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Late Quaternary Environmental Changes and Human Interference in Africa

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1. Introduction

The African continent has been subjected to several changes in its environmental conditions in the past. These changes have affected vegetation patterns, soil development and earth- surface processes. Repeated change has caused the development of a complex pattern of inherited features in the present-day landscape that regulate its susceptibility towards modern change in environmental conditions. Since the late Pleistocene and early Holocene, human interference in ecodynamics has increased dramatically. Humans have been altering the environment since they first controlled fire and invented agriculture. However, the exponential growth of population in the last 100 years has brought with it an accelerated rate of landscape degradation. The superimposition of anthropogenous sources of interference and climatic factors has often changed the type and intensity of earth-surface processes. This results in an imbalance and often triggers an array of self-reinforcing processes. These processes operate on different spatial scales and in different time frames and are discussed in chapter 2.

In many areas of Africa, intensified use of land has induced serious soil erosion. Particularly in the semi-arid tropical and subtropical zones of Africa, soil-erosion processes are supported by the variable nature of rainfalls, the strong seasonal contrasts in the availability of moisture and the poor vegetation cover and soils and sediments, which are characterised by a high level of erodibility. Chapter 3 provides an attempt to summarise some of the processes and impacts which are associated with soil erosion in Africa. Extreme events played and play an important role in the African morphodynamic system and may pose a threat to humans. The spatial and temporal distribution of extreme events and factors which determine the magnitude, frequency and the impact of such events are discussed in chapter 4. The increasing demand for arable land has resulted in the enlargement of those areas affected by biomass burning. Chapter 5 provides an overview of the impact of savanna fires on the vegetation and the emission of greenhouse gases into the atmosphere.

The objective of this paper is to present a synthesis of the recent research on the influence of human interference on earth-surface processes and the differing reaction paths in the African landscapes.

2. Environmental changes in Africa

2.1 Long-term environmental change

In the course of the Cenozoic period, the African continent experienced several phases characterised by very different environmental conditions. On a time scale of 10^6 to 10^8 years, these changes were associated with the break-up and fragmentation of the ancient continent Pangaea into individual continents in the Mesozoic. The plate tectonic motion of the continents initiated a number of different processes, including the drift of continental masses into polar areas, the uplift of the Tibetan Plateau, the formation of new mountain belts and the establishment of a new ocean circulation system. These changes resulted in a climatic system that was completely different from that of the Cretaceous and early Cenozoic periods (Haq, 1981; Seibold and Berger, 1995; Goudie, 1999; Skinner and Porter, 2000). In the Neogene, the new position of the continents supported the growth of the Antarctic ice sheet and the drop in sea-surface temperatures. This induced a trend towards drier climatic conditions. In the Quaternary the changes culminated in relatively rapid climatic fluctuations. The progression of different climates brought with it changes in the vegetation cover and in the denudation rates. Further processes which influenced the Cenozoic evolution of Africa are slow epirogenic crustal movements, which were responsible for the development of large basins and swells (Summerfield, 1999, Römer, 2004). Periods of local uplift produced elevated continental margins, and intense rift processes promoted intense volcanism and block faulting ((Petters, 1991; Summerfield, 1999). The cumulative effects of these processes ranged from the development of new drainage systems, the rejuvenation of old erosion surfaces, to the development of uplifted and highly dissected plateaux along the continental margins, and the relatively young volcanic areas and uplifted block-faulted mountain zones along the rift valleys in eastern Africa. The different landscapes tend to respond in different ways and at different rates of environmental change. The distinctive response of the landscapes and geomorphic forms, however, depends not only on the lithological and structural conditions in the different geotectonic domains. The strength, propagation and prolongation of the response is also modulated by the different coupling strength between hillslopes, major river systems and oceans (Wirthmann, 2000; Römer, 2012).

Africa encompasses rain forest, savanna, desert and Mediterranean environments. The present pattern of bioclimatic zones is the result of the climatic changes that occurred after the Pleistocene. However, large areas of Africa are covered with sediments and soils that are derived from the Pleistocene period. The sediments and soils are an integral part of the present ecosystem and exert influences on the rate at which earth-surface processes progress, the physico-chemical processes in the soils and the distribution of the vegetation.

2.2 Climatic change in the last millennium

Even within a timescale of 10^2 to 10^3 years, the environmental conditions in Africa are highly variable. In southern Africa, the climate was as warm or warmer from 900 to 1300 (medieval warm period) than at present, but became colder than present from 1300 to 1810 (Tyson and Partridge, 2000). The transitions from the medieval warm period to the period of the "Little Ice Age" (1300 to 1850), and from the end of the "Little Ice Age" to the recent period are well documented in southern Africa. Historical reports, studies of lake levels and dendroclimatological analyses show that it is likely that some of the climatic changes at the end of the "Little Ice Age" occurred synchronously in the northern and the southern hemisphere. According to Nicholson (1999, p. 69), a trend towards increasing dryness is indicated in the droughts that occurred from the 1780's until the 1830's for the northern and the southern hemisphere. However, in the Sahelian zone, the droughts appear to have lasted for two decades whilst in southern Africa they lasted only for a few consecutive years (Nicholson, 1999, p. 80). In southern Africa, the wet to dry conditions are related to tropical

