Subtropical Droughts and Cross-Equatorial Energy Transports

E. B. KRAUS

Cooperative Institute of Marine and Atmospheric Studies, Miami, Fla. 33149

(Manuscript received 19 November 1976, in revised form 16 February 1977)

ABSTRACT

The spatial coherence of subtropical rainfall anomalies is documented by variance analysis. Major droughts repeatedly were felt at the same time around the globe along the arid margins of the tropical rainfall belt. The persistence of anomalies becomes apparent in precipitation time series which combine data from relatively large areas and in streamflow records. These can be used to demonstrate autocorrelations and unexpectedly long runs of wet or dry years.

The latest drought episode culminated in 1972 not only in the Sahel and the Sudan, but also along the borders of the Indian desert and in Central America. It is shown to have been accompanied by relatively low temperatures in the southern subtropics and by abnormally high temperatures in the antarctic. The meridional temperature gradient and the meridional slope of the 500 mb surface were correspondingly reduced. It is suggested that this was associated with a reduced demand for energy (and zonal momentum) exports from the tropics and therefore relatively weak direct tropical circulations. As a result, these circulations—which tend to straddle the equator—did not deliver the normal amount of precipitation along their northernmost borders in the monsoonal fringe area.

1. Introduction

There are no absolute criteria for drought; the dryness of the desert does not wilt the thorn bush or the cactus and it is not experienced as a disaster by Bushmen or Bedouins. By definition droughts are anomalies —deviations from a rainfall regime to which people, plants and animals have adapted as the local norm.

Seasonal rainfall variations in the subtropics are determined by the behavior of the intertropical convergence zone (ITCZ). Compared to its oceanic counterpart, the continental ITCZ moves seasonally much further north and south. In monsoon regions it also affects a much broader area. Summer rainfall failures can be associated with a smaller seasonal shift of the ITCZ, with a reduction of the area covered by the rain-producing synoptic systems or with a reduction of the moisture convergence into that area. The evidence presented here suggests that the big droughts of this century were associated partly at least with the first of these mechanisms. There also are indications that at least during the latest great subtropical drought, the reduced northward excursion of the ITCZ was accompanied by a relatively weak southward heat flux into the Southern Hemisphere. This is in keeping with an interpretation of the ITCZ as a thermal equator or as the locus of a divide which separates areas of zero integral heat balance. The northward displacement of the ITCZ would then be reduced if the demand for heat transport to southern latitudes is relatively small.

2. Characteristics of regional rainfall records

Rainfall records are more "noisy" than those of other geophysical variables. This applies particularly to regions which receive their water through convective storms. A few isolated perturbations can, and often do, produce transient excess precipitation locally without necessarily breaking a continuing drought. In general, the overall duration of a subtropical drought period can therefore not be documented from the analysis of single station rainfall records. Streamflow records are more suitable. Alternatively, the records from several stations have to be combined.

Such a combination cannot be carried out by simple addition or arithmetic averaging of local rainfall values. The number of stations which are operative in a given region may vary from year to year. Some may have useful records which are shorter, however, than the overall period which is the subject of analysis. At the same time one must expect large and systematic variations in the average record of rainfall stations which are only short distances apart. The isohyetal gradient can be particularly sharp in subtropical hill country. A simple averaging of total precipitation amounts would produce in this case a combined time series which would be dominated entirely by stations with the highest rainfall intensity. To minimize this effect some form of normalization procedure must be adopted.

Symbols used below have the following meaning:

r_{ij} annual rainfall at station i during year j
J number of years in period chosen for analysis
J_i number of record years of station i during the period J
I number of stations in the region to be analyzed
I_j number of regional stations operative in the year j.

The mean annual rainfall at station i and its variance are

$$\bar{r}_i=\frac{1}{J_i}\sum_j r_{ij},\quad \sigma_i{}^2=\frac{1}{J_i}\sum_j r_{ij}^2-\bar{r}_i{}^2. \tag{1}$$

Two normalizations were tried in the present investigation:

$$x_{ij}=(r_{ij}-\bar{r}_i)/\sigma_i \quad\text{and}\quad x'_{ij}=(r_{ij}-\bar{r}_i)/\bar{r}_i. \tag{2}$$

Both these formulations have zero means, i.e.,

$$\bar{x}_i=\overline{x'_i}=0. \tag{3}$$

Though they give essentially analogous results, the first is preferred here because it gives somewhat less weight to desert stations which have a relatively large variance, but contribute little to the water resources of a region. The values of x_{ij} as distinct from x'_{ij} also provide immediate information about the significance of particular deviations from the mean, and their variance in time has the very simple form

$$\frac{1}{J_i}\sum_j x_{ij}^2=1. \tag{4}$$

The area-averaged value of x_{ij} for the year j, i.e.,

$$a_j=\frac{1}{I_j}\sum_i x_{ij}, \tag{5}$$

provides a useful index of regional rainfall for that year.

Before one can use this index with confidence for the discussion and analysis of regional year-to-year fluctuations, it has to be demonstrated that it is indeed representative for the region as a whole. In other words, it has to be shown that the geographical variations of the x_{ij}'s between different places within the given region are small, compared to the temporal year-to-year variations of the whole region as represented by the a_j's. This can be done with a variance analysis which partitions the total sum of all x_{ij}^2 into a spatial and a temporal part.

The I_j station values which are available for the year j can be considered a particular sample or subset of a set of data which consists of all the annual rainfall values from all contributing stations. The total number of these station years is

$$N=\sum_j I_j=\sum_i J_i. \tag{6}$$

It follows from the four preceding expressions that

$$\sum_j I_j a_j=0,\quad \sum_i\sum_j x_{ij}^2=N. \tag{7}$$

Allowing for (7) and for the fact that I_j may vary from year to year, one can now estimate the variance in time, which characterizes the year-to-year regional fluctuations in the form

$$V(\text{time})=\frac{\sum I_j a_j{}^2}{J-1}. \tag{8}$$

An estimate of the mean geographical variance between the rainfall anomalies within the region is given by

$$V(\text{area})=\frac{N-\sum I_j a_j{}^2}{N-J}. \tag{9}$$

The relative importance of these two variance estimates can be assessed by an F test (see e.g., Panofsky and Brier, 1958).

