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What Are the Key Components of Climate as a Driver of Desertification?

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ABSTRACT

Climate and land surface are inextricably linked. Desertification must therefore necessarily be evaluated within the context of the climatic background. Climatic variables such as temperature and rainfall determine land surface character to a first approximation; the characteristics of the surface, in turn, affect the fluxes of energy, moisture, and particulates that modulate meteorological processes. This chapter provides an overview of these interactions and an examination of the aspects of climate which influence surface vegetation. The essential question considered is: what determines the availability of surface moisture? This chapter also compares and contrasts the meteorological characters of the world's drylands and discusses the extent to which the differences influence sensitivity to climatic change. The feedback between desertification and climate is considered in detail, and an appropriate model for evaluating the relationship between desertification and climate is described. The aspects which are probably most important involve changes in the water retention capability of the land as a result of changes in surface soils and vegetation and large-scale generation of aerosols.

Many of the global drylands are very sensitive to climatic variability, so that there is much concern about these regions in the context of global change. A particularly relevant question is whether global change might make the Earth's drylands more susceptible to desertification. Unfortunately, this is a question that can only be answered via numerical simulations of climate. At present, climate models are not sufficiently advanced to be able to answer this question with any great confidence. In view of these inadequacies and because of the links between climate and desertification, it is important to first conduct analytical studies that focus on understanding the causes of climatic variability on a regional basis. Following this, numerical simulations via regional climate models coupled to global models become appropriate.

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INTRODUCTION

Climate and land surface are inextricably linked. A dramatic case in point is the African Sahel, a region which was the focus of many of the earliest assessments of desertification. The first United Nations (UN) Conference on Desertification in 1977 was accompanied by extensive literature discussing the "marching sands" and the "advancing desert." In the Sahel, the quantification of the process was based mostly on two measurements giving the limits of the Acacia tree in the Sudan, one in 1958 and the other in 1975. Figure 3.1 shows the dramatic change of rainfall occurring in the Sahel as a whole over this time period. Clearly, any changes of the land surface must be assessed in the context of a major change in the region's climate.

Climatic variables such as temperature and rainfall determine land surface character to a first approximation and the characteristics of the surface, in turn, affect the fluxes of energy, moisture, and particulates that modulate meteorological processes. To understand the meteorological aspects of desertification, both the impact of climatic variability and the accompanying feedbacks to the atmosphere must be considered. I begin with an overview of the interactions in the context of factors regulating the interannual variability of climate, then continue with an examination of the aspects of climate which influence surface vegetation. For most dryland regions, this is equivalent to asking what determines the availability of surface moisture. Next I compare and contrast the meteorological character of the world's drylands and discuss the extent to which the differences influence sensitivity to climatic change. The feedback between desertification and climate is thereafter considered. An appropriate model for evaluating the relationship between desertification and climate is described, which draws on the discussions in the earlier sections. Finally, the various aspects of the meteorological dimensions of desertification are then synthesized within the discussion of issues concerning modeling, assessment, and prediction of climate and global change.

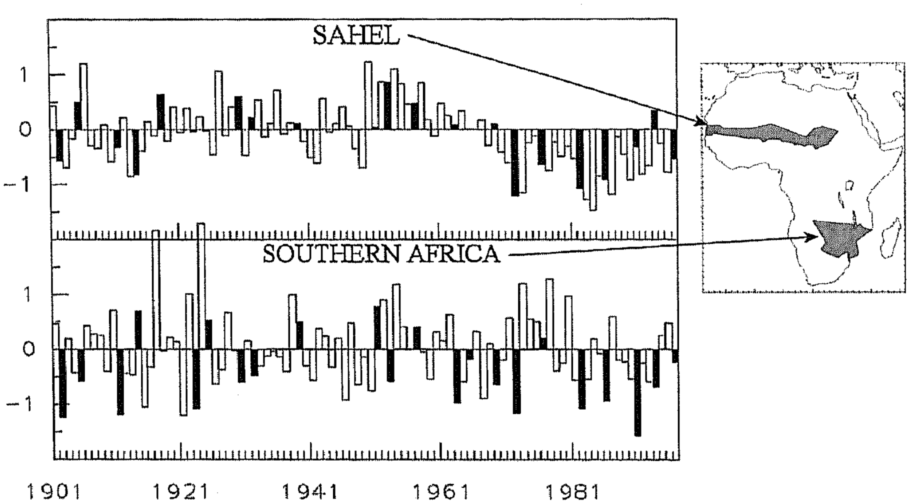

Figure 3.1 Rainfall in the Sahel and in southern Africa from 1901 to 1997, expressed as a percent of the standard deviation from the long-term mean (based on Nicholson et al. 2000). El Niño years are shaded. For southern Africa, rainfall is calculated from July of the indicated year to June of the following year. (Departures calculated from a series of individual station means than averaged for each region.)

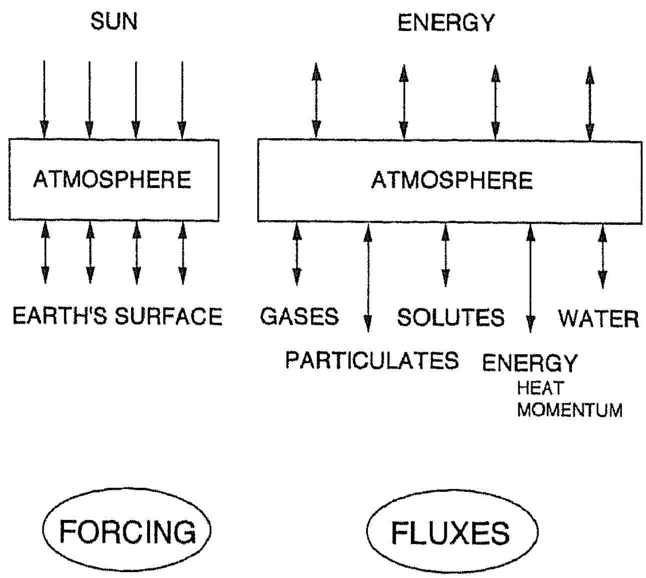

Figure 3.2 Schematic of interactions and fluxes between land surface and atmosphere. Atmospheric processes are forced via the flux of gases, particulates, solutes, energy, and water from the Earth and energy from the sun. Double arrows indicate fluxes both to and from the atmosphere (Nicholson 1999).

