is made to explore maintenance mechanisms of these two midtropospheric circulation elements. Major findings of this effort are summarized as follows:

1) In terms of the velocity potential maintenance equation, it is shown that the North African divergent center over the Chad–Sudan region is maintained by the vertical differential heating established by the Saharan thermal-low heating in the lower troposphere and the Saharan radiative cooling in the upper troposphere.
2) The Saharan high is spatially in quadrature with the North African divergent center. It is inferred from the streamfunction budget analysis that the Saharan high is maintained by the east–west circulation, which is formed by the east–west differential heating between the Saharan thermal-low heating and the eastern North Atlantic cooling. This inference is further substantiated by forced barotropic model simulations.
3) The AEJ around the tropical periphery of the Saharan high is almost perpendicular to the equatorward divergent northerlies spilling out of the North African divergent center. The energetic interaction between divergent and rotational flows reveals that this jet is maintained by the Coriolis acceleration associated with these divergent winds.

These findings not only reveal the maintenance mechanism of the Saharan high and the associated AEJ, but also facilitate the search for answers to some problems of the North African summer weather/climate system.

1. Introduction

The basic features of major monsoon circulations [including the Indian monsoon (e.g., Krishnamurti 1979), the North American monsoon (e.g., Barlow et al. 1998), the South American monsoon (e.g., Chen et al. 1999), and the Australian monsoon (e.g., McBride 1987)] observed from a planetary-scale perspective (Chen 2003) may be characterized as follows:

1) A monsoon circulation comprises a monsoon high in the upper troposphere and a monsoon (or continental thermal) low in the lower troposphere,
2) the monsoon high (low) and the upper (lower-level monsoon divergent circulation are spatially in quadrature, and
3) the coincidence between the diabatic heating (cooling) center and the upper-level divergent (convergent) center of the monsoon circulation indicates that the divergent circulation in a monsoon system is driven by east–west differential heating.

Does the West African monsoon circulation possess these circulation features? Cook (1999) gave a depiction of the North African summer circulation. Unlike

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other major monsoon systems, this summer circulation consisting of the midtropospheric Saharan high and the lower-tropospheric Saharan thermal low lies underneath the western part of the Tibetan high. Around the tropical periphery of the Tibetan high exists the tropical easterly jet. Despite its smaller horizontal dimension, the North African summer circulation has a circulation structure similar to the Asian monsoons: similar to the Asian monsoon, the African easterly jet (AEJ) appears around the southern rim of the Saharan high. However, another interesting feature of the African monsoon is the occurrence of monsoon rainfall along the West African seaboard south of the AEJ, rather than over the Saharan thermal low. This rainfall region is located south of the southern periphery of the Saharan thermal low where the moist, cold southwesterlies from the Gulf of Guinea meet the dry, warm Saharan air from the north and form the ITCZ.

The most conspicuous elements of the North African summer circulation linked to the West African monsoon are the Saharan low, the Saharan high, and the AEJ. Because of its vital role in the genesis of African easterly waves (Burpee 1972) and the occurrence of West African monsoon rainfall, the maintenance of the AEJ has been a research focus of this monsoon. It was demonstrated by previous studies that the AEJ is maintained by (i) a strong meridional surface temperature and moisture gradient and (ii) thermally forced meridional circulations. Burpee suggested that the AEJ was a response of the lower-tropospheric flow to the surface baroclinic zone substained by strong meridional surface temperature gradients and the reversal of temperature gradients in the middle troposphere. Performing numerical simulations with the general circulation model developed by Manabe (1969) at the Geophysical Fluid Dynamics Laboratory, Cook (1999) showed that without soil moisture the positive meridional temperature gradients were not able to produce the AEJ with the observed intensity. Modeling the ITCZ and Hadley circulation, Schubert et al. (1991) stressed that the meridional circulation forced by the ITCZ convection was important to maintain the AEJ. Their suggestion was expanded further by Thorncroft and Blackburn (1999) to include a thermally forced shallow meridional circulation induced by the Saharan thermal-low heating.

Being an integral part of the Saharan high, the AEJ owes its existence to this midtropospheric high. The maintenance mechanism of this anticyclone system should be included in the search for the maintenance mechanism of the AEJ. Rodwell and Hoskins (1996) pointed out that the anticyclone center of North Africa was located over the high ground of northwest Africa, but this anticyclone's maintenance was not explored. While examining the maintenance of the Tibetan high and the Indian monsoon low by east–west differential heating, Chen (2003) showed that the Asian monsoon–western tropical Pacific region and North Africa were linked by the Walker-type east–west circulation. Because North Africa is covered by the descending branch of this east–west circulation, can the descending motion of this circulation contribute to the formation/maintenance of the Saharan high?

Following Gill's (1980) analytic solution of a monsoon circulation in response to a tropical forcing, Rodwell and Hoskins (1996) used the dry general circulation model developed by Hoskins and Rodwell (1995) to search for the desertification mechanism in North Africa. This model, imposed with a South Asian monsoon heating source, was able to generate the Gill-type monsoon circulation. Since the Walker-type east–west circulation was not generated by this simulation, a negative cell of the westward-propagating Rossby wave emerging over North Africa was suggested as the cause of the Saharan desertification. In contrast, the Saharan thermal low was regarded by Cook (1999) as a westward extension of the Indian monsoon low. According to Thorncroft and Blackburn (1999), the maintenance of the AEJ is linked to upward motion driven by the Saharan surface heating. In view of the findings in these studies, one may question what role the Saharan upward motion plays in the development of the Saharan high.

A brief review of previous studies related to the formation of the three conspicuous elements of the North African summer circulation leads to the following hypothesis:

> The upward vertical motion driven by the Saharan thermal-low heating and the downward branch of the Indian monsoon east–west circulation form a strong divergent center in the midtroposphere and maintain the North African anticyclone. Along the southern rim of the Saharan high, the AEJ can be accelerated by the Coriolis force induced by downgradient divergent flows on the tropical side of this anticyclone.

An effort was made in this study to test this hypothesis. The National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalyses (Kalnay et al. 1996) for the period of 1979–2002 were used to portray the three-dimensional structure of the North African summer [June–August (JJA)] circulation in the context of the global tropical circulation with streamfunction and velocity potential in section 2. The formation of the North African divergent center by the Saharan thermal heating from the lower troposphere and the radiative cooling associated with the downward branch of the Asian monsoon east–west circulation from the upper troposphere will be presented in terms of the heat and velocity potential budgets in section 3. The maintenance mechanism of the Saharan high will be illustrated with the streamfunction budget analysis and simulations by a forced barotropic model in section 4. The effect of the divergent circulation (involved with the formation of the Saharan high) on the AEJ will be shown in section 5, while some concluding remarks and suggestions for a future study will be offered in section 6.

2. Summer circulation of North Africa

a. Summer circulation

A salient feature of the summer subtropical circulation in the upper troposphere is the Tibetan and Mexican highs juxtaposed with the North Pacific and North Atlantic oceanic troughs (Krishnamurti 1971). Underneath these circulation elements are the Indian and North American monsoon lows aligned with the North Pacific and North Atlantic anticyclones (Godbole and Shukla 1981). As inferred from the contrast between these circulation components in the upper and lower troposphere, the vertical structure of the summer subtropical circulation exhibits a phase reversal (White 1982). The major features of the summer subtropical circulation are clearly portrayed by the ψ(200 mb) and ψ(925 mb) fields, and the longitude–height cross section of ψ(20°N) (shown in Fig. 1. If the midtropospheric circulation depicted by ψ(600 mb) (Fig. 1b) is ignored, the North African summer circulation described by ψ(200 mb) and ψ(925 mb) seems to be only a westward extension of the Indian monsoon, including western parts of the Tibetan high in the upper troposphere and the Indian monsoon low in the lower troposphere.

Because of its effect on the West African monsoon climate and the genesis of African easterly waves, research attention on the maintenance of the AEJ was called for by Cook (1999) and Thorncroft and Blackburn (1999). In addition to the AEJ, the Saharan high is a conspicuous feature of the North African summer circulation in the midtroposphere. The AEJ core (indicated by a thick solid line in Fig. 1b) is located around the southern periphery of this midtropospheric anticyclone. The comparison of streamfunction fields at the three levels in Fig. 1 reveals that the Saharan high is sandwiched between western parts of the two most pronounced components of the Asian monsoon circulation. The relationship between these three circulation elements may be further illustrated by the longitude–height cross section of ψ(20°N) in Fig. 1d. The Saharan high in the midtroposphere is a distinct independent high system east of the North Atlantic anticyclone. Just like a major monsoon circulation, the Saharan thermal low lies below the Saharan high. This monsoon-like North African summer circulation in the lower half of the troposphere is overlaid by the western part of the Tibetan high. A different perspective of the three-dimensional structure of the North African summer circulation may be obtained by the latitude–height cross section of ψ(5°E) shown in Fig. 2a. The Saharan high centered at 25°N is juxtaposed with a Northern Hemisphere high-latitude trough and a Southern Hemisphere anticyclone. As indicated by the monsoon westerlies close to the surface and the midtropospheric easterlies (AEJ) in Fig. 2a, the West African monsoon is basically formed by the southern part of the North African summer circulation in the lower half of the troposphere. It is also revealed from this vertical structure of the North African summer circulation (Figs. 1d and 2a) that the West African monsoon is unique in the global monsoon system in such a way that it resides underneath the western part of the upper-level monsoon high of another major monsoon. Since the midtropospheric North African summer circulation consists of the Saharan high and the AEJ, the formation/maintenance of the Saharan high should be included in the search for the maintenance mechanism of the AEJ.

b. East–west circulation

It was suggested by Thorncroft and Blackburn (1999) that the AEJ may be maintained by the local Hadley circulation forced by deep cumulus convection along the ITCZ and the shallow meridional circulation driven by surface heating along the Saharan thermal low (Thorncroft and Blackburn 1999). Are Thorncroft and Blackburn's meridional circulations well depicted by the reanalysis data? The meridional circulation (v_D, $-\omega$) (5°E; where v_D is meridional divergent wind) superimposed with diabatic heating (stippled areas) and cooling (dotted areas) is shown in Fig. 2c. The local Hadley circulation south of 10°N is driven by the north–south differential heating established by the tropical ITCZ heating centered at 10°N (maintained by tropical rainfall shown at the bottom of Fig. 2b) and diabatic cooling south of the equator. The midtropospheric cooling over North Africa (Fig. 2c) may decelerate the lower-tropospheric upward motion excited by the surface Saharan heating and confine the double-cell Saharan meridional circulation below 500 mb. The southern Saharan cell is overlaid by a weak upper meridional cell formed by the ITCZ heating and the midtropospheric Saharan cooling.

