The Atmospheric Heat Source over the Tibetan Plateau: May–August 1979

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(Manuscript received 2 February 1985, in final form 23 May 1985)

ABSTRACT

Estimates of the time and space variability of the atmospheric heat source over Tibet are presented for the summer of 1979. These estimates rely on new data from the People's Republic of China allowing a better assessment of the surface heat fluxes, and on new satellite data from Nimbus-7 giving the radiation balance at the top of the atmosphere. Our estimates of the atmospheric heat source turned out to be considerably smaller than those provided earlier in the literature, mainly because of different assumptions of the drag coefficient. The atmospheric heat source over Tibet is mainly modulated by the release of latent heat. Over the southeastern and southwestern plateau regions the heat source appears to be in phase with the precipitation yield of the Indian summer monsoon, whereas central Tibet reveals an out-of-phase behavior. Over western Tibet there appears to be hardly any net import of moisture from outside the region, whereas the maintenance of the hydrological cycle over eastern Tibet requires moisture flux convergence from outside the region of up to 40% of the mean rainfall, in agreement with what is known about the surface hydrology of Tibet.

1. Introduction

The role of the Qinghai-Xizang (Tibet) Plateau as an elevated heat source has been recognized by many scientists (Staff Members Academia Sinica, 1958; Flohn, 1968; Yeh and Gao, 1979; Reiter and Gao, 1982). Yeh *et al.* (1957) were among the first to describe the winter heat sink and summer heat source characteristics of the plateau region. From aerological and surface observations Chen *et al.* (1965) arrived at the conclusion that maximum heating of the atmosphere during summer actually does not occur over the center of the plateau but is shifted toward its southeastern part where latent heat input assumes considerable significance. Sensible heating dominates over the western part of the plateau, as has been confirmed recently by Feng *et al.* (1985).

In searching through the literature for estimates of the magnitude and effectiveness of this heat source, one finds many discrepancies. In Table 1 we have attempted to summarize the major findings by other researchers.

Much of the variability evident from past analyses can be ascribed to insufficient data, different parametric assumptions in the calculations and different computation schemes. During the 1979 Qinghai-Xizang (Tibet) Plateau Meteorological Experiment (TIPMEX) six additional surface radiation balance stations and several other conventional surface and upper-air stations were established in Tibet. From these data Chinese meteorologists (Pan, 1984; Chen and Weng, 1984; Weng *et al.*, 1984) obtained new empirical formulae for calculations of the surface radiation balance and of the sensible heat exchange between ground and air. These new data and computational methods were not available to the studies listed in Table 1. They have, however, been incorporated in the present investigation, as have data from the Nimbus-7 Earth Radiation Budget Experiment (ERB), especially from its narrow field of view (NFOV) with an effective space resolution of about 500 × 500 km (the exact grid size varies with latitude). These satellite data were obtained through the efforts of Dr. Alan Arking, NASA Goddard Space Flight Center. Thus, the net radiation at the top of the atmosphere can now be estimated relatively reliably.

Using these, as well as the FGGE Level II-B data, we were able to calculate the strength and variability of the atmospheric heat source over the Qinghai-Xizang plateau for the period 19 May–31 August 1979. For a correct interpretation of our results it should be pointed out that 1979 was not a typical monsoon year. The monsoon arrived late over most of India and had several prolonged break periods. The overall precipitation yield of the monsoon season during that year remained

TABLE 1. Comparison of estimates on the atmospheric heat source over Tibet.

Author	Region/season	Heat source (W m^{-2})	Heating rate (°C day^{-1})
	Whole plateau, 660–100 mb		
Yeh and Gao (1979)	June	109	1.64
	July	101	1.52
	August	74	1.11
	Western plateau, 630–100 mb		
	June	138	2.19
	July	120	1.91
	August	90	1.43
	Eastern plateau, 690–100 mb		
	June	89	1.27
	July	92	1.31
	August	67	0.96
Chen *et al.* (1982)	Whole plateau, summer		<1.0
Yao *et al.* (1984)	Bay of Bengal and Assam, summer		max: 6–8 avg: 3–4
Nitta (1983)	Eastern plateau†		
Luo and Yanai (1984)	Assam-Bengal and Northeast India, June 1979	250	2.34*
	Eastern plateau	113	1.90*
	Western plateau, June 1979	107	1.80*

* Estimated by present authors.
† Similar to Yeh and Gao (1979).

below normal over the Indian subcontinent (Fein and Kuettner, 1980) but above normal over the Tibetan Plateau, especially in July.

2. Computational methods

Figure 1 shows the distribution of surface stations, 44 altogether, used in our computations, together with the 14 "target squares" for which Nimbus-7 satellite data were available. Over southwestern Tibet we have adopted some of the computational results reported by Chen *et al.* (1985). We assumed that the net radiation for all stations within the same target area is that given for this particular target area.

The following equations were used to compute the vertical integrals of the apparent heat source $\langle Q1 \rangle$ and moisture heat source $\langle Q2 \rangle$ (see e.g., Nitta, 1983; Luo and Yanai, 1984):

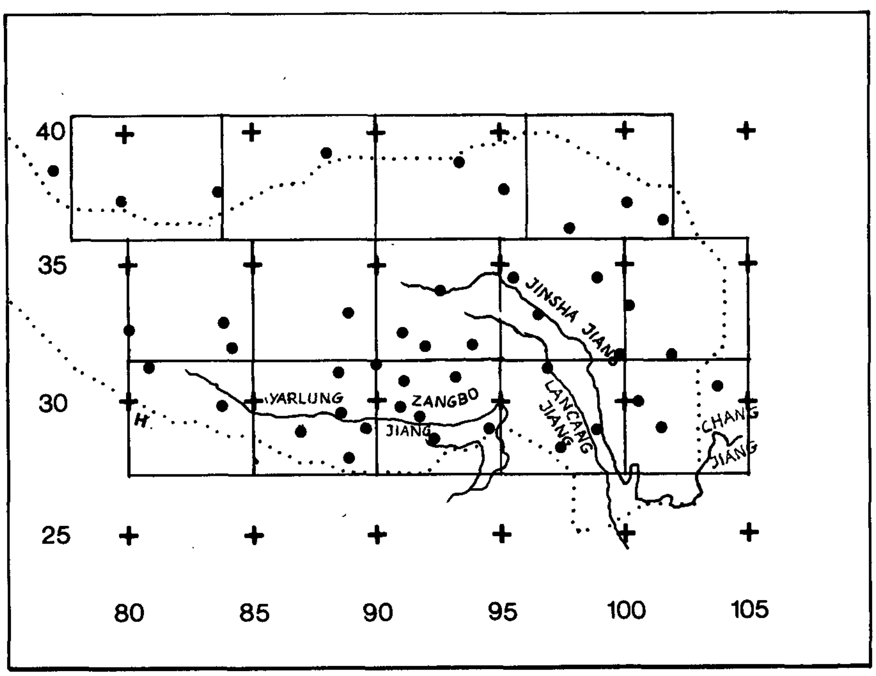


FIG. 1. Dots indicate surface stations. Squares: target areas for which Nimbus-7 data were available. Crosses show latitude–longitude intersection points.

TABLE 2. Empirical coefficients in Eq. (5) (after Pan, 1984; Weng *et al.*, 1984; Chen and Weng, 1984).

	Coefficient				
Region	A	B	a	b	c
Western and central plateau	0.285	0.535			
Yarlung Zangbo valley	0.334	0.546			
Southeastern plateau	0.164	0.615			
Northeastern plateau	0.121	0.612			
Western and northern plateau			0.548	0.164	0.0220
Southern and eastern plateau			0.631	0.200	0.0084

$$\left.\begin{aligned}\frac{1}{g}\int_{p_T}^{p_s} \mathrm{Q1}\,dp &\approx \frac{1}{g}\int_{p_T}^{p_s} Q_R dp + \mathrm{LP} + \mathrm{SH} \\ \frac{1}{g}\int_{p_T}^{p_s} \mathrm{Q2}\,dp &\approx \mathrm{LP} - \mathrm{LE}\end{aligned}\right\}, \quad (1)$$

where p_s and p_T are the surface pressure and the pressure at the top of the atmosphere, Q_R symbolizes radiation heating, LP is the latent heat released to the atmosphere by precipitation, SH is the sensible heat exchange with the ground, and LE is the heat used for evaporation of moisture from the ground. In the subsequent discussion we will use the notation

$$\left.\begin{aligned}\langle \mathrm{Q1}\rangle &= \frac{1}{g}\int_{p_T}^{p_s} \mathrm{Q1}\,dp \\ \langle \mathrm{Q2}\rangle &= \frac{1}{g}\int_{p_T}^{p_s} \mathrm{Q2}\,dp\end{aligned}\right\}, \quad (2)$$

and assume that

$$\mathrm{RC} = (R_\infty - R_0) = \frac{1}{g}\int_{p_T}^{p_s} Q_R dp, \quad (3)$$

where R_∞ and R_0 are the net radiation values at the top of the atmosphere and the earth's surface, respectively. Both values are defined as positive if directed

TABLE 3. Mean values of atmospheric heat source over Tibet, 1979 (W m^{-2}).*

	Whole plateau		Eastern plateau				Western plateau		
	Herein	Y&G	Herein	Y&G	Nitta	L&Y	Herein	Y&G	L&Y
June									
SH	68	141	59	102		(104)	80	219	(126)
LP	53	63	83	86		71	14	17	58
RC	−55	−95	−53	−99		−62	−57	−98	−77
LE	35		46			(27)	20		(34)
⟨Q1⟩	66	109	89	89		113	37	138	107
⟨Q2⟩	18		37			44	−6		24
July									
SH	55	116	45	79			69	190	
LP	80	78	101	106			52	23	
RC	−59	−93	−59	−93			−59	−93	
LE	51		60				39		
⟨Q1⟩	76	101	87	92			62	120	
⟨Q2⟩	29		41				13		
August									
SH	52	96	45	64			61	158	
LP	60	78	76	102			38	31	
RC	−64	−100	−67	−99			−60	−99	
LE	52		58				42		
⟨Q1⟩	48	74	54	67			39	90	
⟨Q2⟩	8		18				−4		
June–August									
SH	59	118	50	82	(105)		70	189	
LP	65	73	87	97	90		35	24	
RC	−60	−96	−60	−97	−75		−59	−97	
LE	47		56		65		34		
⟨Q1⟩	64	94	77	82	120		46	116	
⟨Q2⟩	19		31		25		1		

* Rc = $R_\infty - R_0$; Y&G = Yeh and Gao, 1979; Nitta, 1983; L&Y = Luo and Yanai, 1984.