As an example, Table 1 presents combined rainfall data for two regions. The basic data set for this table consisted of all rainfall records which were available at the National Center for Atmospheric Research (NCAR) and which provided at least 30 annual rainfall values within the 64-year period 1911–74. The left-hand side of the table represents the area-averaged rainfall anomaly for all stations in the region 30°W–60°E and 10°N–25°N. On the right side of the table the figures in the last column are based on data from 10 stations in the northwestern fringe area of the Indian monsoon. The selected stations (Lahore, Kalat, Hyderabad, Bikaner, Agra, Jodhpur, Jaipur, Ahmadabad, Dwarka, Veraval) are all within a distance of 300 km from the Indian desert.

For the African data we have:

Variances:	$V(\text{time})=4.079$	$V(\text{area})=0.832$
Degrees of freedom:	$J-1=63$	$N-J=1340$

The ratio of the variance estimates is 4.90 which is highly significant, as the limiting value of F for 1% probability is only 1.49. In other words, the probability that differences between years in the combined regional rainfall record can be accounted for by random fluctuations at a few individual stations, is substantially less than 1:100. Rainfall anomalies all across the North African summer rainfall region tend to vary indeed in a statistically coherent manner. For the 10 monsoon stations which provided the basis for the averaged normalized anomalies on the right-hand side of Table 1, the ratio $V(\text{time})/V(\text{area})=5.71$, which is even more significant and suggests a corresponding regional coherence.

TABLE 1. Area-averaged normalized rainfall anomalies (a_j) for two monsoonal rainfall regions. (The two regions in Africa and in the northwestern part of the Indian subcontinent are defined in the text.) The last column indicates the years when the rainfall anomalies in the two regions had either the same (=) or the opposite (≠) sign.

Year	Africa		India			Year	Africa		India		
(j)	I_j	a_j	I_j	a_j		(j)	I_j	a_j	I_j	a_j	
1911	10	−0.234	10	−0.901	=	1943	31	0.331	9	0.068	=
1912	10	−0.073	10	0.077	≠	1944	31	−0.341	9	1.105	≠
1913	10	−0.930	10	−0.147	=	1945	31	0.090	9	0.473	=
1914	10	0.200	10	0.348	=	1946	32	0.436	9	−0.343	≠
1915	10	−0.461	10	−1.008	=	1947	32	−0.374	10	−0.083	=
1916	12	0.365	10	0.424	=	1948	32	−0.405	10	−0.067	=
1917	12	−0.663	10	1.750	≠	1949	30	−0.442	10	−0.045	=
1918	13	−0.181	10	−1.463	=	1950	31	0.739	10	0.544	=
1919	13	−0.588	10	−0.005	=	1951	32	0.308	10	−0.298	≠
1920	13	0.267	10	−0.492	≠	1952	32	0.685	10	−0.139	≠
1921	15	−0.038	10	−0.104	=	1953	32	0.667	10	−0.108	≠
1922	15	0.106	10	−0.497	≠	1954	32	0.551	10	0.467	=
1923	15	0.567	10	−0.519	≠	1955	32	−0.461	10	0.176	=
1924	15	0.292	10	0.339	=	1956	32	0.445	10	1.227	=
1925	15	−0.127	10	−0.653	=	1957	32	0.482	10	0.095	=
1926	14	−0.902	10	0.768	≠	1958	32	0.291	10	0.582	=
1927	14	0.532	10	0.182	=	1959	32	0.108	10	1.196	=
1928	14	0.071	10	−0.361	≠	1960	31	−0.329	10	−0.341	=
1929	14	0.466	10	−0.102	≠	1961	29	−0.828	9	1.055	≠
1930	14	0.324	10	−0.387	≠	1962	28	0.129	8	0.423	=
1931	14	0.095	9	0.140	=	1963	27	−0.212	8	−0.392	=
1932	14	0.045	9	−0.196	≠	1964	27	0.543	8	0.864	=
1933	14	0.255	9	1.169	=	1965	27	0.125	8	−0.323	≠
1934	14	0.007	9	−0.053	≠	1966	27	−0.349	8	−0.280	=
1935	14	0.124	9	0.044	=	1967	27	−0.138	9	0.968	≠
1936	15	0.201	9	0.103	=	1968	26	−0.413	9	−0.975	=
1937	15	−0.123	9	−0.035	=	1969	26	−0.395	9	−0.681	=
1938	14	0.269	9	−0.825	≠	1970	26	−0.461	9	0.444	≠
1939	14	0.251	9	−0.664	≠	1971	26	−0.616	9	−0.388	=
1940	14	−0.382	9	−0.463	=	1972	26	−0.857	9	−0.770	=
1941	30	−0.512	9	−0.888	=	1973	26	−0.724	9	0.126	≠
1942	31	−0.340	9	0.596	≠	1974	26	−0.350	9	−0.778	=

The NCAR data set with the same limiting criteria (30 years of record between 1911 and 1974) was used to compute also area-averaged normalized anomalies for the following other 90° sectors of longitude:

Subtropics

10°N–25°N	30°W–120°W	(N. America)
10°N–25°N	140°W–130°E	(N. Pacific)
10°S–25°S	140°W–130°E	(S. Pacific & Australia)
10°S–25°S	60°E – 30°W	(S. Africa)

Equatorial Zone

10°S–10°N	30°W–120°W	(America)
10°S–10°N	120°W–150°E	(Pacific)
10°S–10°N	150°E – 60°E	(S. Asia)
10°S–10°N	60°E – 30°W	(Africa)

The resultant ratios of variance estimates exceeded the 1% limiting values of F in each of the subtropical regions, though they were not quite as high as those which characterized the North African and Indian values listed in Table 1. In the equatorial zone the ratio was generally lower, but still significant on the 1% level in the American and African sectors. It was significant on the 5% level in the Indian/Indonesian sector of South Asia. In the equatorial Pacific sector, the ratio was not formally significant. The number of stations available there, however, was too small for any definite conclusion.