OVERVIEW OF INTERACTIONS BETWEEN THE LAND SURFACE AND CLIMATE

Figure 3.2 schematically depicts the "forcing" of global climate. The climate system is driven by a combination of fluxes from the upper and lower boundaries and the internal dynamics of the atmosphere. The flux from the upper boundary is essentially ultraviolet, visible, and infrared radiation from the sun and is termed shortwave radiation. The fluxes from the lower boundary, the Earth's surface, include water vapor, gaseous molecules such as methane and carbon dioxide, momentum, radiative energy, latent heat associated with evaporation, and particulates. Internal dynamics denotes interactions that are purely within the atmospheric domain, such as the interactions of atmospheric waves and jet streams or the interactions between storm systems, clouds, and the general atmospheric circulation. Interannual variability of climate is a response to both boundary forcing and internal dynamics, but the latter acts on a much shorter time scale. Thus, boundary forcing is determined by the distribution of land, water, and ice as well as the physical characteristics of these surfaces (such as roughness, temperature, and, in the case of land, moisture content). Boundary forcing generally serves to stabilize meteorological processes, but it can trigger change, as in the case of the El Niño.

CLIMATE AS A DETERMINANT AND RESPONSE TO THE GLOBAL LAND SURFACE

The climate system operates in such a way that external energy input from the sun is converted to the kinetic energy of the atmosphere. Key to this conversion are processes at the Earth's surface that determine the amount of energy available to the atmosphere via radiation and latent heating. These processes can be described through the consideration of surface radiation and energy balance and surface water balance (see below).

The relationship to vegetation is complex. The global distribution of vegetation, including type and degree of cover, is largely determined by the availability of energy and moisture at the surface. However, surface cover is also a large determinant of this availability. To better understand the relevant feedback processes, the radiation, energy, and water balance of the surface are considered in the context of how the surface character (particularly, vegetation, soil type, and soil moisture) both responds to and influences the processes that control energy and moisture availability.

Radiation and Energy Balance

The close link between the land surface and climate can be demonstrated by considering the ultimate driver of climate, solar energy. Atmosphere is relatively transparent to solar energy (shortwave radiation) but readily absorbs the radiation emitted by the Earth's surface (longwave radiation). The general circulation of the atmosphere, including storm systems, represents a conversion of this radiant energy to kinetic energy. A critical component in this process is the conversion of shortwave radiation to longwave radiation and latent heat that takes place at the Earth's surface.

The net amount of radiant energy available at the surface to drive the climate system is termed net radiation, defined as:

$$R_{net} = R_{sw}\downarrow - R_{sw}\uparrow + R_{lw}\downarrow - R_{lw}\uparrow = R_{sw}\downarrow(1-a_s) + R_{lw}\downarrow - R_{lw}\uparrow, \tag{3.1}$$

where sw refers to shortwave or solar radiation, a_s is surface albedo or reflectivity, and lw refers to the longwave radiation emitted by the Earth, atmosphere, and clouds. $R_{lw}\uparrow$ is that emitted spaceward from the Earth's surface; $R_{lw}\downarrow$ is that emitted earthward by clouds and the atmosphere. When R_{net} is defined for the surface, the two downward fluxes represent the shortwave radiation reaching the surface and the longwave radiation emitted by the atmosphere and the two upward (negative) fluxes represent the shortwave radiation reflected by the surface back to the atmosphere and the longwave radiation emitted by the surface.

Downward fluxes are determined largely by time and location, cloudiness, the atmospheric temperature, and the amount of greenhouse gases and particulates in the atmosphere. Upward fluxes are determined by the nature of the surface, which determines the surface albedo and the surface temperature. Albedo is particularly important because vegetation and soil moisture produce large variations in this term. It ranges from about 7–12% for tropical rainforests to 15–25% for grasslands. Surface albedo can be as high as 45–50% in extreme deserts with light-colored, sandy, or rock surfaces. Hence, the amount and type of surface vegetation cover, its seasonal phenology, and the degree of wetness of the underlying soil play a major role in determining the net radiation available to drive both surface and meteorological processes.

Most studies of the bioclimatic limits of specific vegetation types or formations have considered primarily temperature or potential evapotranspiration plus precipitation. Budyko (1986) reformulated this concept to link vegetation directly to net radiation (Figure 3.3) via a parameter referred to as the " dryness ratio." This ratio compares the amount of energy

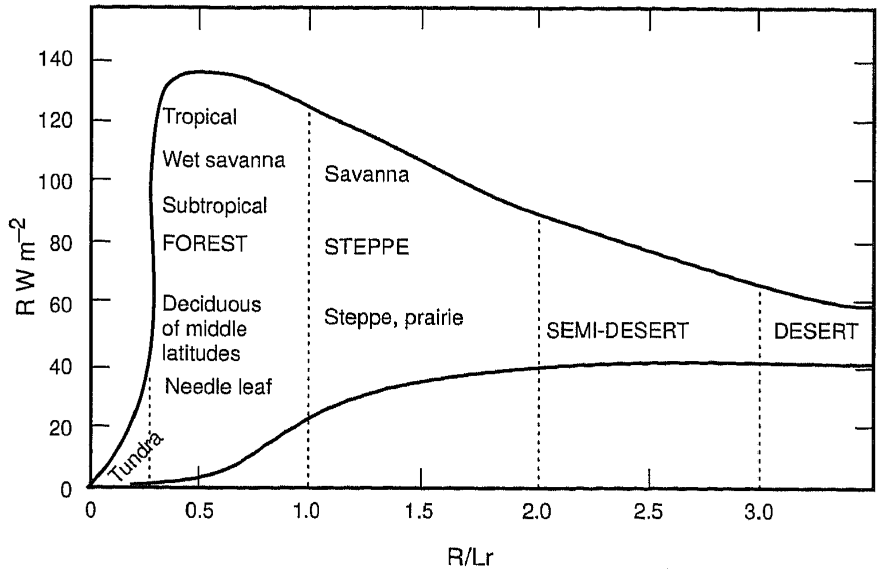

Figure 3.3 Budyko's (1986) concept of geographical zonality: the principal ecosystems as a function of net radiation and the dryness ratio (ratio between annual net radiation and annual average precipitation multiplied by the latent heat of evaporation). The upper and lower bounds of the figure are intended to represent the approximate limits of net radiation conditions in natural environments.