easterlies and westerly disturbances. The strengthening of the tropical easterlies is associated with warm, wet spells and high rainfall amounts in the summer rainfall regions of southern Africa. Conversely, dry spells result from the more frequent westerly disturbances (Tyson and Partridge, 2000). According to this model, atmospheric adjustments resulting in cooler conditions imply a decrease in rainfalls over summer rainfall areas of southern Africa whilst a warming caused by adjustments in the tropical circulation is associated with a general increase in rainfalls. In the Sahelian zone, the process of aridification during the transition from the eighteenth to the nineteenth century appears to have been related to a later northward advance of the ITCZ. This implies a shorter rainy season in the Sahel but wetter conditions on the coast of Guinea (Nicholson, 1999). Based on reports of prevailing winds and hydrological records, an earlier advance of the ITCZ seems to explain the wetter conditions in the Sahel and West Africa during the seventeenth and eighteenth century (Nicholson, 1999, p.69). However, even these wet periods were interrupted by severe droughts, which lasted one or two decades (Nicholson, 1999). Some pronounced events appear to coincide with global climatic trends in the atmospheric circulation pattern that are associated with the "Little Ice Age". The severe droughts of the 1820's and 1830's in Africa correspond with the last intensification of the "Little Ice Age" and may be related to a global period of anomalous circulation (Nicholson, 1999). The cooler and drier conditions in Southern Africa during the "Little Ice Age" are clearly documented from several tree ring analyses (Tyson and Partridge, 2000).
More recent changes are associated with links between sea surface temperatures and rainfall amounts over tropical Southern Africa. Studies of Richard et al. (2001) imply a higher likelihood between El Niño events, higher water temperatures in the Indian Ocean and reduced rainfall amounts over tropical Southern Africa since 1970. A similar direction of change is expected by Landman and Mason (1999) for Namibia and South Africa. According to their investigation, wetter conditions in north-eastern Southern Africa and northern Namibia tend to be associated more often with warm events in the Indian Ocean, whilst prior to the late 1970's, there was a stronger correspondence between warm events in the Indian Ocean and dry conditions over Namibia and South Africa.

2.3 Environmental change and human interference

In Africa, the superimposition of environmental changes and human interference is not a recent phenomenon. In the savannas, humans appear to have used fire for over 1.5 million years (Gowlett et al., 1981, Goldammer, 1993). Colluvial deposits containing artefacts of the late Pleistocene and early Holocene periods point towards an increase of erosion resulting from deterioration of the vegetation cover (Lewis, 2008) caused by human interference. Extensive woodland clearance is also documented from Tanzania, where charcoal production for iron smelting in the last 900 years has led to an increase in soil erosion (Schmidt, 1997, Eriksson et al., 2000). Agriculture and fires that were ignited by humans appear to have played a key role in the development of the vegetation pattern and in the composition of the plant communities in most parts of the savanna areas.
The highly variable environmental conditions in Africa have, at all times, increased the pressure to expand the utilisation of land into areas which do not support large populations on a subsistence basis. The semi-arid tropics of Africa, in particular, have been subjected on several occasions to marked environmental degradation (Seuffert, 1987, Mensching, 1990). As the amount of summer rainfall in these areas determines the quantity of forage and crop yield, rainfall amounts also determine the economic base of the human population. An

example of the complex interaction of the various processes is the Sahel zone, where the rapid growth of the population over the last century coincides with an enlargement of areas for agriculture (Seuffert, 1987). Varying rainfall amounts and consecutive years with low rainfall amounts in conjunction with intensified cultivation methods, vegetation clearance and increased livestock husbandry resulted in a degradation of many areas. In the Sahel zone, the carrying capacity of the land was exceeded and the southward expansion into more humid areas caused further environmental degradation (Mensching, 1990).

3. Soil erosion

3.1 Factors contributing to soil erosion

Land degradation resulting from inappropriate cultivation practices, high grazing intensities and clearance of the vegetation is often associated with an increase in erosion by water and wind. In the tropical and subtropical areas of Africa, the effects of soil erosion appear to be correlated with a decline of the productivity of the cultivated land and of the per capita ratio of cultivated area (Beckedahl, 2002). Soil erosion may induce irreversible damage to arable land, and tends to produce ecologically unstable landscapes and socioeconomic problems (Scoones et al., 1996). Human interference contributes to soil erosion in a direct and indirect way by modifying the topography for buildings, clearance of vegetation for pasture and cultivation, or by compacting the soil by the use of machines. Although the extrapolation of anthropogenously induced soil loss over longer periods remains a challenge, most authors agree that land degradation by soil erosion has affected large areas of Africa (Reading, et al., 1995; Scoones et al., 1996; Valentin et al., 2004; Bork, 2006; Dahlke and Bork, 2006; Nyssen et al., 2004). Beckedahl (2002) states that about 85% of the area of Africa north of the equator is potentially endangered by soil erosion and that the area of the arable land has decreased from 0.3 ha/person (1986) to about 0.23 ha/person (2000). A further reduction of the arable area per person to 0.15 ha is predicted for the year 2050 (Beckedahl, 2002, p.18).

The process of soil erosion is a function of a number of interrelated factors. These include the climatic conditions, the soil, the relief, the density and type of the vegetation and the land use and agricultural techniques. The on-site effects of soil erosion range from soil loss to a decrease in nutrients in the soil, to changes in the water balance and runoff. Off-site effects are the pollution of fresh water by delivering eroded, nutrient and heavy metal-laden sediments to rivers and lakes (Zachar, 1982, Reading et al., 1995, De Meyer et al. 2011).

Soil erosion is not only an African problem, but environmental and socioeconomic conditions provide a specific set of factors which appear to be different from those of other continents. Extensive areas of Africa encompass pericratonic and cratonic terrain-types. These areas are characterised by old landscapes, which presumably originated in the late Mesozoic or early Cenozoic periods (Wirthmann, 2000; Römer, 2007). Deeply weathered rocks, escarpments and deeply incised valleys with steep valley side slopes feature high susceptibility to soil erosion and slope failure. In some of these areas, weathering mantles and soils have survived for millions of years. This has supported the depletion of the weathering mantles and the development of sandy materials that are prone to erosion processes (Dingle et al., 1983; Areola, 1999; Partridge and Maud, 2000). In semi-arid areas of Africa such as the Sahel, the Sub-Sahelian regions or the Kalahari large tracts are covered by sand dunes and sediments that were mobilised during the Pleistocene. These materials are generally liable to soil crusting and erosion by both water and wind (Valentin, et al., 2005).

High rates of soil erosion are associated with laterised weathering layers and soils, which are characterised by clay-enriched argic horizons with weak microstructure (e.g. Acrisols) or low aggregate stability (Lixisols).