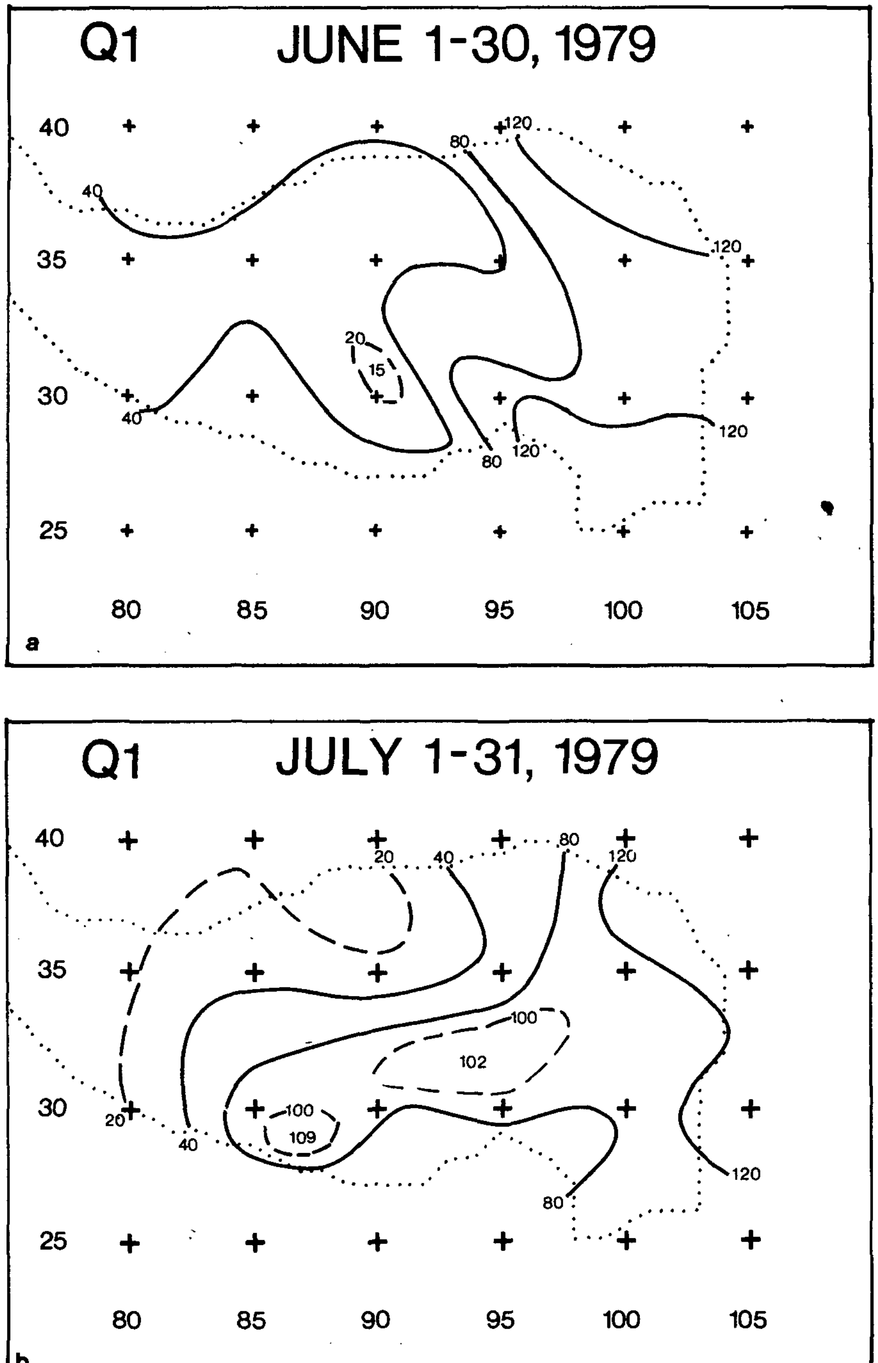


FIG. 2. Atmospheric heat source, $\langle Q1 \rangle$, in W m^{-2} during (a) June, (b) July, and (c) August 1979.

downward. If $RC = R_\infty - R_0 > 0$, radiative heating occurs in the atmosphere. Since, however, in our investigation $RC < 0$ over the plateau, this term effectively symbolizes radiative cooling. If $\langle Q1 \rangle > 0$, diabatic processes cause heating of the atmosphere, hence we will refer to this term as the "atmospheric heat source." If $\langle Q2 \rangle > 0$, precipitation exceeds evaporation, thus the atmospheric component of the hydrological cycle serves as a "moisture heat source" which has to be in balance with horizontal moisture flux convergence in the atmosphere.

In an approach different from that chosen by Nitta (1983) and Luo and Yanai (1984), who calculated the values of Q1 and Q2 at individual pressure levels and

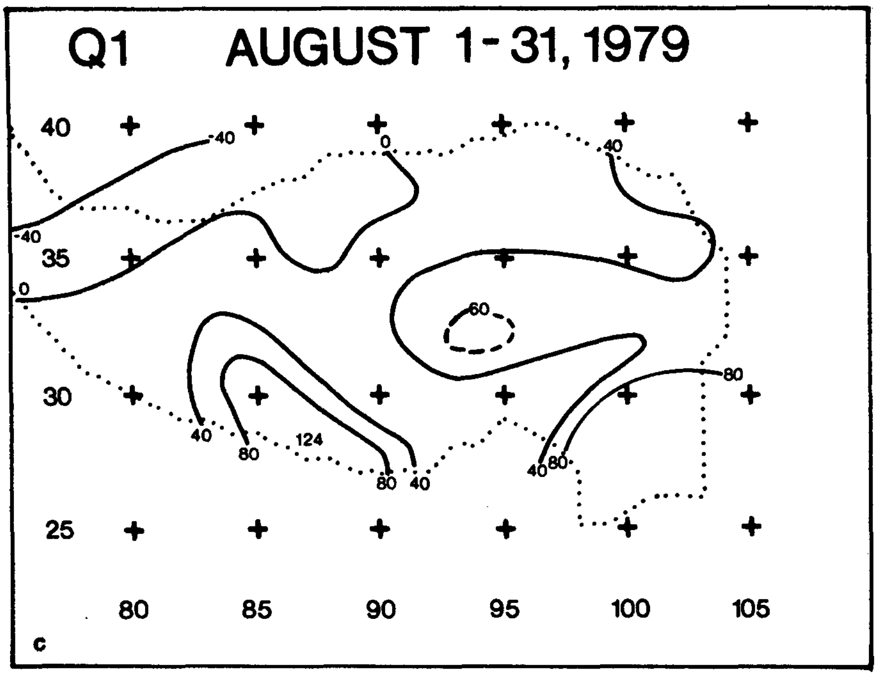


FIG. 2. (*Continued*)

then performed a vertical integration, we computed $\langle Q1 \rangle$ and $\langle Q2 \rangle$ directly from the balance equations (1), (2) and (3), abbreviated in the form

$$\left.\begin{aligned}\langle Q1 \rangle &= (R_\infty - R_0) + \mathrm{SH} + \mathrm{LP}\\ \langle Q2 \rangle &= \mathrm{LP} - \mathrm{LE}\end{aligned}\right\}, \qquad (4)$$

referring to these computations as having been done by the "direct" method. As will become apparent from the discussion of our results, the discrepancies between earlier findings and our own cannot be blamed entirely on the use of different computational procedures.

Nimbus-7 provides ERB data with a narrow (NFOV) and wide field of view (WFOV). Arking and Vemury (1984) estimated that the WFOV values for R_∞ were larger on the average by about 14 W m^{-2} than those from the NFOV. Over the Tibetan Plateau during July 1979 analysis of the data revealed this discrepancy to be of the order of 11 W m^{-2}, and only 3 W m^{-2} during August 1979. Arking and Vemury suggested that the discrepancy observed by them should be apportioned as a positive error (i.e., too high a reading) of 10 W m^{-2} to the WFOV, and as a negative error of 4 W m^{-2} to the NFOV. Apparently the albedo is overestimated slightly in the NFOV. In our calculations we have not made any corrective adjustments to the Nimbus-7 NFOV data, because of our reduced estimates of the discrepancies and the uncertainty as to how these discrepancies should be apportioned. Given the error estimates by Arking and Vemury, our estimates of R_∞ might be too low by about 4 W m^{-2}, or by approximately 10%.

The following empirical relationship for the surface radiation budget was used in our calculations:

$$R_0 = S_0(A + BH/H_0)(1 - \alpha) - \delta\sigma[T_g^4 - T_a^4(a + b\cdot N + cE)], \qquad (5)$$

where S_0 is the total amount of solar radiation received by the "ideal" atmosphere; H signifies the observed, H_0 the maximum possible hours of insolation at a given station; α the surface albedo; $\delta = 0.95$, the coefficient of emissivity; and σ the Stefan–Boltzmann constant. The ground surface temperature, T_g, is measured directly at Chinese weather stations, and does not have to be parameterized; T_a is the air temperature at meteorological screen height, N the cloud amount and E the water vapor pressure, also measured in the meteorological screen. The empirical coefficients A, B, a, b, and c are listed in Table 2 (obtained from Pan, 1984; Weng *et al.*, 1984; Chen and Weng, 1984). Ten-day mean values of the surface radiation balance calculated by means of Eq. (5), according to the Chinese literature cited above, had a mean error of about 6% when compared against actual measurements of that balance. We calculated daily values of the radiation budget and found reasonable agreement with the earlier 10-day mean results quoted above and taken from the atlas published by the First Research Group (1984) for each of the station locations shown in Fig. 1.