available to evaporate the annual rainfall (R_{net}) to the amount of energy required for this. The latter is calculated by multiplying the annual rainfall by the latent heat of evaporation. Because of limited sampling and imprecise measurements of net radiation, Budyko's analysis was quantitatively in error but conceptually valid. His diagram succinctly showed the bulk control of surface vegetation by climate and, via the R_{net} term, the feedback exerted by the vegetation.

Feedback of the vegetation to the atmosphere is illustrated by relating the radiation balance to the surface energy balance. For a thermal equilibrium to exist, net radiation must be balanced by other forms of heat transfer. This balance is expressed as:

$$R_{net} = LE + S \, , \tag{3.2}$$

where S is sensible heat transfer and LE is latent heat exchange, the product of the latent heat of condensation L and evapotranspiration E. The surface conducts heat to and from the atmosphere and ground, so that S includes both heat conducted to or from the subsurface and conduction and convection between the surface and atmosphere.

The partitioning of the surface contribution to atmospheric heating into sensible and latent heat depends on surface moisture and vegetation cover. The wetter the surface, the greater the utilization of heat for evaporation, the lower the temperature and the lower the Bowen ratio (ratio of sensible to latent heating). Water, with its high specific heat and thermal conductivity, also moderates temperature by absorbing radiation with little increase in temperature and by transfering heat to the subsurface. Similarly, vegetation moderates temperature by spreading absorbed radiation over a large and distributive surface area and by retarding loss of surface heat. Thus, a wetter, vegetated surface results in a higher proportion of energy used for latent heating. This is important because latent heat is released high in the atmosphere, where it can influence meteorological processes. When the balance is mainly via sensible heat, the heating is largely confined to the lower boundary layer and has little influence on large-scale meteorological processes.

Surface Water Balance

Because rainfall is the limiting factor in vegetation growth in most of the global drylands, the character of the rainfall, i.e., its distribution in time and space, has the ultimate influence on growth. This distribution is related to the critical question of what determines the overall availability of moisture to the ecosystem, i.e., the amount available for "storage" after the incoming precipitation is depleted by runoff and evaporation. It can be expressed by rearranging the basic equation for the surface water balance, such that:

$$\frac{dm}{dt} = P - E - N \ , \tag{3.3}$$

where m is soil moisture stored in the root zone, P is precipitation, E is evapotranspiration, and N is runoff. E includes both transpiration from plants and direct surficial evaporation from plant, soil, and water surfaces.

Runoff and evaporation are influenced by the intensity and duration of rainfall, which is a largely localized characteristic of rainfall. To some extent, these can be generalized on the basis of whether rainfall is convective or frontal in nature. In the case of convective precipitation, which is dominant in summer and in the tropics, rainfall is highly concentrated in both time and space. This results in intense runoff and high evaporation. In the case of frontal precipitation, which is dominant in mid-latitudes in winter and in high latitudes throughout the course of the year, the distribution is fairly uniform in space and individual precipitation events are protracted in time. This reduces both runoff and evaporation. Evaporation is also influenced by such factors as temperature and incoming radiation, so that it is higher in the tropics and lower in areas where the rainfall is associated with prolonged cloud cover.

The importance of these factors is illustrated by comparing two locations in Africa, one in the West African Sahel (Niamey) and one in southern Africa (Gaborone). The mean annual rainfall is similar in the two (559 mm for Niamey vs. 531 mm for Gaborone), but in southern Africa it is spread over a 5- to 7-month period, compared to 3 to 4 months in West Africa (Figure 3.4). The southern African location has much higher runoff but much lower evaporative loss and, therefore, higher soil moisture. Consequently, the vegetation productivity sustained by the annual total is much higher over southern Africa than over the Sahel.

The degree and type of vegetation cover also plays a critical but complex role in determining the proportions of precipitation which go into evapotranspiration and runoff. Plants intercept and retain water on their surface, retard evaporation from the soil beneath them, and remove soil moisture via transpiration. Vegetation moderates temperatures and reduces surface wind speeds, effects which reduce evaporation. Both above- and below-ground plant material retard lateral and vertical water movement, thereby reducing runoff and altering the soil texture and organic matter content to promote infiltration and soil moisture retention. These effects, which are sometimes compensatory, must be accounted for when determining the overall impact of a vegetation cover on moisture availability and loss.