Studies from several areas in Africa imply that even under undisturbed conditions natural erosion rates may exceed the rate of soil delivery (Shakesby and Whitlow, 1991; Idike, 1992, Braun et al., 2003). However, comparative measurements are rare. In disturbed areas, soil erosion rates are estimated to range from 0.5 to 110 t ha^{-1} a^{-1}, largely depending on the eksystemic and ensystemic conditions at the site, the intensity of disturbance and the measurement methods (Reading et al., 1995; Stocking, 1995; Beckedahl, 2002). These rates appear to exceed the natural erosion rates by an order or several orders of magnitude (Thomas, 1994; Reading et al., 1995). The predicted increase in soil erosion in future is considered to have devastating consequences on the soil system, the productivity of arable land and the natural habitat.

3.2 Soil erosion and environmental conditions

Soil erosion by water is the result of a combination of several processes, which increase in intensity with the amount, duration and intensity of the rainfall events. The processes involved in soil erosion include splash, interrill, rill and gully erosion. Another group of erosion processes includes gravitative processes such as landslides and erosion by wind.

Soil erosion in Africa is not a phenomenon of a distinctive physiographic zone, though some factors support soil erosion. Soil erosion has been documented from nearly all physiographic zones of Africa (Rapp, 1976; Elwell and Stocking, 1982; Biot, 1990; Mensching, 1990; Bork, 2004). A characteristic feature of the tropical and subtropical regions of Africa is that rainfall varies in amount, intensity and structure within a year, and between years and decades (Seuffert, 1987; Hulme, 1999; Tyson and Partridge, 2000). In semi-arid tropical areas, the decrease in the annual rainfall is often associated with an increase in the intensity and variability of rainfall events and a decrease in vegetation cover of natural and cultivated areas (Seuffert et al., 1999). While rainfall intensity provides the energy for soil erosion, the vegetation cover controls the amount of rainfall that reaches the surface. Important parameters are the high spacing and the structure of the vegetation, the density of the under storey vegetation, and the ground cover. A change in the vegetation cover may result in a marked increase in the rate of soil erosion. According to Lal (1998), rainfall intensity determines the distribution of raindrop sizes while interception is a function of the total rainfall amount and of the high spacing and the structure of the vegetation cover. A decline in the under storey vegetation and the ground cover is associated with a decrease in the litter at the soil surface. This causes a decrease in surface roughness. The decline of the under storey vegetation results in a change in the size distribution and the terminal velocity of raindrops. This involves an increase in the impact energy of the raindrops as a function of the height of the canopy. The decreased roughness at the soil surface and the increase in the impact energy of raindrops promotes splash erosion and the generation of overland flow. A further effect of a decline in the vegetation cover is the depletion of organic matter at the soil surface. This enhances the physical, chemical and biological degradation processes in the soils and results in a reduced stability of the soil structure. The reduced stability of the soil structure increases the susceptibility to soil-crusting, which, in turn, affects the runoff production and susceptibility to soil erosion (Valentin et al., 2005). The decrease in the density and depth of the roots weakens the mechanical reinforcement of the soils as the root-binding effects and the apparent cohesion are reduced (Greenway, 1987).

As soil erosion is caused by individual rainfall events, the total amount and the temporal distribution of rainfall intensity are more important than the mean annual rainfall. The rate of overland flow generation depends on the infiltration rate which is a function of the physical characteristics of the soil, the vegetation cover, the relative relief, the slope gradient, the roughness of the surface and the moisture content of the soil (Bork, 2004). Susceptibility of soils to soil erosion is a results of several interacting components. At microscale level, these include the content of organic matter and physical properties such as grain-size distribution, mineralogical composition, water-holding capacity and shear strength (Zachar, 1982; Grabowski et al., 2011). The amount of water-stable aggregates appears to control the generation of overland flow and the detachability of soil aggregates (de Vleeschauwer et al., 1978). Raindrop impact during high-intensity rainstorms can have profound effects on soils with silty and clayey composition. The impact of large raindrops promotes the compaction of the soil. Small particles that have been moved by the impact block the soil pores and air is imprisoned in the pores. This impedes the infiltration of water into the soil. Even small changes in the textural composition of the surface soil can induce change in susceptibility to erosion. In the Sahel zone, the deposition of dust and the colonisation of the soil surface with blue-green algae during fallow periods promoted the development of soil crusts (Valentin et al., 2004).

Soils and deposits such as colluvium that are characterised by a high exchangeable sodium percentage are highly prone to soil erosion. Susceptibility to erosion is associated with highly expansive clays (Botha and Partridge, 2000; Grabowski et al. 2011). Clay minerals tend to adsorb more water at a high sodium-adsorption-ratio and water infiltrating between the clay units causes an expansion of the clays. Consequently, the clay minerals are pushed apart. The expansion reduces the attraction between the clay particles. Dispersive soils are prone to piping processes, when the seepage water causes the development of subsurface drainage conduits (Bryan and Jones, 1997). Piping may support the development of slope failures by undermining the slope base and the collapse of pipe roofs is frequently associated with an increase in gully erosion (Heinrich, 1998; Singh et al., 2008).

On a hillslope and drainage basin scale, the response to a change in the eksystemic components is modified by the steepness of the hillslopes, the coupling strength of hillslopes to major rivers and by the density and degree of development of the drainage net. On a regional scale lithological and structural controls, neotectonic activities and rainshadow effects caused by large escarpments or mountain chains may exert considerable influence on the rate of soil erosion. Recent studies of Fubelli et al. (2008) on the Ethiopian highlands indicate that increased rainfalls and neotectonic activity are likely to be responsible for the high rates of river incision and the frequent occurrence of landsliding, and Singh et al. (2008) emphasise the association between palaeolandslides and active seismic zones in the KwaZulu-Natal area of South Africa.

A consequence of the high number of interacting factors is that erosion rates are highly variable and tend to vary even between areas which are characterised by identical structural, lithological and geomorphological settings within the same bioclimatic zone. Accordingly, this results in a highly variable response to changes in the eksystemic components, which often obscures the distinction between the effects of human interference and naturally induced fluctuations.