The calculation of the sensible heat transfer between ground and atmosphere follows the standard procedure

$$\mathrm{SH} = \rho C_p C_d V_0(T_g - T_a), \qquad (6)$$

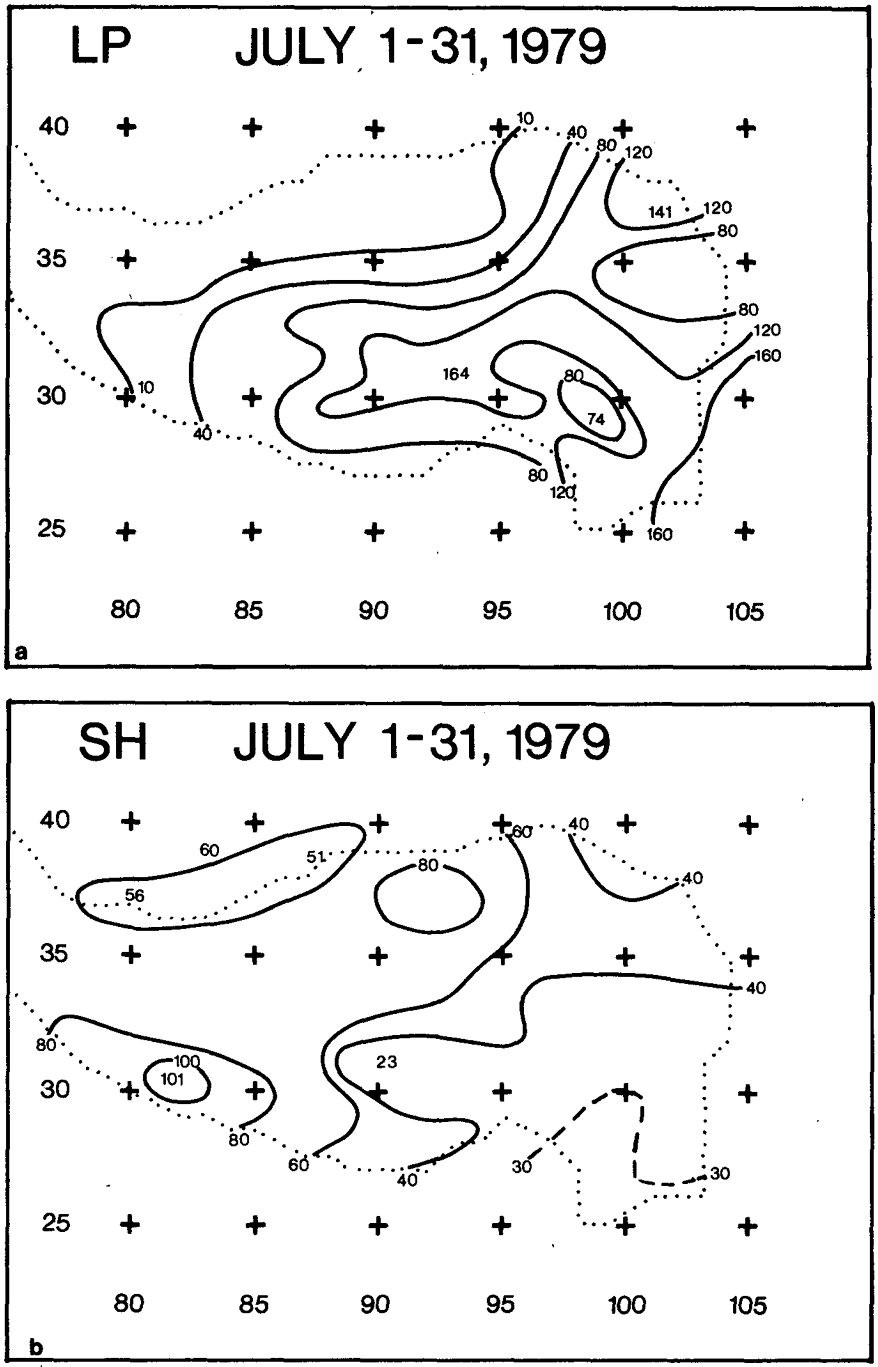


FIG. 3. Components (W m^{-2}) entering into the computation of $\langle Q1 \rangle$ and $\langle Q2 \rangle$, for July 1979. (a) Latent heat of precipitation; (b) sensible heating from ground; (c) radiative cooling, Rc = $R_\infty - R_0$; (d) latent heat of evaporation from surface.

where ρ is the air density at ground level, C_p the specific heat at constant pressure, C_d the drag coefficient and V_0 the mean wind speed measured at 10 m above ground. This equation has been used widely in parameterizing surface heat fluxes over the oceans ($C_d \approx 1 \times 10^{-3}$) and over land (see e.g., Yeh and Gao, 1979). Obviously, the values obtained for SH are rather sensitive to the assumptions made regarding C_d. Cressman (1960) suggested a value of 8.0×10^{-3} to be used over Tibet. This value remained unquestioned for many years and was used in most of the earlier calculations of the atmospheric heat budget (Yeh and Gao,

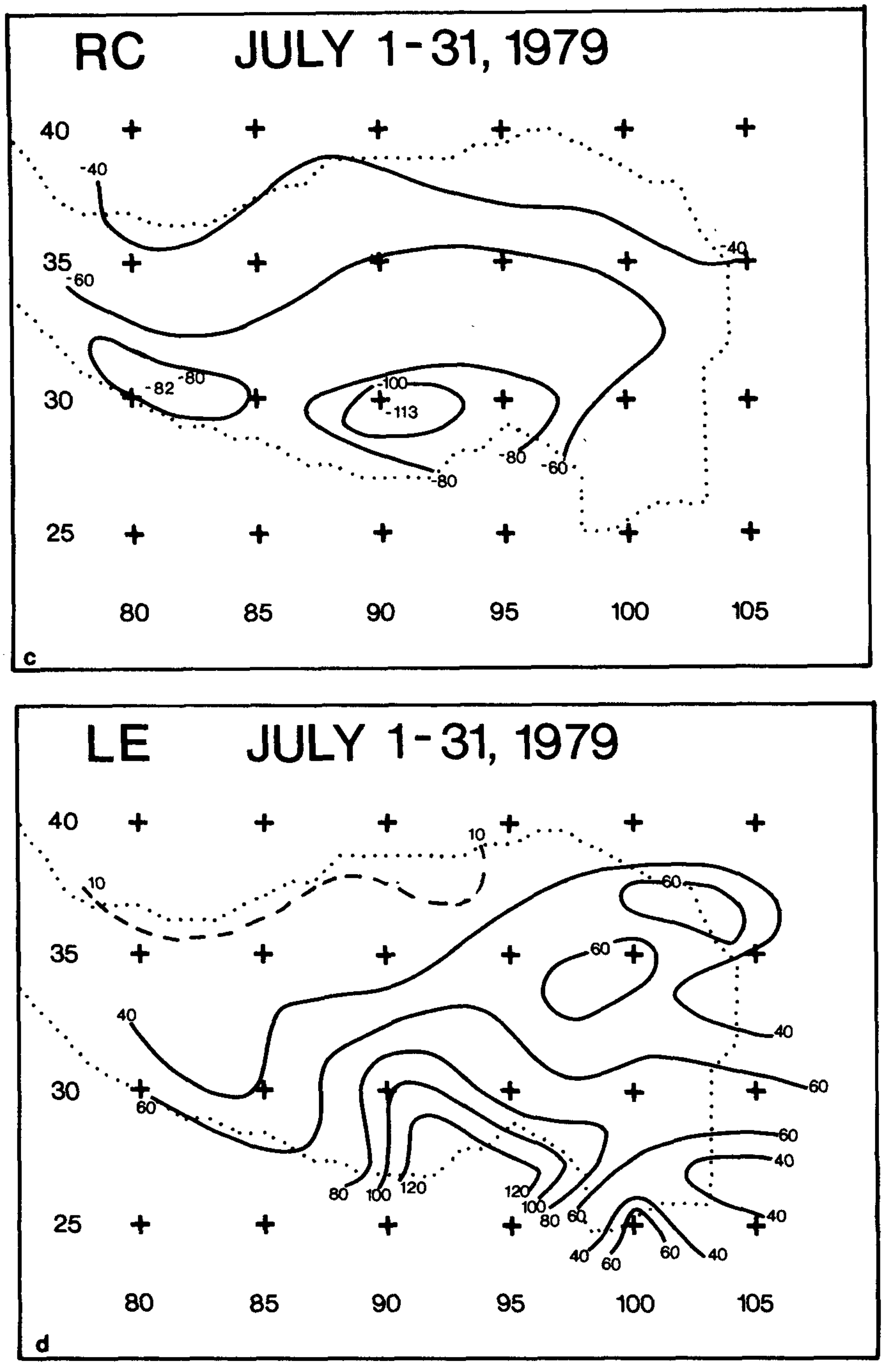


FIG. 3. (*Continued*)

1979). New evidence presented by Chen and Weng (1984), however, casts some doubt on the validity of this assumption. From six heat balance stations erected during TIPMEX, SH was computed from

$$\mathrm{SH} = R_0 - \mathrm{LP} - G, \tag{7}$$

where G is the heat flux into the ground, a quantity also measured at the six Tibetan stations. Only data from relatively dry periods were used with the assumption that LP = LE during these periods. Then regressions were obtained for C_d, using Eq. (5), and yielding the following empirical results

$$C_d \equiv \begin{cases} 0.00112 + 0.01/V_0, & \text{for } z > 2800 \text{ m} \\ 0.00112 + 0.01/V_0 - 0.00362(p_0 - 720)/280, & \text{for } z \leq 2800 \text{ m}, \end{cases} \tag{8}$$

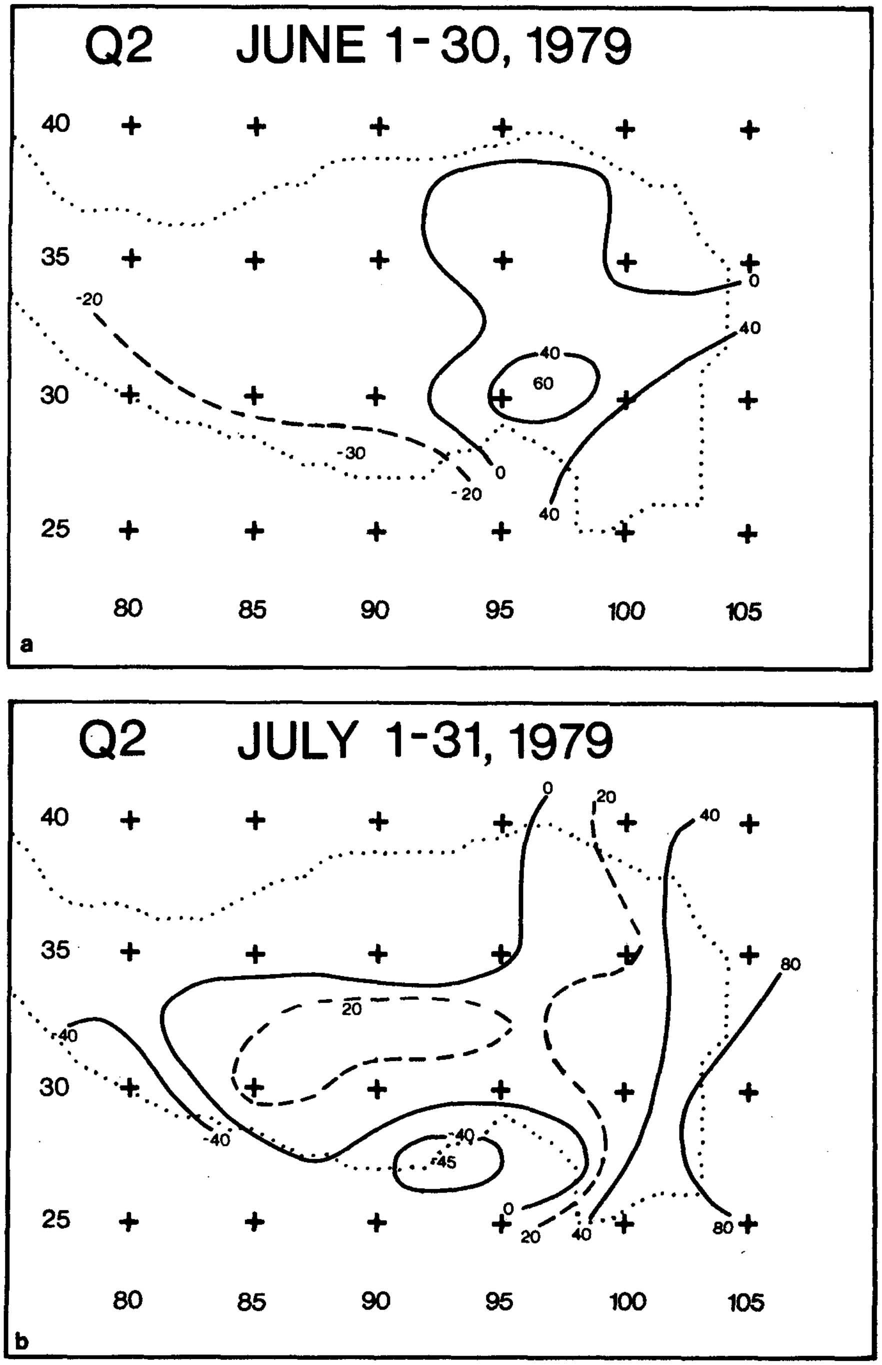


FIG. 4. Atmospheric moisture heat source, ⟨Q2⟩, in W m^{-2} during (a) June, (b) July, and (c) August 1979.