It follows from these analyses, that within regions of relatively large extent, one year differs from the next in a significant way. Area-averaged normalized anomalies can be used with some confidence as an index for the study of these year-to-year fluctuations.

3. The duration of droughts

The most striking feature of the African data listed in Table 1 is the length of runs of anomalies of either sign. For example, there was a deficit of rain during every year from 1966 to 1974. Though the year 1972 had the largest deficit, the disastrous famines of that year were due largely to the cumulative legacy of the preceding run of dry years.

Even more surprising, in arid regions where station rainfall statistics tend to be positively skew, are the long runs of positive anomalies. Relatively wet years followed each other without interruption from 1927 to 1936 and again with even larger excesses from 1950 to 1959. These

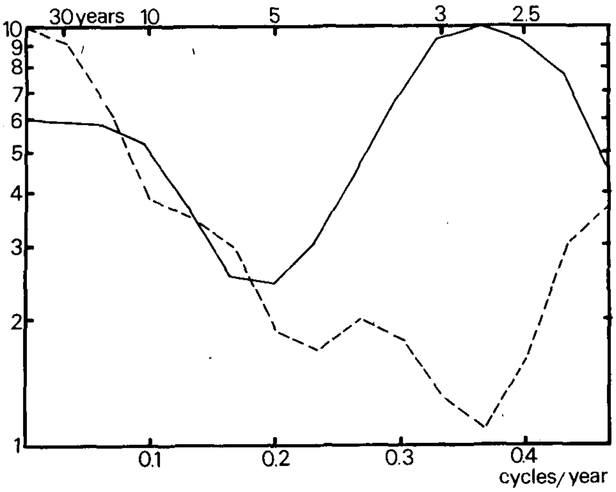

FIG. 1. Normalized power spectra for the time series presented in Table 1 (Africa, full line; India, dashed line).

runs become apparent only in the area-integrated precipitation or in streamflow records; they were not found by Bunting *et al.* (1976) in their analysis of individual African station records.

The probability of finding long runs of either positive or negative values in a limited record of a randomly distributed variables has been discussed by Julian (1970). Application to the African record listed in Table 1 of the formula quoted by Julian indicates that the chance of finding two runs of 10 wet years in 64 years of record would be about $1:10^6$. This leads to the conclusion that the African rainfall series is almost certainly not random. The clumping or persistence of dry and wet anomaly values which are exhibited by the record must be assumed to be the manifestation of a real physical process. The biblical story of the seven fat and the seven lean years in Egypt may well have had a basis in fact.

There are of course also other methods to measure persistence in the African rainfall record. In particular, the series of a_j's have an autocorrelation coefficient $r(j, j+1)=0.32$. With $64-3=61$ degrees of freedom, a "Students" t-test indicates that the significance of this value is almost exactly 0.99. The computed autocorrelation is depressed by the flip-flops in the record; for example from -0.902 in 1925 to $+0.532$ in 1926 or from -0.442 in 1949 to $+0.739$ in 1950. The departure from randomness which is suggested by the value of the 1% probability level is still very persuasive.

The record from the Indian monsoonal fringe area is positively skew and much noisier than the African record. There are no unusually long runs, perhaps because the area is smaller with fewer stations, but this does not imply that there were not regimes which were either predominantly wet or predominantly dry over relatively long periods. For example, in the 10-year interval 1965–74, simultaneously with the great African drought, one finds seven years of negative anomaly in the monsoonal fringe area. In the 1950s when North Africa had its long wet spell, a run of six wet years (1954–59) occurred also in the Indian region. Earlier on, both records show only four years with a positive anomaly in the period 1911 and 1922.

In the Indian monsoonal fringe area, dry spells are liable to be interrupted by storms or series of storms which can cause a short-lived heavy excess of precipitation. This produced some wet years like 1917 or 1967 which both show large positive anomalies in the middle of otherwise predominantly dry periods. As a result there is little autocorrelation in the Indian record and the numerical covariance between the African and the Indian time series is insignificant.

That does not mean that the two areas have completely independent precipitation regimes. One can reduce the search for a parallel behavior to the question whether the area-averaged anomaly values have the same sign in the two areas during a significant number of years. This can be established by the computation of a tetrachoric correlation. The area-averaged normalized anomalies had the same sign in the two regions in 40 years; they had opposite signs only in 24 years. The associated tetrachoric correlation coefficient $r^1=0.34$, significant at the 0.99 level. In other words there is only a 1% probability that the relatively frequent, simultaneous occurrence of dry or wet years in the two regions could have been produced by chance.

The nature of the relationship in the climatic development of the two regions becomes clearer by a spectral analysis. The two graphs in Fig. 1 show the spectra for the time series listed in Table 1. Linear trends, though insignificant, were removed before analysis. Each spectrum can be seen to be bimodal and this seems to be characteristic for many spectra of annual tropical and subtropical precipitation. There is commonly a high-frequency peak which is probably associated with the quasi-biennial oscillation. It has a characteristic frequency which corresponds everywhere to a period between two and three years. As the Nyquist boundary for annual data is two years, the present spectra cannot be used to identify the actual frequency of this peak with much accuracy. The quasi-biennial peak is separated by something like a spectral gap from a second peak which demonstrates the power in fluctuations with periods of a decade or more. The existence of these two spectral peaks in the climatic record was noted also by Mitchell (1976).