Since the structure of the North African meridional

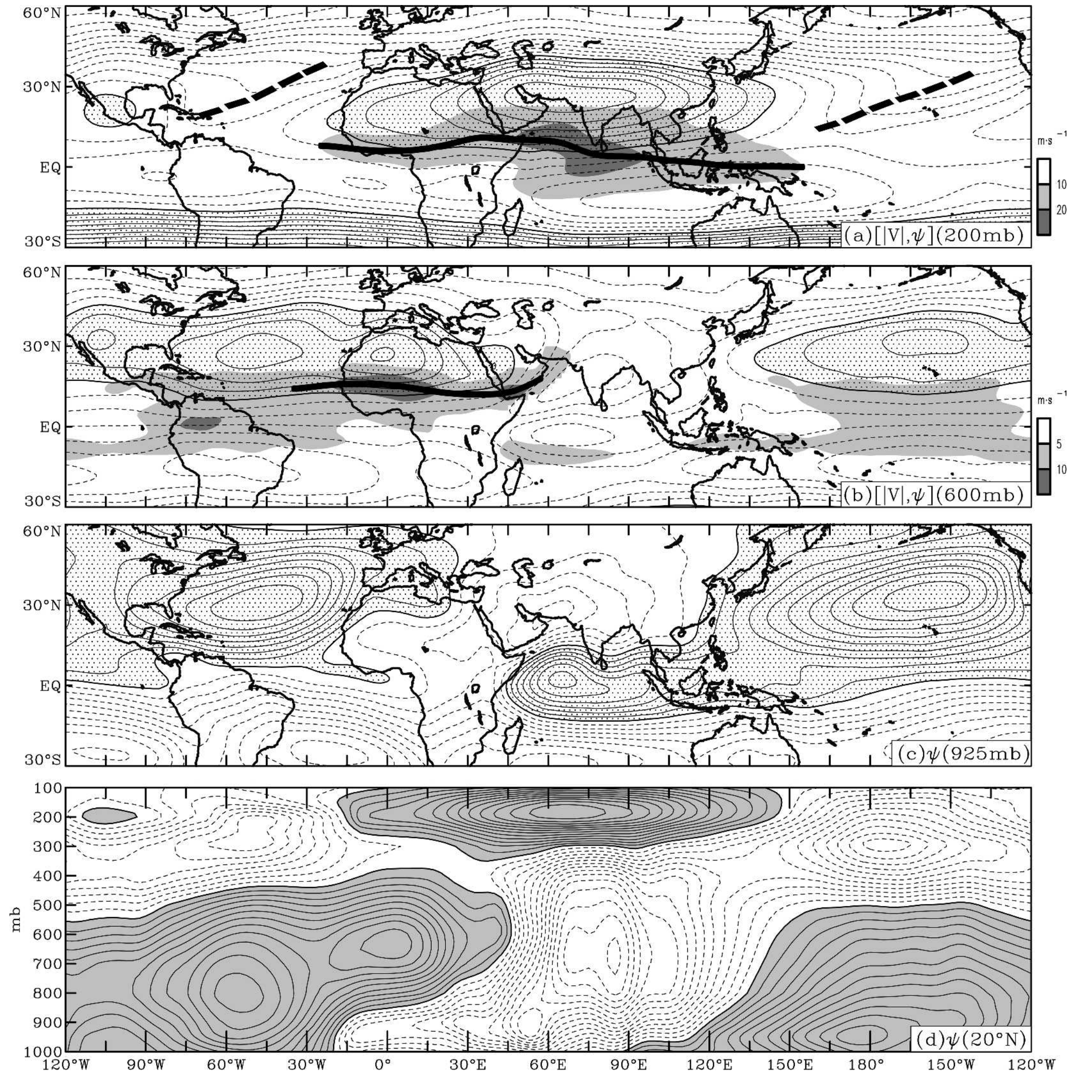

FIG. 1. Observed summer (JJA) streamfunction (ψ) at (a) 200, (b) 600, and (c) 925 mb and (d) longitude–height cross section at 20°N. Positive values of ψ are dotted in (a)–(c) and stippled in (d). Isotachs (stippled areas), oceanic trough lines (thick dashed lines), and cores of the tropical easterly jet (thick solid line at 200 mb) and the African easterly jet (thick solid line at 600 mb) are superimposed on the horizontal ψ fields. Contour intervals of ψ are 5×10^6 m^2 s^{-1} at 200 mb, 3×10^6 m^2 s^{-1} at 600 mb, 2×10^6 m^2 s^{-1} at 925 mb, and 2×10^6 m^2 s^{-1} above 300 mb and 10^6 m^2 s^{-1} below 300 mb in (d).

circulation resembles Thorncroft and Blackburn's numerical simulation, some aspects of the North African summer circulation may be related to this meridional circulation:

1) As revealed from Fig. 2c, the lower-tropospheric monsoon southerly flow of West Africa is basically established by the lower branch of the Hadley circulation. The West African monsoon rainfall (Fig. 2b) is generated by the upward branch of the Hadley circulation along the ITCZ, instead of the Saharan meridional circulation.
2) The tropical easterly jet in the upper troposphere (Figs. 2a,b) is located aloft over the *upward* branch of the Hadley circulation, while the AEJ is situated over the *downward* branch of the southern Saharan circulation, somewhat more equatorward than Thorncroft and Blackburn's simulation. What is the role of the North African summer divergent circulation in the formation and maintenance of the AEJ?

It was shown in Fig. 1b that the AEJ is a part of the midtropospheric Saharan high. The formation/

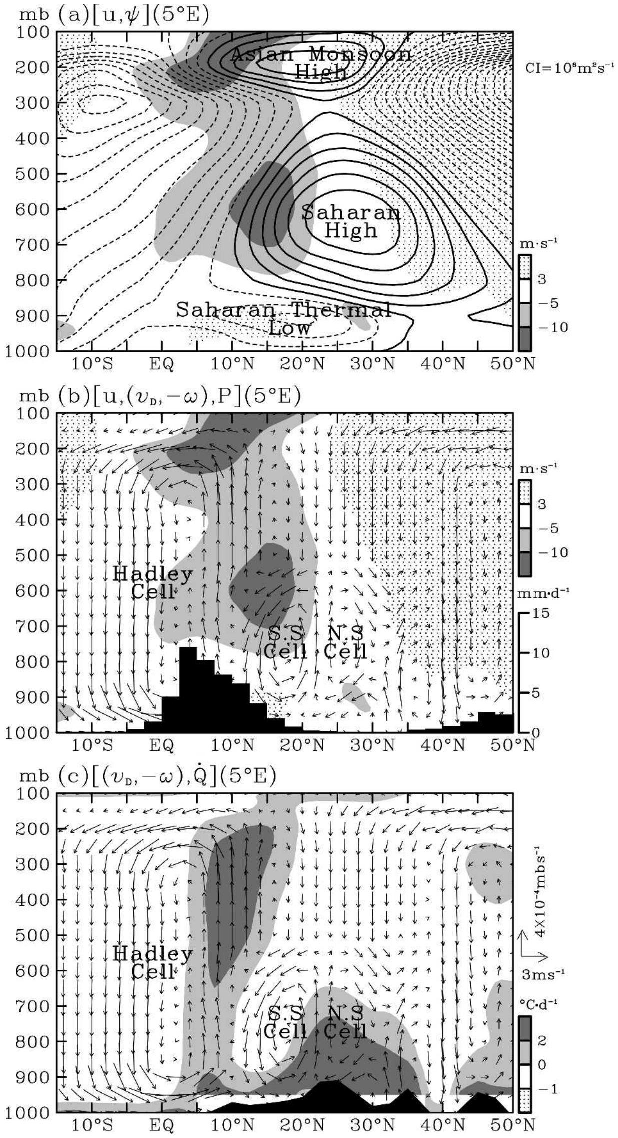

FIG. 2. Summer latitude–height cross sections of (a) streamfunction ψ (5°E) superimposed with zonal wind u (5°E) (stippled areas), (b) meridional circulation (v_D, $-\omega$) (5°E) superimposed with zonal wind u(5°E) (stippled area) and precipitation (histograms), and (c) meridional circulation superimposed with diabatic heating $\dot{Q}/c_p$(5°E) ($\dot{Q}/c_p \geq 0$ stippled areas, while $\dot{Q}/c_p \leq 0$ dotted areas) and topography (dark areas). These cross sections were averaged over the longitudinal zone of 0°–10°E.

maintenance of this midtropospheric jet should be linked to the Saharan high. How is this North African high maintained? As will be shown below, the North African divergent circulation in the lower troposphere includes the North African local Hadley circulation and the eastern Atlantic–North African east–west circulation. In spite of Thorncroft and Blackburn's (1999) emphasis of the possible maintenance of AEJ by the meridional circulation, can the east–west circulation play a role in maintaining the Saharan high?

In the summer Northern Hemisphere, the Asian and North American monsoons are maintained by the east–west differential heating through the Walker-type east–west circulation (Chen 2003). Can the North African summer circulation in the lower half of the troposphere be maintained by the same mechanism? To answer this question, the divergent circulation at 200 and 925 mb and the east–west circulation (u_D, $-\omega$) at 17.5°N (along the Saharan thermal low) superimposed with diabatic heating $\dot{Q}$ are displayed in Figs. 3a,c,d, respectively. As shown in Fig. 3d, the upward (downward) branches of the east–west circulation associated with the Indian and North American monsoons coincide with the upper-level divergent (convergent) and diabatic heating (cooling) centers. These east–west circulations penetrate the entire troposphere. In contrast, the North African summer circulation (including the Saharan high and thermal low; Fig. 1d) is overlaid by the downward branch of the Indian monsoon's east–west circulation (Fig. 3d). Actually, three interesting features of the divergent circulation over North Africa are revealed from a careful inspection of Fig. 3:

1) A well-organized 600-mb divergent center (Fig. 3b) in the Chad–Sudan region is located aloft over the 925-mb convergent center along the Saharan thermal low (Fig. 3c). It may be inferred from the east–west circulation over North Africa (Fig. 3d) that this 600-mb divergent center is formed by the downward branch of the Asian east–west circulation and upward motion along the Saharan thermal low.
2) An east–west circulation between the eastern subtropical Atlantic and North Africa (Fig. 3d) is established by the Saharan upward motion below 600 mb and the downward branch of the east–west circulation associated with both the North Atlantic oceanic trough in the upper troposphere and the North Atlantic anticyclone in the lower troposphere. As indicated by diabatic heating, the east–west circulation described here is developed by the east–west differential heating between the surface heating (heavily stippled areas) over the Saharan thermal low and the diabatic cooling (dotted areas) over the eastern subtropical Atlantic.
3) A spatial quadrature relationship exists between the Saharan high at 600 mb (Fig. 1b) and the east–west circulation over the eastern subtropical Atlantic–North African region (Fig. 3d).