in which z is the station elevation, p_0 is the mean surface pressure in millibars at each station, and V_0 the 10-day mean wind speed in m s^{-1}, 10 m above ground. Since over any 10-day period the average wind speed always differed from zero, the correction terms which contain this parameter never became excessively large. The scatter of data points from which these regression values were derived suggested an error of 10.6 percent in the application of Eq. (7). During TIPMEX the average wind speed was approximately 4 m s^{-1}, suggesting an average value of $C_d = 3.62 \times 10^{-3}$ over the Tibetan Plateau. This value is considerably smaller than the one suggested by Cressman and used widely in the past. In our computations we assumed that the regressions expressed in Eq. (7) would also be valid for daily data.

It should be pointed out, however, that the stations

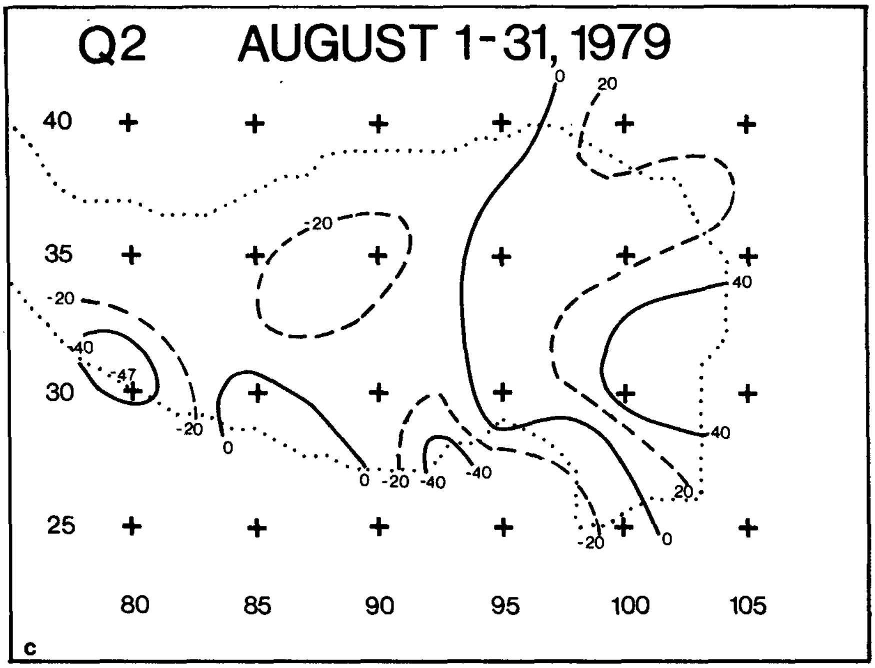


FIG. 4. (*Continued*)

from which the empirical relationships for C_d were obtained were biased towards valley locations and hence did not encompass mountain ridge effects. Whereas Cressman's estimate of a drag coefficient appears to be a gross overestimate, our adoption of the Chinese recommendations may have produced somewhat of an underestimate.

The latent heat release into the atmosphere, LP, needed in Eq. (4), is determined easily from precipitation data. It is also used to calculate the "moisture heat source," $\langle Q2 \rangle$. It is difficult to estimate LE over Tibet because of the paucity of data. For our calculations we have adopted the values taken from the atlas published by the First Research Group (1984), which calculated LE as a residual in the heat balance equation. Unfortunately, there are no river runoff and groundwater storage data available to test these estimates.

3. Mean heat source, summer 1979

Table 3 summarizes the monthly means for June through August 1979, as well as the means for the whole period, of $\langle Q1 \rangle$ and $\langle Q2 \rangle$ and their components. Values are shown for the whole plateau, as well as for its eastern (90–102.5°E, 27–40°N) and western (80–90°E, 27–38.5°N) parts. This table also provides comparisons with earlier results by Yeh and Gao (1979), Nitta (1983) and Luo and Yanai (1984).

The atmospheric heat source values given in Table 3 for the total plateau for June, July and August (66, 76 and 48 W m^{-2}, respectively) translate into heating rates for the 660–100 mb air column of approximately 0.99, 1.14 and 0.72°C day^{-1}, respectively. The average heating rate for the whole summer period is 0.96°C day^{-1}. These values are much lower than the ones obtained by Yeh and Gao (94 W m^{-2} or 1.42°C day^{-1} for the summer period). The heat source over the eastern plateau (77 W m^{-2}) dominates over that of the western plateau (46 W m^{-2}). Largest values of $\langle Q1 \rangle$ were observed during July. Over the eastern plateau LP dominates, whereas over the western plateau SH > LP. Both terms are of significance when averaged over the whole plateau, but also in the two individual regions, with the exception of June, when the western plateau experienced a dry spell and LP became negligibly small there. Qualitatively, these results are identical to those obtained by Flohn (1968), Yeh and Gao (1979) and Luo and Yanai (1984).

The main quantitative discrepancies between our results and those obtained by Yeh and Gao (1979) turn up over the western plateau (46 W m^{-2} or 0.73°C day^{-1} versus 116 W m^{-2} or 1.84°C day^{-1}), mainly due to the reduced drag coefficient values assumed in our calculations. There is, however, also a discrepancy of 38 W m^{-2} in the atmospheric radiative heat loss between the two sets of computations. The Nimbus-7 data yielded somewhat lower values of R_∞ than those used by other authors. Even if one allowed for the error quoted by Arking and Vemury (1984), this discrepancy could not be resolved. Neither could it be resolved if one allowed for a relative error of up to 6% in the ground level radiation budget R_0. The values of $\langle Q1 \rangle$ derived for summer 1979 by Nitta (1983) and those for June by Luo and Yanai (1984) are even larger than those found by Yeh and Gao.

In the light of the foregoing discussion we have to conclude that previous estimates of the magnitude of

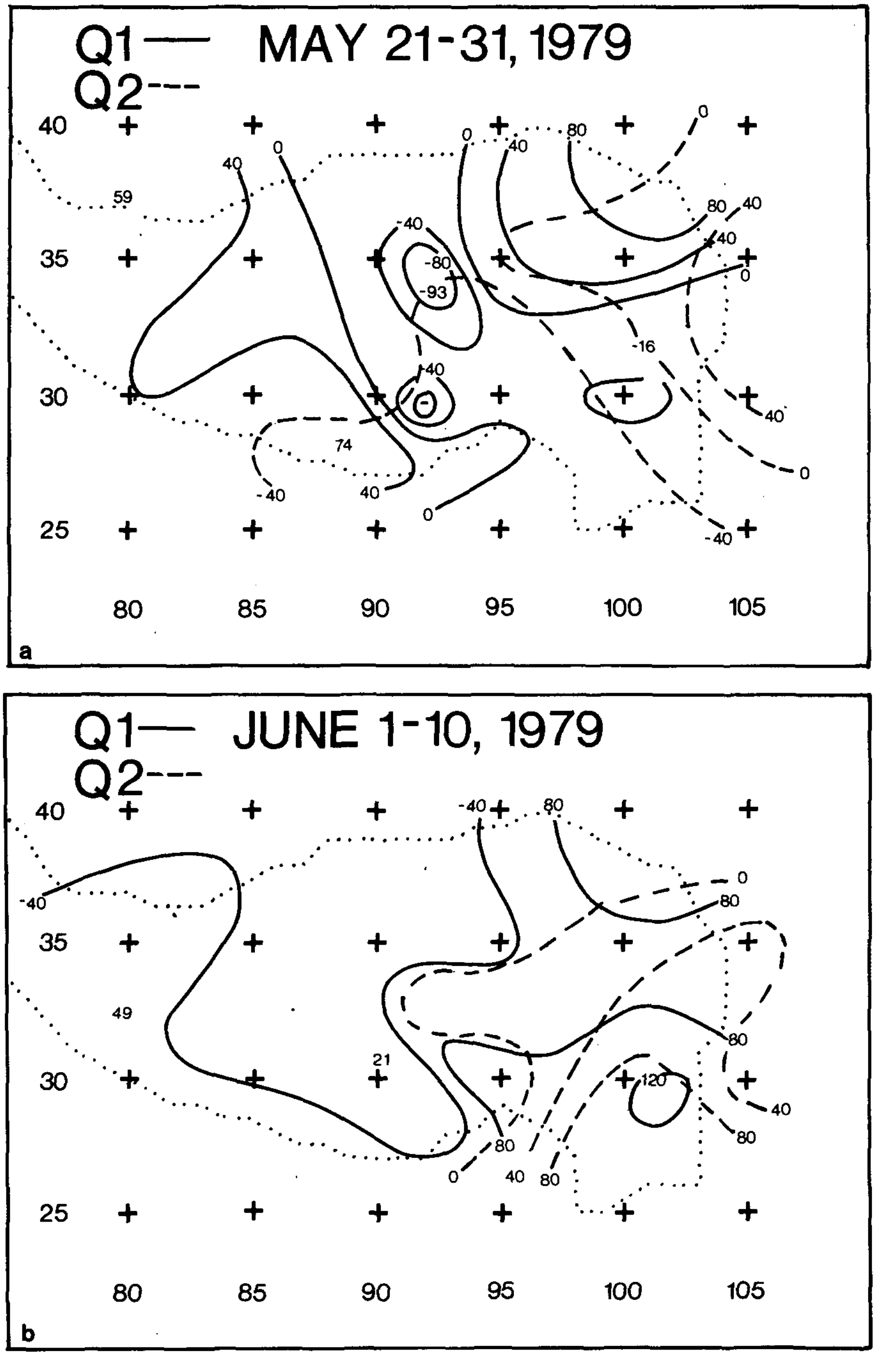


FIG. 5. Ten-day averages of ⟨Q1⟩ (solid lines) and ⟨Q2⟩ (dashed lines) between 21 May and 30 August 1979, in W m^{-2}.

the atmospheric heat source over Tibet using different approaches might have been too large.

