African rainfall and its relation to the upper air circulation

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Summary

The rainfall distribution over Africa analysed for the period 1951–1975 shows a downward trend at 15°N not present at other latitudes. A widespread reduction in rainfall is observed during the drought years 1972 and 1973. Upper air analyses have been used to construct mean monthly vertical motion patterns and water vapour budgets but the results are not sufficiently precise to examine year-to-year differences. Changes in the August flow patterns show that low rainfall in the Sahel is associated with the virtual disappearance of the 850-mb trough near 8°N and weakening of the easterly jet above it. An empirical orthogonal function analysis of the 850-, 500- and 200-mb temperatures from 150 stations shows that these changes are global in extent. The downward trend in the Sahel rainfall has evidently been paralleled by a weakening of the northern hemisphere circulation.

These results suggest that local modification of surface conditions is not the principal cause of the decline in Sahel rainfall over the last two decades.

1. Introduction

The widespread severe droughts in the early 1970s over the Sahel region on the southern fringe of the Sahara have attracted world-wide attention as a result of the extensive stock losses which occurred and the starvation suffered by many thousands of people. While unusually low rainfall was experienced in these years, these consequences seem to have been an inevitable outcome of a number of factors leading to destruction of vegetation. Picardi (1975) and Picardi and Seifert (1976) have presented results from a computer simulation model which show how overgrazing, which became chronic in the early 1960s, triggered a rapid depletion in the rangeland. While most of the factors involved are determined by the social structure of herdsmen who populate the area, the decline in rainfall from the ‘above-normal’ values in the 1950s was undoubtedly of importance. During these years with above-normal rainfall the herds were built up to a point where overgrazing became inevitable as rainfall and plant growth declined. It has been suggested by Charney (1975) that the decline in rainfall was in fact brought on by the loss in vegetative cover, due to overgrazing, which raised the surface albedo. In an analytical model, for which the results were subsequently supported by numerical integration, he shows that an increase in surface albedo will lead to an increase in net radiative heat loss. Increased subsidence, necessary to maintain thermal equilibrium, further suppresses the rainfall so that the arid conditions are reinforced.

This bio-feedback process may have some influence on the cyclic pattern of overgrazing and subsequent regeneration of the rangelands described by Picardi, although it was not included in his model. Rainfall was treated as an exogenous variable and simulations with different rainfall series drawn from the same population gave similar results. Low-rainfall years could hasten the end of the overgrazing cycle but were not its principal cause.

The main question of concern here is whether the present rainfall regime will continue or whether a permanent decline in rainfall will cause much of the Sahel region to be lost to the Sahara. Extreme changes in rainfall over Africa have occurred in the past in association with global climatic changes (Street and Grove 1976) and the possibility that the present decline in rainfall is due to a change in global circulation patterns cannot be ignored.

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So far, several workers have attempted to relate changes in Sahel rainfall to changes in the general circulation but the results are not particularly convincing. Winstanley (1973) has attributed the decline in Sahel rainfall to an expansion of the circumpolar vortex, but while this holds for the subpolar pressure minimum, Miles and Folland (1974) show that the westerly maximum and the subtropical high over the North Atlantic have both moved polewards. Tanaka et al. (1975) found some evidence for a slight change in the morphology of the subtropical high pressure belt and the midlatitude westerlies during 1968–73 but none which supports a global southward expansion of either. Namias (1974) has presented preliminary results linking Sahel rainfall to blocking patterns in the 700-mb flow over the North Atlantic and western Europe.

In the present study the rainfall distribution over Africa has been examined in detail for the period 1951–75. The mean vertical motion and water vapour flux patterns are derived

Figure 1. Mean rainfall in mm for the period 1951–73.
from monthly wind component and specific humidity analyses, and year-to-year changes in rainfall have been related to those occurring in the upper wind flow. Evidence has also been found for global-scale changes in tropospheric temperature fields which parallel those in Sahel rainfall. Although these changes suggest a trend to a weaker zonal circulation, their physical causes are not understood and no rational basis has been found for predicting whether the trend to drier conditions in the Sahel will continue.

