Desert climate and its dynamics

D. R. Sikka

40 Mausam Vihar, Delhi 110 051, India

Basic characteristics of desert climate are discussed in relation to aridity indices based on water balance and energy balance consideration. The dynamical aspects of the underlying causes in terms of tropical Hadley cell, orographically forced subsidence and large scale ocean–atmosphere interactions are elaborated. Climatological aspects of the Thar desert (Rajasthan) in India and a review of the present and past climates of the region are presented. Interannual fluctuations of arid conditions are discussed in terms of several parameters. Local and regional dynamical aspects of the desert climate as revealed by recent field experiments are discussed. Anthropogenic causes of desertification are discussed in terms of the biogeophysical feedback processes such as albedo, soil moisture and roughness length. Recent work of several workers, employing general circulation modelling experiments to study the sensitivity of land surface processes to desert climate are briefly reviewed.

DESERTS or arid zones are one of the important components of the Earth's environment. They are fragile ecosystem where vegetation is absent or scarce as moisture is insufficient to support it. They are progeny of global climate system under the influence of large scale land–ocean–atmosphere–cryosphere interactions.

Two systems of climatic classification are usually followed after Koppen and Thornthwaite. In the Koppen's[1] system desert or arid climate is characterized by regions where mean annual precipitation is low (< 250 mm) and surface air temperature is high. The Thornthwaite's[2] system is based on water balance or hydrological approach and compares the mean annual rainfall against mean annual potential evapotranspiration (PET). Deserts (arid and semi-arid regions) occupy nearly 30% of the land area of the Earth and nearly one sixth of world's human population live in these environments. Obviously extensive scientific work on various aspects of the desert environment has been done. Our objective is to consolidate briefly information on desert climate, mostly restricted to near-surface climate.

The following indices are usually used to define aridity or moisture availability:

$$\text{Thornthwaite's Index } I = \frac{\text{Mean annual precipitation } (R)}{\text{Mean annual potential evapotranspiration}}$$

Different classes of arid zones are described as – (i) Hyperarid: $I < 0.05$ (almost absence of vegetation); (ii) Arid: $0.05 > I < 0.20$ (dry zone where perennial agriculture is not possible); (iii) Semi-arid: $0.20 < I < 0.50$ (shrub lands where rainfed agriculture is practiced).

Moisture index (I_m) is usually defined as:

$$I_m = \frac{100\,S - 60\,d}{\text{PET}},$$

where S is water surplus and d is water deficiency. A region is considered having desert climate if I_m is below – 60 and this limit on I_m generally agrees with Koppen's classification BW for the deserts.

An entirely different method of quantification of aridity is after Budyko[3] and follows the energy balance approach in which net radiation (Q^*) received at the surface of the earth on annual basis is compared against the mean annual precipitation (R). (Q^*) is given by the well-known energy balance equation

$$Q^* = (Q + q)(1 - x) - (\text{LWu} - \text{LWd}) = H + LE + G,$$

in which Q and q are the direct and diffuse solar radiation received at earths surface, x is the albedo of the surface, LWu and LWd are the upward and downward long wave emissions by the earth–atmosphere system respectively. The net radiation is used in three major ways, viz. (i) in transporting sensible heat (H) upward by turbulent eddies, (ii) in removing water/moisture from surface by evaporation and transpiration (LE) and, (iii) in conducting heat into the soil (G). The small portion of energy used in photosynthesis is neglected for the energy balance purpose. Lettau[4] defines a dryness ratio by comparing Q against heat required to evaporate rainwater

$$\text{Dryness ratio} = \frac{Q^*}{LR},$$

where L is latent heat of vaporization.

For hyperarid regions dryness ratio is > 10, for arid regions it is 7 to 10 and for semi-arid regions it is 2 to 6.

Global distribution and mechanisms of deserts

In Koppen's classification the desert climate is designated as BW and these regions are surrounded by the climate type BS. The marine deserts, where rainfall is also scanty, are designated as BM. The world's major deserts in different continents are: (i) Sahara, Saudi Arabia, East Africa, Namib, Kalahari in African continent, (ii) the Thar, Taklamakan and Gobi in the Asian continent, (iii) the Great Sandy and Simpson deserts in Australia, (iv) Mojavet, Sonaran and Chiliuahan in North America, and (v) Peruvian, Atacama, Patagontan in South America. Oceanic deserts exist in the equational central and eastern Pacific where the SST is < 26°C and under the sub-tropical high pressure belts in north and south Atlantic and Pacific Oceans, in the South Indian Ocean and over Western and Central Arabian Sea. Coastal marine deserts exist off the major ocean upwelling regions or where cold ocean current prevail. Any geographical atlas depicts the distribution of major deserts over land. From this distribution it becomes apparent that most of the deserts over the Earth are located within the descending limb of the direct Hadley circulation whose ascending limb on annual average basis lies in the near equational region. The ascending and descending limbs shift pole-ward (equatorward) during summer (winter). Thus in the sub-tropical belts some rains occur in the desert regions during summer due to moisture incursion with the poleward excursion of the Intertropical Convergence Zone (ITCZ) or Monsoon Trough. A few rain spells may also occur in winter under the equatorward extension of baroclinic disturbances. The subsidence motion under the sub-tropical high pressure belts does not favour convective motion or cloud formation. The sub-tropical deserts are further desiccated if a region lies under the influence of cold currents or off oceanic upwelling regions such as the Namib desert in South Africa, the Atacama and Peruvian deserts in South America and the East African desert of Somalia. The inner continental deserts within or close to the sub-tropical belts are also sustained by the downward motion on the leeward side of major mountain barriers such as the Taklamakan desert in Asia and Patagontan desert in South America. The aridity of the Interior Karnataka, parts of Maharashtra and Andhra Pradesh States of India is also ascribed to the rain shadow effect of the Western Ghats.

Characteristics of desert climate

Low rainfall and high temperature are the most conspicuous characteristics of desert climate. The mean annual aridity indices are different for different land deserts of the world and the aridity is highest for the core Saharan and Peruvian deserts, followed by the Saudi Arabian, East African and Gobi, Australian and South African deserts. It is low for India's Thar desert and the North American deserts of California and Mexico. It is observed (Berkofsky[5]) that there are regions in the Saharan, the Australian and the Namibian deserts where the annual rainfall is almost negligible < 50 mm. Compared to that most of the Thar desert in India receives considerable rainfall (100–300 mm).

Present and past climates of Thar desert of India

The Thar desert of Sind Province of Pakistan and Western Rajasthan region of India is a meteorologically homogeneous region where physiographic and anthropogenic conditions are somewhat comparable to the contiguous Saharan region to its West. The Aravallis range in Rajasthan defines the eastern limit of the Thar. Most spectacular amongst the land forms in West Rajasthan are the sand dunes which contribute about 40% of area of the region. The dunes are not contiguous and dune-free corridors exist. Some of the dunes are stabilized whereas others are less stable and wind erosion is a common feature from dunes to dune-free areas. The Thar region is under the influence of the Indian South-west monsoon (or summer monsoon) during July and August and occasionally rain-producing weather disturbances like 'depressions' and 'lows' penetrate up to parts of Rajasthan triggering rain events in the Thar. The precious rain events in Thar desert support a large population with above average growth rate than the average growth rate of India.

Climatology over the Western Rajasthan

The climate of the Western Rajasthan is controlled during the summer monsoon (June–September) by the seasonal heat low over Sind–Western Rajasthan region and during winter (December–March) by the Siberian anticyclone at the surface level. Thus the wind flow near the surface is from South West/West during the summer monsoon and from a general NW direction during the winter. At the upper tropospheric level wind flows from the east during the summer monsoon and from the west in the winter season. In the transition months of October and November in autumn, the pressure gradient is very slack. In the spring/early summer during April and May intense heating during the day, coupled with existence of pressure gradient in the formative stage of the summer time heat low, supports strong dust-raising winds. The almost clear weather pattern during late January to June is interrupted by the passage of a western disturbance which occasionally affects the weather of Western Rajasthan such that a few mild rain events occur during January–March and dust storms from April to June.