The water vapor heat source ⟨Q2⟩ over the western plateau, averaged over the whole summer period, was equivalent to only 0.034 mm of water per day, indicating that precipitation and evaporation in this region were essentially balanced and no major water vapor flux convergence occurred over the region. The eastern plateau, on the other hand, with 31 W m^{-2} or 1.07 mm day^{-1}, enjoyed an excess of precipitation which amounted to approximately 36% of the total precipitation in that region. This excess precipitation requires an appropriate amount of moisture flux convergence and feeds, in turn, several major river systems.

Over the whole plateau (⟨Q2⟩ = 19 W m^{-2} or 0.65 mm day^{-1}) there is a net surplus of precipitation over evaporation of about ⅓ of the total rainfall. There is, however, considerable variability from month to

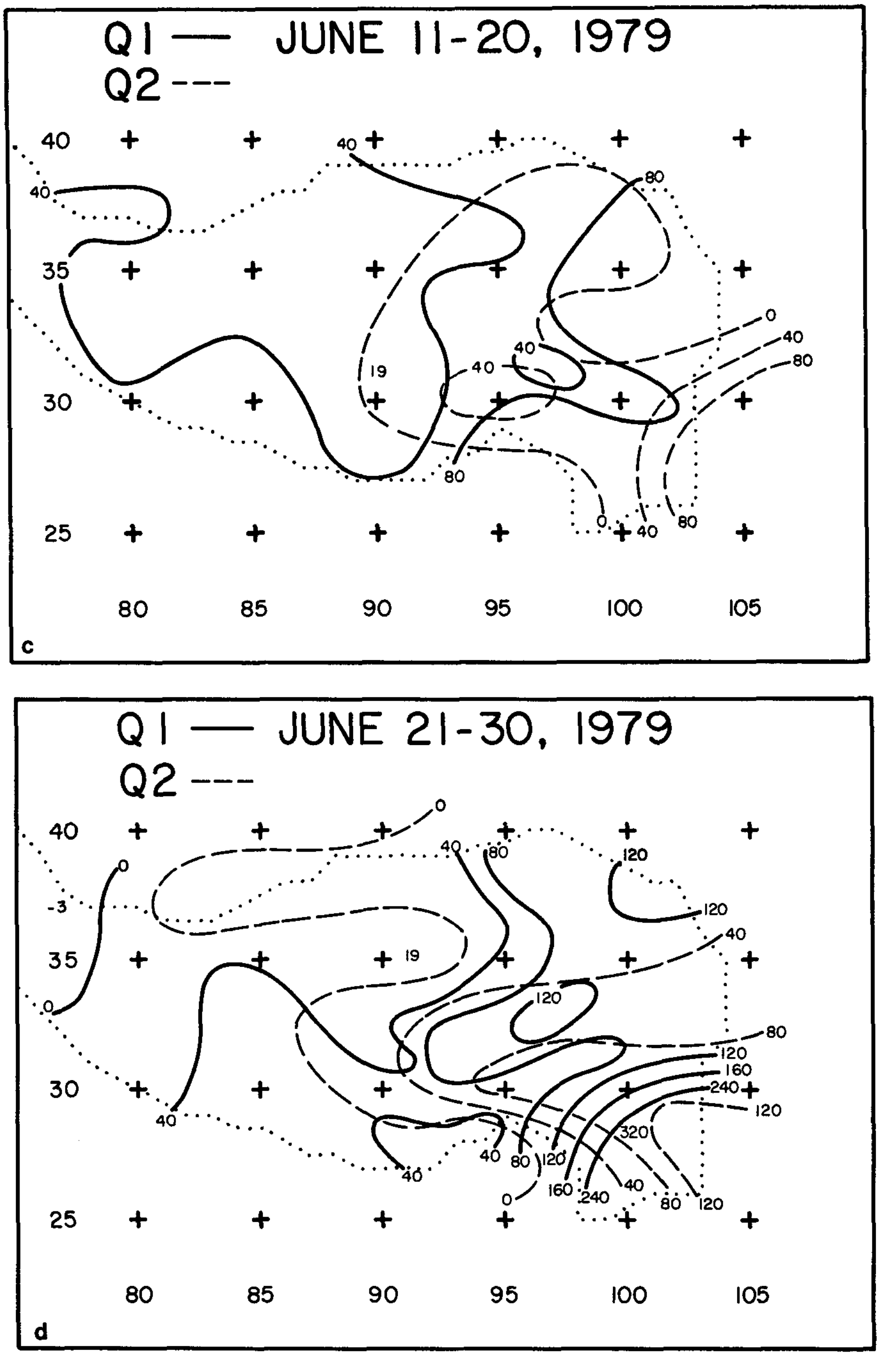


FIG. 5. (*Continued*)

month. For instance, the western region during July showed a positive hydrological balance, $\langle Q2 \rangle > 0$, whereas during June and August the balance was negative. The eastern plateau showed a positive balance throughout the summer.

Even though our computational results appear to reflect well the large-scale, climatic summer conditions in western and eastern Tibet, a word of caution is in order. The values estimated for R_0, SH, LP, and LE used observations collected from valley stations, although some of them were located at considerable elevation above sea level. The interior valleys of central and western Tibet are quite dry, many of them dotted with sand dunes (Reiter and Reiter, 1981), while some of the major mountain ranges show massive glaciation that would require much more than the reported mean annual precipitation to maintain itself against ablation. Unfortunately, no concrete estimates are available on

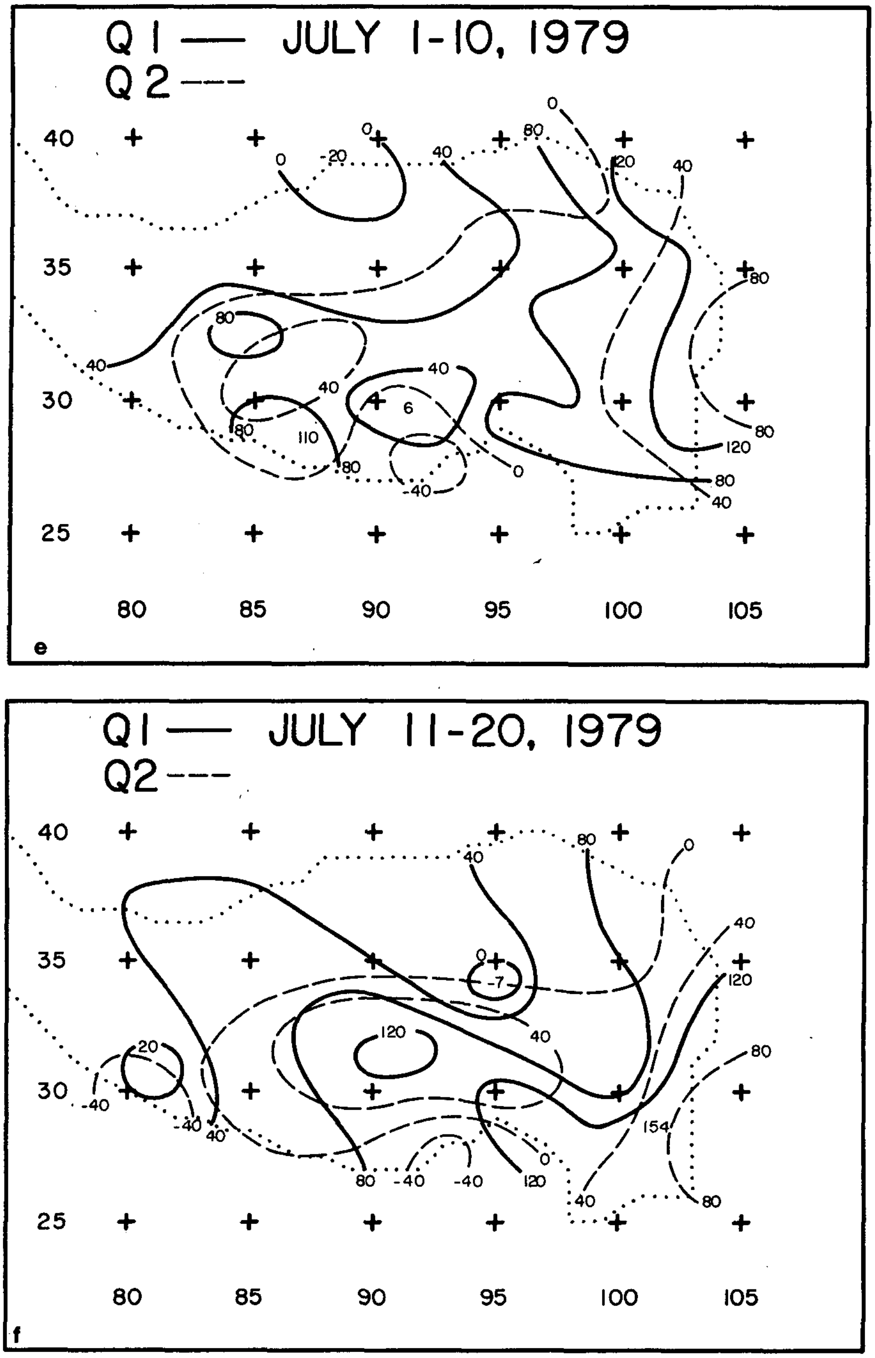
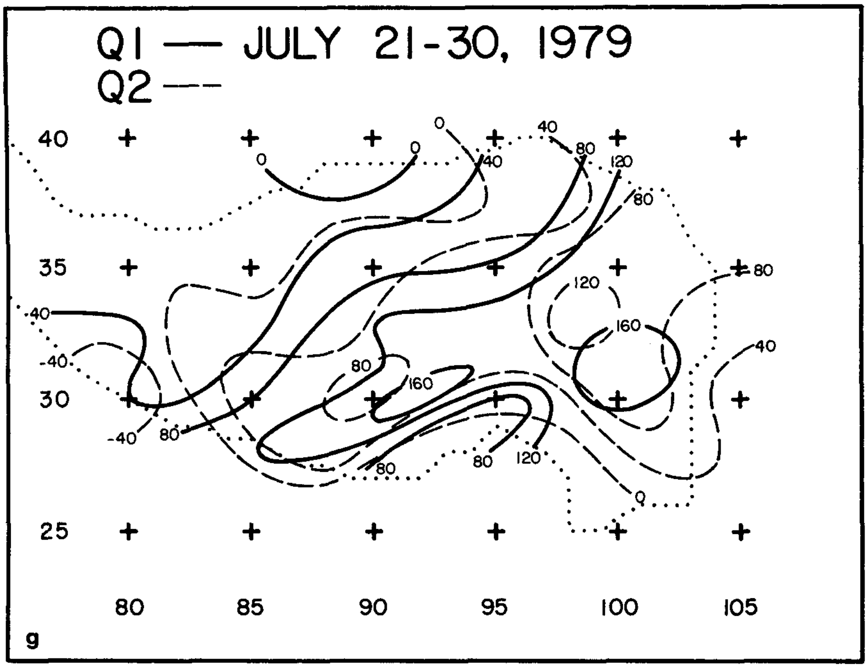


FIG. 5. (*Continued*)

the altitude dependence of precipitation in the interior of Tibet. We believe, however, that such numbers, if they were available, would not drastically alter our findings for the summer season because of the relatively small area covered by these glaciers. A slightly greater role of moisture flux convergences from outside the study region than that revealed by our analyses might have to be conceded.