2. Rainfall Analysis

Rainfall variation in the Sahel beginning from the early part of this century has been examined by Jenkinson (1973), Bunting et al. (1976), and others. Because precipitation in this area is primarily convective in nature, statistics derived from only a few stations may not be representative of the region as a whole but in general the following features are observed.

The annual-mean rainfall fluctuates appreciably from year to year and although periods of 20–30 years appear to be present, Bunting et al. find none that are significant at the 95% level. The main feature of recent years is a fairly steady decline in annual rainfall from the peak values of the early 1950s to those in 1972 and 1973 which are among the lowest recorded this century. Jenkinson shows, furthermore, that the rainfall deficits for periods of up to 10 years ending in 1972 are comparable to those of the earlier drought years centred on 1913 and 1941.

Tanaka et al. have used as large a data set as possible to examine year-to-year changes in rainfall over the entire African continent for the period 1951–73. August alone was selected for this pilot study as it is the month in which peak rainfall occurs in the Sahel. The present study includes an extension of this earlier work to cover all months of the year and additional data for 1974 and 1975 that have since become available.

The main source of data has been the Monthly climatic data for the world (available

![Figure 2. Mean rainfall integrated over the African continent for 1951–73 and the latitude of its 'centre of gravity'.](attachment:image.png)
in the NCAR world surface climatology tapes), and the station distribution in the Sahel region is essentially that shown in Fig. 1 of Tanaka et al. These monthly-mean data have been supplemented with station data obtained directly from most of the countries which were shown in their Fig. 2 as having no data during 1961–70. The analyses were performed on a $13 \times 15$ $5^\circ$ grid on a Mercator projection, with boundaries at $33^\circ$N, $33^\circ$S, $15^\circ$W and $45^\circ$E, covering most of the African continent. The analysis procedure followed Cressman (1959) with a first guess of zero and scan radii of $10^\circ$ and $5^\circ$ so that the values are representative of a $5^\circ$-grid square.

The long-term monthly means for January, April, July and October, which show the annual march of the equatorial rainbelt, are presented in Fig. 1. They are similar to those of Thompson (1965) but with less small-scale detail. The annual movement of the rainbelt is shown in Fig. 2, together with the integrated rainfall. Its mean position is at $0.5^\circ$S and the extreme displacements are to $8.7^\circ$N in August and $9.7^\circ$S in January, so that stations lying within this range will normally experience a double rainfall maximum. The integrated rainfall also reaches its peak in August and this may be attributed to the larger land area over which rain falls rather than a significant change in rainfall intensity. No evidence has been found for a southward trend in the rainbelt position in recent years.

The interannual variation of the yearly totals in the equatorial rainbelt can be seen in Fig. 3 where the grid-point values have been accumulated over the land area west of $27.5^\circ$E. A downward trend prior to 1970 is seen only at $15^\circ$N but the minimum at $15^\circ$N in the early 1970s also appears at most other latitudes. Integration over the entire equatorial rainbelt shows that the 1972–73 rainfall deficit was approximately 14% of the 1951–73 mean, corresponding to a departure of 2.5 standard deviations. The downward trend in the rainfall at $15^\circ$N is approximately $7.7$ mm yr$^{-1}$ and is equivalent to a southward displacement of

![Figure 3](image_url) **Figure 3.** Mean annual rainfall (in mm) for the African continent west of $27.5^\circ$E for 1951–75.
the desert margin of around 5 km yr\(^{-1}\). An examination of the time series for the individual grid points at 15°N shows this trend to be present at all longitudes from 5°W to 30°E. The downward trend is not present at 10°W but lower-than-normal values are experienced in the early 1970s.

The rainfall analysis therefore shows a declining trend at 15°N throughout the period of this study which is not evident at other latitudes. The extreme departures in 1972 and 1973 are, however, seen to be part of a widespread reduction in African rainfall. They do not appear to be outside the range of previously observed rainfall minima from which subsequent recoveries to a normal rainfall have been made.