In areas of relatively sparse vegetation cover, the nature of the soil is an important factor in determining moisture availability because of its influence on runoff and on moisture uptake

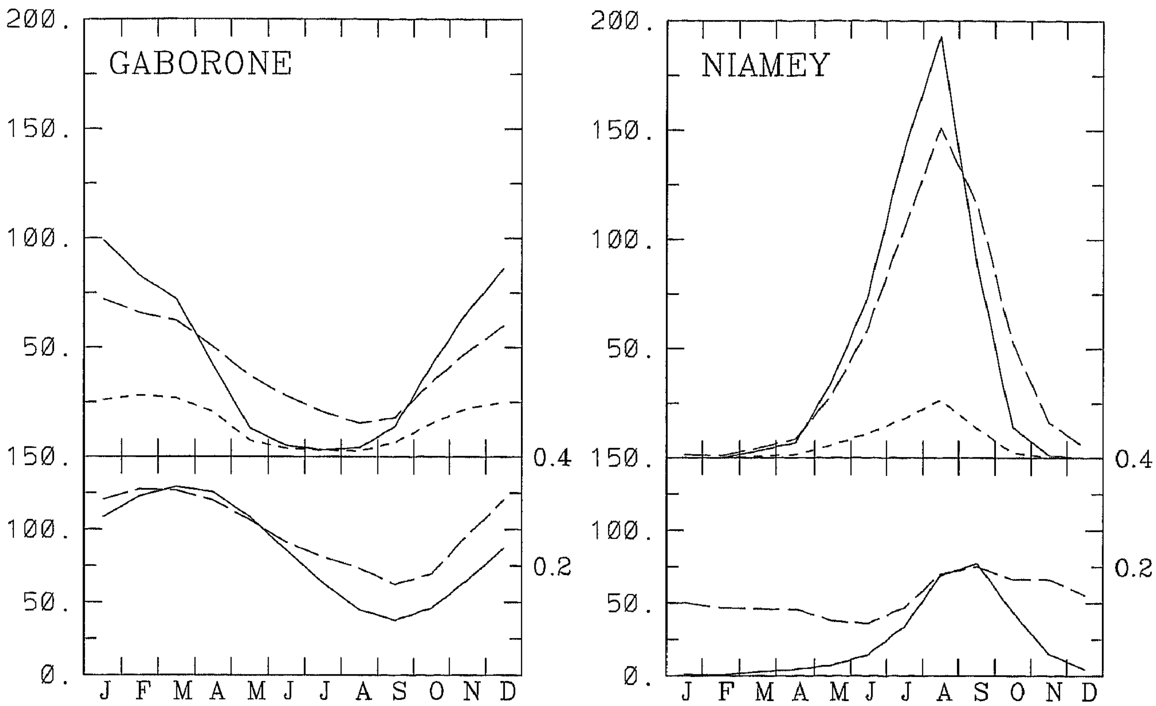

Figure 3.4 Top: Monthly precipitation (solid line), runoff (dotted line), and evaporation (dashed line) for Gaborone and Niamey, in mm. Runoff is multiplied by a factor of ten. Bottom: NDVI (right hand axis, dashed line) vs. soil moisture (left hand axis in mm, solid line). Note that background noise increases NDVI in the dry season in West Africa (Nicholson et al. 1997).

by the vegetation. Figure 3.5 illustrates this influence by showing the value of the normalized difference vegetation index (NDVI) as a function of rainfall and soil moisture for five soil types in Botswana. For these types, the higher the clay content, the more efficient the rate of growth per unit rainfall or per unit soil moisture.

COMMONALITIES AND INDIVIDUALITY OF THE GLOBAL DRYLANDS

Except for coastal deserts, the climates of drylands are characterized by low rainfall that is highly variable in time and space, severe moisture deficits during some or all of the year, localized rainfall events of short duration but usually high intensity, and generally thermal extremes with high diurnal and annual temperature ranges. These characteristics tend to be more extreme in low-latitude drylands. Semi-arid regions generally share these characteristics, but to a less extreme degree. Other commonalities of the semi-arid regions are a pronounced seasonality of precipitation, an abrupt transition between extreme dryness and an often brief rainy season, high sensitivity to climatic fluctuations and climatic change, and high risk of drought or flood.

The dissimilarities of these world's drylands are perhaps more notable. Figure 3.6 shows the thermal classification and seasonality of rainfall in low-latitude deserts. The degree to which temperature is a limiting factor for the ecosystem depends on whether the region is a

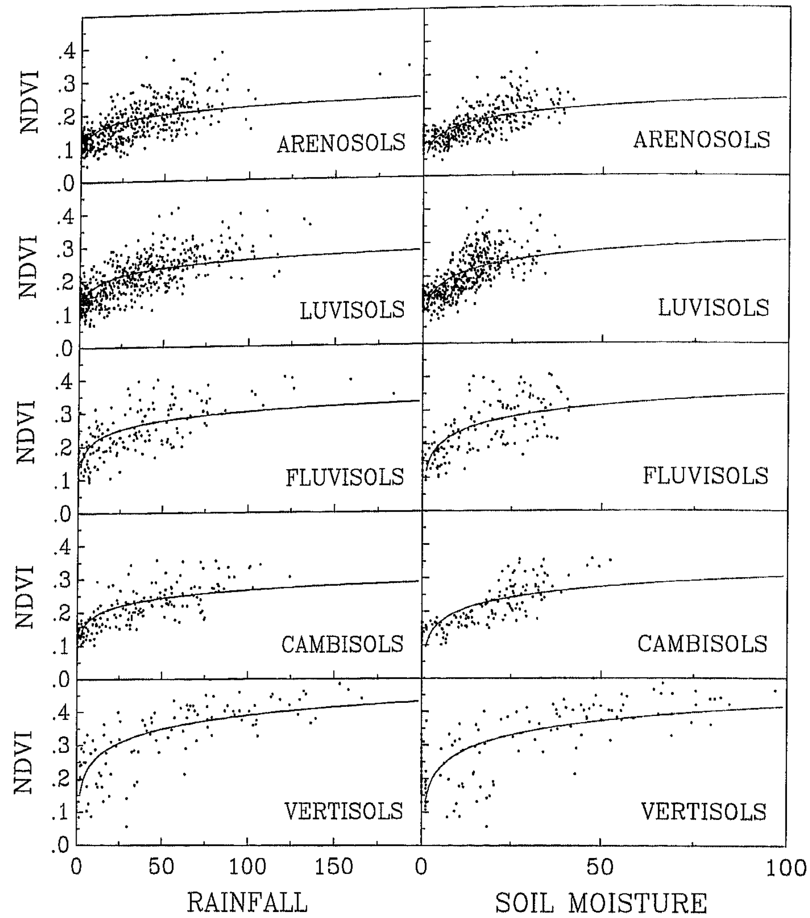

Figure 3. 5 Scatter diagrams of monthly normalized difference vegetation index (NDVI) versus rainfall and monthly NDVI versus soil moisture for five soil types in Botswana (from Farrar et al. 1994). Rainfall is a monthly value averaged for the concurrent plus two previous months; soil moisture is for the month concurrent with NDVI; the line of regression is indicated.