Rainfall

The main rainy season is during July and August when almost 80% of the annual rain occurs. Figure 1 shows the annual distribution of rains for West Rajasthan along with the coefficient of variation. The monsoon seasonal rainfall follows the annual rainfall and decreases from about 500 mm over East Rajasthan to about 100 mm over extreme western portion of West Rajasthan. The rainy days also decrease from East to West Rajasthan. The coefficient of variation is also higher over West Rajasthan. The rainfall is mostly convective in nature such that an intense weather system even in one day produces 50% or higher of the annual rainfall in one day. The highest one-day rainstorm of Rajasthan has occurred over Banswara in East Rajasthan (559 mm). Figure 2 shows the one-day probable mean precipitation (PMP) for Rajasthan. Dhar *et al.*[6] have studied the depth–area–duration of some of the severe most rainstorms that have occurred over Rajasthan for 1- to 5-day and have reported that for 100 km^2 area the 1-day rain depth has been approached in the rainstorm on 19 July 1981. The same rainstorm produced a rainfall of 265 mm at a station in Nagaur district in West Rajasthan. The mean annual rainfall of West/East Rajasthan is given in Table 1.

Length of the monsoon rainy season

The monsoon rains begin in West Rajasthan by 10–15 July and end by 7 September. In the mean, the season lasts for about 50–60 days. Inter-annual variations in the duration of the season are marked due to early arrival and late cessation of the rains and in extreme cases the season can last between 40 and 70 days. This large fluctuation has marked effect on the quantum of rainfall in this desert.

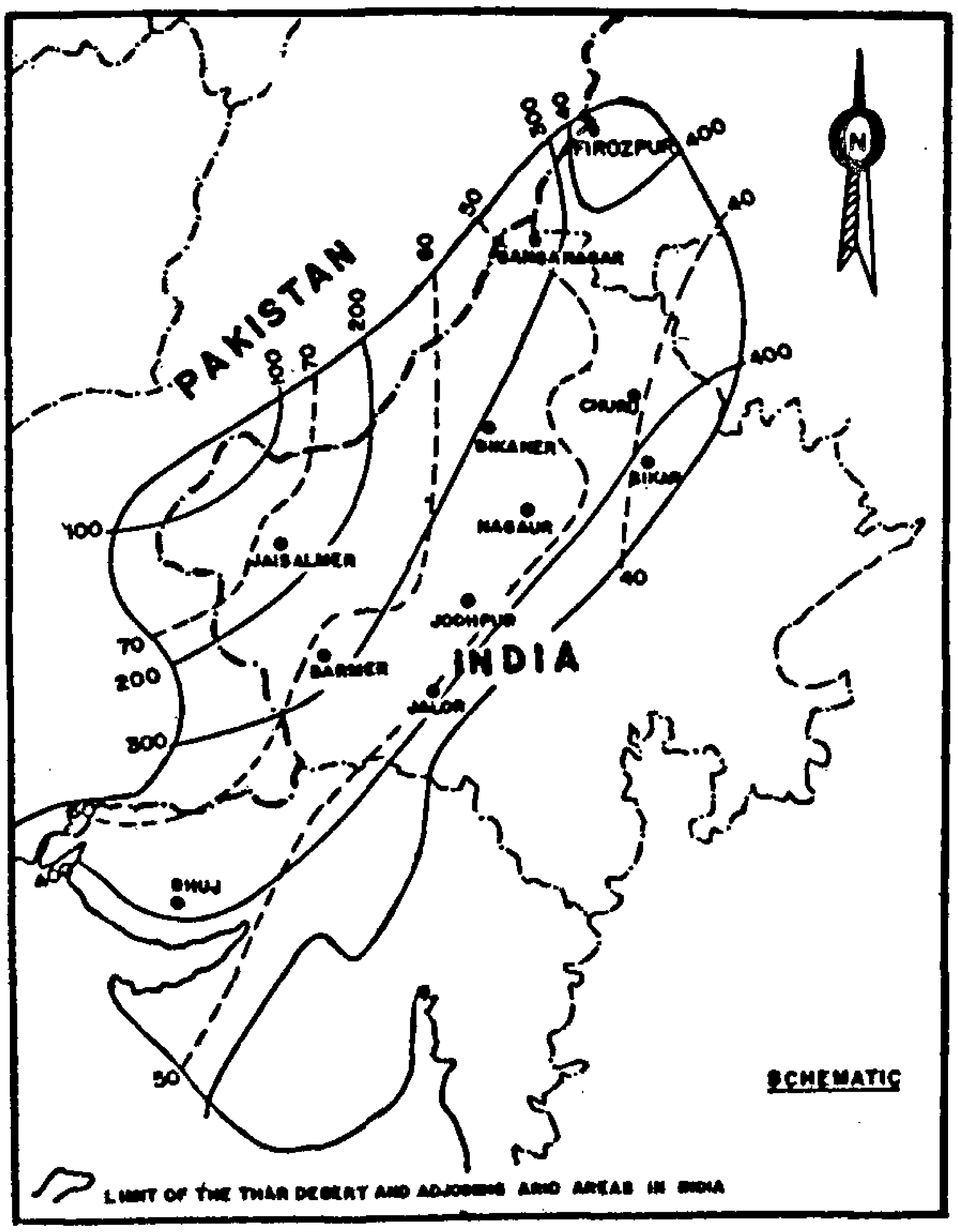

Figure 1. Mean annual rainfall (in mm, full lines) and its coefficient of variation (in %, dotted line) over West Rajasthan.

Fluctuation in aridity index

Nityanand Singh *et al.*[7] have examined the year-to-year fluctuations of the area under less than 560 mm of annual rains, which they call as arid area, over India, North India and Peninsular India. For North India (north of 20°N), the arid area, as defined above, covers 43×10^4 km^2 with a standard deviation of about 14.8×10^4 km^2 (coefficient of variation, 34.3%). This area shows expansion and contraction on inter-annual basis (Figure 3) and in an extremely dry year like 1918 or 1987 major part of north India experienced arid conditions whereas the area contracted to cover only parts of West Rajasthan in extremely good monsoon year like 1917 or 1961. In recent years there is a declining tendency in the arid area with a westward shift in heavy rainfall events. Figure 4 after the work of Ramakrishna[8], shows that Thornthwaite's mean moisture index line of – 80% has shown east-west fluctuations around its long term normal position – the protrusion to west being somewhat large in the decade of 1971–80 (wet decade). This and the rainfall analysis of Rajasthan suggest that in recent decades the Thar desert appears to have shrunk instead of expanding to the east. Mean annual aridity index calculated by Parthasarathy and Rakhecha[10] is given in Table 2 along with annual rainfall and annual maximum and minimum temperatures.

Droughts in Rajasthan

Table 3 shows the individual drought years (annual rainfall departure from long term normal < 25%) over the meteorological sub-divisions of West and East Rajasthan as well as the drought years which are common to both the sub-divisions.

It is clear from Table 3 that West Rajasthan is more prone to drought and several of the drought years are not common between the two sub-divisions even though they are contiguous zones. Besides, the common drought years are mostly those in which the country as a whole had suffered from bad monsoon (rainfall < one standard deviation for the country as a whole). These drought years are those in which India is affected by the adverse effects of planetary scale ocean–atmosphere coupled system ENSO phenomena. Sikka[11] was among the first to suggest ENSO–monsoon relationship and since then many researchers have further investigated this relationship. Rakhecha[12] has listed the years of excess rainfall (> 50% normal) and deficit rainfall (< 50% normal) years for the period 1871–1990.

Stability and moisture content

The Thar desert of India witnesses extreme surface air temperature during the year. The maximum temperature in May–June touches 50°C and the minimum temperature in January–February reaches –3°C or even less in extreme cold wave conditions. The lowest temperature recorded at Jaisalmer is – 5.9°C with the mean monthly extreme lowest temperature of – 2°C in January. Whereas the diurnal temperature range in summer and winter is normally about 20°C, it reaches 25°C under extreme conditions.