3. Mean upper air circulation patterns

In order to examine more fully the reasons for the rainfall variations described in the previous section, analyses were made of the monthly wind component and specific humidity fields at standard levels from the surface to 100 mb. The analyses had to be limited to the two periods May 1958 to April 1963 and December 1969 to December 1973 for which data were readily available. For the first period, during which near-normal rainfall was observed, wind component data were taken from the tropical study of Newell et al. (1972) but the specific humidity data were incomplete. For the later period, covering the drought years, all parameters except the surface wind were available from the Monthly climatic data for the world upper air tape.

The analyses were made on the same grid as for rainfall, and the station distribution for both periods is shown in Fig. 4. The analysis scheme was similar to that for rainfall but differed in two respects: the first guess used was the long-term mean for the particular

Figure 4. Distribution of stations used in the upper air analyses. Open circles show those present during 1958-63, crosses those present during 1970-73 and solid circles both periods.
month; and an anisotropic scan radius was found to be necessary for the zonal wind component, $u$, above the surface level. The modifications for the $u$-component analyses were needed to preserve adequate continuity in regions where the station coverage was poor. Distances along the west-east axis were scaled by a factor of 3 so that the influence radii in this direction were effectively 30° and 15° longitude for the two scans.

The mean flow patterns are similar to the three-month means in Newell et al. (1972) and will not be presented here. Attention will be concentrated instead on the diagnostic fields derived from them. The vertical motion field has been computed using the equation of continuity evaluated over 10° squares centred on each of the interior grid points. The boundary conditions imposed were no net divergence in the column between the surface and 100 mb, and the vertical velocity at the surface set equal to the orographic component. The 600-mb patterns, which were taken as representative of the mid troposphere, are shown

Figure 5. Vertical motion at 600 mb in units of $10^{-5}$ mb s$^{-1}$, based on mean analyses for 9 years. Negative values indicate upward motion.
in Fig. 5 for the months of January, April, July and October. In general, they correspond quite well to the rainfall analyses in Fig 1, showing a seasonal movement which parallels that of the equatorial rainfall belt. Difficulties were experienced in the wind component analyses outside the border of the continent and over Sudan and Egypt, accounting for the main areas of discrepancy. The correlation between individual monthly vertical motion patterns and rainfall is shown in Fig. 6 and magnitudes exceeding the 95% confidence limit of 0.2 are observed over most of the area affected by the seasonally migrating rainbelt. With the annual cycle removed, the correlations are barely significant but still in the same sense with above-normal rainfall occurring in areas of ascent. It appears that the vertical motion field is not sufficiently well determined on a monthly basis to relate rainfall departures to departures from the seasonal vertical motion.

The transport of water vapour over the African continent has been shown by Peixoto and Obasi (1965) and Flohn et al. (1965) to be due primarily to the mean wind, with eddy fluxes an order of magnitude smaller. This enables water vapour budgets to be constructed from the monthly-mean wind and humidity analyses obtained in this study even though no eddy transports are available. The fluxes at individual levels were computed using the mean wind components modified in such a way as to produce no net mass divergence in the column. They were then integrated vertically to give the net water vapour transport vector, which was separated into its divergent and nondivergent parts. Details of the calculations are given in the appendix.

The streamlines and divergence patterns obtained from the vertical integrals of the long-term means for January, April, July and October are shown in Fig. 7. In general, the
water vapour flux divergence patterns resemble those of the vertical motion while the non-divergent part is quite similar to the streamlines of the 850-mb flow and shows that the net transport is predominantly from the east. The maximum divergence values observed are around $-5 \times 10^{-6} \text{ ms}^{-1}$ which corresponds to a rainfall rate of 4.4 mm per day. If an evapotranspiration rate of 80 mm per month is assumed to be typical of the equatorial rain forest (Bultot and Griffiths 1972, p. 272) the overall rainfall rate of around 210 mm per month would be comparable to the values in Fig. 1 for areas without substantial orographic enhancement.