Despite being limited below 500 mb, the monsoonlike North African summer circulation may be maintained

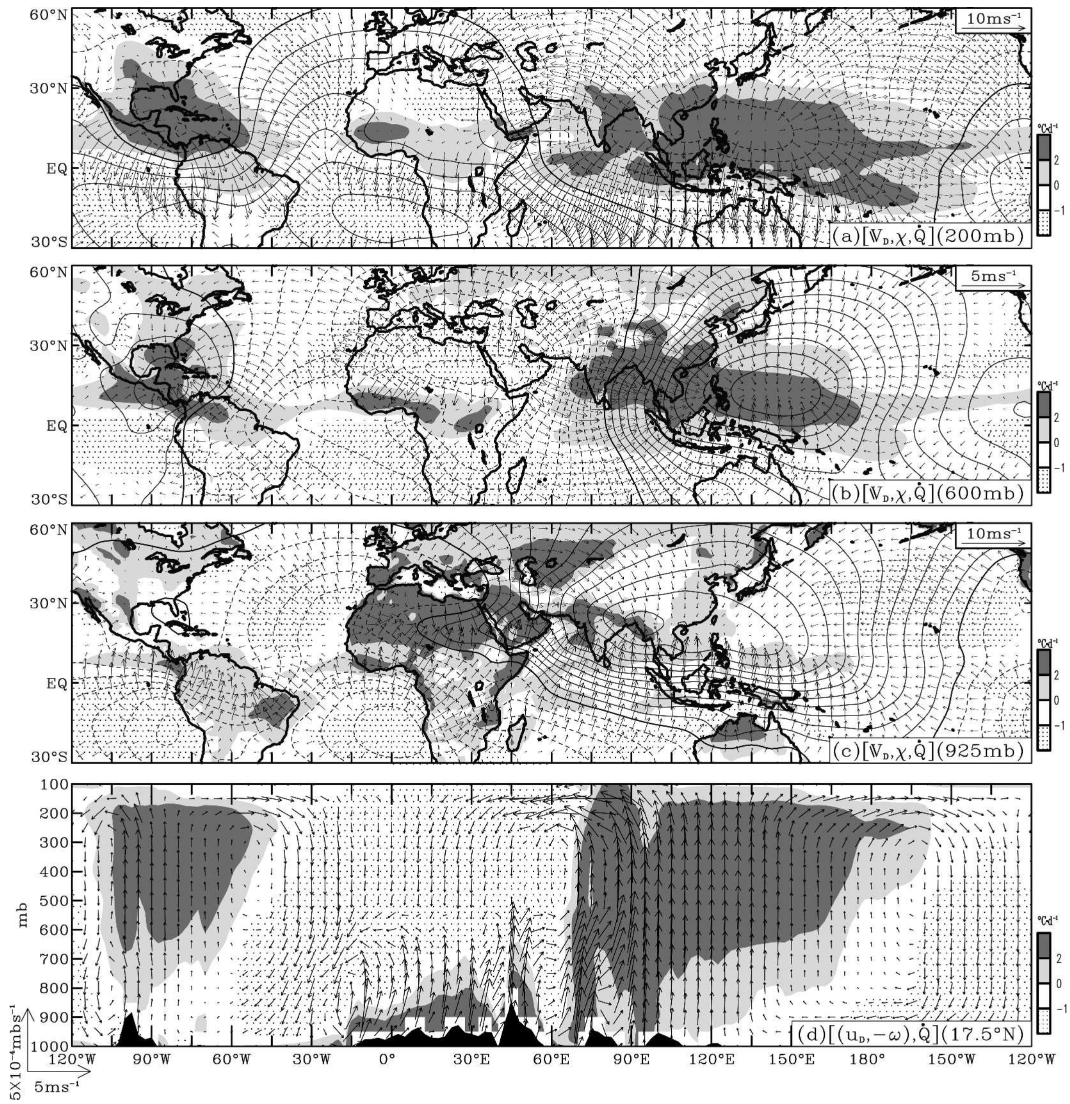

FIG. 3. Observed summer velocity potential (χ) and divergent wind vector (V_D) at (a) 200, (b) 600, and (c) 925 mb and (d) the east–west circulation (u_D, $-\omega$) at 17.5°N superimposed with diabatic heating/cooling (shaded/dotted areas). Contour intervals of χ is 2×10^6 m^2 s^{-1} at 200 mb, 5×10^5 m^2 s^{-1} at 600 mb, and 10^6 m^2 s^{-1} at 925 mb.

by a mechanism similar to the maintenance mechanism of major monsoons illustrated by Chen (2003).

3. Maintenance of the North African divergent circulation

The midtropospheric divergent center over North Africa (Fig. 3b) is a unique feature of the global divergent circulation. It was suggested in section 2 that the formation of this divergent center over North Africa was attributed to the convergence formed by upward motion induced by the Saharan surface heating and downward motion accompanying the upper-tropospheric cooling. This hypothesized formation mechanism of the North African divergent center is tested with the following diagnoses, which include analyses of the heat budget and the velocity potential maintenance equation.

a. Heat budget

The long-term mean heat budget equation may be written as

$$\mathbf{V} \cdot \nabla T - \sigma\omega = \frac{1}{c_p}\dot{Q}, \tag{1}$$

where $\mathbf{V}$, T, σ, ω, $\dot{Q}$, and c_p are wind vector, temperature, static stability, p-vertical velocity, diabatic heating, and specific heat with constant pressure, respectively. Thermal advection may likely be significant when the ambient flow is strong. During the northern summer, the upper-level zonal winds near the jet in East Africa [inferred with ψ(200 mb) in Fig. 1a] and the low-level Mediterranean northerlies in central North Africa [e.g. Cook 1999; inferred with ψ(925 mb) in Fig. 1c] may make thermal advection significant in these two regions. Actually, it was revealed from our area-averaged heat budget analyses over North Africa (equator–45°N, 20°W–45°E) at 850, 700, and 500 mb (not shown) that thermal advection is always less than 10% of diabatic heating. Thus, vertical motion, which is a part of divergent circulation, over North Africa may be primarily induced by diabatic heating,

$$-\sigma\omega \simeq \frac{1}{c_p}\dot{Q}. \qquad (1)'$$

b. Velocity potential maintenance equation

The link between divergent circulation depicted by velocity potential (χ) and diabatic heating ($\dot{Q}$) can be established by combining the heat and continuity equations (Chen and Yen 1991a,b),

$$\chi = \nabla^{-2}\left[\frac{\partial}{\partial p}\left(\frac{1}{\sigma}\mathbf{V}\cdot\boldsymbol{\nabla}T\right)\right] + \nabla^{-2}\left[-\frac{\partial}{\partial p}\left(\frac{1}{c_p\sigma}\dot{Q}\right)\right]. \qquad (2)$$

Following the approximation adopted by Eq. (1), one may attain

$$\chi \simeq \nabla^{-2}\left[-\frac{\partial}{\partial p}\left(\frac{1}{c_p\sigma}\dot{Q}\right)\right] \equiv \chi_{\dot{Q}}. \qquad (2)'$$

It is indicated by Eq. (2)′ that the χ field (with which divergent circulation is portrayed) is essentially maintained by *vertical differential heating.* According to Eq. (2)′, the upper-level diabatic cooling over North Africa and the Saharan surface heating provide the necessary vertical differential heating to maintain the North African divergent center. Replacing $\dot{Q}$ in Eq. (2)′ by ω expressed in Eq. (1)′, one obtains

$$\chi_{\dot{Q}} \simeq \nabla^{-2}(\boldsymbol{\nabla}\cdot\mathbf{V}) = \chi, \qquad (3)$$

after application of the continuity equation.

Because the ω and $\dot{Q}$ fields were not initially specified, these variables generated by the global data assimilation system would be biased by the model spinup during the data assimilation. In order avoid the effect of spinup, the ω and $\dot{Q}$ fields produced by the reanalysis were not directly used in our diagnoses. These two variables were computed by the adjusted kinematics method (O'Brien 1970) and the residual method of the heat budget (e.g., Chen and Baker 1986), respectively. The ω field and the heat budget at 500 and 850 mb over North Africa are shown in Figs. 4a–c and 4d–f, respectively, without heat advection, which is much smaller in magnitude than other quantities in the heat budget. As expected by Eq. (1)′, both $-\sigma\omega$ (adiabatic warming/cooling; Figs. 4b and 4e) and $\dot{Q}/c_p$ (diabatic heating/cooling; Figs. 4c and 4f) are very much alike in their spatial distributions and comparable in their magnitudes. As shown in Fig. 4, vertical motion over North Africa is basically induced by diabatic heating/cooling: ascending motion along the Saharan heat low is excited by surface heating, while descending motion north of the ITCZ is driven by the upper-level cooling. Upward motion over the Saharan heat low is capped by the downward branch of the east–west circulation of the Indian monsoon. Actually, the lack of moisture supply to maintain deep cumulus convection is perhaps another major cause of the Saharan shallow convection.

Can the Saharan surface heating and the upper-level cooling over North Africa form the 600-mb divergent center, as expected with Eq. (2)′? To answer this question, $\chi_{\dot{Q}}$ was computed by the following form:

$$\chi_{\dot{Q}}(600\,\text{mb}) = \nabla^{-2}\left[\frac{\left(\frac{1}{\sigma}\dot{Q}\right)(500\,\text{mb}) - \left(\frac{1}{\sigma}\dot{Q}\right)(700\,\text{mb})}{-c_p(500\,\text{mb} - 700\,\text{mb})}\right], \qquad (2)''$$

The $\chi_{\dot{Q}}$(600 mb) field generated with Eq. (2)″ is shown in Fig. 5a. The comparison of $\chi_{\dot{Q}}$(600 mb) with χ(600 mb) shown in Fig. 3b is measured by the ratio rms [$\chi_{\dot{Q}}$(600 mb) $-$ χ(600 mb)]/rms[χ(600 mb)] (where rms is root mean square value), which is about 2%. The small difference between $\chi_{\dot{Q}}$(600 mb) and χ(600 mb) strongly indicates that the 600-mb Saharan divergent center is formed and maintained by vertical differential heating.

Cook (1999) emphasized the importance of meridional gradients of soil moisture between the West African coast and the Saharan desert to the AEJ intensity, while Thorncroft and Blackburn (1999) demonstrated the maintenance of the AEJ by the ITCZ and Saharan thermal-low heatings. What these studies stressed is the effect of the north–south differential heating on the formation of the AEJ through the local Hadley circulation. In contrast, the lower-tropospheric east–west differential heating between North Africa and the eastern Atlantic Ocean, shown in Fig. 3d, also maintains a shallow east–west circulation. Both the east–west and meridional circulations over North Africa are parts of the global divergent circulation of this continent. What impact would diabatic heating along the ITCZ and