The methods for predicting the rates of soil loss range from the extrapolation of test plots to the calculation by empirical and theoretical formulations. Of the latter, the "Universal Soil Loss Equation" (USLE) or related formulations of the soil erosion process are used (Stocking, 1995, Seuffert et al., 1999). However, the determination of erosion rates applying these methods remains a challenge. Beckedahl (2002) mentions a prediction error of 55% by applying the USLE on African soils. Apart from the complex interaction of factors which has to be determined on test sites, the inclusion of erosive rainfall events in soil erosion formulations remains a problem. Indices of rainfall, such as mean rainfall, wettest month or other indices, which are derived from rainfall data are rarely capable of predicting soil-erosion events and may be misleading (de Ploey et al., 1991). Methods based on the determination of magnitude-frequency relationships of individual rainfall events that are beyond the threshold of erosive rains may provide an alternative (de Ploey et al., 1991). Such methods enable the determination of the likelihood of erosive rains and provide information on the cumulative effects of individual rainfall events. This information makes it possible to deduce the Cumulative Erosion Potential (CEP) (de Ploey et al., 1991). The CEP-Index is based on the magnitude-frequency concept of Wolman and Miller (1960). According to this concept, the impact of extreme events is compensated by its lower frequency whilst the cumulative effect of more frequent events of a certain magnitude results in higher output. Figure 1 shows the magnitude-frequency relationship and the CEP-Index (table 1) for some stations in Lesotho, Kenya and Zimbabwe. The magnitude-frequency relationship and the CEP have been calculated from data provided by de Ploey et al. (1991), Calles and Kulander (1996) and Römer (2004). At stations where recurrence intervals of erosive rains are very short, the CEP-Index indicates a high potential of soil erosion. However, even if the concept of the cumulative effects of discrete rainfall events provides a reasonable approach to erosion events, the problems involved in a numerical calculation of the complex repercussions between seasonal effects, vegetation growth periods, rainfall structure and short-term clusters of intense rainfall events require further research.

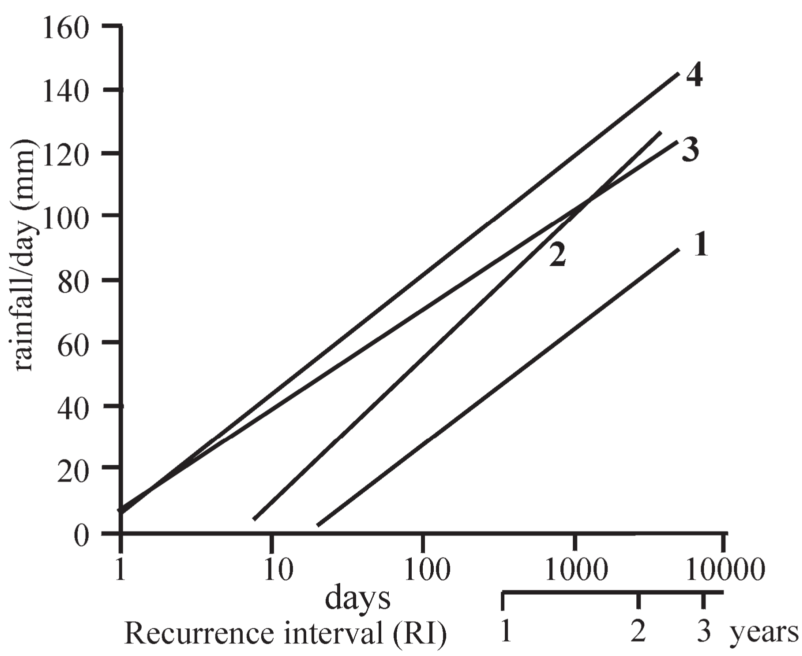

Fig. 1. Magnitude frequency relationship for stations in Zimbabwe, Lesotho and Kenya. 1 – West Nicolson (Zimbabwe) y = 36.46 log RI – 46.17; mean annual rainfall 579 mm (Römer, 2004, p. 21). 2 – Harare (Zimbabwe) y = 48.48 log RI – 32.07; mean annual rainfall 867mm (Römer, 2004, p.21). 3 – Machakos (Kenya) y = 15.85 log RI + 64.2; mean annual rainfall 1050mm (de Ploey et al., 1991, p. 404). 4 – Leribe (Lesotho) y = 26.09 log RI + 48.34: mean annual rainfall 795mm (Calles and Kulander, 1996, p. 163).

*h**	*Harare* *q = 153506*	*Machakos* *q = 9890*	*Leribe* *q = 34358*
0	26061	189306	73091
10	18939	100485	49822
20	14031	53235	33940
30	10189	28117	23135
40	7473	13779	15770

CEP = $m!\,\beta m \exp((\alpha - h^*)/\beta)\,\lambda$; $\lambda = \exp((\ln m! + m \ln ß + h^*/\beta) - h^*/\beta)$
The constants α and β are calculated from regression analysis according to the equation $y = \alpha + \beta x$. The constants correspond to the constants in the equations of the magnitude-frequency analysis in figure 1.
Constants: Harare α = -32.07; β = 48.48; Machakos α = 64.2; β = 15.85; Leribe α = 48.34; β = 26.09.
The CEP was calculated with m = 2.5 (silty to sandy soil) and different values for h*, which is a parameter for water storage. h* ranges from 0 to 10 on bare soils to 100 in areas with dense vegetation cover (de Ploey et al., 1991, p. 407, 408); q = dominant sediment transport amount. The CEP has been determined from the magnitude-frequency relationships published in de Ploey et al., (1991, p. 406); Calles and Kulander (1996; p.163), and Römer (2004; p. 21).

Table 1. Cumulative Erosion Potential (CEP) according to de Ploey et al., (1991). The CEP has been calculated for stations in Zimbabwe (Harare), Kenya (Machakos) and Lesotho (Leribe).