The soil temperature at surface in summer is even 10°C warmer than the maximum surface air temperature near noon and it is about 3°C colder than the minimum air temperature in winter. This indicates that super adiabatic lapse rate (highly unstable lapse rate) prevails during the day time in summer and a strong inversion (highly stable lapse rate) exists in winter near the surface from dusk to early morning. Ramakrishna *et al.*[13] conducted a special experiment from May 1987 to April 1988 at stable and unstable sand dunes in the area to measure soil temperature and surface air temperature. The data showed that the maximum soil temperature of 55°C or more was recorded in May 1987 on all faces and again temperature even over 60°C, were recorded in September 1987. During November there was a considerable fall in the maximum temperature for the soil and air. Thus, maximum difference in soil to surface air maximum temperature recorded during the experiment reached a value of 21.2°C at the crest during June 1987, showing high super adiabatic lapse rate over the dune surface during day time. Minimum soil temperature during early morning was highest in May 1987 (about 28°C) and it showed considerable fall in November 1987. The minimum air temperature was higher than the minimum soil temperature, indicating inversion lapse rate (stable) from the soil to the surface air layer during night/early morning. The lapse rate above the surface layer is nearly adiabatic up to about 600 hPa in May and June during day hours under the influence of intense solar heating coupled with thermal and mechanical turbulence. During the monsoon months of July–August the lapse rate is mostly between dry-adiabatic and moist adiabatic, favouring conditional instability, which promotes moist convection under favourable conditions. Rajkumar *et al.*[14] while analysing MONTBLEX data at Jodhpur have found that even during the monsoon season, when the ground is wet and the sky is cloudy, the lapse rate in the lowest part of the atmosphere in the PBL is mostly stable during day and night hours. Usually most of the moisture resides in the mixed layer or planetary boundary layer during the monsoon period and it decreases sharply above the low level inversion layer which acts as a lid for the moist processes and suppression of convective clouds. During winter the lapse rate in the lower troposphere is statistically stable under normal conditions except during rain episodes under the influence of western disturbances.

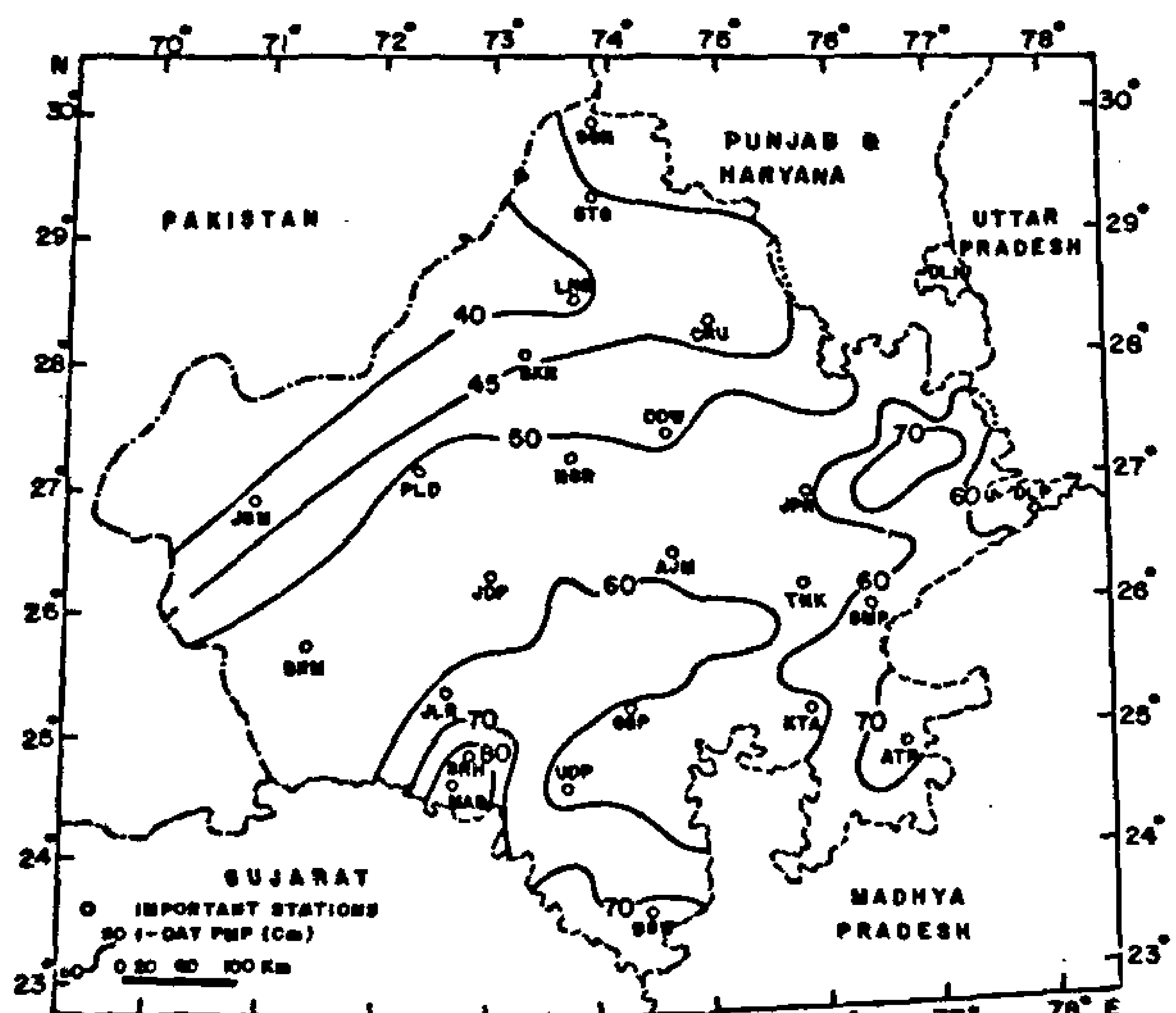

Figure 2. One day probable maximum precipitation over Rajasthan (after Rakhecha[12]).

Solar radiation and sunshine

Thar desert region receives intense solar heating during April to June and again in September and October. The

Table 1. Mean annual and monsoon season rainfall statistics for Rajasthan

Region	Annual rainfall (mm)	C.V. (%)	Monsoon season rainfall in mm (% of annual in parenthesis)	No. of rainy days
West Rajasthan	309	40	275 (88)	17
East Rajasthan	679	26	623 (92)	34
Rajasthan as a whole	468	31	425 (91)	29