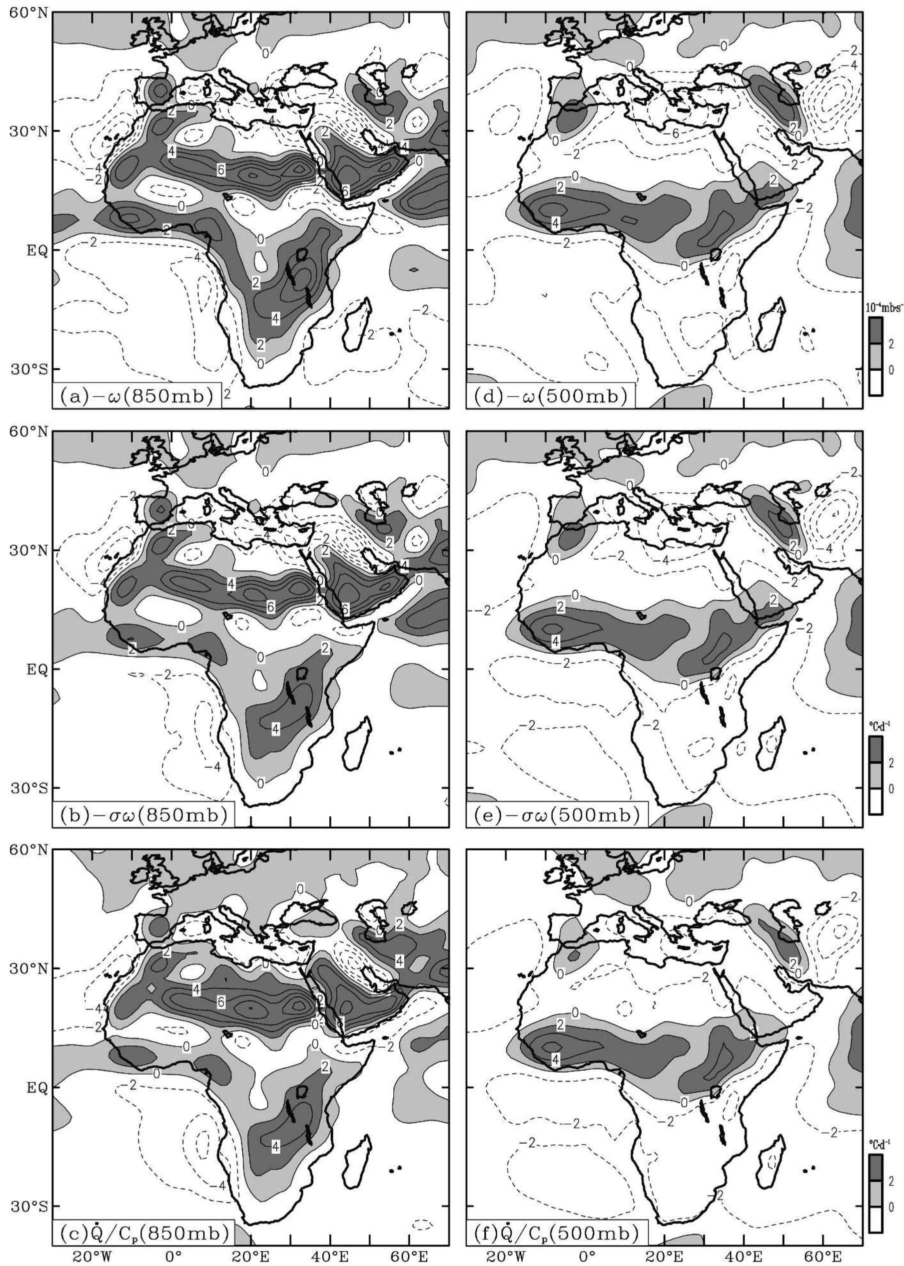

FIG. 4. The summer ω (p-vertical velocity) fields at (a) 850 and (d) 500 mb and the summer heating budget at (b), (c) 850 and (e), (f) 500 mb over all of Africa. Positive values of all variables are stippled.

the Saharan thermal low have on the midtropospheric North African divergent center? To answer this question, the $\chi_{\dot{Q}}$(600 mb) field *without* these heatings is computed. For convenience of discussion, let us designate $\chi_{\dot{Q}S}$ to be $\chi_{\dot{Q}}$ without the contribution of the Saharan thermal-low heating, and $\chi_{\dot{Q}I}$ without the contribution of the ITCZ heating. These two variables are shown in Figs. 5b and 5c, respectively. The comparison of $\chi_{\dot{Q}}$(600 mb) with these two variables reveals the following salient features:

1) Without the Saharan thermal-low heating, the east–west differential heating across North Africa is formed between the continent and the Saudi Arabian peninsula. The North African divergent center disappears from $\chi_{\dot{Q}S}$(600 mb) (Fig. 5b), and the

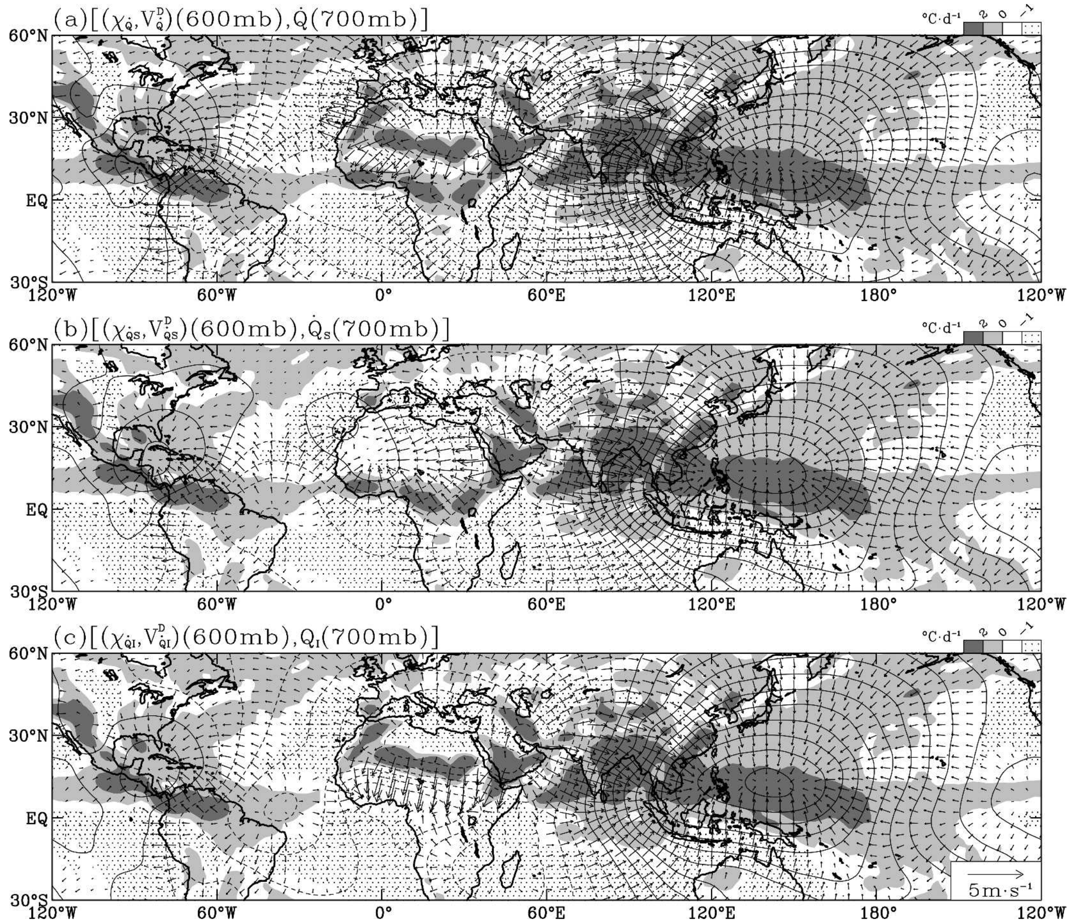

FIG. 5. The summer 600-mb thermal velocity potential ($\chi_{\dot{Q}}$) and divergent wind vectors ($V_{\dot{Q}}^D \equiv \nabla\chi_{\dot{Q}}$) superimposed with the 700-mb diabatic heating [$\dot{Q}$(700 mb)]: (a) total diabatic heating, (b) the Saharan thermal-low heating removed, $\chi_{\dot{Q}S}$, and (c) the ITCZ heating removed, $\chi_{\dot{Q}I}$. The contour interval of $\chi_{\dot{Q}}$, $\chi_{\dot{Q}S}$, and $\chi_{\dot{Q}I}$ is 5×10^5 m^2 s^{-1}.

midtropospheric divergent circulation breaks into a dipole with a convergent center at the northwestern coast of North Africa and a divergent center over the southern Saudi Arabian peninsula. The disappearance of the North African divergent center from $\chi_{\dot{Q}S}$(600 mb) clearly indicates that this divergent center is formed and maintained by the Saharan thermal-low heating.

2) Without the ITCZ heating, the North African divergent center of $\chi_{\dot{Q}I}$(600 mb) (Fig. 5c) moves slightly northward. The weak divergence zone of $\chi_{\dot{Q}}$(600 mb) along the ITCZ is replaced by a West African convergent center of $\chi_{\dot{Q}I}$(600 mb). This change in the north–south differential heating (by removing the ITCZ heating) eliminates the divergent flow from the ITCZ heating zone and lets the North African air converge toward the equator.

These salient features of the $\chi_{\dot{Q}S}$(600 mb) and $\chi_{\dot{Q}I}$(600 mb) fields over North Africa support the argument that the midtropospheric North African divergent center is formed by the Saharan thermal-low heating and fixed aloft along the Saharan thermal low by the ITCZ heating. To strengthen this argument, we also generate the velocity potential fields with *only* the Saharan thermal-low heating $\chi_{\dot{Q}}^S$ and the ITCZ heating $\chi_{\dot{Q}}^I$ in Fig. 6. To warrant computational accuracy, both $\chi_{\dot{Q}}$(600 mb) = $\chi_{\dot{Q}S}$(600 mb) + $\chi_{\dot{Q}}^S$(600 mb) and $\chi_{\dot{Q}}$(600 mb) = $\chi_{\dot{Q}I}$(600 mb) + $\chi_{\dot{Q}}^I$(600 mb) were verified. The $\chi_{\dot{Q}}^S$(600 mb) field (Fig. 6b) exhibits a well-organized *divergent* center with an intensity comparable to the North African χ (600 mb) center in Fig. 3b and $\chi_{\dot{Q}}$(600 mb) center in Fig. 5a. The comparison between $\chi_{\dot{Q}}^S$(600 mb) and $\chi_{\dot{Q}S}$ (600 mb) in Fig. 5b clearly shows that the Saharan thermal-low heating is the major agent responsible for the formation of the North African divergent center. In contrast, the ITCZ heating alone generates a *convergent* center over this heating. The slight northward shift of

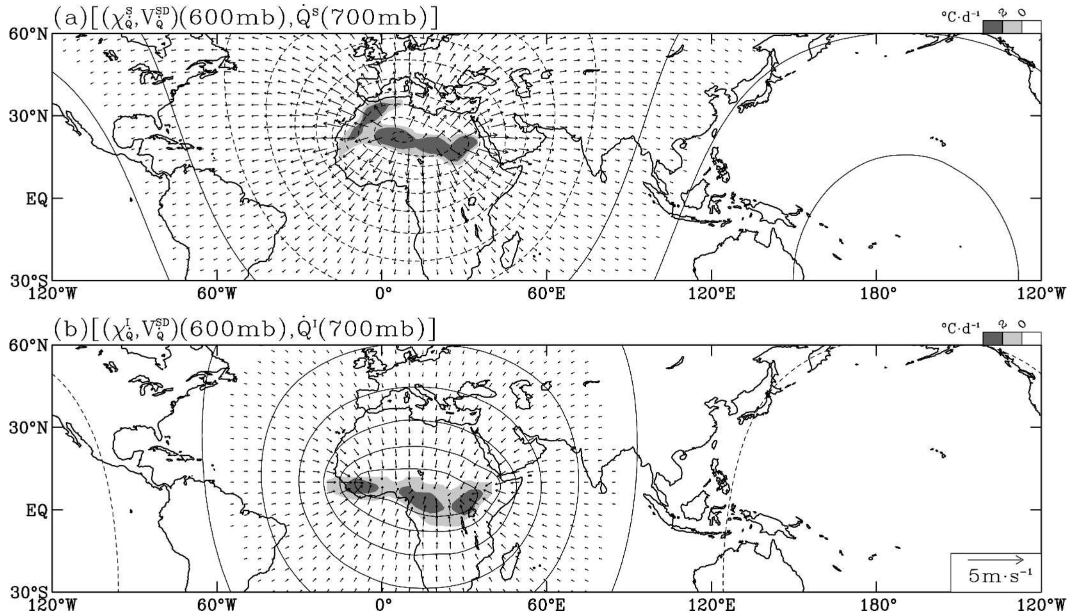

FIG. 6. As in Fig. 5 except for the thermal velocity potential generated with only (a) the Saharan thermal-low heating $\chi_{\dot{Q}}^{S}$(600 mb) and (b) the ITCZ heating $\chi_{\dot{Q}}^{I}$(600 mb).

the North African $\chi_{\dot{Q}I}$(600 mb) center in Fig. 5c is caused by the absence of the $\chi_{\dot{Q}}^{I}$(600 mb) center (Fig. 6b).