It is unfortunate that the incompleteness of the humidity data has prevented a comparison of the water vapour budgets for wet and dry years.
4. Year-to-year changes over Africa

The examination of year-to-year changes in the circulation patterns has been concentrated on the month of August, when the rainfall in the Sahel is at its peak. Mean conditions for this month derived from the analyses of the previous section are shown in Fig. 8. The mean 850-mb trough lies near 8°N as does the axis of the 200-mb easterly jet. This latitude also serves as the dividing line between the two streams of water vapour flux: from the Mediterranean and Red Sea in the north and from the Indian Ocean south of the equator. The upward motion is mainly confined to the area between the equator and 15°N and the water vapour flux convergence pattern, which is not shown in the figure, is similar to that of the vertical motion. The westerly flow to the south of the 850-mb trough does not carry sufficient water vapour to give a net flux from the west, although it does provide an important low-level moisture source. The ITCZ is, therefore, located near 8°N in the mid troposphere during August, although its surface location is considered to be the Sahara heat low at 18–20°N (Flohne 1965).

A comparison of the conditions prevailing during wet and dry months shows that the 850-mb trough and the 200-mb jet are both more pronounced during Augusts with higher rainfall at 15°N. The patterns for the two wettest years (1959 and 1961) and the two driest years (1972 and 1973) are shown in Fig. 9 where the differences can be clearly seen. The 850-mb trough is well developed during the wet years but has completely disappeared in 1972 and 1973. The 200-mb easterlies are much weaker during the dry years as shown also in Fig. 10 where the mean wind speed from 0° to 15°N at 10°E is plotted together with the mean rainfall along 15°N. The correlation between these two parameters is -0.79, which is significant at the 95% level.

From Fig. 11 it can be seen that the 200-mb easterly jet core lies in the region of northerly flow to the south of 15°N. Momentum budget calculations using the wind component and vertical motion analyses obtained here, together with eddy flux data from Newell et al. (1972), indicate that the Coriolis torque acting on this flow is primarily responsible for the maintenance of the jet.

Figure 8. Mean conditions over Africa during August. A streamline spacing of 5° corresponds to a wind speed of 2.5 m s⁻¹ at 850 mb or 20 m s⁻¹ at 200 mb or to a water vapour flux of 100 kg m⁻² s⁻¹. The vertical motion units are again 10⁻⁵ mb s⁻¹.
The year-to-year changes described in this section are, therefore, consistent with a stronger meridional circulation in wet years with increased upward motion in the low-level trough and stronger southward flow in the upper troposphere.

5. **Global circulation changes**

Changes in the global circulation patterns which might be related to those in the Sahel rainfall were examined with the aid of an upper-level temperature data set containing monthly-mean standard-level temperatures from almost 200 stations for the period 1958–74. It was hoped that the temperature data would be more accurate than wind data and would thus be more likely to provide evidence of small systematic changes from year to year. The
Figure 10. Mean 200-mb wind speed in m s$^{-1}$ from 0° to 15°N at 10°E and mean rainfall in mm along 15°N from 5°W to 30°E.

Figure 11. Cross-section of the mean wind for August averaged at 5, 10 and 15°E. The isopleths of the zonal ($u$) component are shown as dashed lines and those of the meridional ($v$) component as continuous lines.

levels investigated here were at 850, 500, and 200 mb to represent conditions throughout the troposphere.

Firstly, correlations between monthly rainfall departures from the mean at 15°N and temperatures at each level were obtained for the rainy season months July–September. The results showed evidence of large-scale patterns but were difficult to analyse because of lack of continuity between adjacent values. In order to eliminate much of this small-scale variation, an analysis of the temperature fields into empirical orthogonal functions (EOFs) was carried out. The number of stations selected for EOF analysis was set at 150 for each level with the selection procedure taking the most uniform geographical coverage possible from stations with at least 80% of the data present. The annual cycle was removed and the departures normalized for each station in order to avoid undue weighting of the EOFs by the high-latitude stations. At each level, 10 EOFs were selected for further study and accounted for 42-3, 43-7, and 48-2% of the normalized variance at the 850-, 500- and 200-mb levels respectively. The correlations of the 51 normalized rainfall departures for July through September at 15°N with these 10 EOFs proved to be significant in a number of cases. A linear stepwise