3.3 Soil erosion and human interference

In recent decades, intensified agriculture, livestock husbandry, clearance of forests and the increased density of settlements have contributed to the enlargement of areas affected by soil erosion. In several parts of Africa, human interference has accelerated natural erosion processes to a degree that influences the economics of extensive regions. The effects of human disturbance are generally most pronounced in hilly and mountainous terrains where steep hillslopes and high relief are conducive to high levels of erosion and a rapid response. In a study on the effects of interrill and rill erosion, Kimaro et al. (2008) demonstrated that soil loss due to deforestation and cultivation in the Uluguru Mountains of Tanzania exceeds 200 t ha^{-1} a^{-1}. The high degree of soil loss results from the steepness of the slopes, the high rainfalls but is also a consequence of continuous shallow and fine cultivation and tillage practices (Kimaro et al., 2008, p.42). In the Ethiopian highlands, changes in land use induced gully enlargement and gully incision. This resulted in a lowering of the groundwater. A concomitant effect was the decrease in soil moisture which was associated with a decline of crop yield (Nysson et al., 2004).
However, even in areas with low relief, the effects of slope gradient on soil erosion are noticeable. In an investigation conducted over a period of six years, soil erosion on maize-covered fields in Zimbabwe, Hutchinson and Jackson (1959) observed an average increase in soil loss of 3.1 t ha^{-1} a^{-1} at a slope gradient of 1.5° to 6.7t ha^{-1} a^{-1} at slope gradient of 3.5°. In the Middle Veld of Swaziland, threshold slope gradients seem to control the gully initiation on valley side slopes, whilst differences in land use or vegetation are subordinate (Morgan and Mngomezulu, 2003).
Although slope gradient is an important factor, the decline in ground cover may cause an increase in soil loss by several orders of magnitudes (Thomas, 1994, p.143,144; Reading et al., 1995). Studies of Nearing et al.,(2005) indicate that rainfall intensity and ground cover are likely to have a greater effect on soil erosion than changes in runoff and in the canopy cover alone. High soil losses of more than 200 t ha^{-1} a^{-1} are also indicated in studies of erosion in villages and on roads and in areas where heavy machines are used (Nyssen et al., 2002, de

Meyer et al., 2011). The increased runoff on roads, unpaved roads, pathways and landing sites promotes the concentration of overland flow into rills and the development of gullies by crossgrading and micropiracy. According to a study in Uganda, the soil losses range from 34 to 207t ha^{-1} a^{-1} (de Meyer et al., 2011). Despite the small percentage of total area of only 2.2 percent, de Meyer et al. (2011) emphasise that these areas are the major source areas for sediment delivery to Lake Victoria and that the total soil loss corresponds to an erosion rate of 2.1 t ha^{-1} a^{-1} (de Meyer et al., 2011, p. 97).

Patches of bare ground may induce soil erosion even on the low sloping surfaces of the basement regions of the African savannas (fig. 2). These landscapes are often characterised by a discontinuous soil cover that is interrupted by flat rock pedestals and small protrusions of bedrock. Fine-grained colluvial sediments that have been transported from the residual hills onto the gentle sloping pediments are often more prone to soil erosion than the coarser weathering products of the basement and may promote the development of large gully systems (fig. 3). However, serious and presumably irreversible effects seem to be more often the result of the interplay of several factors. This includes climatic fluctuation over a time-scale of several consecutive years or of decades, human activities and the role of inherited materials and forms in the landscape. Such conditions prevail in the Sahel, Sub-Sahelian zone, and in other transitional areas to the savannas, where extensive areas are covered with (fossil) sand dunes, sandy sediments and depleted weathering products. Highly susceptible to erosion are also savanna areas with dry seasons which last for six to eight months, where soils with a low aggregate stability or weak microstructure have been exposed by changes in the vegetation cover.

Fig. 2. Gullies formed during a heavy rainstorm at the start of the rainy season in southern Zimbabwe. Splash erosion and rill erosion affect the small slopes of the "badland" area. (Photo. Römer)

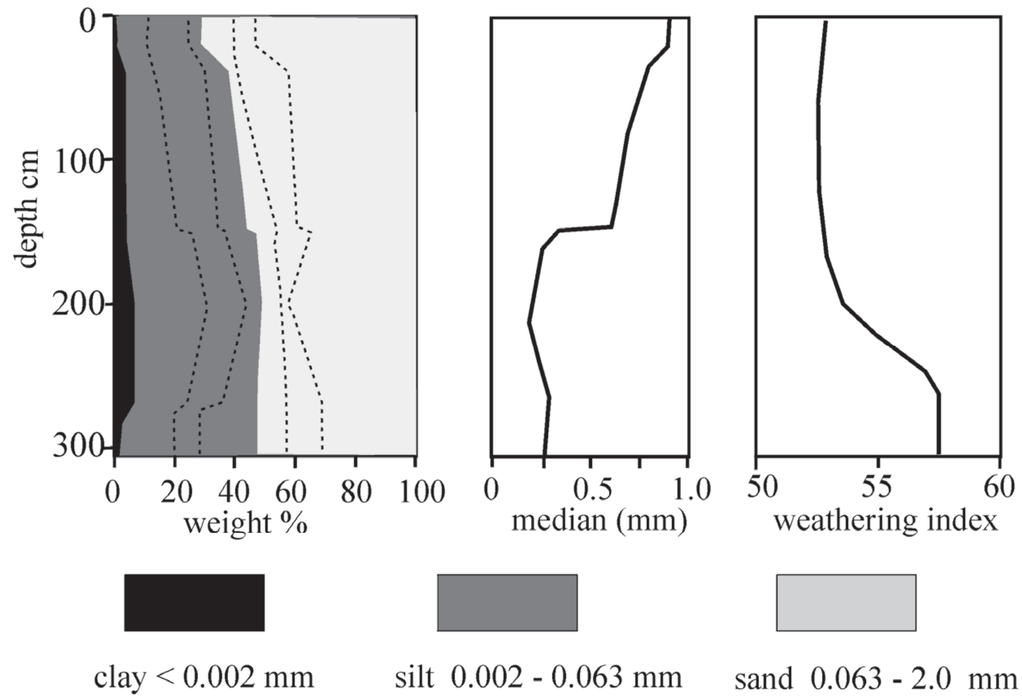

Fig. 3. Grain-size distribution, median grain-size and weathering index (CWI) of colluvium in southern Zimbabwe. The relatively fine-grained material formed the wall of a more than 3m deep gully. The decrease of the median grain-size corresponds to the increase in the weathering index and may indicate an non-conformity within the deposit (modified after Römer, 2004).