After this lengthy demonstration, we confidently claim that the midtropospheric North African divergent center is formed by the vertical differential heating between the Saharan thermal-low heating at lower levels and the Saharan radiative cooling at upper levels. The effect of the Saharan thermal-low heating on the midtropospheric Saharan high is achieved through the dynamic interaction between the North African divergent center and the Saharan high. This dynamic interaction will be illustrated in section 4.

4. Maintenance of the Saharan high

What is the formation/maintenance mechanism of the midtropospheric Saharan high? The midtropospheric high is centered over the high ground of Northwest Africa (Fig. 1b), while the 600-mb North African divergent center is located (along the Saharan thermal low) over the Chad–Sudan region. This spatial quadrature relationship between these two midtropospheric circulation elements suggests that the maintenance mechanism of major summer monsoon circulations by east–west differential heating (Chen 2003) may be applicable to the Saharan high. A two-step diagnosis is adopted to test this suggestion: (i) the streamfunction budget analysis and (ii) simulations by a forced barotropic model.

a. Streamfunction budget analysis

The vorticity equation is often used to illustrate dynamical processes maintaining the atmospheric circulation. Since the atmospheric circulation can be well depicted by the streamfunction, the maintenance of the atmospheric circulation may be elucidated in terms of the streamfunction budget, which is a Laplace inverse transform of the vorticity equation (Sanders 1984):

$$\underset{-\psi_{A1}}{\nabla^{-2}\left(u_Z \frac{\partial \zeta}{\partial x}\right)} + \underset{-\psi_{A2}}{\nabla^{-2}(v\beta)} = \underset{\psi_\chi}{\nabla^{-2}(-f\nabla \cdot \mathbf{V})}. \qquad (4)$$

The Saharan high is well depicted by ψ(600 mb) in Fig. 1b. The budget analysis of Eq. (4) will be used to explain this midtropospheric high maintenance. As revealed from the ψ and χ fields in Figs. 1 and 3, respectively, the tropical summer circulation is dominated by planetary-scale waves 1–2 (e.g., Holton and Colton 1972) and the global divergent circulation by wave 1 (e.g., Krishnamurti 1971). It was previously pointed out that the North African summer circulation is overlaid by the western part of the Tibetan high. Thus, the depiction and maintenance mechanism of the North African summer circulation can be easily masked by the Indian monsoon. To avoid this masking in the diagnosis, it is revealed from our analysis that the best wave regime to portray the Saharan high is waves 2–15. For this reason, we shall focus the ψ budget analysis of Eq. (4) on this wave regime, which is denoted

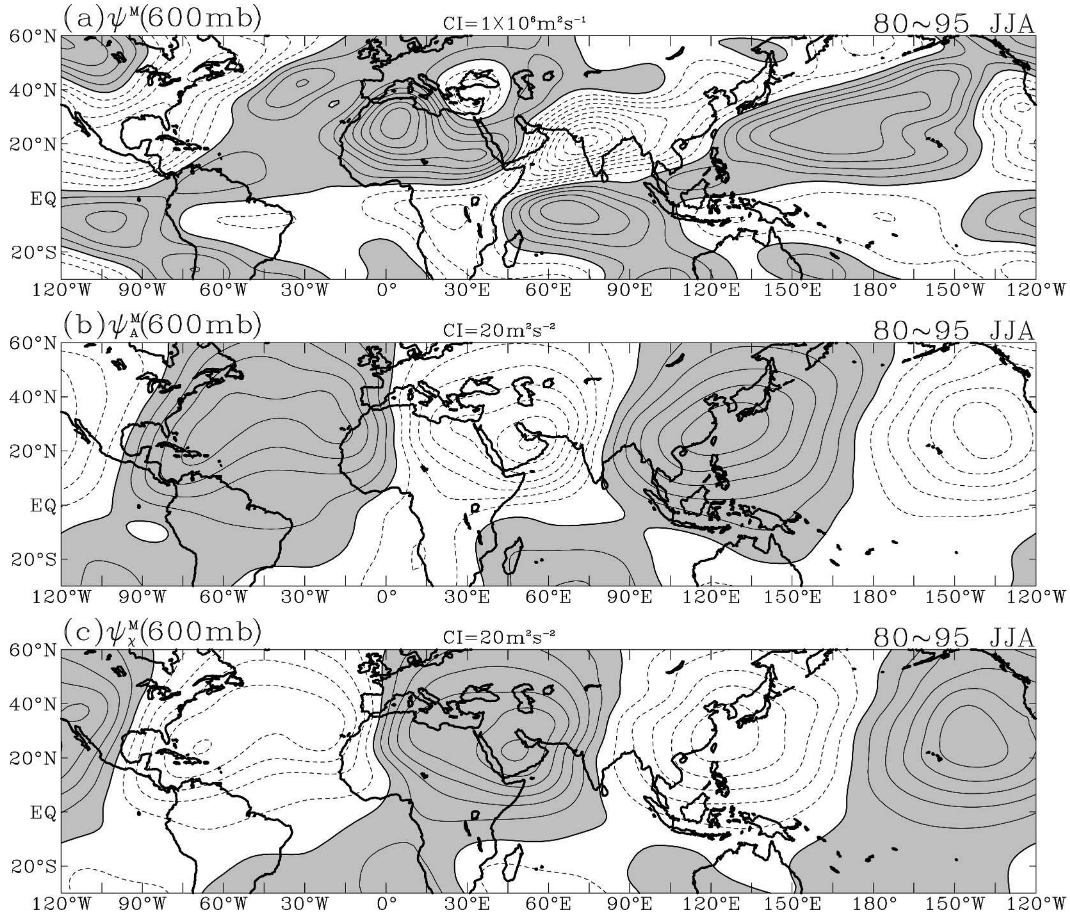

FIG. 7. The 600-mb summer streamfunction budget in the M-wave (waves 2–15) regime: (a) ψ^M(600 mb), (b) ψ_A^M(600 mb), and (c) ψ_χ^M(600 mb). Contour intervals of ψ^M, ψ_A^M, and ψ_χ^M are 10^6 m^2 s^{-1} and 20 m^2 s^{-2} for the latter two fields, respectively.

by $(\ )^M$. Equation (4) in this wave regime may be written as

$$0 \simeq \psi_{A1}^M + \psi_{A2}^M + \psi_\chi^M. \tag{5}$$

In low latitudes, the dynamics of the planetary-scale summer circulations are represented by the Sverdrup vorticity balance (i.e., $v\beta = -f\nabla \cdot \mathbf{V}$). Note that the tropical easterly jet and AEJ can reach their maximum speed of 25 and 13 m s^{-1}, respectively, and ψ_{A1} may not be completely negligible in low latitudes. However, the low-latitude easterlies enable the westward relative vorticity advection to work collaboratively with the north–south advection of planetary vorticity in a way opposite to wintertime stationary waves in midlatitudes. Consequently, the inclusion of both ψ_{A1} and ψ_{A2} in Eq. (4) (may be denoted by $\psi_{A=}\psi_{A1+}\psi_{A2}$) does not alter the basic dynamics of the Sverdrup vorticity balance.

The ψ^M(600 mb) budget is presented in Fig. 7. Recall that the Saharan high [indicated by the positive center of ψ(600 mb) over northwest Africa in Fig. 1b] and the North African divergent center [indicated by the negative cell of χ(600 mb) over the Chad–Sudan region in Fig. 3b] are spatially in quadrature. A spatial quadrature relationship is expected to exist between the Saharan high cell of ψ^M(600 mb) and ψ_χ^M(600 mb). The positive (negative) cell of ψ_χ^M(600 mb) over North Africa corresponds to upward (downward) motion as shown in Fig. 3d. This contrast between vertical motion and ψ_χ^M(600 mb) substantiates our hypothesis that the Saharan high is maintained by the east–west differential heating (between the Saharan thermal-low heating and the eastern Atlantic cooling) through the east–west circulation (u_D, $-\omega$) (17.5°N) over North Africa and the east Atlantic shown in Fig. 3d. The resemblance of spatial structure and comparable magnitude between ψ_χ^M (600 mb) (Fig. 7c) and ψ_A^M(600 mb) (Fig. 7b) strongly indicates that the dynamics of the Saharan high is well illustrated by the Sverdrup vorticity balance. This balance provides the dynamical foundation for simulations of the Saharan high by a forced barotropic model in section 4b.

b. Simulations by a forced barotropical model

The velocity potential maintenance analysis [Eq. (2)′] in section 3 showed that the Saharan thermal-low heating is the primary agent in maintaining the midtropospheric North African divergent center. The 600-mb streamfunction budget analysis [Eq. (4)] revealed that the Saharan high is maintained by a counterbalance between vortex stretching and vorticity advection. The former dynamical process provides the impact of the Saharan thermal-low heating on the Saharan high through the North African divergent center. This interaction can be established by a combination of Eqs. (2)′ and (4):

$$\nabla^{-2}\left(u_Z \frac{\partial \zeta}{\partial x} + v\beta\right) \cong \nabla^{-2}\left[-f \frac{\partial}{\partial p}\left(\frac{1}{c_p \sigma}\dot{Q}\right)\right]. \quad (6)$$

To substantiate our inference, the possible link between the Saharan thermal-low heating and the Saharan high is tested with a linearized forced barotropic model on a sphere introduced by Branstator (1983):

$$\frac{u_Z}{a\cos\varphi}\frac{\partial}{\partial\lambda}(\nabla^2\psi_E) + \frac{1}{a^2\cos\varphi}\left(\frac{\partial\psi_E}{\partial\lambda}\right)\frac{\partial}{\partial\varphi} \times \left[f - \frac{1}{a\cos\varphi}(u_Z\cos\varphi)\right] = F - \alpha\nabla^2\psi_E + \gamma\nabla^2\psi_E, \quad (7)$$

where a, λ, φ, α ($= 1.57 \times 10^{-6}$ s^{-1} $\sim$ 7 days) and γ ($= 2.34 \times 10^{16}$ m^4 s^{-1}) are the earth's radius, longitude, latitude, drag, and diffusion coefficients, respectively. Mathematical and numerical details of the model can be found in Branstator (1983). Equation (7) is solved in terms of spherical harmonics with a 20-wave triangular truncation. The 600-mb zonal-wind flow used in this model [Eq. (7)] is shown in Fig. 8. The tropical region is dominated by easterlies. Following Eqs. (2)″ and (6), the forcing F in Eq. (7) is expressed by the following form:

$$F = -f\left[\frac{\left(\frac{1}{\sigma}\dot{Q}\right)(500\text{ mb}) - \left(\frac{1}{\sigma}\dot{Q}\right)(700\text{ mb})}{c_p(500\text{ mb} - 700\text{ mb})}\right]. \quad (8)$$

The Saharan high is an integral part of the North African summer circulation. It may be misleading to simulate this midtropospheric North African anticyclone in terms of a forced barotropic model. Therefore, clarification is in order. A barotropic model of the atmosphere conventionally represents the conservation of total vorticity. In this model, vorticity can be changed only by total vorticity advection. In contrast, the vortex stretching term is included in Eq. (4), which is actually a linearized Rossby potential vorticity equation (Holton 2004). By virtue of Eq. (1)′ or Eq. (2)′, vortex stretching can be expressed in terms of vertical differential heating in Eq. (6). Therefore, the forced barotropic model of Eq. (7), which does not portray the evolution of the convectional barotropy of an atmosphere, is basically the same as the one used by Gill (1980) to depict analytically a monsoon circulation, except no ambient flow was included in the Gill's model.