4. Geographic distribution of the atmospheric heat source over Tibet

Figure 2 shows the monthly patterns of ⟨Q1⟩ during June, July, and August 1979. Maximum values of the heat source during June were located over the northeastern and southeastern parts of the plateau, while a minimum of 15 W m^{-2} resided over the central region.



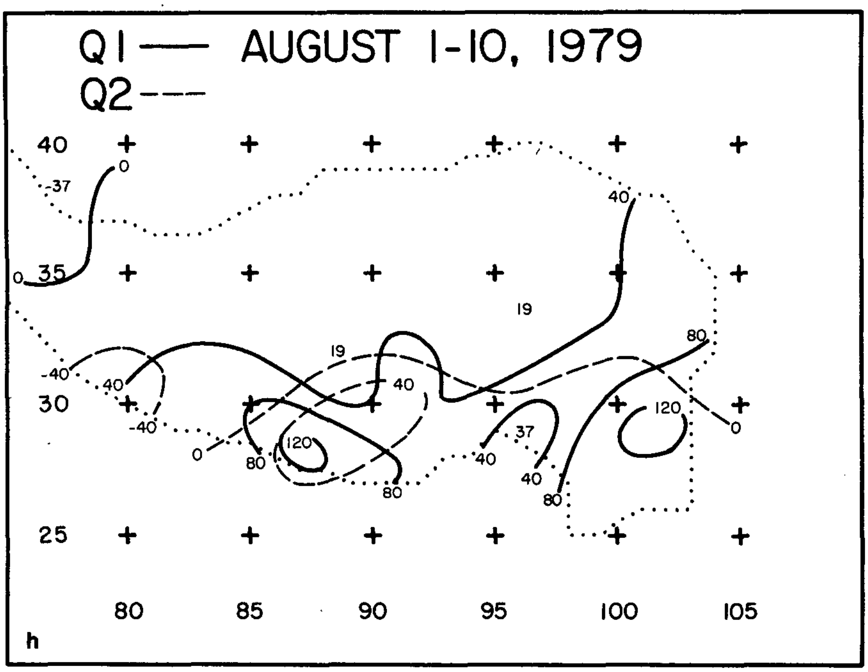


FIG. 5. (*Continued*)

During July the high value area of ⟨Q1⟩ moved westward and two high value centers were present, one over the central and the other over the southwestern region of the plateau. In August the maximum was found over the southern plateau, while the northwestern regions became a heat sink.

The components entering into the computation of ⟨Q1⟩ and ⟨Q2⟩ are shown in Figs. 3a–d for July 1979. Most of the latent heat input during that month (Fig. 3a) occurred over the southeastern and southcentral parts of the plateau, with maximum values along the valley of the Yarlung Zangbo Jiang (Brahmaputra River). At the same time, in this region, the sensible heat input remained low (Fig. 3b). As can be seen from these two diagrams, the roles of SH and LP as major heat sources for the atmosphere were reversed between

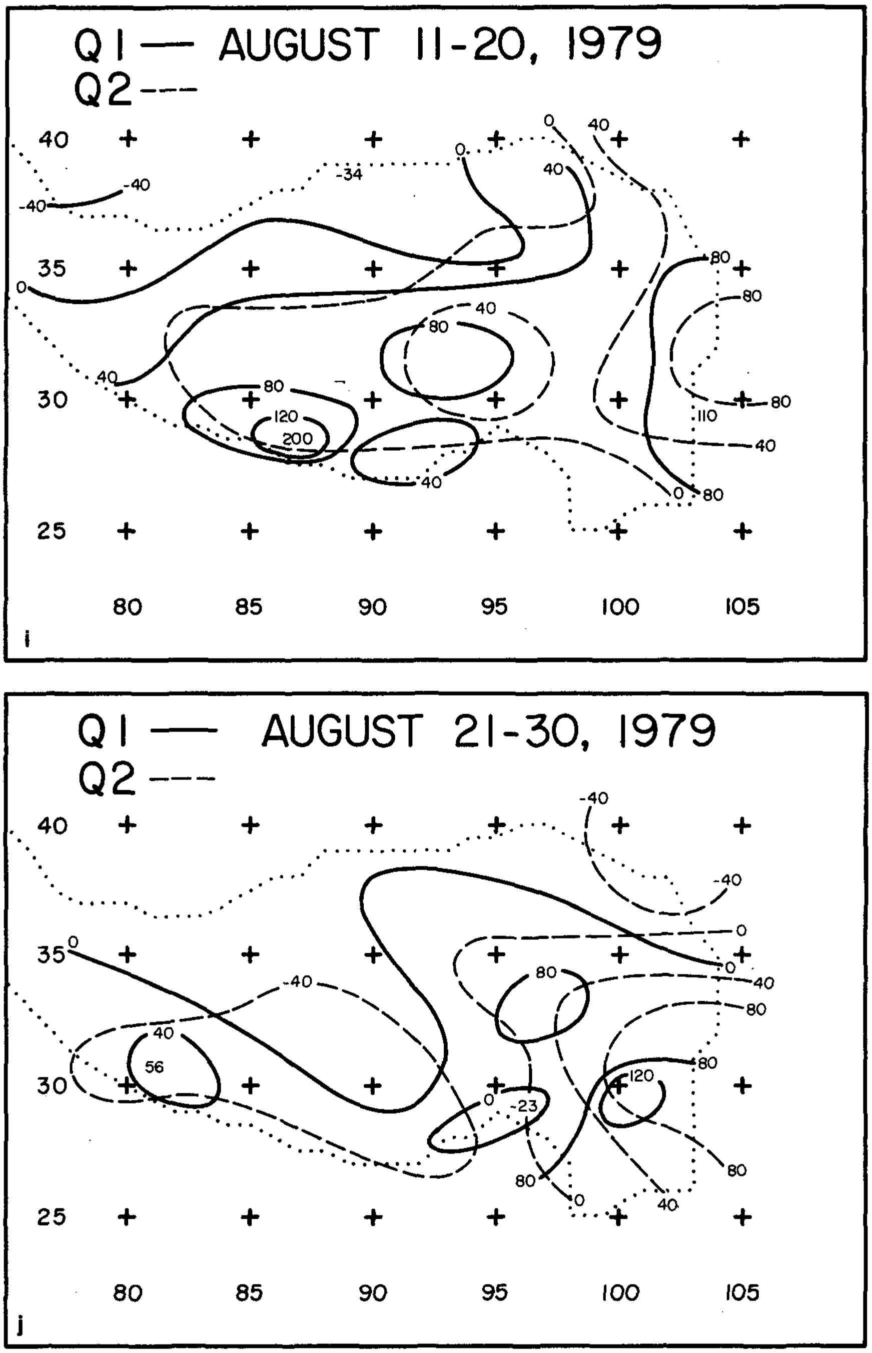


FIG. 5. (*Continued*)

the western and eastern plateau regions. The centers of radiation cooling during July (Fig. 3c) overlay the maxima of LP and SH in the middle and upper reaches of the Yarlung Zangbo Jiang where the surface albedo is relatively low. Surface evaporation, LE (Fig. 3d), reached a maximum where the Yarlung Zangbo Jiang and several smaller rivers break southward through the Himalayas. Here, in the region of rain forests (see Reiter and Reiter, 1981), evaporation plays a major role in supplying moisture to the atmosphere so that precipitation and evaporation are nearly in equilibrium, as can be seen from the distributions of ⟨Q2⟩, presented in Fig. 4.

The set of diagrams shows a negative atmospheric moisture source balance in western Tibet during June, while over eastern Tibet the influx of moist monsoon air generated a positive balance. During July most of the central and western plateau regions began to show a positive balance. The western plateau dried up again during August. For the summer average, positive values of ⟨Q2⟩ were maintained over the eastern part of the plateau where precipitation exceeded evaporation. The western part of the plateau, during the same summer period, showed an average balance between evaporation and precipitation. Apparently much of the moisture precipitating over western Tibet is recycled locally and does not come from the monsoon systems to the south, since the Himalayas provide a formidable moisture barrier. These conclusions are confirmed by the local geography: Western Tibet is dotted by numerous lakes, most of them salt lakes with no outflow. Only the tributaries to the Yarlung Zangbo Jiang provide outflow from southern Tibet.

5. Time variability of the atmospheric heat source

Figure 5 presents 10-day mean patterns of ⟨Q1⟩ and ⟨Q2⟩, starting with 21 May 1979. One immediately recognizes a great deal of variability between these relatively short time periods, obviously caused by the changing synoptic conditions. The progression of the moisture intrusion from the southeast into central Tibet between the last days of May and the end of July is clearly indicated in the patterns of ⟨Q2⟩. The monsoon burst over the Bay of Bengal during the first ten days of June which heralded the beginning of the rainy season over the southeastern part of the plateau. The atmospheric heat source, ⟨Q1⟩, shifted over that region during the same time period in early June (Fig. 5b). The first 10 days of August were characterized by a monsoon break over the plateau, indicated by a dramatic pattern shift and reduction in the values of ⟨Q2⟩ and ⟨Q1⟩ (Fig. 5h). A new monsoon surge intruded from the southwest during the second half of August.

The systematic variability of the heat source characteristics of the Tibetan Plateau are brought to light even better in the time series plots of Fig. 6 which presents overlapping five day means of ⟨Q1⟩ and its components (sensible heat, SH; latent heat, LP; and radiative cooling, RC) for the whole plateau (Fig. 6a), for the eastern region (Fig. 6b) and the western region (Fig. 6c). Nitta (1983) also plotted the time series of ⟨Q1⟩ for the eastern plateau (not shown here), which appears very similar to Fig. 6b except for greater absolute values and amplitudes. Sensible heating over each of the two plateau regions reaches maximum values before the onset of the rainy season in early June and, at that time, exceeds the contribution of LP towards the atmospheric heat source. The burst of the monsoon brings a dramatic rise in LP and a fall of SH, first over eastern Tibet and almost one month later over western Tibet. Of all components, LP reveals the largest amplitudes of variation and, thus, becomes the main modulator of ⟨Q1⟩, with oscillation periods between 10 and 15 days. Spectrum analysis (Feng *et al.*, 1985) confirms this quasi-periodicity whose prominence is next to that of the seasonal cycle.