3.4 Inherited forms and materials and environmental change

Marked changes in rainfall distribution have been responsible for crisis-situations in the Sahel zone. Meteorological records of the Sahel zone show a relatively continuous decline of rainfall-levels from the 1960's to the 1970's when rainfall records attained a first minimum (Mensching, 1990; Warren, 1999). A second rainfall minimum occurred in the mid 1980's. The direct effect of these changes was an extension of areas of bare ground and a reduction in the biological diversity of the vegetation (Warren, 1999). At the same time, the variability of rainfalls increased, while there was a tendency towards a relative increase in short, high-intensity rainfall events (Gießner, 1989; Pflaumbaum et al., 1990). Studies indicate that the extension of the population and the increase in agronomic activities into a belt of formerly fixed dunes (Ooz-belt) in the west of the White Nile caused irreversible changes in the ecosystem (Mensching, 1984, 1990).

The dunes of the Ooz belt formed in the late Pleistocene and early Holocene when the climatic conditions were drier (Mensching, 1984, 1990). The development of the dune belt was interrupted by a more humid period that was characterised by greater availability of soil moisture and a denser vegetation cover. Weathering processes resulted in an encrusting of sand grains by iron-oxides. These sand layers are more resistant against wind erosion and are more impermeable than unweathered dune sands (Mensching, 1984). A further shift to dry climatic conditions provided the sands of the younger dunes of the Ooz dune belt. These dunes became inactive during the Holocene.

In the 1970's and 1980's, increased aridity, overgrazing and intense cultivation advanced the degradation of vegetation cover in the Ooz belt. This initiated the development of drifting dunes. At sites where the iron-impregnated sands of the older dunes became exposed, the lower infiltration rate increased runoff production during heavy rainfalls, which, in turn, enhanced the erosion processes (Mensching, 1984). The exposed older dune sands provide only poor soil material for further agricultural use. This caused a decline in the harvest, which was counteracted by enlarging areas for cultivation and grazing, thereby inducing a further degradation of the area. The high sensitivity of inherited sediments and soils to changes in the rainfall regime and vegetation cover is documented for several African countries.

Heinrich (1998) reports on changes in the Gongola Basin in Nigeria, where deforestation, a relatively high population density and intensified agricultural activity have caused serious erosion on Pleistocene aeolian sediments and hillwash sediments. Late Holocene (2000 BP) shifting cultivation in the Gongola Basin resulted in wash erosion on hillslopes and in the accumulation of sediments in small valleys with gallery forests, whilst the increase in population in the 20th century promoted the degradation of the savanna vegetation (Van Noten and de Ploey, 1977). Severe soil erosion resulted in a number of changes in the hydrologic regime and the nutrient balance of the soils (Heinrich, 1998). Apart from an increase in overland flow and diffuse wash erosion on low sloping surfaces, the higher runoff initiated a deeper dissection of gully-systems. The deep dissection was accompanied by a lowering of the ground water in the areas surrounding the gullies and a decline in the number of trees. High seepage gradients increased interflow and particle transport in the bedded subsoil. This resulted in the development of subsurface pipes and in increasing edaphic aridity, which strengthened the diffuse surface wash processes (Heinrich, 1998). The collapse of subsoil routes provided sites for the development of new gullies, whilst on low sloping surfaces wash erosion caused a marked decline in the clay content of the soils, which reduced the nutrient storage of the soils.

4. The role of high magnitude events

4.1 Extreme events in Africa

Climatic and hydrologic regimes on the African continent are highly variable in terms of both space and time. Rivers show the highest average extreme flood index of all continents, whilst the runoff ratios are lowest (McMahon et al., 1992). The temporal and spatial variability of rainfalls and rainstorms, as well as the repeated occurrence of periods of extreme droughts in semi-arid tropical and subtropical areas, indicate that extreme events play an important role in the African morphodynamic system. Relatively little is known about the relative work done by rare events of high magnitude when these events are compared with more frequent events with a low magnitude (Gallart, 1995). Although studies indicate that the impact of erosion increases with increasing amounts of rainfall and rainfall intensity, such relationships are not without ambiguity, as events of similar magnitude may have different effects, whilst events with a higher frequency and lower-magnitude are capable of inducing similar effects (Gallart, 1995). While the incidence of drought and rainfall events is determined by the present-day climatic system, human activities may change the magnitude of the impact by changing the vegetation cover, the hydrologic regimes and characteristics of the surface materials and forms. Therefore, an increase in the impact of smaller events with shorter recurrence intervals and lower magnitude is likely. The intensity of the response to these events is a function of the intensity of interference with the ecosystem, coupling of the subsystems and the sensitivity of the subsystems affected. However, changes in environmental conditions frequently bring with them a non-linear behavioural pattern caused by feedbacks (Thomas, 2004). These feedbacks weaken or reinforce the response to changes in different subsystems. In semi-arid areas, a decrease in vegetation cover may reinforce the decline of rainfalls as the degradation of the vegetation decreases the surface roughness and soil moisture. Consequently, the evaporation and transpiration rates decrease, which, in turn, reduces transport of vapour into the atmosphere (Warren, 1999). Further effects are likely to involve changes in the cloudiness, the spatial and temporal distribution and intensity of the rainfalls and the lower inflow of rainwater to the ground water. A decrease of the vegetation cover causes a decrease in the water-retention capacity of the soils, which, in turn, may reduce the threshold of runoff-producing storms (Gallart, 1995). The effects of such changes point towards a higher preparedness of landscape components to react to events of lower magnitude and higher frequency. This appears to bring with it an increase in the impact of events of lower magnitude. The latter is corroborated in studies of sediment yield in Kenya, which indicate that an increase in land use is associated with an increase of the relative work of events of higher magnitude (Dunne, 1979).