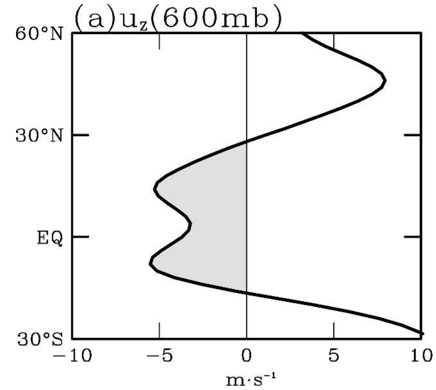

FIG. 8. The zonal-mean summer flow at 600 mb, u_Z(600 mb), used in the forced barotropical model.

Following our examination of the possible effects of the Saharan thermal-low and ITCZ heating on the North African divergent center, three experiments were performed with the forced barotropic model: 1) *Control* experiment (EC: with the full-diabatic heating imposed), 2) *Saharan thermal-low heating* experiment (ES: with the Saharan thermal-low heating removed), and 3) *ITCZ heating* experiment (EI: with the ITCZ heating removed). Eddy streamfunctions generated by these three experiments are denoted by ψ_{EC}, ψ_{ES}, and ψ_{EI}. To accompany the last two experiments, two additional experiments are preformed with Gill's (1980) approach to further our understanding of the effect of the Saharan thermal low and ITCZ heating on the formation of the midtropospheric North African anticyclone. These two extra experiments are designated as ψ_G^S and ψ_G^I, respectively.

1) CONTROL EXPERIMENT (EC)

The ψ_{EC}(600 mb) field in Fig. 9b captures all major features of the observed ψ_E(600 mb) field [Fig. 9a; ob-

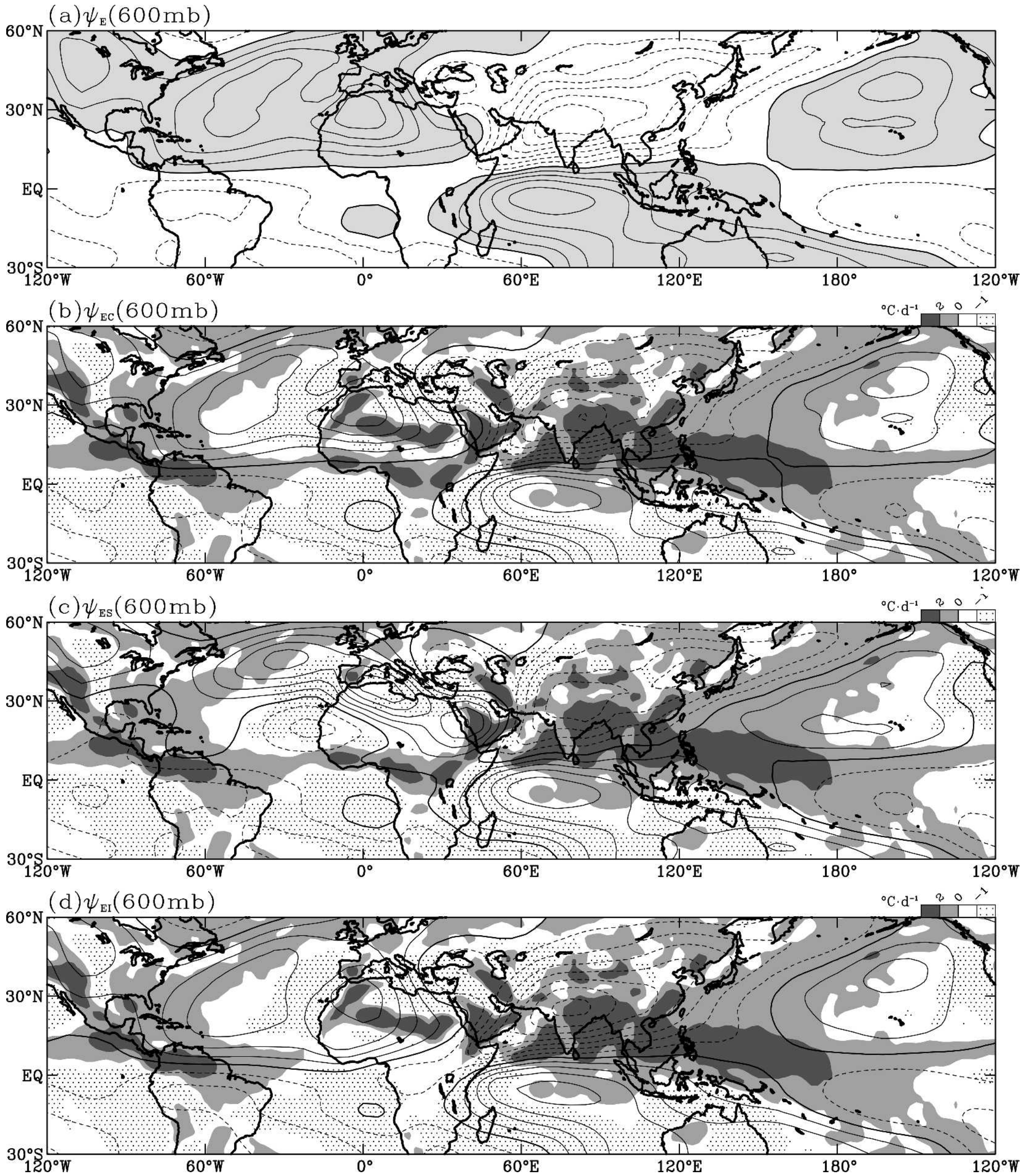

FIG. 9. Summer eddy streamfunction at 600 mb: (a) observed (ψ_E), (b) EC (ψ_{EC}) with total diabatic heating, (c) ES (ψ_{ES}) with the Saharan thermal-low heating removed, and (d) EI (ψ_{EI}) with the ITCZ heating removed. The contour interval of ψ_E, ψ_{EC}, ψ_{ES}, and ψ_{EI} is 2×10^6 m^2 s^{-1}.

tained by removing the zonal component of ψ(600 mb) in Fig. 1b]: the Saharan high, the two oceanic anticyclones, and the Indian monsoon trough. The bias of ψ_{EC}(600 mb) against the observed, measured by the ratio rms[ψ_{EC}(600 mb) $-$ ψ_E(600 mb)]/rms[ψ_E(600 mb)], is about 1.5%. The simulation of the Saharan high with such a small model bias strongly suggests that the forced barotropic model [Eq. (7)] can be used to perform simple sensitivity experiments of the Saharan thermal low and the ITCZ heating.

2) SAHARAN THERMAL-LOW HEATING EXPERIMENT (ES)

It was shown in Fig. 5b that the removal of the Saharan thermal-low heating results in a pronounced change in the midtropospheric divergent circulation over North Africa. The disappearance of the North African divergent center leads to the emergence of a convergent center over northwest Africa and the direction reversal of some divergent flows out of the midtropo-

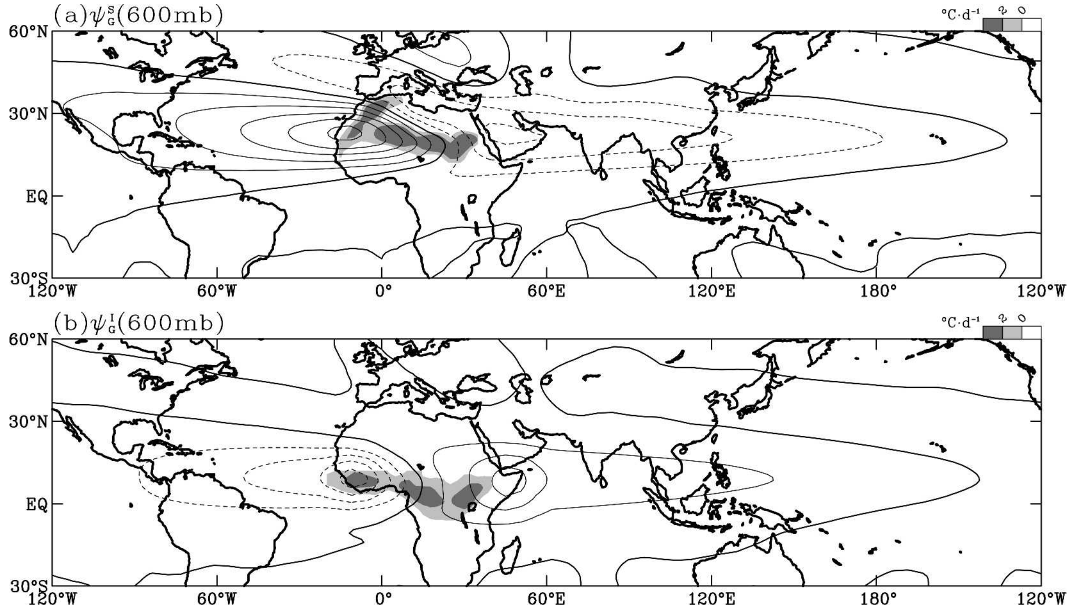

FIG. 10. As in Fig. 8 except for responses of the forced barotropic model to (a) the Saharan thermal-low heating ψ_G^S(600 mb) and (b) the ITCZ heating ψ_G^I(600 mb). The contour interval of ψ_G^S and ψ_G^I(600 mb) is 2×10^6 m^2 s^{-1}.

spheric Saharan divergent center. These changes in the midtropospheric divergent circulation have *two* important impacts on the Saharan high:

(i) As inferred from the ψ(600 mb) budget, the Saharan high depicted by ψ_E(600 mb) should be spatially in quadrature with the two $\chi_{\dot{Q}S}$(600 mb) centers over North Africa. The simulated ψ_{ES}(600 mb) over northwest Africa shown in Fig. 9c exhibits such an expected structure;

(ii) another interesting feature of ψ_{ES}(600 mb) over western North Africa is the northward shift of the positive ψ_{ES}(600 mb) center along with the appearance of a minor negative ψ_{ES}(600 mb) center in West Africa. This change in ψ_E(600 mb) results in the northward shift of the AEJ and a short wave train around North Africa.

Apparently, the reversal of the 600-mb divergent flows south of the Saharan divergent center produces an eastward Coriolis acceleration to slow the AEJ. In their numerical experiment, Thorncroft and Blackburn (1999) showed that the lower-tropospheric easterlies centered at 17°N can be produced by the ITCZ heating. The West African zonal flows simulated by the forced barotropic model are consistent with Thorncroft and Blackburn's result. What may cause the northward shift of the positive ψ_{ES}(600 mb) cell over West Africa? This question may be answered by a Gill-type experiment. The response of the forced barotropic model to the Saharan thermal-low diabatic heating in an experiment designated as QS is shown in Fig. 10a. In addition to a short wave train emanating from West Africa to Europe, a well-organized dipole of ψ_G^S(600 mb)anomalies emerges between West Africa and Saudi Arabia. By removing ψ_G^S(600 mb) anomalies from ψ_{EC}(600 mb), it is not surprising to have the appearance of a negative ψ_{ES}(600 mb) cell over West Africa and the weakening of the negative ψ_{EC}(600 mb) cell centered over northern India.