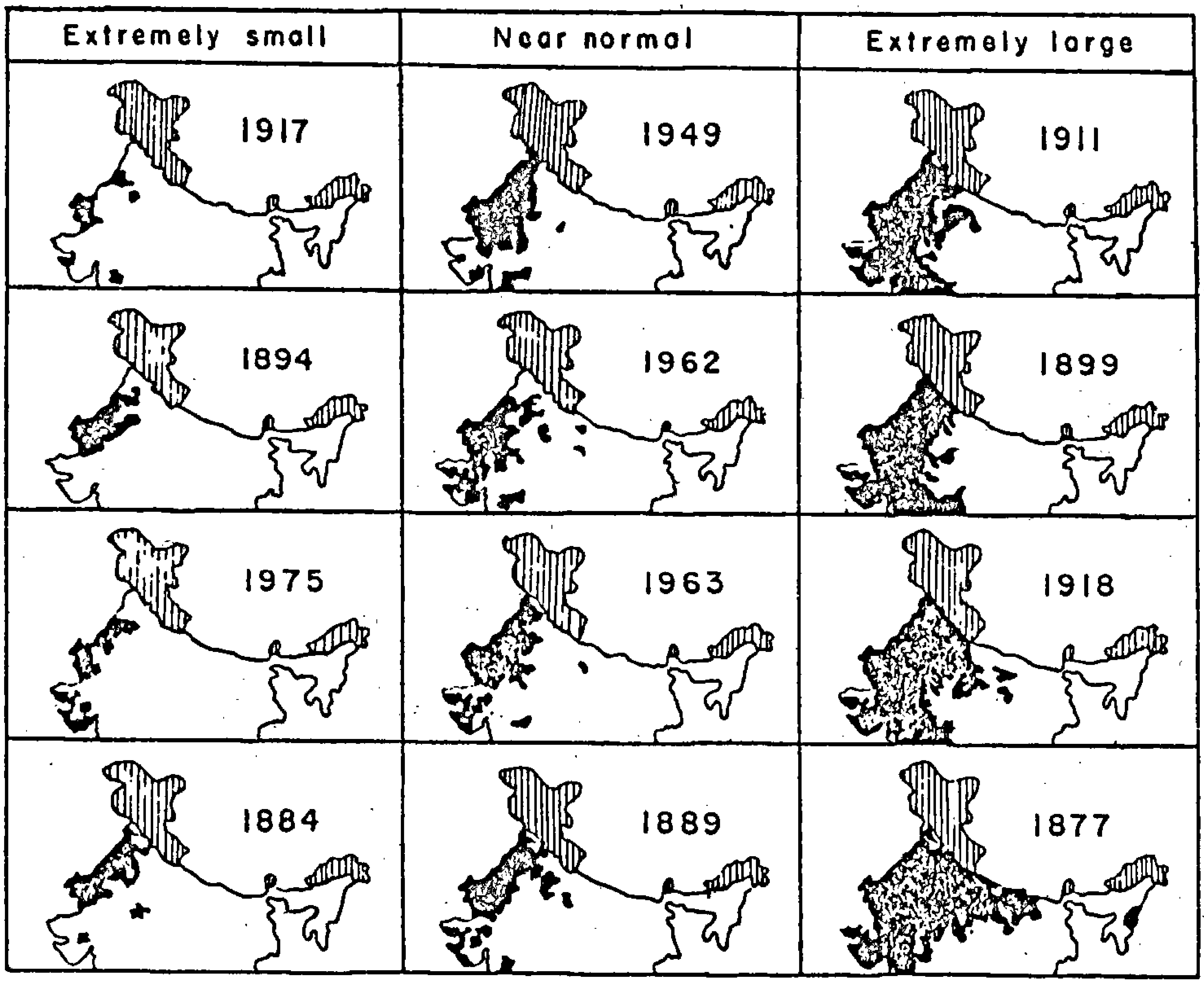

Figure 3. Fluctuations in arid area of North India in some typical years (after Nityanand Singh *et al.*[7]).

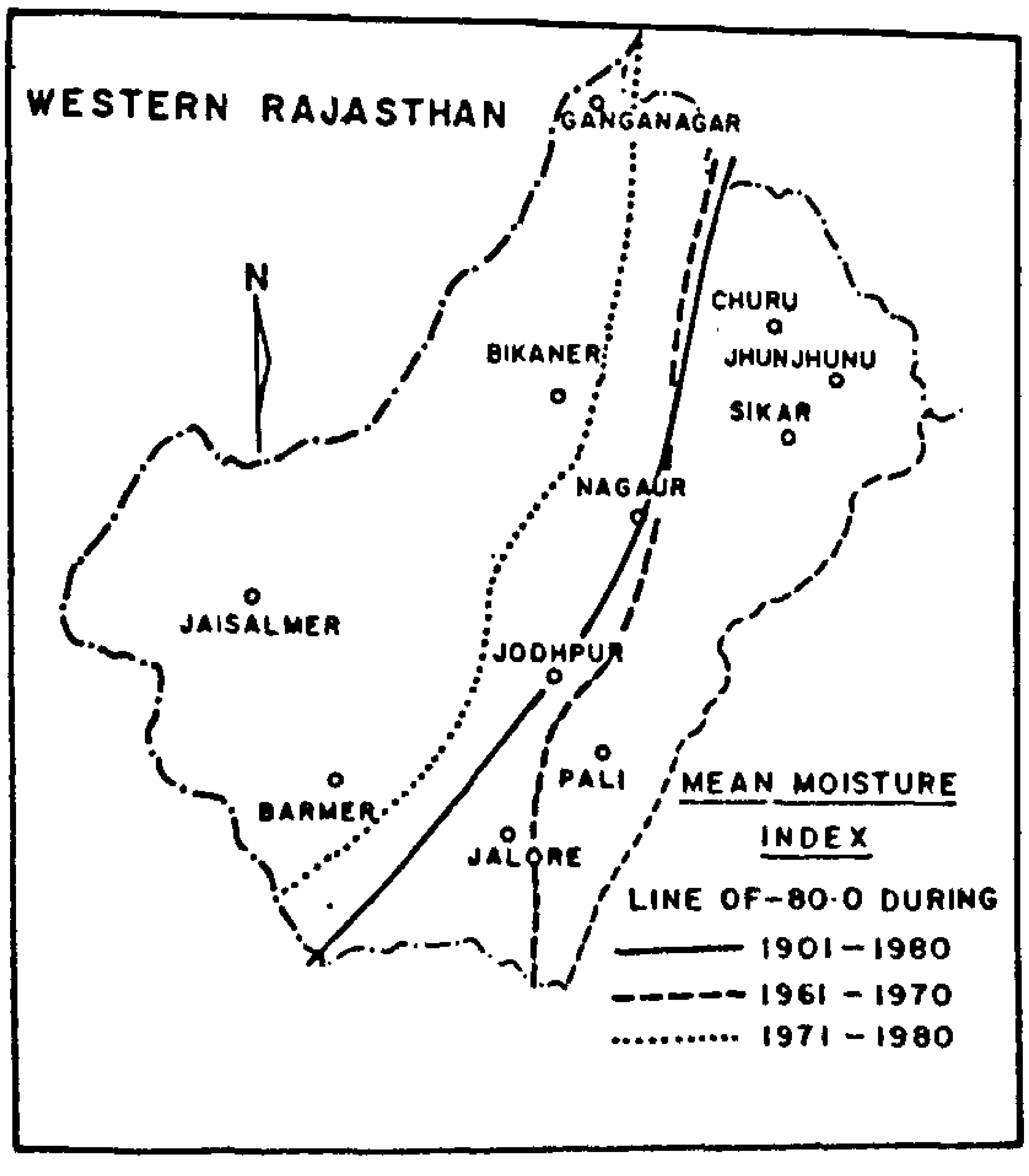

Figure 4. Shifts in the mean moisture index line of –80.0 over western Rajasthan (after Ramakrishna[8]).

maximum global radiation (direct and diffuse sky) per day is 700–800 cal/cm^2 per day in May and minimum of 400 cal/cm^2 in January with annual average of 560 cal/cm^2 per day. Under overcast sky conditions on a particular day the solar radiation is reduced to 190 cal/cm^2 per day. Solar radiation decreases in July and August and a secondary maximum occurs in October. Dust and clouds are major factors in attenuation of solar radiation. Maximum hours of bright sunshine, reaching 10 h per day occur in May and October[15,16]. According to their estimates[15,16] global radiation of 60 cal/cm^2/h is available for 4 to 5 h per day (1100 to 1600 h), indicating high potential for solar energy in the Thar desert. Desikan *et al.*[17] showed that there is a decreasing trend in solar radiation from 1965 to 1985 over Jodhpur, perhaps due to higher dust load.

Surface winds

Generally the winds near the surface have westerly component during major part of the year. They are stronger over the West Rajasthan than over the East Rajasthan. Surface winds reach a value of 25–30 km/h over West Rajasthan during May–June. Such strong

Table 2. Mean annual aridity index rainfall maximum and minimum temperatures for some stations in West Rajasthan

Station	Aridity index	C.V. (%)	Annual rainfall (mm)	C.V. of rainfall (%)	Annual max. temp. (°C)	Annual min. temp. (°C)
Barmer	−80.5	18	268	58	46.0	6.0
Jodhpur	−75.7	25	361	53	45.0	3.0
Bikaner	−77.5	22	294	55	46.0	2.0
Ganganagar	−77.5	22	214	43	46.0	3.0

Table 3. Drought years for West and East Rajasthan and the common droughts (period 1871–1990)

West Rajasthan drought years	East Rajasthan drought years	Common drought years
1877, 1883, 1885, 1887, 1891, 1901, 1902, 1904, 1905, 1911, 1913, 1915, 1918, 1920, 1925, 1938, 1939, 1941, 1946, 1948, 1951, 1963, 1968, 1969, 1972, 1974, 1979, 1982, 1986, 1987.	1877, 1895, 1898, 1899, 1901, 1905, 1907, 1911, 1913, 1915, 1918, 1925, 1938, 1939, 1941, 1951, 1965, 1966, 1972, 1979, 1982, 1986, 1987.	1877, 1892, 1901, 1905, 1911, 1913, 1915, 1918, 1925, 1938, 1939, 1941, 1951, 1972, 1979, 1982, 1986, 1987.
(30 years)	(23 years)	(18 years)

winds raise tremendous dust from the loose sandy soil of the region and the dust is transported eastward over neighbouring states of India.