6. The distribution of the Tibetan heat source in relation to the Indian summer monsoon

The classic monsoon theory postulates that heating of the Tibetan Plateau promotes the establishment of an upper tropospheric anticyclone. The summer monsoon circulation, then, is maintained by the release of latent heat which occurs mainly on the southeastern flank of this anticyclone (for a survey of literature see Reiter, 1963, 1969). There still exists, however, some controversy regarding the details of the involvement of the Tibetan heat source in the evolution of the Indian summer monsoon system.

Nitta (1983), for instance, correlated the heat source over eastern Tibet with the monsoonal rainfall yield over India during summer 1979 and found a reasonably good out-of-phase correlation between these two parameters. Huang (1984) pointed out that dry spells over Tibet often happen in conjunction with active Indian monsoon periods, whereas Indian monsoon breaks coincide with wet spells over Tibet (see also Chang, 1981). Since the release of latent heat has a major impact on the intensity of the heat source over Tibet, as we have demonstrated in the foregoing sections, these two conclusions appear to be in agreement.

With the data on hand we were able to further shed some light on this problem. Taking Nitta's (1983) rainfall data over the Indian peninsula as an index for the intensity of the Indian summer monsoon of 1979, we defined four phases of monsoon behavior whose dates are listed in Table 4. Three distinct rainfall oscillations during summer 1979 provided us with three cases in each of the four phases. Granted, one cannot call this sample statistically significant. Nevertheless, the results of our analysis merit discussion because they give at least a preliminary indication of the processes involved in the aforementioned correlation.

In Fig. 7 the total atmospheric heat source, ⟨Q1⟩ in W m^{-2}, is shown in its composite, spatial distribution for the four monsoon phases listed in Table 4. Over the southeastern and southwestern plateau regions ⟨Q1⟩ achieved maximum values during the peak rainfall days over India, opposing to Nitta's (1983) conclusion and suggesting that monsoon disturbances intrude over these regions. However, over the central plateau the heat source maximized during the transition from Indian rainfall maximum to minimum (Phase 2) and reached extreme negative values during Phase 4 (transition from Indian rainfall minimum to maximum), in agreement with Nitta's (1983) and Huang's (1984) statements. Our analyses suggest, therefore, that Nitta's and Huang's conclusions are not necessarily applicable simultaneously to different parts of the plateau.

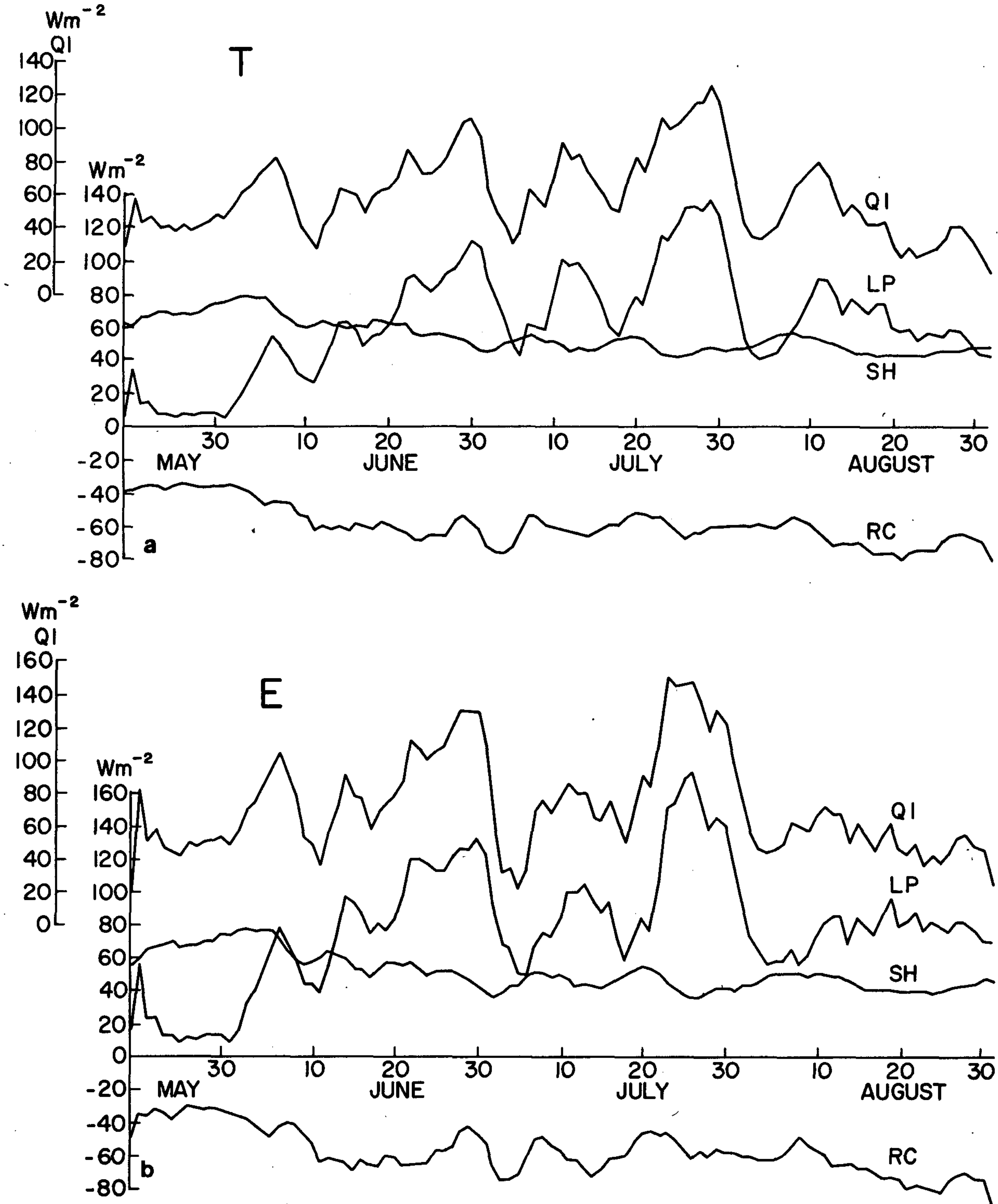


FIG. 6. Time series of ⟨Q1⟩, LP, SH and Rc for the summer of 1979. (a) Averaged over the whole plateau; (b) over the eastern plateau; and (c) over the western plateau.

7. Conclusions

In our present study of the atmospheric heat source over Tibet we employed the "direct" method of estimating heat budgets from local station observations and from Nimbus-7 satellite data, thus avoiding some of the uncertainties commonly associated with the "indirect" method of budget estimates involving the cal-

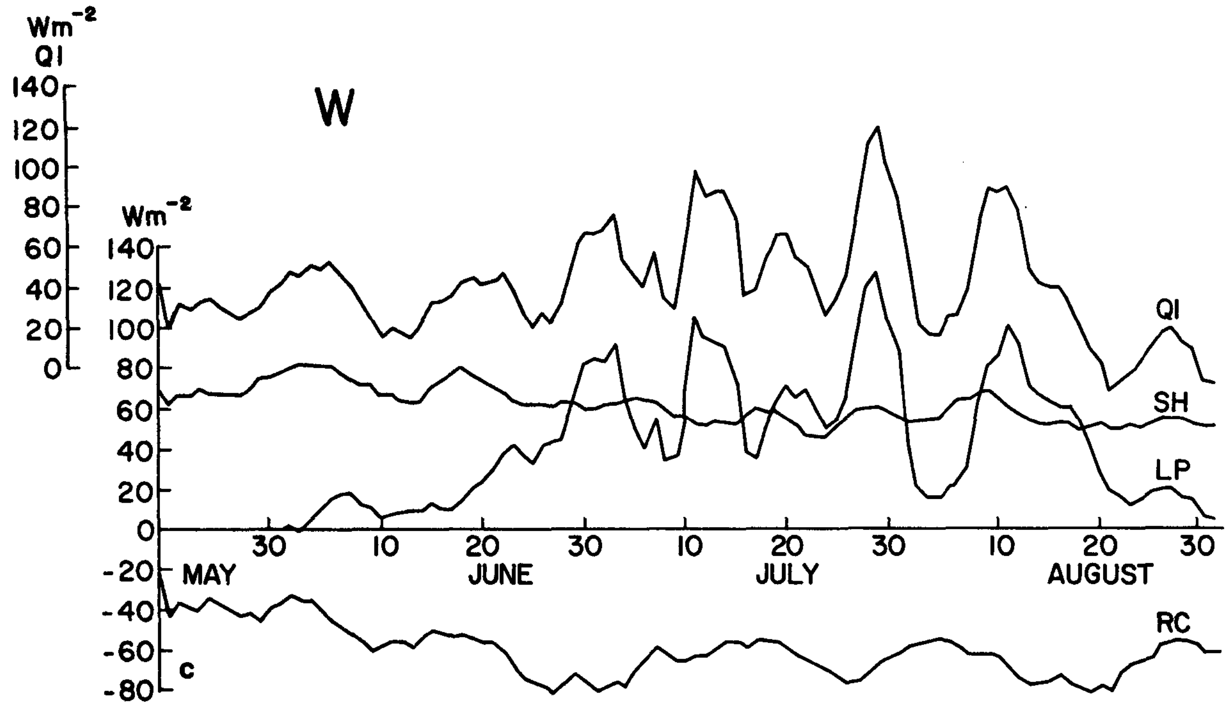


FIG. 6. (*Continued*)

culation of vertical velocity and horizontal advective terms. Our surface budget estimates hinge upon the application of empirical relationships developed from field data from the 1979 TIPMEX. Our analyses, for the first time, provide reasonable estimates for the atmospheric heat budget and its components over western Tibet, heretofore considered a data-void region. These estimates are consistent with the large-scale surface hydrological characteristics of that region. Nevertheless, several caveats had to be raised during the discussion of our research results.

One of them concerned the application of new, altitude-dependent drag coefficient values derived from six surface energy budget stations established in Tibet during the summer of 1979. These values, approximately half as large as those originally proposed by Cressman (1960) and since adopted in most of the previous heat budget calculations and numerical modeling experiments, tend to reduce drastically the strength of the sensible heat source over Tibet. In favor of these reduced drag coefficient values is the fact that our own numerical model runs (Shen *et al.*, 1985) performed considerably better with the smaller than with the larger coefficients. We expect, however, that the book on drag coefficients over mountainous terrain and high plateaus is not yet closed.

Of somewhat less concern, but still of importance to the reliability of our estimates, is the problem of "valley biases" in the precipitation data that formed the foundation for the estimates of the latent heat source, LP, and of the atmospheric moisture source, $\langle Q2 \rangle$. Such valley biases apply to most mountainous regions of the globe. The few mountain observatories which collect precipitation data at peak or ridge locations bring their own set of problems to a solution of the question of accurate precipitation estimates. Under the usually prevailing strong wind conditions, rainfall catchment becomes a function of rain gauge design and exposition.