As inferred from the ψ^M(600 mb) budget analysis (Fig. 7) in section 4a, the Saharan high is maintained by the east–west differential heating through the North African divergent circulation with a divergent center over the Chad–Sudan region and a convergent center over the eastern Atlantic. The maintenance mechanism of the Saharan high by the east–west differential heating is further substantiated by simulations with the simple forced barotropic model.

3) ITCZ HEATING EXPERIMENT (EI)

Noticeable changes in the midtropospheric North African divergent circulation caused by the removal of the ITCZ heating are that (i) the North African divergent center not only becomes weaker, but also migrates northward from central North Africa to the Mediterranean Sea, and (ii) the weak divergent center along the ITCZ is replaced by a strong convergence zone along the equator. This equatorward divergent wind zone is driven/maintained by the meridional differential heating between the Saharan thermal-low heating center and the equatorial African cooling zone. Changes in the divergent circulation of North Africa seem to fill the western European trough and to expand the Atlantic and the Saharan high equatorward (Fig. 9d). Consequently, the enhanced meridional gradients of streamfunction on the tropical side of this North African high intensifies the AEJ (will be shown in section 5). Thorncroft and Blackburn (1999) showed that the Saharan thermal-low heating produces the AEJ. The large meridional gradients of ψ_{EI}(600 mb) over tropical West Africa confirms not only their demonstration, but also the vitality of the Saharan thermal-low heating in maintaining the Saharan high.

The contrast of meridional gradients between ψ_{EI} (600 mb) and ψ_{EC}(600 mb) over West Africa may be illustrated further by the Gill-type experiment with only the ITCZ heating in an experiment designated as GI (Fig. 10b). As shown by Gill (1980), the tropical response to a diabatic heating center located north of the equator is a monsoonlike low with equatorial westerlies west of the heating center. The ψ_G^I(600 mb)anomalies (generated by the ITCZ heating along tropical Africa) are displayed in Fig. 10b; a negative ψ_G^I(600 mb)cell appears over West Africa, while a positive one exists over east Africa. By removing the ITCZ heating in the EI experiment, meridional gradients of ψ_{EI}(600 mb) over West Africa are strengthened [compared to ψ_{EC}(600 mb)].

In summary, it is revealed from the ES and EI experiments that the midtropospheric Saharan high is formed/maintained by the east–west differential heating between the Saharan thermal low and the eastern Atlantic Ocean and modulated by the ITCZ heating.

5. Maintenance of the AEJ

Using a dry general circulation model forced by the prescribed forcing, Thorncroft and Blackburn (1999) demonstrated numerically that the AEJ is possibly maintained by the shallow meridional circulation driven by the Saharan thermal-low heating and the deep Hadley circulation induced by the ITCZ heating. These secondary meridional circulations in West Africa are parts of the divergent circulation [depicted by (χ, V_D) (600 mb) in Fig. 3b], while the AEJ is a part of the North African summer circulation [portrayed by ψ(600 mb) in Fig. 1b]. The $\chi - \psi$ interaction may be illustrated by the ψ_χ tendency through vortex stretching [Eq. (4)], but the maintenance of the AEJ may not be directly elucidated from this dynamical interaction. However, the release of available potential energy to maintain kinetic energy is accomplished through the warm-air rising and cold-air sinking. Thus, Burpee (1972) suggested that the secondary meridional circulation in the lower half of the troposphere may be an important agent in supplying kinetic energy to the AEJ. Burpee's suggestion was echoed by Cook (1999) from her demonstration of the existence of ageostrophic flow across the Guinea coast and the Sahara desert. How does the secondary circulation release the available potential energy to maintain the AEJ? As shown in Fig. 3b, the divergent winds that spill out of the east–west oriented divergent center over North Africa are almost *perpendicular* to the rotational flow inferred from east–west oriented contours of the streamfunction (Fig. 1b). Burpee's (1972) suggestion may be substantiated by the energetic interaction between divergent and rotational flow introduced by Chen and Wiin-Nielsen (1976).

Following Wiin-Nielsen and Chen [1993, their Eqs. (13.10–13.11)], kinetic energy equations of divergent (V_D) and rotational (V_R) flows can be written as

$$\frac{\partial k_D}{\partial t} \cong \underbrace{-\nabla \cdot (\mathbf{V}_D k)}_{B(k_D)} \underbrace{- (\zeta + f)(u_D v_R - u_R v_D)}_{-C(k_D,\, k_R)} \underbrace{- \mathbf{V}_D \cdot \nabla \phi}_{G(k_D)}, \tag{9}$$

$$\frac{\partial k_R}{\partial t} \cong \underbrace{-\nabla \cdot (\mathbf{V}_R k)}_{B(k_R)} - \underbrace{(\zeta + f)(u_D v_R - u_R v_D)}_{C(k_D,\, k_R)} \underbrace{- \mathbf{V}_R \cdot \nabla \phi}_{G(k_R)}, \tag{10}$$

where $k_D = (u_D^2 + v_D^2)/2$, $k_R = (u_R^2 + v_R^2)/2$, $V_D = (u_D, v_D)$, and $V_R = (u_R, v_R)$. $B(\,)$, $G(\,)$, and $C(k_D, k_R)$ are divergence of the kinetic energy flux associated with divergent (V_D)/rotational (V_R) flow, generation of ( ), and conversion between k_D and k_R, respectively. It is generally true that $|V_D|/|V_R| \approx O(10^{-1})$, $k_D/k_R \approx O(10^{-2})$, and $|\zeta|/f \approx O(10^{-1})$. Thus, the residence time of k_D is usually short, and the link between k_D and k_R can be depicted by the following energy diagram:

$$\xrightarrow[G(k_D)]{} \boxed{k_D} \xrightarrow[\substack{C(k_D,\, k_R) \\ \iint \\ f u_R v_D}]{} \boxed{k_R},$$

where $C(k_D, k_R)$ is approximated by $f\, u_R v_D$ because $u_D \ll v_D$ and $v_R \ll u_R$ over West Africa.

Energetic analysis of $G(k_D)$ and $C(k_D, k_R)$ at 600 mb is shown in Figs. 11a and 11c, respectively. The complete energetic analysis of Eqs. (9) and (10) was performed, but we shall focus our discussion on the energy diagram to avoid any sidetrack from the AEJ maintenance by meridional circulations. Both $G(k_D)$ and $C(k_D, k_R)$ are similar in their spatial distributions and comparable in their structures. These particular features of $G(k_D)$ and $C(k_D, k_R)$ validate the approximation adopted in the energy diagram that k_D is generated by divergent flow spilled out of the east–west oriented North African divergent center along the meridional

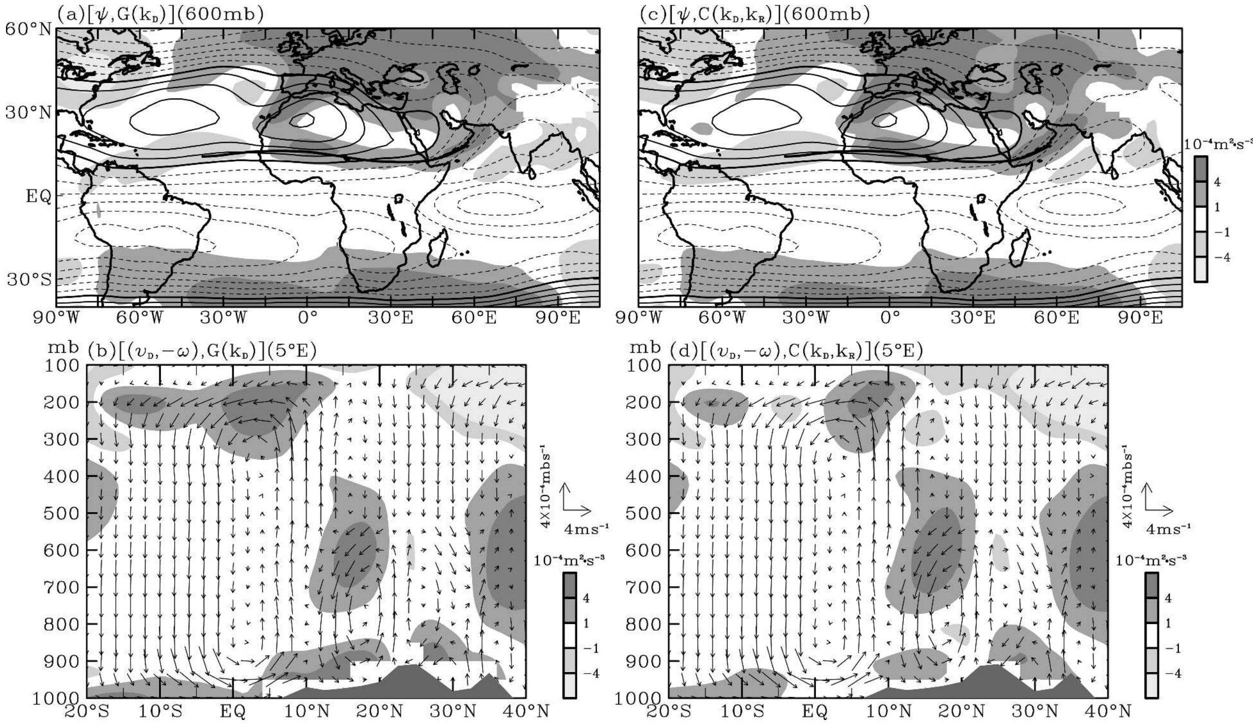

FIG. 11. The AEJ energetics at 600 mb: (a) [ψ, $G(k_D)$](600 mb), (b) [ψ, $C(k_D, k_R)$](600 mb), and the latitude–height cross sections of the AEJ energetics at 5°E: (c) [(v_D, $-\omega$), $G(k_D)$](5°E) and (d) [(v_D, $-\omega$), $C(k_D, k_R)$](5°E). Values of $G(k_D)$ and $C(k_D, k_R)$ are stippled according to scale shown in the right of each panel.

slope of the Saharan high. The Coriolis force induced by the downgradient divergent flows (fv_D) accelerates the AEJ (represented by u_R) formed by large north–south gradients of ψ(600 mb) along the Guinea coast.

Note that the v_D field across the Guinea coast is a part of the local Hadley and southern Saharan meridional circulations. In view of the acceleration of the AEJ by the Coriolis force associated with v_D, how is the AEJ maintained by the these meridional circulations? To answer this question, both energy variables, $G(k_D)$ and $C(k_D, k_R)$, are superimposed on the meridional circulation at 5°E, (v_D, $-\omega$) (5°E), in Figs. 11b and 11d, respectively. The tropical easterly jet is located at the top of the upward branch of the Hadley circulation where both $G(k_D)$ (Fig. 11b) and $C(k_D, k_R)$ (Fig. 11d) exhibit centers of positive values. The upper-tropospheric positive center of $C(k_D, k_R)$ implies that the Coriolis force associated with these divergent northerlies accelerates easterlies to maintain the tropical easterly jet. The midtropospheric AEJ is located almost over the downward branch of the *southern* Saharan meridional circulation. Like the tropical easterly jet, the southern slope of the Saharan high maintains the downslope divergent northerlies. The Coriolis acceleration caused by the equatorward northerlies of the meridional circulation maintains the AEJ, as indicated by the midtropospheric positive center of $C(k_D, k_R)$. Results of the energetic analysis presented in Fig. 11 show that the AEJ owes its existence to the interaction between the North African divergent circulation and the Saharan high.