The spectral basis of the relationship in the rainfall regime of the two areas is illustrated by the coherence graph in Fig. 2. It shows two peaks with moderately significant coherence for fluctuations with low and with relatively high frequencies. A cross-phase analysis indicates further that fluctuations at these frequencies tend to be in phase in both areas.

The nature of the subtropical annual rainfall spectra,

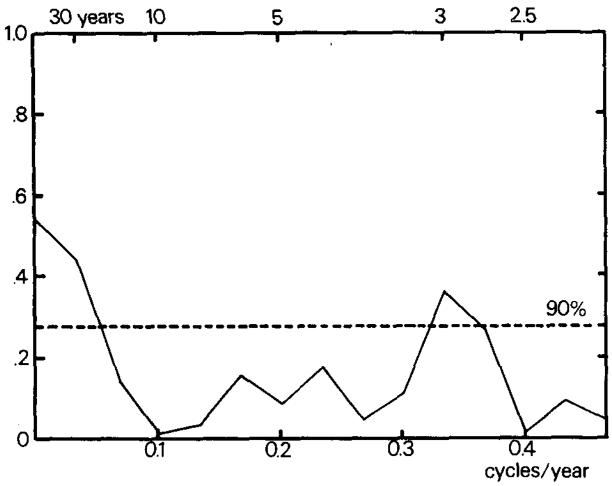

Fig. 2. Coherence of the two time series presented in Table 1. The dashed line represents the 90% significance level.

with their two rather marked peaks, implies a measure of potential predictability. Whether this is enough for operational purposes is not certain, but it does suggest further exploration.

4. The great subtropical droughts

Having established that the African and Indian monsoon regimes exhibit some evidence of parallel behavior, particularly along their arid margins, one wants to know whether that is also reflected in other subtropical regions.

Fig. 3 is based on actual station values of normalized rainfall anomalies (x_{ij}) in the whole belt between 30°S and 30°N. In the Northern Hemisphere and in the equatorial zone the anomalies are for the calender year 1972. In the monsoonal summer rainfall areas of the Southern Hemisphere, they were computed for the September 1971–August 1972 rain year. No isohyets were drawn, because this would have been rather arbitrary in the absence of oceanic information.

The chart indicates that drought conditions were prevalent in 1972 all through the subtropics; not only all across Africa and northern India, but also in America, along the high pressure belt in the South Pacific and in Australia. The widespread lack of rainfall in the subtropics was compensated or perhaps even over-compensated by excess rainfall closer to the equator in southern India, the Philippines and the western Pacific including the Coral Sea. There also were excess monsoon rains in the highlands of southern Africa.

The value of Fig. 3 as a drought indicator may be questioned on the grounds that rainfall statistics in arid areas are positively skew. More dry years are therefore expected *a priori*. Though this is true, the expected excess is much smaller than the actual excess of black symbols in Fig. 3. The skewness of annual rainfall distribution is not very pronounced. Even in such an extreme location as Khartoum which is close to the desert and has no winter rainfall at all, there were 29 years of above average rainfall against 35 years of below average in 64 years of record. The standard deviation of a normal distribution with 32 dry and 32 wet years is $(64\times0.5\times0.5)^{\frac{1}{2}}=\pm4$ years. The excess three dry years in the Khartoum record represent therefore a departure from normal which is less than one standard deviation. For North Africa as a whole the number of wet anomalies is actually somewhat larger than the number of dry years as indicated by Table 1. It follows that Fig. 3 can be accepted as representing truly an exceptional prevalence of drought conditions in the subtropics. The picture is compatible with a reduced northward excursion of the ITCZ during the summer of 1972 and possibly also with a reduced southward excursion (except for Rhodesia) in the preceding southern summer.

Similar widespread and long lasting drought episodes are not particularly rare events in the climatic regime of this century. One did occur in the early 1940's, culminating with famine in northern India, disastrous crop failures in Mexico and a 29 million decline in the sheep population in subtropical Australia (Director of Meteorology, 1960). Stratified values of the normalized rainfall anomalies for 1940 (rain year 1940/41 in the monsoon areas of the Southern Hemisphere) are charted in Fig. 4. The distribution of symbols on this figure

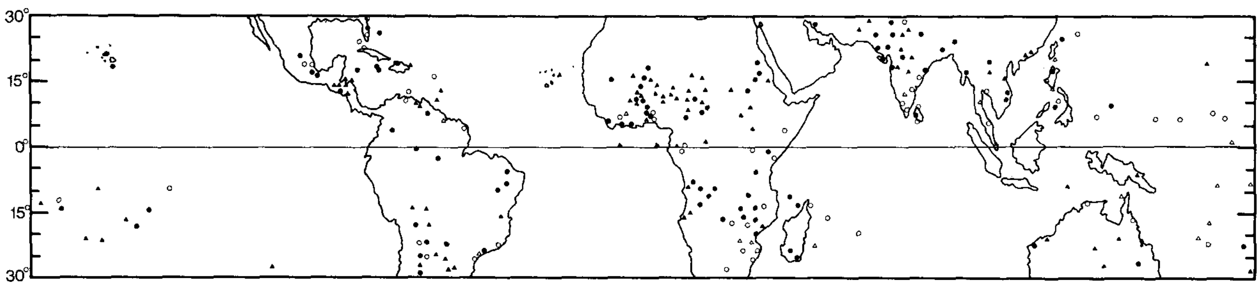

Fig. 3. 1972 precipitation anomalies (June 1971–May 1972) for summer rainfall stations of the Southern Hemisphere. The black symbols indicate local rainfall deficits; the white symbols represent surplus rainfall. Triangles denote deviations in excess of one standard deviation.