With respect to the impact of meteorological events on erosion-processes, the effects of continuous rainfall and short-term high-intensity rainfall events must be distinguished. Long- lasting rainfall events of exceptional magnitude determine the saturation of soils and induce saturation overland flow and liquefaction of the soil layers. High-intensity rainfalls, on the other hand, are capable of inducing Hortonian overland flow causing a rapid increase of runoff. However, the impact of such events depends strongly on the antecedent state of the ground. The role of heavy downpours increases towards the semi-arid and arid tropical areas, where daily rainfall events may exceed the mean annual rainfall by more than 40% (Starkel, 1976). Rainfall intensities ranging from 250mm to more than 400mm have been reported from Mauritania and Tunisia, and daily maximum rainfalls exceeding the annual rainfall by 50mm appear to occur several times within a decade ((Mensching et al., 1970; Starkel, 1976). Such rainstorm events are often accompanied by high discharges and floods (Starkel, 1976). Continuous rainfall events are associated with the adduction of humid air masses, which often occurs in tropical, tropical-monsoonal areas or in areas where air masses are impeded by mountains. Tropical cyclones such as the Mauritius cyclone in the Mozambique channel are also associated with high rainfall events. According to Weischet and Endlicher (2000) about 520 cyclones have been registered in 70 years, and most cyclones deposit large volumes of rainfall along the coast. An extreme event accompanied the cyclone Donoina, which occurred in the year 1984. This cyclone crossed southern Africa, and rainfall intensities achieved about 900mm in a few days. This resulted in severe flooding and intense erosion in Mozambique, Swaziland and South Africa (Goudie, 1999).

4.2 Extreme events and complex response

The response to extreme events depends not only on the magnitude of the event. Studies on flood frequencies at the Orange River in South Africa indicate that the rate of change and antecedent environmental conditions play an important role. During the last 5500 years, the lower Orange River has experienced marked changes in terms of its hydrologic regime. Zawada (2000) was able to distinguish four periods with different flood magnitudes and frequencies. Although there is a close association between high levels of discharge and warm and wet periods, the most extreme discharge events occurred during a warm interval of the "Little Ice Age", in the period 1500 to 1675 AD (Zawada, 2000). The maximum flood discharge during this brief period exceeded any historically gauged floods by a factor of three. According to Zawada (2000), the high floods cannot be attributed to the increase in rainfall, as during earlier, more humid periods the flood discharge was significantly lower, though these paleoflood discharges exceed all documented floods since the end of the 18th century. Zawada (2000) argues that the sudden onset of warming caused an intense change in the hydrologic regime. Apparently the change affected hillslopes as well as rivers within a time interval that was shorter than the time that is necessary to achieve a full adjustment of the vegetation cover to the changed conditions.

Singular events of high magnitude may result in serious damage. Rapp (1976) has documented the effects of a rainstorm in the Mgeta mountains of Tanzania,. The rainfall event achieved an intensity of 100.7mm in less than three hours and triggered more than 1000 shallow landslides in the highly weathered soils. Landsliding affected about 47% of the cultivated land, 46% of the grasslands but less than 1 % of the wooded areas (Rapp, 1976, p.92). The results highlight the link between slope stability, soil properties and changes in the vegetation cover. Trees lower the water table in the soils by transpiration and reduce the amount of rainfall reaching the slope surface as a part of the rainwater is intercepted in the canopy. Both processes counteract soil saturation and delay the development of high pore water pressure. Once deforestation takes place, these positive effects are lost. In combination with the loss of tree roots, this results in a reduced shear strength of the soils, a higher probability of high-pore water pressure and a lower threshold of stability against landsliding (Rapp, 1976).

The sensitivity to change is a further factor which appears to exert an important influence on the magnitude of events. Most landscapes in Africa have suffered progressive change through time and tend to accumulate the imprints of different environmental conditions. These imprints range from deposits and weathering layers formed during periods with different climatic conditions to hillslope forms and polyphase landscape elements. In the KwaZulu area, Singh et al. (2008) investigated extensive landslide complexes which seem to have been active in the middle and late Holocene. The volume of large individual landslides ranged from $1 \cdot 10^7$ to $2 \cdot 10^7$ m^3. Some smaller, secondary occurrences of slope failure were apparently reactivated on the larger landslide masses. However, the large landslide complexes appear to be stable. According to Singh et al. (2008), these landslides resulted from a combination of long–term rock-weathering and the location in a seismic active zone.

High-intensity rainfall events in 1987 and 1997 in Natal (Southern Africa) indicated a strong association between landsliding and colluvial deposits (Bell and Maud, 2000; Singh et al., 2008). The colluvial deposits in this area are characterised by several non-conformities resulting from differences in the intensity of weathering, the variable thickness, texture and permeability. According to Bell and Maud (2000), landslides on the hillslopes of the Natal group are closely associated with the specific behaviour of the colluvial deposits. The weathered colluvium consists of an upper sandy layer (topsoil) and an illuvial horizon, which lies above a clayey weathered layer. During heavy rainstorms, the silty and clayey layers impede the downward percolation of the water. This promotes the development of high-pore water pressure and of saturated conditions in the upper soil layers. The lateral throughflow in the more permeable layers of the colluvium and weathering layers on the upper hillslope-segments increases the flow of ground water to the middle and lower hillslope-segments. This causes the development of excess pore water pressure and artesian conditions on the lower hillslope segments, which, in turn, is accompanied by viscous flow movements and liquefaction of the soils (Bell and Maud, 2000). During the 1987 event most landslides were triggered by an extreme rainstorm episode with an intensity of 576 mm in 72h (Bell and Maud, 2000, p. 1034).

However, antecedent moisture conditions seem to play an important role, as prior to 1987 no records of larger landslide events are documented, while it is likely that rainfall events of similar magnitude have occurred several times in the past. The importance of antecedent moisture conditions and of the properties of the colluvial layers is indicated in the critical precipitation coefficients for slope failure that were calculated by Bell and Maud (2000). According to their investigations, major landslides and landslide episodes will occur when rainfall intensities exceed the mean annual precipitation by 20%. However, most landslides were triggered in the latter months of the rainy season when the colluvium was almost saturated with water. Accordingly, occurrences of slope failure in this area depend on the rainfall intensity and on the antecedent moisture conditions (Bell and Maud, 2000). On the other hand, the investigations emphasise the important role of permeability non-conformities in the colluvium and at the weathered-unweathered rock boundary. This indicates that the occurrence of landslides depends strongly on local conditions and that several factors must be kept in mind in the analysis of landslides. These factors include the association between slope parameters, mechanical parameters of the soils, rocks or sediments, and the presence of palaeolandslides.