"cold," "warm," or "foggy" desert. The latter class represents the coastal deserts, which are characterized by a stable temperature regime with little change during the season or between day and night. Temperature is a much more important consideration in cold deserts. Another dissimilarity is the seasonality of rainfall. In some cases, there is a distinct winter or summer rainfall regime, but many drylands are areas of transition between these regimes. In such regions, the seasonal cycle of rainfall is complex and irregular or may be completely lacking.

This seasonality has a profound effect on the ecosystem. The drylands of the southwestern United States provide one example. The distinct vegetation types of the Mojave, Sonoran, and Chihuahuan deserts are functions of their respective winter, transition season, and summer rainfall regimes. Vegetation is particularly distinctive in the Sonoran desert, which experiences two meager rainy seasons during the course of the year (MacMahon and Wagner 1985). The seasonality of rainfall in South Africa, which has both winter and summer rainfall areas, also has a profound influence on the dominant biomes (Ellery et al. 1991).

Additional dissimilarities of the various dryland regions are related to the factors governing the interannual variability of rainfall, and hence governing the sensitivity of the region to large-scale climatic change. It is important to bear in mind that factors forcing the variability

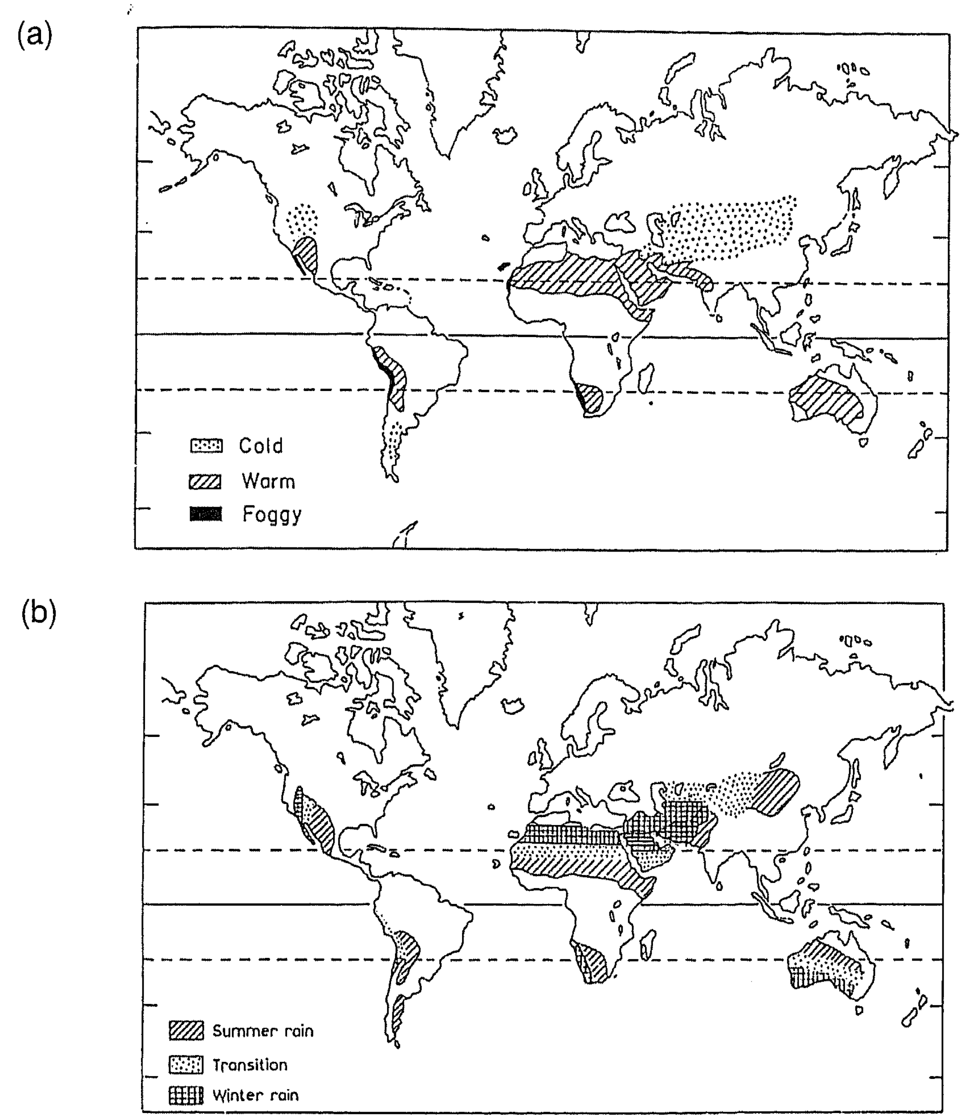

Figure 3.6 (a) Classification of desert regions into cold, hot, and foggy deserts (from Shmida 1985). (b) Deserts with summer, winter, and transition season rainfall (from Evenari 1985).

may be completely different than those determining the overall mean climate. In East Africa, for example, the patterns of aridity and rainfall are strongly influenced by local factors, such as topography and its influence on local winds. However, the same factors influence the interannual variability in both arid and humid regions, and the year-to-year fluctuations of rainfall are almost identical throughout the region, although the mean state of climate is so diverse that even the rainfall seasonality varies throughout the region. Thus, it is not possible to generalize the climatic response of, say, winter rainfall or summer rainfall drylands or low-latitude versus mid- or high-latitude drylands. Other dissimilarities include the proximity to and degree of influence by maritime effects, the location with respect to the climatic equivalent of ecotones, and the precipitation gradients within a region. The latter two differences strongly influence the degree to which a region is sensitive to climatic change.