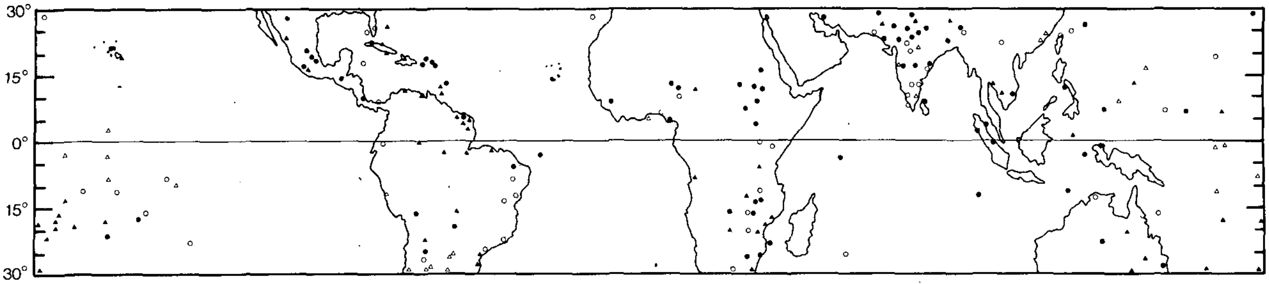

FIG. 4. 1940 precipitation anomalies (June 1940–May 1941) for summer rainfall stations of the Southern Hemisphere. Symbols as in Fig. 3.

resembles to some extent that of Fig. 3. It shows the same widespread rainfall deficits in the subtropics which again did not extend to the equatorial flank of the Indian monsoon or to very low latitudes in the eastern Pacific. In fact, in the Pacific one finds a rather striking wedge of excess rainfall between apparent drought conditions further north and further south.

In 1940 there were not many rainfall stations in the Sahel region of Africa and the evidence for drought conditions there, as presented in Fig. 4, is not exactly striking. River discharge records, however, show that the average five-year discharge of some major rivers reached its lowest level during the 1940–44 interval. The period 1970–74 turns out to be the second driest period.

The cumulative deficit for the 1940–44 period amounted to 66.5×10^9 m^3 for the Niger and 50×10^9 m^3 for the Senegal. Though the five-year deficit was somewhat smaller in the 1970s, lower mean annual discharges were recorded then. In particular, during 1972, the average discharge of the Senegal was only 33% of the normal for the whole record period and in 1974 the river dried up altogether for some time.

The recent discharge deficits, however, are not exceptional extremes. The lowest mean annual discharge ever recorded on the Niger occurred in 1913. The flood stage of the Senegal only reached 17% of normal during 1913. The same year was also characterized by the lowest annual discharge (49% of normal) recorded for the Nile at Aswan during the record period 1871–1945.

The low African river flows of 1913 can be associated with a prolonged worldwide drought which had begun much earlier (at the end of the 19th century) with a major break in subtropical rainfall regimes. This global change was documented in a series of papers in the *Quarterly Journal of the Royal Meteorological Society*. A condensation of these results together with the appropriate references can be found in a short paper in *Nature* (Kraus, 1958). It suggests that the hydrological cycle in the subtropics did not fully recover for another 50 years, from the change which occurred at the turn of the century. As an example, the long-term averages for two periods before and after 1898 of the Nile discharge at Aswan (Hurst *et al.*, 1946) are listed in Table 3. The fractional change between the two periods amounts to about 33%. It occurred with dramatic suddenness (Kraus, 1958) and the social impact of the dry years after 1898 following the preceding wet period was probably as severe as the corresponding change in the 1960s.

As on later occasions, precipitation changes on the fringes of the African monsoon area were paralleled by similar changes in the northwestern fringe area of the Indian monsoon and in other subtropical regions. For example, during the period 1895–1905 the monsoon yielded annually a below average volume of water during ten years in Bikaner and during nine years in Lahore, Hyderabad, New Delhi, Agra, Jaipur and Veraval. Pueblo, Mexico had nine dry years during the same period. Christobal, Panama had eight and Havana, Cuba was exceptionally dry in 1898–1900. The official Australian publication referred to earlier (Director of Meteorology, 1960) states with reference to the period 1895–1906 that ". . . It is difficult for most present day Australians to realise the magnitude of the effects of this drought on the economy of the country. Sheep numbers which had reached 100 million were reduced by half and cattle numbers by over 40 per cent. Average wheat yields exceeded 8 bushels per acre in only one year of the nine." Lake George, east of Canberra, dried out completely in 1897 and remained dry for some 20 years.

TABLE 2. The mean discharge (m^3 s^{-1}) of the Niger and the Senegal during different periods (after Roche *et al.*, 1975).

River	Record period	Normal for whole record	1940–44	1970–74
Niger at Koulikoro	1907–74	1540	1120	1150
Senegal at Bakel	1903–74	770	455	467

TABLE 3. Mean discharge (m^3 s^{-1}) of the Nile at Aswan.

Record Period 1871–1945	1871–98	1898–1945
Mean discharge 2948	3487	2627

To summarize, one can document the existence of at least three quasi-worldwide droughts separated by somewhat wetter periods in the subtropics during the last century. The drought conditions in each case extended with possible interruptions over a time period which was of the order of one decade or longer. There is reasonably strong evidence that these droughts were in phase all across North Africa and northwestern India. There is weaker, but still suggestive evidence that droughts affected the subtropics simultaneously in America and Oceania.

5. Conditions in the Southern Hemisphere

The plots in Figs. 3 and 4 suggest that during major drought years, tropical convective rainfall systems fail to penetrate into those marginal regions which normally get most of their water yield during a relatively short summer precipitation season. For 1972 the reduced northward excursion of the ITCZ was documented in more detail by Tanaka *et al.* (1975). On the Indian subcontinent both Figs. 3 and 4 show the concentration of severe rainfall deficits along the margins of the arid zone. The same happened in America. Riehl (1973) points out that the exceptional dryness of the year 1972 in northern and central Venezuela could be related to an abnormally low latitude of the ITCZ. As a result, subsidence and strong upper westerlies prevailed over regions which normally experience a succession of rain-producing easterly waves.