The role played by the Hadley circulation in the AEJ maintenance is not directly revealed from the energetic analysis shown in Fig. 11. However, to clarify contributions from both the Hadley and southern Saharan meridional circulations to the AEJ maintenance, we examine the AEJ intensity in the experiments performed in section 4b. Let us denote eddy components of 600-mb zonal wind at 5°E by u_E(5°E) (Fig. 12a). The comparison between u_E(5°E) and u_{EC}(5°E) of the control experiment (Fig. 12b) verifies the realistic simulation of the AEJ at 600 mb by the forced barotropic model. The AEJ in experiment ES (equivalent to without the shallow Saharan meridional circulation) shifts northward, away from tropical West Africa (Fig. 12c). In contrast, the AEJ in experiment EI (equivalent to without the Hadley circulation) shifts equatorward (Fig. 12d). The eddy component of zonal winds in these two experiments is verified by comparing ($u_{\mathrm{ES}} + u_G^S$) (Fig. 12e) and ($u_{\mathrm{EI}} + u_G^I$) (Fig. 12f) against u_{EC}. Bias of the first two eddy zonal winds at the AEJ core against u_{EC} is not recognizable. The impact of the ITCZ heating/the

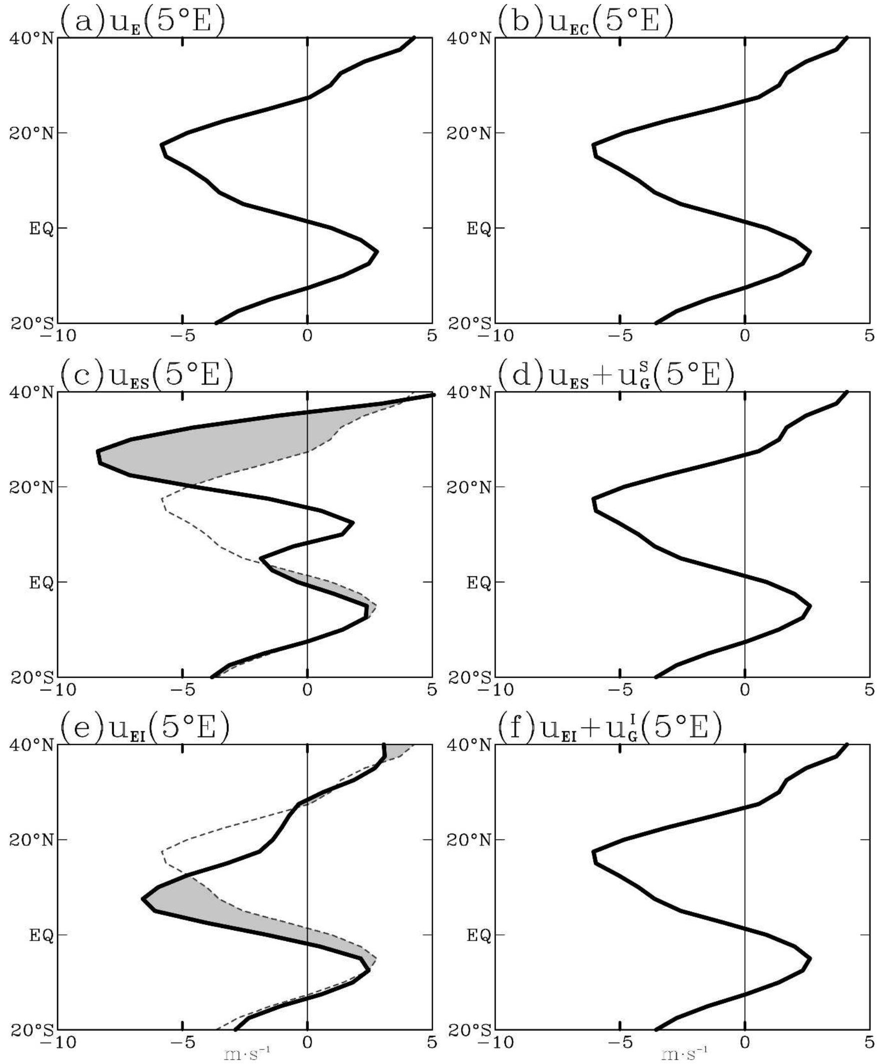

FIG. 12. Summer eddy component of the 600-mb zonal wind at 5°E: (a) observed [u_E(5°E)], (b) EC [u_{EC}(5°E)], (c) ES [u_{ES} (5°E)], (d) (ES + GS)[(u_{ES} + u_G^S(600 mb)] (5°E), (e) EI [u_{EI} (5°E)], and (f) (EI + GI) [u_{EI} + u_G^I(5°E)].

Hadley circulation on the AEJ is realized through the dynamical response/adjustment of the North African summer circulation.

6. Concluding remarks

The North African summer circulation is basically formed by the following pronounced components: 1) the western part of the Tibetan high and the tropical easterly jet in the upper troposphere and 2) the Saharan thermal low (often considered as the westward extension of the Indian monsoon trough) and the West African monsoon southwesterly flow (coupled with the ITCZ) in the lower troposphere. These North African summer circulation components are separated by the midtroposheric *Saharan high,* which is surrounded by the *African easterly jet* along its southern periphery. On the other hand, the upward motion generated by the Saharan thermal heating and the downward branch of the Indian monsoon's east–west circulation form the midtropospheric North African divergent center over the Chad–Sudan region. Therefore, a link between the upper and lower tropospheric circulation over North Africa can be established through this divergent center. Because this divergent center is located east of the Saharan high, the lower-tropospheric east–west circulation between North Africa and the eastern North Atlantic is spatially in quadrature with the Saharan high. This spatial relationship is typical in any major monsoon circulation. The AEJ, which is an integral part of the Saharan high, owes its existence to the midtropospheric Saharan anticyclone. In view of the interrelationship among the midtropospheric North African divergence center, the Saharan high, and the AEJ, an effort was made in this study to explore maintenance mechanism of these North African circulation elements. Major findings of this study are summarized as follows:

1) *Maintenance of the North African divergent center*: The midtropospheric divergent center of the North African divergent circulation is located over the Chad–Sudan region. Using the χ-maintenance equation, we are able to demonstrate that this divergent center is formed by upward motions induced by the Saharan thermal-low heating and the downward motions caused by the radiative cooling associated with the downward branch of the Indian monsoon's east–west circulation.
2) *Maintenance of the midtropospheric Saharan high:* It is revealed from the contrast between the 600-mb velocity potential and streamfunction in the 2–15 wave regime, χ^M(600 mb) and ψ^M(600 mb), that the Saharan high and the North African divergent center are spatially in quadrature. It is inferred from the χ^M–ψ^M interaction through vortex stretching that the Saharan high is maintained by the east–west circulation formed by the Saharan thermal-low heating and the east Atlantic cooling. The inference from the ψ^M(600 mb) budget analysis was further substantiated by simulations of the Saharan high with a forced barotropic model.
3) *Maintenance of the AEJ*: Applying the energetics scheme of atmospheric divergent and rotational flows developed by Chen and Wiin-Nielsen (1976), it was shown that the AEJ is maintained by the Coriolis acceleration associated with the equatorward downgradient divergent northerlies. Cook (1999) demonstrated the importance of the north–south thermal and moisture gradient to the AEJ through the thermal wind, and Thorncroft and Blackburn (1999) stressed the role of the meridional circulations associated with the ITCZ and the Saharan thermal-low heating in the AEJ maintenance. The north–south differential heating emphasized by these studies is consistent with the downgradient divergent northerlies of the Saharan divergent circulation.

It was pointed out by numerous studies (e.g., Lamb and Peppler 1991; Nicholson 2000; Le Barbi et al. 2002; and others) that West Africa has undergone a major drought period since the 1970s. The results obtained by this study may be biased by this interdecadal variation of the West African climate system. Despite this possible bias of climate change, findings summarized above not only enhance our understanding of the maintenance mechanisms of the Saharan high and the AEJ, but also facilitate our search for answers to some outstanding weather/climate problems of North Africa. Some examples are described here. The barotropic–baroclinic instability of Charney and Stern (1962) was suggested by Burpee (1972) as the genesis mechanism of African easterly waves (AEWs). Therefore, the cyclonic shear side of the AEJ is a favorable region of the AEW genesis. After numerous efforts were made to explore Burpee's AEW genesis mechanism, Thorncroft and Hoskins (1994) numerically demonstrated that the AEW genesis is most likely caused by a mixed barotropic/baroclinic instability. Burpee's (1972) AEW genesis mechanism was evolved from Carlson's (1969a,b) depiction of AEW, but Carlson's AEWs originated in the North African desert underneath the Saharan high, instead of the cyclonic shear side of the AEJ. In view of this environmental contrast of AEW genesis, Burpee's mechanism may be only one of multiple AEW genesis mechanisms. The better depiction of the North African summer circulation and the maintenance mechanisms of the Saharan high and the AEJ in this study provide a more accurate continental-scale environment for the search of these AEW genesis mechanisms.

The North African divergent center is formed by the downward branch of the Asian monsoon east–west circulation and the upward motion induced by the Saharan thermal-low heating, while the Saharan high is maintained by the divergent cell over eastern North Africa and the convergent cell over the eastern North Atlantic. The possible effect of the El Niño–Southern Oscillation (ENSO) on the North African climate system may well be established through the effect of ENSO on the Asian monsoon's east–west circulation and in turn on the North African anticyclone. Since the AEJ is a part of the Saharan high, its interannual variation of this midtropospheric jet may follow that of the Saharan high. On the other hand, the West African monsoon rainfall occurs along the ITCZ, which is formed by the upward branch of the Hadley circulation and blocked by the downward branch of the southern Saharan meridional circulation cell. Without the midtropospheric Saharan high, the downward branch of the southern Saharan cell may be weakened or may not exist. Therefore, it is likely that the ITCZ and the West African monsoon rainfall are shifted somewhat northward. The Saharan high and the AEJ seem to play important roles in the African climate variability, which is one of the Climate Variability and Predictability (CLIVAR) principal research areas (WCRP 1998). The spatial structure and maintenance mechanism of the midtropospheric North African summer circulation presented in this study provide a crucial basis to accomplish this CLIVAR research initiative of the African climate system.

An important approach to answer the African weather/climate problems described above is numerical simulation with a full-physics global climate model. However, the accurate simulation of the midtropospheric North African summer circulation is a prerequisite for this exercise. The North African summer circulation comprises several complicated features: the separation of the AEJ and tropical easterly jet by the Saharan high, drastic transition between the cold moist monsoon of the West African seaboard along the Guinea coast and the dry warm Saharan desert, and the merger between the upward branch of the local Hadley cell along the ITCZ, the inland Saharan meridional circulation, and the downward branch of the Asian monsoon east–west circulation. These vertical and meridional structures of the summer circulation post a real challenge to the climate model simulation over the North African continent. However, findings in this study give not only verification of these numerical simulations, but also possible directions of model performance, particularly couplings between the AEJ and the Saharan thermal low through the southern Saharan meridional circulation and between the tropical easterly jet and the West African monsoon rainfall through the local Hadley circulation.

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