Dust layer and dust storms

A dust layer from ground up to 8–9 km height is built up gradually from March to June which affects the horizontal and vertical visibility. Dust depletes solar radiation by 30%. During intense episode of dust-raising winds or following the dust storm day the sun is obscured from view even during the mid-day. The dust (sand) raising capacity of the surface air is proportional to the cube of the wind speed and hence the capacity of the winds to raise sand from the desert increases sharply after a wind speed of 10 km/h. The average soil loss in the Thar desert varies between 30 and 60 kg/m^2 per day during April to June[18].

Dust storms also occur more frequently in West Rajasthan than over the East Rajasthan during April to July, the frequency, however, being highest over Ganganagar (17 days), followed by Bikaner (15 days), Jodhpur (6 days) and Jaisalmer (2–3 days). Dust storms are often accompanied by thunderstorms (dry and wet). The annual number of thunderstorm days is about 10 over West Rajasthan and increases to about 40 over East Rajasthan. Singh *et al.*[19] have studied the hygro-thermal regime during a duststorm period in May 1990 under the MONTBLEX Programme.

Parameters of the hydrologic cycle

The annual PET over West Rajasthan varies from 2000 to 2500 mm and reaches a maximum of over 3000 mm in May–June. PET for monsoon months is about 600–800 mm. As the annual and monsoon seasonal rainfall figures are below the PET, moisture deficit exists over West Rajasthan throughout the year under normal conditions. However for the East Rajasthan moisture surplus does occur during July and August, particularly in years of excess rainfall. Since water at the surface is available only during rain spells in monsoon season and in winter, the evapotranspiration is high only during periods of rain spells. Soon thereafter the water gets evaporated due to high winds and high temperature. In the long dry period, the evapotranspiration is low as the vegetation closes its stomata to restrict transpiration in order to survive itself against moisture stress. Thus the average annual evapotranspiration over the Indian part of the Thar desert is about 400 mm. Since the soil in the desert is sandy, most of the rainfall percolates below the soil and is stored as soil moisture and very little is left for runoff. Even the run off percolates below the soil and the streams dry off down the course rather fast. Studies on soil moisture by Abichandani *et al.*[20] and Gupta[21] show that soil reaches field capacity in intense convective episodes and is charged with water up to even 100 cm depth. While natural hardy vegetation near the surface soil dries up in the dry season, deeper parts of its roots are sustained by moisture supply from the deep soil and with the first rains of the monsoon the desert springs up into life. Ramakrishna *et al.*[22] have discussed in detail about the annual moisture regime variability and its impact on agriculture in Rajasthan desert.

Climatic fluctuations over Rajasthan during instrumental period

Several workers have examined the question of climatic fluctuations over Rajasthan since the first major drought occurred in 1951 after India's independence[23-30]. All the studies suggest that there has been no major shift in the rainfall or temperature climate of Rajasthan. Pant and Hingane[29] showed that there is very little trend in Rajasthan rainfall over the last 100 years. However a small positive trend exists over East Rajasthan and neighbouring region of West Madhya Pradesh, Haryana and Punjab for the monsoon season rainfall and annual rainfall. Based on their analysis Pant and Hingane have rejected the results of Winstanley[31], suggesting that the rainfall of NW India was declining due to the equatorward shift of polar circulation and consequent equatorward shift of the ITCZ. They also conclude that a gradual decreasing trend in mean annual temperature of NW India is observed. The increasing rainfall and decreasing temperature trends are consistent and are attributed cautiously by the authors, to a progressive increase in the irrigated area. Ramakrishna and Rao[32] have also shown trend toward more rainfall, lower temperature and lower incidence of dust-storms in the command area of Indira Gandhi Canal.

Climate change scenario for Rajasthan due to anthropogenic activity

There is considerable worldwide concern about the possibility of climate change in desert regions due to global warming under the increasing greenhouse effect being caused by anthropogenic activity. Several numerical experiments have been performed world-wide for climate change under double CO_2 scenario using equilibrium climate change and employing stand alone atmospheric GCMs and coupled ocean–atmosphere GCMs. Rajasthan is too small a region to be effectively studied through the results of numerical modelling experiments. The result of the experiments could be valid with some certainty for the global climate system and considerable uncertainty exists when they are applied to regions/sub-regions. Since Rajasthan rainfall is mostly determined by the south-west monsoon of India, any change in its intensity under the double CO_2 conditions would impinge upon Rajasthan rainfall and water resources. Sikka and Pant[33] showed that for the stand-alone GCMs the monsoon rainfall of northern India may increase by 5–10% and the surface temperature may increase by 1–2°C. However, in the latest results of Lal *et al.*[34] using a coupled atmosphere–ocean model and incorporating sulphate aerosol due to projected enhancement of industrialization, the rainfall showed decrease by about 5% and the temperature also showed decrease by about 1°C. In any case these scenarios are only suggestive and the magnitude of projected change is rather quite small, much smaller in comparison to the interannual or interdecadal variability of the meteorological parameters witnessed in the instrumental period. Thus there is hardly any cause for concern with regard to climate change due to greenhouse global warming over India or over Rajasthan desert. The real problem in desert climate is the inter-annual variability. Most of India is situated in the deep moist tropics, with major ascent centred over the warm north Bay of Bengal waters and adjoining monsoon convective regions over the Eastern Gangetic plain in the west and moist Bangladesh, SE Asia and South China in the East. So long as this moist convective regime is maintained by natural atmospheric dynamics, it would suppress advancement of the arid conditions of Western Rajasthan farther eastward into the Gangetic plain.

Appraisal of palaeoclimate of the Thar desert

The Thar desert region, like many other deserts has witnessed climate changes due to natural causes. Thus, it is expected that in the historical and pre-historical pasts, monsoon fluctuations due to natural causes such as change in earth–sun geometry and glacial–inter-glacial episodes lasting for hundreds of years would have affected the climate of the Thar desert. Besides tectonic disturbances in the Aravallis would have affected the courses of river systems such as the Chambal–Yamuna river system and the legendary Saraswati and Drishdawati rivers which once flowed through the desert. Many attempts have been made to reconstruct the palaeoclimate of the Thar desert by employing a variety of multi-disciplinary approaches such as physiography[35], pollen data from the desert salt lakes[36], thermoluminescence chronology of fossil sand dunes[37,38], archaeological records and history of past civilizations which flourished in the Ganga–Yamuna–Indus–Saraswati river beds[39], glacial and inter-glacial chronology[40–44] and geomorphologic studies[45–49]. Pant and Maliekal[50] have given good surveys of the past climate changes from last glacial to the quaternary phase of the global climate. According to these studies the past climate history of Rajasthan can be grouped into the following phases:

Phase I. (20000–10000 years BP, last-glacial). This was the period of ice age conditions with cold temperature and extreme aridity in Rajasthan. Several numerical experiments, using atmospheric general circulation models support this inference as the monsoon with ice age boundary condition is simulated to be very weak in the models with its northward limit restricted to about 15°N only.

Phase II (10000 to 5000 years BP). Increase in vegetation and lake water levels suggestive of a continued moist period in Thar desert. Lake pollen and other data support this inference. Onset of deglaciation weakened sand mobilization in the period.

Phase III (5000 to 3000 years BP). This is the most moist period with frequent floods. The rainfall is estimated to be between twice and thrice of the present. This period is, therefore, called as the climate optimum of the region with warm, humid and equable climate when great civilization of Mohanjo Daro and Harappa flourished at its peak and its decay set in toward the end of the phase – possibly by large fluctuations in the monsoon due to climate change from wet to the dry phase.

Phase IV (3000 BP to 18th century AD). This was the warm and dry period when the aridity had established once again and has continued almost unabated to the present with minor interdecadal and centennial scales of climate fluctuations in which somewhat wetter or drier regimes followed each other. This is supported by historical/archaeological evidence of rise and fall of post-Harappan cultures in the region like the painted grey ware (2800–2400 years BP) and the Rang Mahal culture (100 BC to 100 AD). The period between 600 AD and 1000 AD is inferred to be wetter and the period between 1400 and 1700 AD somewhat drier (Little Ice Age).