Summer rainfalls in the Sahel, in the Indian peninsula and to a lesser extent in Central America are all associated with large-scale cross-equatorial monsoon circulations. These circulations provide the main mechanism for the seasonal transport of energy and moisture across the equator. When the energy transport is integrated over the whole depth of the atmosphere and all around the equator, it is found to be always directed toward the winter hemisphere. This is due to the cross-equatorial flux of potential energy which is larger than the atmospheric transport of latent and sensible heat in the opposite direction (Newell *et al.*, 1974). With a seasonal amplitude of $\sim 1.5 \times 10^{15}$ W, the seasonally variable energy flux across the equator makes a significant contribution to the time-dependent part of the separate heat budget of each hemisphere.

Table 4. 500 mb height anomalies during 1972 for all radiosonde stations between 30°S and the equator (D = summer-winter difference).

	Summer	Winter	D
Number of negative anomalies	52.5*	33	42
Number of positive anomalies	2.5	19	9
Averaged anomaly (m)	−13.6	− 4.7	− 9.3

* Zero anomalies were counted as 0.5 positive and 0.5 negative.

Table 5. 700 mb temperature anomalies during 1972 for all stations between 30°S and the equator. (D = summer-winter difference).

	Summer	Winter	D
Number of negative anomalies	54	24	43
Number of positive anomalies	5	30	11
Averaged anomaly (K)	−0.48	+0.11	−0.61

The greater the energy loss in the winter hemisphere, the greater is the demand for an energy transport across the equator and one can expect that this would require a larger area for energy collection in the summer hemisphere. In other words, one might expect a large meridional excursion of the ITCZ whenever the heat loss from the winter hemisphere is particularly large and vice versa. The dependence of the ITCZ excursion upon the seasonally variable forcing in each hemisphere is illustrated with a simple analytical model in an accompanying note in this issue (Kraus, 1977).

As regards actual data, the evidence presented below in Tables 4 and 5 suggests that simultaneously with the 1972 rainfall failure in the north, the seasonal change of energy in the lower subtropical troposphere south of the equator was anomalously small. The tables are based on all records in the NCAR data file which contain comparable data covering at least four years. The winter season includes data for the months of June, July and August. Summer is defined similarly by data for January, February and December of the preceding calendar year. Anomalies are specified by the deviation of the 1972 values from the individual station mean. The most striking feature of the data in Table 4 is the number of negative anomalies. Only two stations out of 55 had above normal 500 mb heights during the summer of 1972; only nine out of 51 had an above average seasonal fluctuation. Some of the stations had either winter or summer records missing. The averaged value of the seasonal change D is based, therefore, on the individual value of this difference for each available station.

The variations in the 500 mb surface are closely related to the temperature of the atmospheric column below, as indicated in Table 5. In a belt of 30° width the actual seasonal temperature difference varies between about 0 K near the equator and 10 K at the southern limits. The averaged value of the anomaly

Table 6. 500 mb height anomalies during 1972 for all antarctic radiosonde stations.

	Summer	Winter
Number of positive anomalies	10	5.5
Number of negative anomalies	1	1.5
Averaged anomaly (m)	+27.9	+34.6

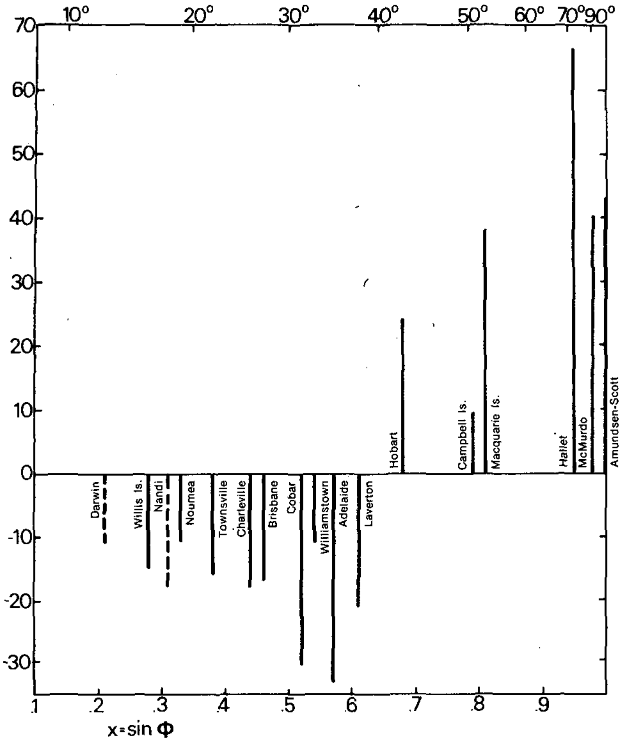

FIG. 5. 1972 anomalies (m) from station record mean of the mean annual 500 mb surface height. Data used for this cross section included all available stations within the longitude sector of 150°E±10° (with Darwin and Nandi also included although they are outside this sector).

difference may not have much value in these circumstances. The relative number of positive and negative anomalies, which is more relevant, is again very suggestive.

Table 4 indicated that during 1972—particularly during summer—the height of the 500 mb surface in the southern subtropics was abnormally low. Table 6 shows that at the same time this surface was abnormally high in the antarctic. This also applied to the 700 mb temperature there.

Comparison of Tables 4 and 6 indicates a considerable reduction in the mean meridional slope of the 500 mb surface. The same inference can be drawn from the north-south cross section in the area of the 150°E meridion which is shown in Fig. 5. This reduction in slope was associated with a weakened temperature gradient and a correspondingly decreased mean meridional baroclinicity during 1972 in the Southern Hemisphere.