Some of these dissimilarities are illustrated by comparing the interannual variability of rainfall during the twentieth century for the Sahel and for Botswana in southern Africa (Figure 3.1). Time series of normalized rainfall departures are given for the two regions shown in the inset map. In both cases, the region's dry character is largely dictated by its subtropical location and its location with respect to a subtropical high pressure cell. Nevertheless, clear differences are apparent in the time scales of the variability, the degree of interannual persistence, and the magnitude of the variations. El Niño (indicated by shaded bars) is generally associated with abnormally low rainfall in Botswana but appears to have no consistent impact in the Sahel, although both are regions of summer rainfall.

The Sahel region is particularly sensitive to climatic change because the rainfall gradients are so high in the region. In much of the region, thirty-year means declined by 30% to 40% between 1931–1960 and 1968–1997. This decline was related to a mere one degree latitudinal displacement of the rainfall isohyets (Nicholson et al. 2000), which might imply about the same degree of shift of the general circulation features governing the region's climate. On the other hand, there is a substantial body of evidence that land–atmosphere feedback modulates the interannual variations in the region's rainfall (Nicholson 2000). This, together with the location in a continental interior, may be one of the reasons that the impact of El Niño is relatively weak.

FEEDBACK BETWEEN DESERTIFICATION AND CLIMATE

The components of climate as a driver of desertification are essentially those that reduce its potential to sustain natural vegetation and crops beyond some critical threshold. This includes any aspect that reduces the amount or regularity of moisture supply regionally or locally. This can include temperature, especially in the higher latitudes, because increased temperature will likely increase evaporation. However, in most dryland regions the interannual variability of rainfall is considerably greater than that of temperature. Hence, discussion here will focus on the rainfall regime.

The nature of rainfall variability can include changes in the overall mean or changes in the temporal and spatial distribution that may or may not be accompanied by changes in the mean. These include increased variability in time, changes in the number and intensity of individual rainfall events, changes in the spatial structure of rainfall (i.e., in the patchiness of individual events), changes in the duration of the season, and changes in the seasonal cycle. All of these will influence the availability of moisture in both space and time. Reduced availability stresses the growing vegetation and crops and makes it more difficult to override degradation by human misuse of the land. It also exacerbates the processes of degradation, such as erosion and salinization. Once degradation occurs, the potential for recovery when moisture availability increases is diminished. This feedback can destabilize the ecosystem (Schlesinger et al. 1990) and might also influence climate in ways that reinforce the degradation.

Numerous studies have examined the potential feedback that desertification or other changes in the surface vegetation cover might have on climate (see reviews in Nicholson 1988; Entekhabi 1995; Nicholson et al. 1998; Nicholson 2000). A few observational studies have examined such characteristics as surface albedo and temperature, making measurements on degraded and nondegraded areas of land. The majority of meteorological studies of this issue have used numerical modeling and are based on an oversimplified model of the associated processes. In these, desertification is generally represented by replacing an existing ecosystem with a desert. Both observational and modeling studies have focused foremost on the question of surface albedo (e.g., Charney et al. 1975, 1977; Chervin 1979; Sud and Fennessy 1982). This is likely a consequence of the late Jule Charney's hypothesis that drought can result from the albedo change associated with denuding the land. Many models have also examined the impact of changes in evaporation and/or soil moisture, sometimes in conjunction with changes in albedo (Sud and Fennessy 1984; Sud and Molod 1988; Xue 1991; Xue and Shukla 1993; Dirmeyer and Shukla 1996; see also Xue and Fennessy 2002; Leemans and Kleidon 2002).

Although these models almost universally agree that desertification can reduce rainfall, they are unrealistic for reasons discussed further in the following section. Far more applicable are models that focus on processes. These have produced several notable results (see review in Nicholson 2000) that underscore the need to focus on the impact of desertification on soil moisture retention, which implies the need to understand the impact of desertification on soil texture and structure better, the most important factors in moisture retention. The influence of soil moisture depends on the areal extent, magnitude, and persistence of the initial anomaly (all of these characteristics can be locally influenced by desertification; Fennessy and Shukla 1996). The retention time is determined by the ratio of field capacity (which can also be influenced by processes of desertification) to potential evapotranspiration (PET) (Delworth and Manabe 1993). The model of Koster and Suarez (1996) shows that precipitation variability is inversely related to the time scale of soil moisture retention. Model simulations for the West African Sahel have also shown that interactive soil moisture influences five-day Sahel rainfall forecasts (Rowell and Blondin 1990) and that soil moisture can be the dominant forcing, overriding that of sea-surface temperatures in some years (Rowell et al. 1995).

Recent modeling studies also reinforce the conclusion stated earlier, namely that the connections between climate and desertification are regionally specific. The strength of the soil moisture effect discussed above is dependent on the availability of a nearby moisture source, hence coastal versus continental location; on the strength of the regional dynamic circulation of the atmosphere; and generally on whether the region is tropical or extra-tropical.

APPROPRIATE SCENARIOS FOR MODELING THE IMPACT OF DESERTIFICATION ON CLIMATE

The common perception of desertification is that of an "advancing desert," a living environment becoming irreversibly sterile and barren. This image was fueled by literature with such titles as "Lethal spread of the sands" (Gwynne 1977) or "Spreading deserts — The hand of man" (Eckholm and Brown 1977) that appeared in the wake of the 1977 UN Conference on Desertification. This serves as the general scenario for modeling studies (e.g., Charney et al. 1975; Xue and Shukla 1993; Dirmeyer and Shukla 1996). Large-scale changes in surface albedo and/or soil moisture are typically used to represent the desertified landscape. However, this scenario is generally incorrect (Mainguet 1991; Warren 1996; Prince et al. 1998; Helldén 1991; Thomas and Middleton 1994). Studies in the Sudan by various scientists at the University of Lund showed, through a combination of field work and analysis of satellite photos, that there was neither a systematic advance of the desert or other vegetation zones nor a reduction in the vegetation cover (Helldén 1984; Olsson 1985; Ahlcrona 1988). Instead, degradation and replacement of forage with woody species was apparent. There was no evidence of a systematic spread of desertified land around villages and waterholes or of reduced crop yield due to cultivation of marginal or vulnerable areas. The noted changes in surface vegetation were largely a response to drought, a conclusion also reached by Akhtar-Schuster (1995) for the Sudan, and by Tucker et al. (1991) as well as Tucker and Nicholson (1998) for the Sahel as a whole (see Figure 3.7).