The interface between phase III and phase IV coincided with tectonic activity in the Aravallis region which brought changes in the courses of rivers Saraswati, Drishdawati, Sutlej, Ghaghar and Chambal. The drying out of the river bed of Saraswati and its feeding channels occurred and the glory of Saraswati was lost by the time of *Mahabharata*, i.e. 1500 BC or so[51]. Sand mobilization during this period correlates well with evidence about the desiccation of lakes[52]. Bryson and Baerries[53] suggest possibility of climate modification by human activity in the region during setting up of the natural dry phase which according to them would have led to more dust load and consequently aridity followed.

Local and regional dynamic studies on desert climate

Structure of summer-time heat low

In the recent years several special field experiments have been undertaken internationally and nationally in several countries to understand the dynamical structure of the atmosphere over desert regions. In India, the IITM Pune and the Gujarat Agricultural University, Anand are currently engaged with a land surface process experiment (LANDSPEX) over Gujarat region which is supported by the DST. Earlier in 1990, IITM carried out MONTBLEX[54] and a special observation on boundary layer processes was studied at Jodhpur. An elaborate experiment was carried out in the Arabian desert during the summer MONEX-1979 Programme[55]. The following results have emerged about the local/regional dynamics of the Thar/Arabian desert as a result of the above experiments and field studies carried out by the CAZARI scientists.

(i) Summer-time 'desert heat low' is a shallow cyclone at the surface which is replaced by a warm anticyclone above 1–2 km height.

(ii) Surface air temperature during most of the summer is 45–50°C and the surface soil temperature is higher by 10–15°C. Diurnal variation of air temperature is 20°C and surface soil temperature is 25 to 30°C. The soil cools rather rapidly with the onset of autumn and the winter minimum temperature reaches below the freezing point. Soil and air temperatures rise rapidly with sun rise and near-maximum temperature is reached by 1100 h and stays so for about 6 h in summer. By early night a surface inversion is set up even in summer.

(iii) The planetary boundary layer (PBL) is unique in summer (April to June) and is highly unstable for dry convection as super adiabatic lapse rate prevails. The thermal and wind-induced mechanical turbulence lasts for about 10–15 h during the day. The convective layer is up to 500 hPa (mixed layer height) in depth and the eddy transport of sensible heat is high from warm ground surface upward into the unstable PBL. This vertical advection of heat in the lower troposphere is about 3°C/day. During monsoon period (July and August) the PBL has sufficient moisture to realize conditional instability under favourable synoptic conditions.

(iv) The summer monsoon thermal 'low' undergoes quasi-periodic intra-seasonal oscillations on 5–7 and 10–15 day scales. These oscillations occur due to cooling above the mixed layer from an imbalance between the sensible heat flux divergence and the ground radiative heating. The dust load is mostly confined between the surface and mixed layer, which cuts off part of the solar radiation reaching the ground, leading to some decrease of temperature near the ground as well as reduction in vertical advection of sensible heat into the PBL. The heat law, thus, decreases in intensity (decay phase) and the mixed layer is somewhat stabilized reaching neutral phase from the convective phase.

(v) The dust layer plays a crucial role by absorbing incoming solar radiation and also by restricting outgoing thermal emission in the IR. Thus there is net radiative heating of 2–4°C/day in the troposphere above 700 hPa and net cooling of 6–8°C in the lower troposphere. This results in net cooling of the atmospheric column by 1–2°C at the top of the atmosphere. Radiation budget studies with *in-situ* and satellite-based observations confirm net tropospheric cooling on local and regional scales. As such, in order to sustain the desert heat low, it is necessary that energy is transported toward the desert by the large scale monsoon vertical circulation. This is accomplished by the moist ascent taking place over the convective regime centred over North Bay of Bengal and dry descent in the peripheral desert heat lows over the Thar and the Arabian deserts.

(vi) A two-cell structure in horizontal divergence prevails during the day in the troposphere in the heat low region with shallow convergence (at best up to 850 hPa) and divergence aloft up to mid-troposphere, topped by weak convergence in the upper troposphere. Maximum downward motion exists near the top of the mixed layer. During night time downward motion exists right up to the surface as the low level convergence is considerably reduced or even replaced by divergence.

(vii) Drag coefficient in the desert region[56] is 3 to 4 times higher than for the oceanic areas.

Bhide *et al.*[57] showed that at 500 hPa, heat sources and moisture sinks propagate westward on 10–15 day and 30–50 day scales from eastern part of the trough over the Bay of Bengal across the monsoon trough region. When a monsoon depression moves into the Rajasthan region, moisture is fed into the entire vertical column and tall clouds form, resulting in considerable rainfall over the region. The region of 'heat low' now becomes a region of heat source due to release of latent heat and moisture sink due to moisture convergence accompanying the weather disturbance. This is a transient phenomenon, which is replaced by the normal heat law pattern in about 3–5 days and the desert region again becomes a heat sink.

Regional and global and circulation cells and the Indo-Saharan desert climate

Two important vertical cells, viz. the Hadley cell in the south-north direction and the Walker cell in the east-west direction, have crucial links with the desert climate over the Indo-Saharan region. Hadley cell is primarily responsible for the year-round aridity as the desert regions fall in its downward limb. Even in summer monsoon season when the traditional Hadley cell has shifted north of India, the desert regions are mostly dry except during occasional travel of the monsoon disturbances to extreme western parts of India. This is as a consequence of two factors. Firstly, Das[58] showed the existence of a regional east-west vertical circulation in which deep ascent occurs over eastern India and the compensatory descent occurs over the Thar desert. This regional cell oscillates on sub-seasonal scale as discussed in the earlier paragraphs. Secondly a global scale Walker circulation prevails with ascent centred over eastern India and descent over the Indo-Saharan desert region in the west. This is supported by divergence at 200 hPa over the Tibetan high region and convergence in the adjoining mid-Pacific trough in the east and mid Atlantic trough in the west on the planetary scale. On interannual scale this global Walker cell along the summer ITCZ belt in NH is influenced by the ocean atmosphere coupled system in the equatorial eastern Pacific, commonly known as the ENSO phenomenon. During the warm phase of the ENSO, the Tibetan high at 200 hPa is displaced more eastward. A zone of convection migrates eastward from Indonesia to the central Pacific and the large-scale convective activity is decreased over India and adjoining tropical belt. This results in major monsoon drought over SE Asia, India, Somalia and sub-Saharan regions.

Numerical atmospheric general circulation experiments on desert climate

Numerical experiments

ACGMs have been extensively used in the last two decades to understand the process of desertification on regional scale. The premise for such studies is on three approaches, all of which have links with the energy balance principle, as applied to deserts. Deserts are regions of net energy deficit, because of high albedo, high long-wave emissions, high sensible heat and low latent heat contribution in the desert–atmosphere system. Otterman[59] and Charney[60] proposed hypothesis for the desertification which was based on the increase in albedo of the deserts by human activity such as overgrazing and loss of vegetation cover with consequent more radiative energy loss and decrease in temperature of the desert environment and reduction in convective overturning, resulting in enhancement of the desert system. The second hypothesis was tested by the numerical experiments of Walker and Rowntree[61], Shukla and Mintz[62] and several other workers since then. It is based on the vegetation–atmosphere interaction involving the role of soil moisture and evapotranspiration in modulating the energy balance, rainfall, temperature and moisture fields. A third approach was suggested by Sud and Smith[63] and Sud *et al.*[64] which rested on changes in surface roughness due to decrease in vegetation affecting the surface momentum balance and hence the efficiency of heat and moisture transports in the desert–atmosphere system.

All the three approaches involve biogeophysical mechanisms and are linked with land–vegetation–atmosphere or the land surface processes. Although some observational studies support the above approaches, they can best be tested by employing controlled and anomaly GCM experiments through sensitivity studies. As such a spate of such experiments on all the hypotheses have been performed by different numerical modelling groups since the first work by Charney *et al.*[65] and Chervin[66], using different parametrization schemes for the land surface processes, some of which even include separate biospheric models. Good reviews on the subject are provided by Dickinson[67], Rowntree[68], Sud *et al.*[69] and Dirmeyer and Shukla[70]. Many numerical modelling studies have been carried out to understand regional droughts such as Sahellian, north American Great Plains, Australian and South African droughts on the above three approaches since 1970s. In a review of 11 such studies, Mintz[71] concluded that soil moisture had a stronger impact on model-simulated climate than did albedo or surface roughness. Some of the models have been used to understand the role of biogeophysical processes on the monsoon rainfall in western Indian desert and the results are mixed[64,72,73]. This may be, among other factors, due to the fact that the numerical models have yet to produce an acceptable simulation of the monsoon rainfall of India and hence at this stage may not be very appropriate tools for sensitivity studies on monsoon. Recently Fennesey and Shukla[74] have provided a review of the role of soil moisture anomaly studies on precipitation and temperature anomalies in the mid-latitude belt of USA.