The duration of the drought in the late 1960s and early 1970s as well as the preceding run of wet years, was discussed in Section 3. One wonders whether the 1972 anomaly pattern is reflected in the averages for these longer periods. Systematic upper air observations in the antarctic began only in about 1958. The following tabulations therefore concentrate on a comparison of the period 1958–64 and 1968–74; the first being normal or wet along the northern margin of the tropical rainfall belt and the second, excessively dry. All the stations in Antarctica which were used for this comparison are listed in Table 7.

The data in Table 7 are remarkable for their uniformity. Each of the 10 listed stations show a higher 500 mb surface during the second period. With the exception of Islas Orcadas which is north of the polar circle, the difference is larger everywhere than its standard error; for most stations it is more than twice as large.

The significance of this information can also be presented in a somewhat different way which allows for easier comparison with conditions in the subtropics. This is done in Tables 8 and 9.

There were no year-round data available from the Hallett and Byrd stations in Antarctica. In the subtropics, records from Tahiti, Noumea and Tananarive (Madagascar) were omitted from the analysis because of an apparent systematic inhomogeneity. These three stations all use French instruments and they may have been affected by a change of equipment or procedure.

The most significant feature of Tables 8 and 9 is the nonrandom distribution of positive and negative anomalies. The results would have appeared even more striking if the years 1964 and 1974—the last year in each period—had not been included in the analysis.

The amplitude of the average anomaly change in the subtropics may appear small compared to that in Antarctica. This appearance is somewhat misleading because the standard deviation of the subtropical data is also smaller than that of the polar stations. A tabulation corresponding to Table 7 (not included here) shows that out of the 28 radiosonde stations between 0° and 30°S with records going back to 1958, only four had a

TABLE 7. Average 500 mb height (m) at all available stations during the periods 1958–64 and 1968–74. (Where less than seven years of data are available in either period, the actual number *n* of used record years is shown in parentheses.)

Station	1958–64	1968–74	Δ*	SE**
Argentine Island	5106.7	5124.7	18.0	7.5
Islas Orcadas	5170.8 (6)	5173.2 (6)	2.4	6.4
SANAE	5005.5 (6)	5013.8	8.3	4.4
Amundsen-Scott	4924.0	4947.8 (5)	23.8	7.8
Halley Bay	4985.7 (4)	5007.7	22.0	7.4
Novolazarevskaya	4975.5 (2)	4993.7 (3)	18.2	9.2
Mirney	5022.0 (4)	5031.4 (5)	9.4	6.1
Casey	5008.4	5028.6	20.2	8.3
McMurdo	4918.0	4942.9	24.9	6.9
Mawson	4995.8	5111.0	15.2	7.8
Hallet (summer only)	5088.9	5137.8 (6)	48.9	18.4
Byrd (summer only)	5066.9	5134.0 (6)	87.1	21.5

* Δ denotes the difference between (1968–74) and (1958–64).

** SE, the standard error of the difference, is equal to $\sigma/\sqrt{n}$, where σ is the standard deviation of the whole record and n = 7 or the smaller number in parentheses where this is appropriate.

higher average 500 mb height during the dry period 1968–74 than during 1958–64. In only one case (Lae, New Guinea) was the positive difference larger than its standard error. (One referee points out that Lae is known to be subject to special local anomalies.) The remaining 24 station records—in South America, Oceania, Africa and in the Indian Ocean—all showed a lower 500 mb surface during the second period. The difference between the two periods was more than double the standard error in 13 cases.

6. Discussion

There seems to be little doubt that subtropical droughts can affect extensive areas simultaneously. In the fringe belt between the deserts and the monsoon regions, they also tend to be rather persistent. The latest run of drought conditions, which started about 1968, appears to have lasted at least until 1974. Another worldwide subtropical drought at the beginning of this century lasted even longer. Upper air information has been available only for the latest of these dry episodes. They show that during the 1972 drought, summer temperatures at 700 mb in the tropics were below normal and the seasonal temperature change was also abnormally low. This was accompanied by a correspondingly low height and seasonal height variation of the 500 mb surface. By contrast, the lower antarctic troposphere was unusually warm and the level of the 500 mb surface there was abnormally high. The mean meridional slope of the 500 mb surface from the tropics to Antarctica was therefore reduced. This was characteristic not only for 1972, but for the whole of the recent subtropical drought period. The opposite happened during the preceding period of persistent positive rainfall anomalies in the north, which ended in the early 1960's. The mean meridional slope of the 500 mb surface was then considerably steeper than normal.

The combination of anomalies is compatible with a working hypothesis which suggests that the latitude reached by the ITCZ over the continents during the northern summer is a direct function of the cross-equatorial energy flux and of the transport of energy from the tropics to high southern latitudes.

Table 8. Anomalies (m) of the average annual 500 mb height over Antarctica.

	1958–64	1968–74
Total number of station years	70	61
Number of positive anomalies	14	41
Number of negative anomalies	56	20
Averaged anomaly	−9.6	+6.4
Summer height anomalies at Hallet and Byrd		
Number of positive anomalies	1	11
Number of negative anomalies	13	1
Averaged anomaly	−60.9	+28.5

Table 9. Anomalies (m) of the average annual 500 mb height between the equator and 30°S.

	1958–64	1968–74
Total number of station years	166	191
Number of positive anomalies	118	75
Number of negative anomalies	58	116
Average anomaly	+3.8	−1.8

Though the data which support this contention are statistically significant, some caveats may be in order. The concurrence of a worldwide subtropical drought with an apparently worldwide reduced meridional slope of the 500 mb surface slope—while striking—may be accidental. It can be argued that we deal with a single pair of periods. This is certainly not a statistically viable sample and the described coincidences cannot, therefore, be accepted as a proof of the physical hypothesis.