| TABLE I. RAINFALL 'PREDICTIONS' SELECTED BY STEPWISE REGRESSION |
|-----------------------------|-------------------|-------|--------|-------|
| Level | Predictors and weights | mult. corr. | std. err. | Var. red. |
| 850 | -1·77 × 850<sub>1</sub> + 1·47 × 850<sub>2</sub> + 1·89 × 850<sub>0</sub> | .71 | 0·73 | 51% |
| 500 | 1·70 × 500<sub>1</sub> - 2·53 × 500<sub>2</sub> | .64 | 0.79 | 41% |
| 200 | 1·62 × 200<sub>1</sub> - 1·35 × 200<sub>2</sub> | .54 | 0·87 | 30% |
| All | -2·12 × 850<sub>1</sub> + 1·87 × 850<sub>2</sub> - 2·27 × 500<sub>3</sub> | .74 | 0·71 | 54% |
| All (12-month lead) | -1·33 × 850<sub>1</sub> - 1·95 × 200<sub>0</sub> | .66 | 0·77 | 43% |

(The population standard deviation of the rainfall series is 1·01)
regression was used to select those combinations of EOFs at each level which best explained the rainfall variation. The results are shown in Table 1. The relations are diagnostic in character except for the last where the EOFs were chosen as 'predictors' of the rainfall departure 12 months later.

The 99% confidence limit for a correlation coefficient of -41 is easily exceeded in each case. The strength of these relations is due in part to the downward trend common to each, which accounts for approximately 32% of the variance of the rainfall series and between 46 and 73% of the predictor series. The time series of the normalized rainfall and the above predictions are shown in Fig. 12 together with their mean values for each 3-month wet season. The resemblance between the series is seen to be much closer when the seasonal averages are considered, and correlations between them in the range -77 to -87 are obtained. These results suggest that the rainfall and temperature departures are more closely related on a seasonal basis, but in view of the limited length of the series, no regression equations were obtained. Predictions 12 months ahead show some skill for 1959-74 but may not hold up on independent data as in 1975 for which a low value was forecast.

The temperature patterns at each level which are most closely related to the rainfall departures are obtained from combining the EOF patterns with the weights given in Table 1. The resulting normalized departure patterns are shown in Fig. 13 for the three levels.

The 200-mb pattern is perhaps the easiest to interpret as it is predominantly zonal in character. While the southern hemisphere pattern is poorly defined over the oceans, it is instructive to compare that for the northern hemisphere with Sadler's (1975) mean 200-mb streamline analyses for August. In general the temperature departure changes sign near the northern boundary of the tropical easterlies. The main exception occurs over the sector from the central Pacific eastwards to the Atlantic where the temperature gradient weakens but does not change sign until some distance polewards of the reversal in wind direction. The sense of the variations in the temperature gradient and the resulting wind shear is such that, in wet years, both the tropical easterlies and the northern hemisphere westerlies are strengthened, with the main effect on the westerlies occurring polewards of 50°N.

At 500 mb the interpretation is less clear, particularly in the southern hemisphere and
Figure 13. (a), (b), (c): Normalized temperature departure patterns at 850, 500 and 200 mb best correlated with those for the rainfall at 15°N. (d): The station distribution at 850 mb. The distribution at the other levels is similar.
at higher latitudes in the northern hemisphere. Elsewhere, however, a resemblance can be seen to the June–August rainfall pattern of Möller (1951). Higher temperature and lower rainfall occur over the eastern Pacific, extending on to North America and over the eastern Atlantic. Lower temperature and higher rainfall coincide over the eastern USA, western Africa and Asia. The main departure from this relation is found over the equatorial eastern Pacific where the temperature pattern is not well defined. A tentative interpretation of this relationship may be given through consideration of the terms in the atmospheric heat budget (Newell et al. 1974, chapter 7). The higher temperature in the dry areas may be due to increased subsidence heating while the lower temperature in the high-rainfall areas may result from increased upward motion and cloud cover and a reduction of the incoming solar energy. The stronger meridional circulation would be necessary to maintain the more intense zonal circulation which appears to be present in the upper troposphere.

The 850-mb pattern is the one most closely correlated with rainfall but a physical interpretation cannot be given at the present time. We note, however, that higher temperature over much of Africa is associated with lower rainfall in the Sahel. A similar result was obtained for surface temperature by Tanaka et al. (1975) and it appears that a reduction in cloudiness leading to increased solar heating is responsible.