There is a variety of observational evidence supporting and even contradicting the expectations based on biogeophysical feedbacks. Recent observational studies have pointed to reduced importance of albedo feedback on desertification and have challenged Charney's hypothesis by showing that surface warming instead of surface cooling occurs during prolonged droughts. Besides they also claim that Charney's hypothesis does not explain synchronous occurrence of major droughts over several regions such as in Sahel, Kalahari, Australian and Thar deserts. It also does not explain the termination of such droughts. Thus albedo model is now considered as interesting but unproven. On the other hand, soil moisture mechanism for desertification by bringing changes in the hydrological parameters through vegetation–atmosphere interactions, has gained importance. The major factor for the synchronous occurrence of droughts on inter-annual and inter-decadal scales is considered to be the coupled ocean–atmosphere ENSO phenomenon. However, it cannot be ignored that human activity has pronounced influence on the local scale desert climate through accelerated soil erosion, deforestation and loss of vegetation and soil organic matter by over-grazing and over-exploitation of the soil resource. In areas without irrigation facilities, desertification can lead to increased temperature and reduced rainfall. Irrigation can reverse these effects, as it has happened in the Indira Gandhi Canal command area, but it is producing increased salination of the soil.

Numerical experiments on expansion of desert climate and global climate

As deserts are important sub-component of the global climate system, there is a concern that expansion of deserts may impinge on the global climate. Observational evidence is mounting that desertification has been occurring extensively over the world for the last several decades[75–77]. In order to specifically address the problem of enhancement in desert areas and their possible impact on regional global climate, Dirmeyer and Shukla[78] have used a GCM and the realistic land surface properties with present day extent of deserts (control run) and doubling their extents (anomaly run) over all the continents. The deserts were extended by altering the land surface properties and type of vegetation over the present day deserts and their margins. The double desert case represents an extreme scenario of desertification which might ensue as a result of natural and anthropogenic causes. The results of the control and anomaly runs for 10 years were compared and showed the following aspects prominently:

(i) Significant cooling over much of the tropics and sub-tropics and the cooler air in the double desert case covered an area nearly six times the size of the current deserts. This impact is, therefore, global in nature.

(ii) Rainfall generally, but invariably, decreases over desertified areas. Over Africa the decrease is largest (20 to 60%). Over the leading edge of the Indian monsoon region such as the Gangetic plains and adjoining Myanmar and China significant decrease in rainfall (about 10%) is also noticed. Some increase in rainfall is noticed over Peninsular India, Sri Lanka and Malaysia. Similarly evapotranspiration anomalies are larger over Africa than over South Asia. Lower response over South Asia is perhaps due to the fact that this core region of the monsoon is influenced by warm SST in Western Pacific and Bay of Bengal regions as well as by the elevated heat source over the Tibetan region.

The results of this experiment more or less followed Charney's hypothesis with regard to cooling over the deserts. The experiment of Dirmeyer and Shukla[78] no doubt points to potential danger of decreased rainfall over much of the tropics and sub-tropics in response to doubling of the present extent of the deserts.

Concluding remarks

We have discussed many aspects of desert climate based on the work of many researchers. Desert climate is harsh for human habitation as these ecosystems are highly responsive to climate variability. However, man has lived in several arid and semi-arid areas for thousands of years in harmony with nature, particularly in the Thar desert which is unique for it lies on the western margin of the Indian Summer Monsoon System and where the population is fairly high. Population pressure and increasing exploitation of the vegetational and soil resources in desert lands are raising serious doubts among experts whether nature can sustain such over-exploitation resulting from human greed. As desertification and desert climate are intimately linked, several science and technology measures would benefit sustainable development of the arid and semi-arid regions and some of them are:

(a) Provision of natural resource information to local people and policy makers. For India such an information system is being developed by DST and the Department of Space in collaboration with NRSA.

(b) Enhancement of regional climate monitoring network.

(c) More numerical modelling studies and their coupling with empirical and observational studies such as land surface field experiments.

(d) Application of seasonal monsoon forecasting for contiguous arid and semi-arid regions separately rather than the present system which addresses India as a whole.

Desert climate is an integral sub-component of the global climate. However, if desertification is allowed to go unabated, numerical modelling studies point to a scenario that it would impact on the climate of regions farther away from deserts themselves. This would be an unfortunate event for the modern world civilization which is based on intensive development culture and over-consumption. There is a silver lining that the problems posed by desert and threatened desert margin climates can be tackled adequately by combining modern science and technology with the traditional rational knowledge of the inhabitants of the regions, who have learnt to live for many generations in harmony with their rather harsh natural environment. They have used its fragile resource on sustainable basis till now with wisdom born out of experience. Science and technology can surely help them to do it for future too.

1. Koppen, W., in *Hardluichder Klimatologie* (eds Koppen, W. and Geiger), Gehruder Bomtraenger, Berlin, 1936, vol. 1, Pt. C.
2. Thornthwaite, C. W., *Geograph. Rev.*, 1948, **38**, 55–94.
3. Budyko, M. I., *The Heat Balance of the Earth's Surface*, Translated by N. Stepanava under US Dept. of Commerce, Washington, D. C., 1958, pp. 259.
4. Lettau, H., *Mon. Wea. Rev.*, 1969, **97**, 691–699.
5. Berkofsky, L., *WCP-75*, World Met. Organization, Geneva, 1984, **97**, 691–699.
6. Dhar, O. N., Rakhecha, P. R., Mandal, B. N. and Sangam, R. B., *PAGEOPH*, 1982, **120**, 483–494.
7. Nityanand Singh, S. S., Mulye and Pant, G. B., *PAGEOPH*, 1992, **138**, 135–150.
8. Ramakrishna, Y. S., in *Desertification: Monitoring and Control* (ed. Tiwari, A. K.), Scientific Publications, Jodhpur, pp. 99–115.
9. Krishnan, A., *Ann. Arid Zone*, 1977, **16**, 302–309.
10. Parthasarthy, B. and Rakhecha, P. R., *Ann. Arid Zone*, 1972, **11**, 161–168.
11. Sikka, D. R., *Proc. Indian Acad. Sci.* (*Earth Planet. Sci.*), 1980, **89**, 179–199.
12. Rakhecha, P. R., in *Natural and Human Resources and Rajasthan* (ed. Chauhan, T. S.), Scientific Publications, Jodhpur, 1993.
13. Ramakrishna, Y. S., Rao, A. S., Singh, R. S., Aamal Kar and Singh, S., *J. Arid. Env.*, 1990, **19**, 25–38.
14. Rajkumar, G., Narasimha, R., Singhal, S. P. and Gera, B. S., *Proc. Ind. Acad. Sci.* (*Earth Planet. Sci.*), 1996, **105**, 1–17.
15. Desikan, V., Swaminathan, M. S. and Chacko, O., *Indian J. Meteorol. Geophys.*, 1968, **19**, 149–158.
16. Ganesan, H. R., *Indian J. Meteorol. Geophys.*, 1970, **21**, 267–270.
17. Desikan, V., Chivate, V. R. and Abhyankar, V. V., *Mausam*, 1994, **45**, 267–270.
18. Gupta, J. P., Agarwal, R. K. and Raikhy, N. P., *J. Arid Environ.*, 1981, **4**, 15–20.
19. Singh, R. S., Rao, A. S., Ramakrishna, Y. S. and Prabhu, A., *Ann. Arid Zone*, 1992, **31**, 153–155.
20. Abichandani, C. T., Krishnan, A., Bhatt, P. N. and Rakhecha, P. R., *J. Indian Soc. Soil Sci.*, 1967, **15**, 7–15.
21. Gupta, J. P., *J. Hydrol.*, 1979, **41**, 153–156.
22. Ramakrishna, Y. S., Victor, U. S. and Ramana Rao, B. V., in *Climate Variability and Agriculture* (eds Yash P. Abrol, Sulochana Gadgil and Govind B. Pant), Narosa, New Delhi, 1966, pp. 242–252.
23. Banerji, S. K., in Proceedings of the Symposium on the Rajasthan Desert, National Institute of Science, New Delhi, 1952.
24. Pramanik, S. K., Hariharan, P. S. and Ghosh, S. K., *Indian J. Meteorol. Geophys.*, 1952, **3**.
25. Rao, K. N., *Indian J. Meteorol. Geophys.*, 1958, **9**.
26. Jagannathan, P., Bhalme, H. N. and Rakhecha, P. R., *Proc. Indian Nat. Sci. Acad.*, 1978, **B44**, 364–380.
27. Ramakrishna, Y. S. and Sastri, A. S. R. S., *Nat. Geogr. J. India*, 1980, **26**, 44–49.
28. Ramakrishna, Y. S., Sastri, A. S. R. S., Rao, G. G. S. N. and Ramana Rao, B. V., *Mausam*, 1984, **35**, 349–354.
29. Pant, G. B. and Hingane, L. S., *J. Climatol.*, 1988, **8**, 391–401.
30. Parthasarathy, B., Sontakke, A., Munot, A. A. and Kothawale, D. R., *J. Climatol.*, 1987, **7**, 57–70.
31. Winstrnley, D., *Nature*, 1973, **243**, 464–465.
32. Ramakrishna, Y. S. and Rao, A. S., in *Prospects of IGNP Command Area* (eds Abrol, I. P. and Venkateswrlu, J.), ICAR, New Delhi, 1992, pp. 1–10.
33. Sikka, D. R. and Pant, G. B., in *Impact of Global Climatic Changes on Photosynthetics and Plant Productivity* (eds. Abrol, Y. P. *et al.*), Oxford and IBH, New Delhi, 1992, pp. 551–572.
34. Lal, M., Srinivasan, G. and Cubasch, U., *Curr. Sci.*, 1996, **71**, 746–752.
35. Allachin, B., Goudie, A. and Hedge, K. T. M., *The Pre History and Palaeogeography of the Great Indian Desert*, Academic Press, New York, 1975, pp. 370.
36. Bryson, R. A. and Swain, A. H., *Quart. Res.*, 1981, **16**, 135–145.
37. Chawla, S., Dhir, R. P. and Singhvi, A. K., *Quart. Sci. Rev.*, 1992, **11**, 25–32.
38. Singhvi, A. K., Banerjee, D., Rajguru, S. N. and Krishnan Kumar, V. S., *Curr. Sci.*, 1994, **66**, 770–773.