Particularly problematic is the albedo scenario proposed by Charney (1975) and Otterman (1974). Charney suggested that overgrazing in the Sahel bared high-albedo soils, resulting in an increase in surface net radiation and hence increased radiative loss to space. Accordingly, this required increased subsidence to produce a thermal equilibrium in the region, thus reducing rainfall and resulting in the drought of the early 1970s. Over the Sahel, where numerous sources (e.g., Glantz 1977) have claimed that desertification is intense, there has been relatively little albedo change over the time period for which measurement is possible, from the late 1960s to the present. Albedo has been relatively stable, despite strong interannual variability of rainfall, progressive intensification of drought, and intensification of human impact on the land surface in this region. Its variations have been on the order of 0.02 to 0.03 during the period 1983–1988 (Nicholson et al. 1998). During 1967–1973, its variations were roughly 0.05 during the dry season and 0.10 during the wet season; however, changes were neither progressive nor permanent (Courel et al. 1984).

The temporal sampling of the various assessments of albedo over the Sahel is admittedly poor and the studies cited above are not conclusive. However, recent analyses by Nicholson et

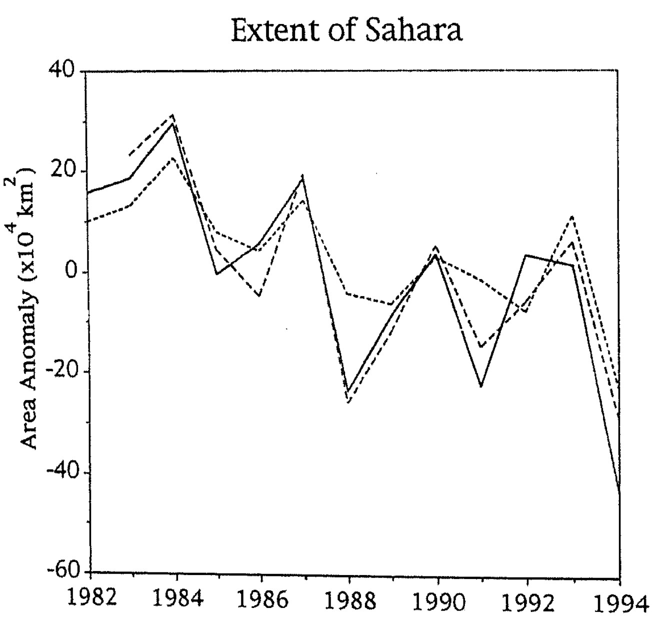

Figure 3.7 The extent of the Sahara Desert, calculated as the area between the 200-mm isohyet and 25° N as assessed from rainfall stations (solid line), Meteosat data (long-dashed line), and the normalized difference vegetation index (NDVI) (short-dashed line). Area is calculated with respect to the mean area during the period 1980–1995 (from Nicholson et al. 1998).

al. (1998) and Ba et al. (2001) strongly suggest that albedo is not a major factor. The results of these studies, partially depicted in Figure 3.8, collectively show that (a) surface albedo is largely linked to vegetation biomes over Africa, (b) seasonal changes in albedo over the Sahel are small and commensurate with changes in surface moisture, and (c) reduction in vegetation cover (e.g., in the woodlands) can actually decrease surface albedo. The results also suggest that in the grasslands surface moisture is the overriding factor determining surface albedo and that major changes in surface albedo are unlikely unless there is a shift from, say, woodland to grassland. Thus, desertification in this region is unlikely to have a major impact on surface albedo.

In simulating the impact of desertification on climate, a more appropriate scenario would include local regions of land degradation by specific processes such as erosion or salinization. In meteorological terms, this translates to a patchwork of surface characteristics such as soil moisture, temperature, and increased atmospheric aerosol content. These are far more likely to influence meteorological processes than the associated changes in surface albedo. These aspects can be important because there is observational evidence of their influence on local precipitation processes as is now discussed.

Numerous modeling studies have indicated that such a patchwork pattern of surface heating and moisture can influence individual weather systems (e.g., Anthes 1984; Pielke and Avissar 1990; Pielke et al. 1991). More importantly, one well-controlled observational study in the Sahel came to a similar conclusion. During the HAPEX–Sahel experiment in 1992, a high-density rain gauge network showed that plots initially receiving high rainfall from the first major rainfall event were more likely than surrounding areas to receive high rainfall during each subsequent event (Taylor and Lebel 1998). Figure 3.9 shows the results for one pair of stations, but a large number of station pairs was analyzed and all indicated similar patterns.

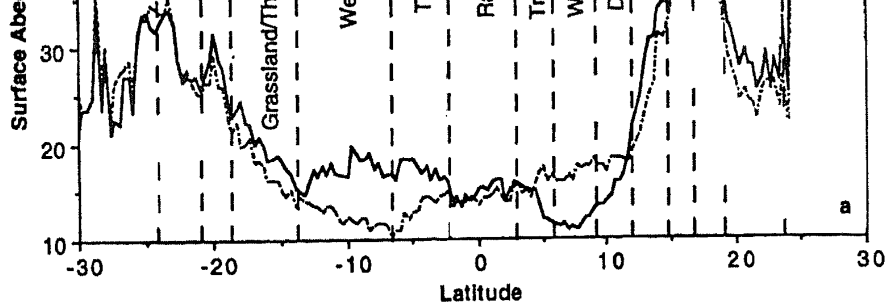

Figure 3.8 Latitudinal variations of surface albedo over Africa between a wet (August, 1987) and dry season (January, 1988) along a north–south transect at 19°E (from Ba et al. 2001). Vertical lines approximately delimit the vegetation types.