6. Conclusions

The rainfall analysis in section 2 provides evidence of a downward trend in rainfall at 15°N since the early 1950s which is not present at 10°N. The unusually dry conditions during 1972–3 were, however, experienced at all latitudes, and it seems unlikely that the widespread reduction in rainfall could have been brought about by a comparatively small change in surface conditions in the Sahel. Indeed, evidence of large-scale perturbations in the circulation over Africa which are related to changes in the Sahel rainfall has been provided in section 4. Low rainfall is found to be associated with the virtual disappearance of the 850-mb trough and a weaker easterly jet at 200 mb, implying a reduction in the strength of the meridional circulation. These changes appear to be part of the global trend towards a weaker circulation in the middle and upper troposphere revealed by the temperature analyses in section 5.

It is difficult at the present time to attribute this trend in the strength of the circulation to a particular physical cause. While air–sea interaction may play an important role in other year-to-year changes such as the Southern Oscillation (Bjerknes 1969) no significant correlations have been found here with the Atlantic or Pacific sea surface temperature EOFs obtained by Weare et al. (1976) and Weare (1977). No indication can therefore be given of future rainfall variations other than extrapolation of the present trend which shows some skill on the dependent data.

It is too early to tell whether a permanent change in the climate of the Sahel is taking place. Examination of station rainfall records from the early part of this century shows that the 1972–3 rainfall was not below previous levels from which a subsequent recovery to 'normal' took place. In view of the links to the larger-scale circulation found here, it seems more likely that any further change would result from natural or anthropogenic perturbations of the global climate rather than from man's activities in the Sahel.

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APPENDIX: WATER VAPOUR FLUX CALCULATIONS

(i) Adjustment of the wind field

The first step in the computation is the removal of spurious divergence from the wind field in a similar manner to the correction of the vertical motion field. The net divergence of the column is calculated from the sum

$$D = \sum_{i=0}^{N} \nabla \cdot V_i \delta p_i$$  \hspace{1cm} (1)

where $V_i$ is the velocity vector at level $i$, and $\delta p_i = \frac{1}{2}(p_{i-1} - p_{i+1})$, is the weight given to this level. (Note that $\delta p_0 = \frac{1}{2}(p_0 - p_1)$, and that $\delta p_N = \frac{1}{2}(p_{N-1} - p_N)$.) The net divergence of the column should ideally be close to zero and to achieve this the wind component fields at each level are adjusted as follows.

Let $D_i = D \times \delta p_i((p_0 - p_N)$ be the divergence to be removed from level $i$, and assume the wind field giving rise to this divergence field to be represented by $V_i' = \nabla \Phi_i$. We can then solve for $\Phi_i$ from the Poisson equation $\nabla \cdot V_i' \equiv D_i = \nabla^2 \Phi_i$.

The corrected wind fields are now obtained by subtracting $V_i$ from the original $V_i$ at each level and these then correspond to the adjusted $\omega$ field described in section 3.

(ii) Water vapour flux

The vertically integrated water vapour flux is given by

$$Q = g^{-1} \sum_{i=0}^{N} q_i V_i' \delta p_i$$

where $q_i$ is the specific humidity at level $i$, and $V_i'$ the adjusted wind vector. $\delta p_i$ is defined as in Eq. (1). Conventionally this flux is represented by its zonal and meridional components, $Q_\lambda$ and $Q_\phi$, but an alternative representation is in terms of its irrotational and nondivergent parts. Let $Q = \nabla \Phi + \mathbf{k} \times \nabla \psi$. Then taking divergence and curl leads to $\nabla \cdot Q = \nabla^2 \Phi$ and $\mathbf{k} \cdot \nabla \times Q = -\nabla^2 \psi$.

The divergent part is best represented by the field of $\nabla \cdot Q$ but the nondivergent flux which is the major part of $Q$ is obtained through solution of the Poisson equation for the stream function $\psi$.

Further discussion of water vapour budget calculations may be obtained from Rasmusson (1972) or Starr and Peixoto (1964).