I have made no effort to establish trendlines for the height of the 500 mb surface in the subtropics or in the antarctic. The study concentrated instead on the comparison of two particular time intervals. These were determined by the peculiarities of the African and northeast Indian rainfall regimes which also show a rather sudden flip from wet to dry conditions as indicated by Table 1. The height anomalies during the two selected periods differ by a significant amount and the difference has opposite signs in low and high southern latitudes. Its magnitude, however, is small compared to the overall drop of some 800 m in the height of the 500 mb surface from the subtropics to the antarctic. While the change in slope is significant, it would still have to be established that it was associated with an equally significant change in the meridional heat flux.

In this context it might be noted also that the hypothesis suggested here is not compatible with the views presented by some other authors. For example, van Loon and Williams (1976) associated polar warming with an increased meridional heat flux. If one assumes, as is done here, that the heat flux is low during a period of reduced meridional baroclinicity, then one has to stipulate changes in the surface boundary fluxes or in the radiation balance, in explanation of the cold lower troposphere in the subtropics and the warm antarctic during the period 1968–74. The question is, whether a weak meridional temperature gradient is caused by a high energy level of the heat transporting eddies or whether, in opposition, the eddy energy level is relatively low because of the decreased forcing. The answer could be established observationally, but suitable data were not available. Theoretically, this is a chicken and egg problem which may not have the same answer for climatic fluctuations of different length and amplitude.

These reservations are cogent, but this does not alter the facts represented by the preceding tables and graphs; and the facts while not exactly conclusive, remain suggestive. It makes sense to associate a relatively weak meridional temperature gradient and an abnormally flat slope of the 500 mb surface with decreased baroclinicity and a reduced generation of extratropical perturbation. This leads in turn to a reduced demand for heat (and zonal momentum) export from the tropics and therefore relatively weak direct tropical circulations. As a result, these circulations do not deliver the normal amount of precipitation along their northernmost borders in the monsoonal fringe areas. The facts (as far as they go) seem to be compatible with such a sequence of events.

If the relationship can be confirmed and quantified, it may open a way for seasonal predictions. In the meantime there are a number of problems to be solved. One would like to know what actually caused the abnormal warmth and high mid-tropospheric pressure in the antarctic during 1968–74 as compared to 1958–64. The difference is largest in summer, though there is no significant difference between the seasons.

Another question concerns the simultaneous variation in the Northern Hemisphere. These were not considered during the present investigation. A preliminary analysis suggests that the northern subtropics, like those south of the equator, were somewhat colder during the second period. Systematic latitudinal differences appear to be much more pronounced, however, in the north and this is brought out also by the van Loon and Williams (1976) investigations.

It would be useful to test the stipulated relationship between the extent of the monsoon region and the thermal conditions in the opposite hemisphere by experiments with sufficiently sensitive numerical general circulation models. Such experiments might also provide more insight into the controversial relationship between the meridional temperature gradient and the meridional heat flux.

In the last resort, the suggested hypothesis will be either strengthened or disproved by the accumulation and analysis of appropriate data after the next worldwide, subtropical drought episode. And for that, we may have to wait another third of a century.

Acknowledgments. The material for this study was collected while I was a summer visitor of the NCAR Climate Project. I am deeply grateful to Roy Jenne and Greg Walters for their help with data processing and also to Paul Julian and Harry van Loon for their courteous, unfailingly available advice. Financially, the research was supported by the National Science Foundation's Climate Dynamics Research Section, under Grant NSF ATM 72-01509A03.

REFERENCES

Bunting, A. H., M. D. Dennet, J. Elston and J. R. Milford, 1976: Rainfall trends in the West African Sahel. *Quart. J. Roy. Meteor. Soc.*, **102**, 59–64.

Director of Meteorology, Melbourne, Australia, 1960: *Selected Tables of Australian Rainfall and Related Data*, Vol. 2, Section 3, Major droughts in Australia. Bureau of Meteorology, Commonwealth of Australia, p. 74.

Hurst, H. E., R. P. Black and Y. M. Simaika, 1946: The Nile Basin. Physics Dept. Pap. No. 51, Cairo University, 96–97.

Julian, P. R., 1970: An application of rank-order statistics to the joint spatial and temporal variations of meteorological elements. *Mon. Wea. Rev.*, **98**, 142–153.

Kraus, E. B., 1958: Recent climatic changes. *Nature*, **181**, 666–668.

——, 1977: Subtropical droughts and cross-equatorial energy transports. *Mon. Wea. Rev.*, **105**, 1052–1055.

Mitchell, J. M., Jr., 1976: An overview of climatic variability and its casual mechanisms. *Quat. Res.*, **6**, No. 4, 481–495.

Newell, R. E., J. W. Kidson, D. G. Vincent and G. J. Boer, 1974: *The General Circulation of the Tropical Atmosphere and Interactions with Extratropical Latitudes*, Vol. 2. M.I.T. Press, 371 pp.

Panofsky, H. A., and G. W. Brier, 1958: *Some Applications of Statistics to Meteorology*. The Pennsylvania State University, 224 pp.

Riehl, H., 1973: Controls of the Venezuela rainy season. *Bull. Amer. Meteor. Soc.*, **54**, 9–12.

Roche, M., J. Rodier and J. Sircoulon, 1975: Meteorological and hydrological aspects of continental droughts. Session Paper IUGG, Grenoble, France.

Tanaka, M., B. C. Weare, A. R. Navato and R. E. Newell, 1975: Recent African rainfall patterns. *Nature*, **225**, 201–203.

van Loon, H., and J. Williams, 1976: The connection between trends of mean temperature and circulation at the surface: Part I. Winter. (Part II. Summer) *Mon. Wea. Rev.*, **104**, 365–380 (1003–1011.)