39. Lal, M., *Settlement Patterns of Rise of Civilization in the Ganga Yamuna Doabs from 1500 BC to 300 AD*, B. R. Publishing Corporation, New Delhi, 1984.
40. Goudie, A. D., Allachin, B. and Hegde, K. T. M., *Geogr. J.*, 1973, **139**, 243–257.
41. CLIMAP Project Members, *Science*, 1976, **191**, 1131–1137.
42. Singh, G. and Aggarwal, D. P., *Nature*, 1976, **260**, 5598.
43. Duptessy, J. C., *Nature*, 1982, **295**, 494–498.
44. Rajguru, S. N., *Environ.*, 1983, **7**, 107–111.
45. Ghose, B., Pande, S., Singh, S. and Lal, G., *Ann. Arid Zone*, 1966, **5**, 10–25.
46. Verstappan, H. T., *Z. Geomorphol.*, 1970, **10**, 104–120.
47. Agarwal, D. P., Dhir, R. P., Krishnamurthy, R. V., Misra, V. N., Nanda, S. C. and Rajguru, S. N., in *Arid Zone Research and Development* (ed. Mann, H. S.), Sai Publishers, Jodhpur, 1980.
48. Wasson, R. J., Smith, G. I. and Agarwal, D. P., *Paleogeogr. Palaeoclimatol. Palaeoecol.*, 1984, **46**, 345–372.
49. Singh, G., Wasson, R. J. and Agarwal, D. P., *Rev. Paleobot. Palyanol.*, 1990, **64**, 351–358.
50. Pant, G. B. and Maliekal, J. A., *Climate Change*, 1987, **10**, 183–194.
51. Ghosh, B. and Kar, A., Proceedings of Indo-US Workshop on Arid Zone Research, Jan 9–14, 1984, pp. 140–149.
52. Singh, G., Joshi, R. D. and Singh, A. B., *Quart. Rev.*, 1972, **2**, 496–505.
53. Bryson, R. A. and Baerries, D. A., *Bull. Am. Meteorol. Soc.*, 1967, **48**, 136–142.
54. Sikka, D. R. and Narasimha, R., *Proc. Indian Acad. Sci. (Earth Planet. Sci.)*, 1995, **104**, 157–187.
55. Blake, D. W., Krishnamurti, T. N., Low-Nam., S. V. and Fein, J. S., *Mon. Wea. Rev. III*, 1992, pp. 1759–1775.
56. Mohanty, U. C., Parihar, P. S., Venugopal, T. and Parashuram, *Proc. Indian Acad. Sci. (Earth Planet. Sci.)*, 1995, **104**, 273–287.
57. Bhide, U. V., Majumdar, V. R., Ghanekar, S. P., Paul, D. K., Chen, T. C. and Rao, G. V., A diagnostic study on beat sources and moisture sinks in the monsoon trough area during active break phases of Indian Summer Monsoon 1979, 1996, Under publication.
58. Das, P. K., *Tellus*, 1962, **14**, 212–220.
59. Otterman, J., *Science*, 1974, **196**, 531–533.
60. Charney, J. G., *Quart. J. R. Meteorol. Soc.*, 1975, **101**, 193–202.
61. Walker, J. and Rowntree, P. R., *Quart. J. R. Meteorol. Soc.*, 1977, **103**, 29–46.
62. Shukla, J. and Mintz, Y., *Science*, 1982, **215**, 1498–1501.
63. Sud, Y. C. and Smith, W. E., *J. Boundary Layer Meteorol.*, 1985, **33**, 14–49.
64. Sud, Y. C., Shukla, J. and Mintz, Y., *J. Appl. Meteorol.*, 1988, **27**, 1036–1054.
65. Charney, J. G., Quirk, W. J., Chow, S. H. and Kornfield, J., *J. Atmos. Sci.*, 1977, **34**, 1366–1385.
66. Chervin, R. M., in *Report of the JOC Study Conference on Climate Models, Performance intercomparison Sensitivity Studies*, Washington D.C., April 1978, GARP Pub. Series 22, 1979, pp. 563–581.
67. Dickinson, R. E., *Adv. Geophys.*, 1983, **25**, 305–353.
68. Rowntree, P., in Proc. U.N. University Workshop on Forest, Climate and Hydrology – Regional Impacts, U.N. University, Tokyo, 1988.
69. Sud, Y. C., Chao, W. C. and Walker, G. K., *J. Arid Environ.*, 1993, **25**, 5–18.
70. Dirmeyer, P. A. and Shukla, J., in *Predictions of International Climate Variations* (ed. Shukla, J.), Springer, NATO Ser. I, vol. 6, 1993, pp. 1–23.
71. Mintz, Y., in *The Global Climate* (ed. Houghton, J.), Cambridge Univ. Press, 1984, pp. 79–105.
72. Sud, Y. C. and Fennessy, M. J., *J. Climatol.*, 1984, **4**, 383–398.
73. Sud, Y. C. and Smith, W. E., *J. Climat. App. Meteorol.*, 1985, **24**, 1015–1036.
74. Fennessy, M. J. and Shukla, J., Rept. No. 34, Centre for Ocean–Land–Atmosphere Studies, Calverton, Md. USA, 1996.
75. Dregne, H. E., in *Desert and Arid Land Technology and Development*, Harwood Academic, New York, 1983, vol. 3, pp. 242.
76. Allan, R. J. and Haylock, M. R., *J. Climate*, 1993, **6**, 1356–1367.
77. William, M. J. and Balling, R. C., in *Interactions of Desertification and Climate*, Arnold Publishers, London, 1996, pp. 270.
78. Dirmeyer, P. and Shukla, J., Rept. No. 03, Centre for Ocean–Land–Atmosphere Studies, Calverton, Md. USA, 1994.

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