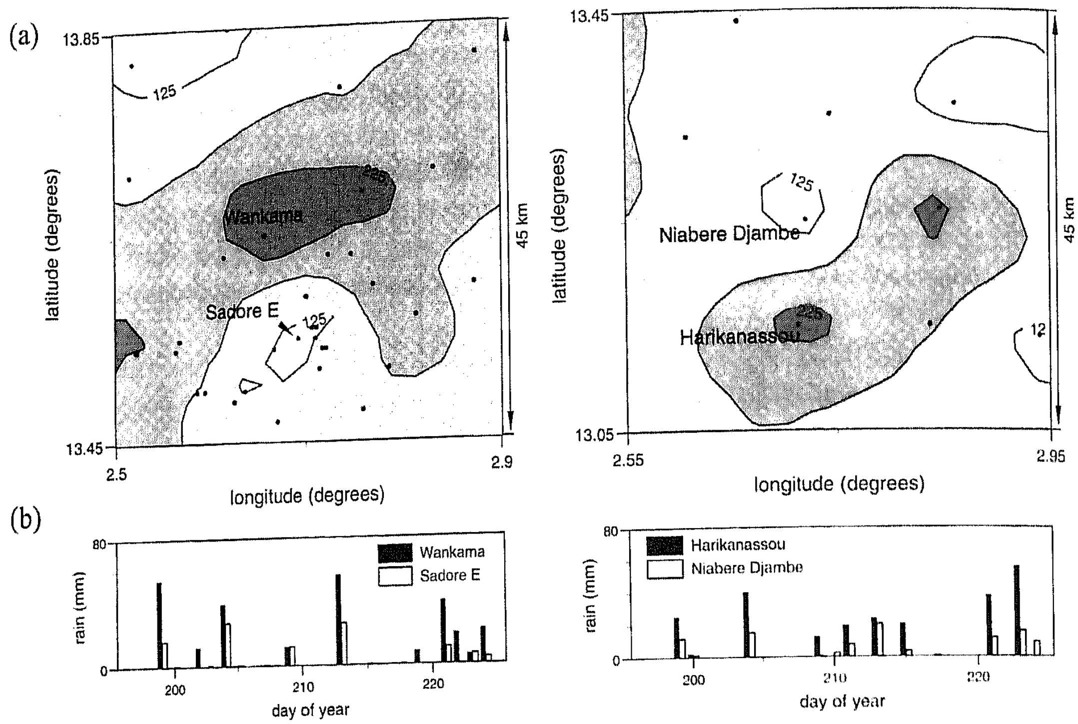

Figure 3.9 (a) Accumulated 30-day rainfall totals in the region and (b) daily rainfall at two pairs of gauges during the HAPEX–Sahel experiment, July 14 to August 12, 1992. In both cases, the gauges are ~10 km apart, and the climatological gradient is small and in the opposite direction of that shown (from Taylor and Lebel 1998).

Relatively new findings from such instruments as the TOMS (Total Ozone Mapping Spectrometer) demonstrate the importance of better understanding the relationship between aerosol generation and desertification. The well-known "dust storms" characterizing drylands are shown not to be localized phenomena, but rather regional-scale events that can travel great distances. There are also specific, well-defined source regions, most of which lie in semi-arid lands rather than true deserts. Roughly half of the global aerosol loading is produced by regional sources in West Africa (Prospero et al. 2002).

More importantly, there is direct but preliminary evidence of the impact of the aerosols on precipitation processes. Using data from the recently launched TRMM (Tropical Rainfall Measuring Mission) satellite, Rosenfeld (1999) has shown that smoke from forest fires essentially shuts off the warm-rain process, which is important in much of the tropics. Thus, clouds must develop to heights above the freezing level in order to precipitate. Aerosol generation is intense throughout the global drylands and appears to be accelerated by climatic change (e.g., N'tchayi Mbourou et al. 1998) and by human disturbance of the soil (Prospero et al. 2002). Over West Africa, aerosols have a major influence on the radiation balance and on the easterly waves associated with rain-bearing disturbances; they also influence the radiation balance globally (Nicholson 2000). Thus, there are numerous mechanisms by which aerosols can influence climate. Hence, an understanding of the relationship between meteorological processes and desertification should also focus on processes related to erosion and dust generation. These effects are influenced by the nature of the surface vegetation cover and the soil, as well as by human disturbance.

SUMMARY AND CONCLUSIONS

The Earth's drylands experience a variable climate that can promote desertification by producing periodic but severe moisture shortages. Many of these regions are also very sensitive to climatic change, so that there is much concern about these regions in the context of global change. A particularly relevant question is whether global change might make the Earth's drylands more susceptible to desertification. Unfortunately, this is a question that can only be answered via numerical simulations of climate.

At the current time, climate models are not sufficiently advanced to be able to answer this question with any great confidence. However, past studies of climate and desertification in the drylands can provide some guidance for relevant modeling studies.An appropriate model of desertification must be used. The one most commonly used in the past, i.e., of productive lands replaced by desert, is invalid. Aspects which are probably most important involve the large-scale generation of aerosols and changes in the water retention capability of the land as a result of changes in surface soils and vegetation.

The drylands where desertification is of concern share some features but are extremely diverse with respect to climatic processes and variability. They are particulary diverse with respect to the causes of and sensitivity to climatic change. The links between climate and desertification are also regionally specific. For this reason, global-scale atmospheric models cannot adequately address regional questions. Instead, analytical studies that focus on understanding the causes of climatic variability on a regional basis and numerical simulations via regional climate models coupled to global models must be conducted to understand climatic change, as well as its links to desertification. This is the new frontier in climate modeling. The incorporation of biogeophysical feedback into these regional models is a critical aspect of such simulations.

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