We believe that the set of radiation budget data at the top of the atmosphere available from Nimbus-7 presents the state of the art for such estimates. The reduced rate of atmospheric cooling in our estimates, as compared to those by other authors, may be the

TABLE 4. Dates of monsoon phases during summer 1979.

Maximum:	June 26, 27, 28; July 14, 15, 16; August 9, 10, 11
Maximum–minimum transition:	June 30; July 1, 2, 17, 18, 19; August 15, 16, 17
Minimum:	July 4, 5, 6; July 28, 29, 30; August 25, 26, 27
Minimum–maximum transition:	June 19, 20, 21; July 9, 10, 11; August 4, 5, 6

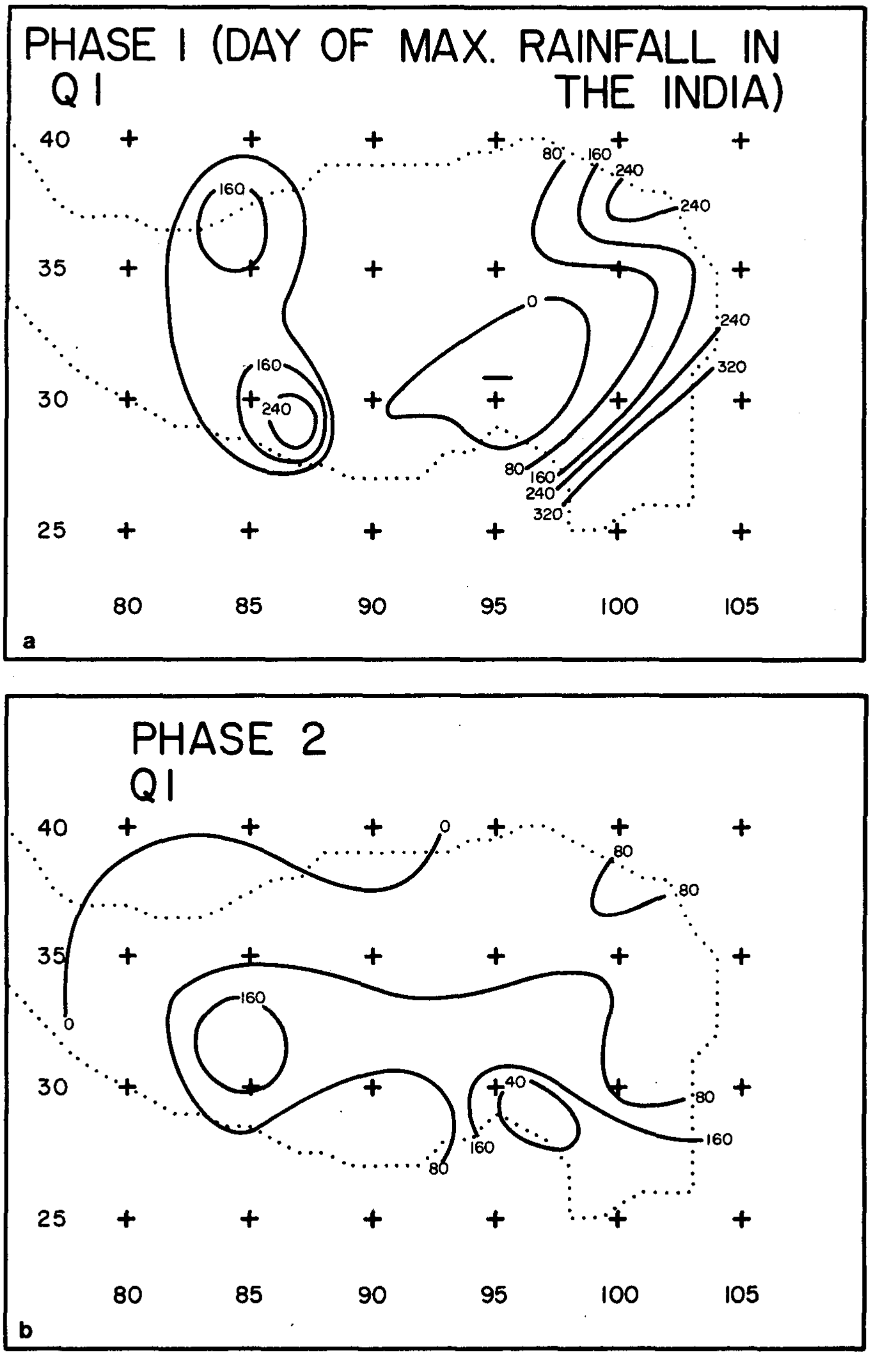


FIG. 7. Total atmospheric heat source, $\langle Q1 \rangle$, in W m^{-2} during the four monsoon phases listed in Table 4. (a) Periods of maximum rainfall in India; (b) transition from maximum to minimum; (c) periods of minimum rainfall in India; (d) transition from minimum to maximum.

effect of improved satellite data, but also may indicate interannual variability, which our data base does not yet permit us to assess.

Evaporation effects (LE) on the atmospheric water vapor source are notoriously difficult to estimate, hence unreliable with the present state of the art. Direct, quantitative assessments of evaporation, under most circumstances, are difficult, especially in the absence of lysimeter, soil moisture and river runoff data which could provide a means to narrow the error brackets of this parameter.

In spite of these shortcomings our analyses allow several interesting conclusions.

1) The sensible heat source, SH, dominated over the plateau before the onset of the rainy season.

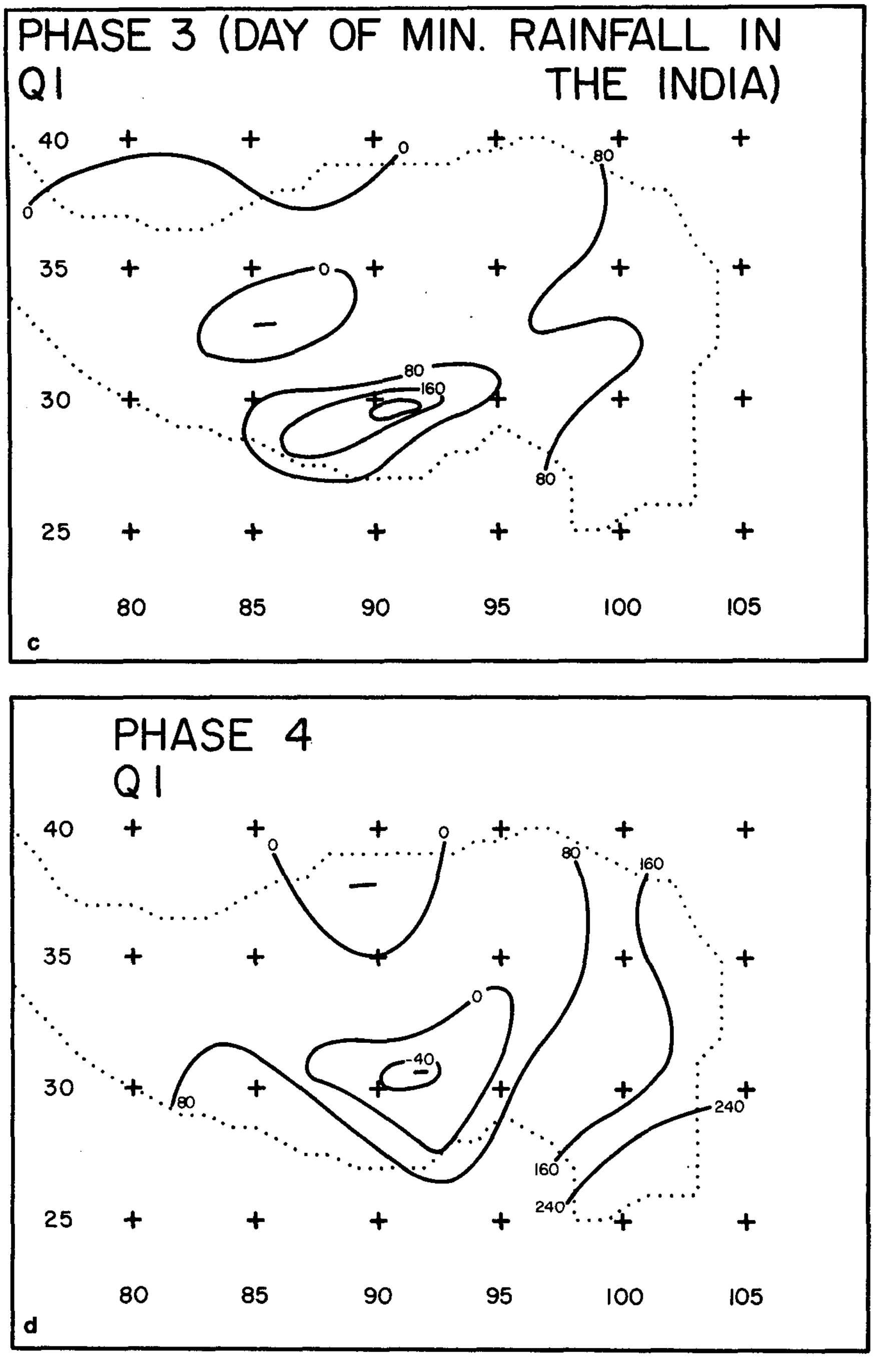


FIG. 7. (*Continued*)

2) The rainy season first came to the eastern plateau and caused a dominance of LP over SH. The western plateau experienced the onset of the rainy season almost one month later, lending equal significance to SH and LP there.

3) The estimates of ⟨Q2⟩, the atmospheric moisture source, indicated that over the western plateau the recycling of water through local evaporation plays an important role in the hydrological balance of the region, whereas over the eastern plateau water vapor amounts close to 40% of the observed precipitation had to be supplied by the convergence of fluxes from outside the region. In agreement with this conclusion, river runoff from western Tibet is, more or less, limited to the headwaters of the Yarlung Zangbo Jiang (Brahmaputra River), whereas eastern Tibet is the source of several major river systems, the Lancang Jiang (Mekong), Chang Jiang or Jinsha Jiang (Yangzi), and Huang He

being among the largest. The large number of salt water lakes over western Tibet affirms the conclusion of the relative importance of local evaporation sources in this region.

4) The atmospheric heat source over Tibet showed a significant oscillation with a period between 10 to 15 days. This oscillation appears to be modulated mainly through LP. It is tempting to assume that it is coupled to a similar oscillation found by Krishnamurti and Bhalme (1976) in monsoon disturbances over India.

5) Whereas the atmospheric heat source over southeastern and southwestern Tibet appears to be in phase with the monsoonal precipitation yield over the Indian Peninsula mainly through the action of LP, the central Tibetan heat source shows a distinct, out-of-phase behavior.

Even though our study incorporated a host of new data made available only recently from Chinese and United States sources, a lot of "whittling away" on the error bars surrounding our estimates remains to be done.

Acknowledgments. This research has been supported by grants from the National Science Foundation, Grant Number ATM 83 13270, Climate Dynamics Program, Atmospheric Science Division; and NASA, Grant Number NAGW-601.

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