5. The role of fires

Fires play an important role in African environments, and few areas in the African savannas appear to have ever escaped fires. In the savanna areas, fires appear to determine the volume of biomass above the ground and the turnover of herbivores and saprophytes.

Wildland fires can be induced by lightning, volcanism and rockfalls. Most fires in savanna environments are ignited in the dry season by lightning. In west Namibia, lightning ignites about 60% of the savanna fires (Held, 2006). However, in mountainous terrains, rockfalls may be also an important factor. Reports from the Cedar Hills in South Africa indicate that rockfalls contribute to the development of about 25% of the fires (Goldammer, 1993). Since the appearance of humans, the impact of fires on vegetation patterns has progressively increased. The modification of the vegetation in the savannas began in an early epoch, when hunters and gatherers used fire to make hunting easier. Evidence of the early use of fire ranges from sedimentary layers in the Swartkrans Cave in South Africa, with an age of about 1.5 Ma BP (Gowlett et al., 1981) to changes in the vegetation pattern on the Nyika Plateau in Malawi, which seems to indicate the repeated burning of the savanna vegetation at the end of the Pleistocene (Goldammer, 1993).

In more recent times, increasing demand for arable land has resulted in a regular burning of larger savanna areas and in the development of extensive grasslands. Within the moist-savanna-zone, this has caused the development of "derived savannas", which consist of grasslands and are a secondary vegetation formed by fires (Goldammer, 1993; Schultz, 2005). The effects of fires decrease from the moist savannas to the dry savannas, largely as a function of the available biomass.

The on-site effects of fires range from the immediate impact of the selective burning on the bio-diversity and vegetation structure to changes in the physical, chemical and biological components in soils (Schultz, 2005). However the impact varies as a function of the composition of the plant communities, the size and shape of the woody species, the frequency of fires, the heating temperature during burning, the length of the period of time since the last fire, the onset of the fire during the dry season, and the land-use techniques applied (Schultz, 2005). Some cultivation techniques appear to reinforce the danger of further fires by changing the composition and structure of the ground cover as in the case of "slash and burn agriculture" (Goldammer, 1988). Biomass burning affects the reserves and storage of organic matter in the ground cover and in the soils and hence induces changes in soil-nutrient levels. An immediate effect of burning is an increase in K, Ca, Mg and the pH (Singh, 1994). However, the baring of ground promotes erosion by wind and water, and the transport of ashes contributes to the distribution of nutrients over a larger area. The change in the surface colour results in a higher absorption of the solar radiation and in an increase in evaporation. A further effect involves the enrichment of condensed volatile organic substances in the topsoil. This causes the development of a thin layer, which impedes the infiltration of water (Cass et al., 1984). Accordingly, these changes tend to increase the likelihood of soil erosion by the first rainfall events.

The off-site effects of fires are changes in the sediment delivery and in the nutrient level of the rivers' draining areas which are affected by fires. Fires tend to increase the content of dust in the atmosphere as they provide aerosols. Aerosols released by smouldering fires exert control on radiation activity as they increase condensation and cloudiness. However, the surplus of condensation nuclei results in small water droplets that remain suspended in the cloud. Consequently rainfalls are less likely. A further consequence of fires is the emission of oxides of carbon and nitrogen as well as of ozone and halogenides (Helas et al., 1992; Andreae et al., 1996). Particularly methyl chloride and methyl bromide emissions appear to support ozone depletion in the upper atmosphere, though the residence-times of these compounds are shorter than 2 years (Andreae et al., 1996). The estimated amounts of methyl-chloride and methyl-bromide emissions range from 1.8 Tg a^{-1} to 7 Gg a^{-1} (Andreae et al., 1996). This indicates that these compounds are capable of contributing significantly to ozone depletion in the upper atmosphere.

With respect to the extensive areas which are affected by fires, the question arises whether fires increase the level of greenhouse gases in the atmosphere. Andreae (1991) reports that in each year about 75 % of the African savannas are affected by fires. However, during the savanna fires, parts of the biomass are converted into elementary carbon (e.g. black carbon, charcoal). The estimated amount of charcoal formed during a fire appears to account for 5 to 10 % of the total biomass (Goldammer, 1993; Kuhlbusch et al., 1996). This fraction remains in the soil or sediments or is transported by rivers to the ocean, but cannot reenter the atmospheric carbon cycle (Goldammer, 1993). Consequently, this deficit in carbon has to be compensated for by consumption of atmospheric carbon. According to this concept, savannas may become a carbon sink when the processes are balanced through vegetation regeneration. Studies of the annual gas emissions of fires indicate that in the dry savannas the emissions of carbon dioxide, ammonia and nitric oxide do not exceed the amount dictated in the biomass by processes of nitrification and photosynthesis (Schultz, 2000; 2005). However, in the moist savannas, the changes in the vegetation are more pronounced, particularly if there is no regeneration of woody plants and the vegetation structure is destroyed. Accordingly, this may counteract the compensating effects of regeneration.

However, we have a poor understanding of the turnover of carbon in quantitative terms in the savannas due to the complex interaction of weathering, soil formation, vegetation and litter production and different reaction-times. Finally, a full assessment of the climatic impact of biomass-burning depends also on the reliability of the data and on the quality of case studies.

6. Prospect and conclusions

In Africa the superimposition of climatic changes and human activities is accompanied by a serious degradation of environmental conditions generally. The impact of this change involves certain thresholds which depend on the intensity and duration of meteorological events, the condition of the vegetation cover, the physical and chemical properties of the soil system and of the geomorphic settings. High rates of change occur in regions where large areas are affected by human intervention and where factors such as a high relief, steep slopes and a strong coupling between hillslopes and rivers support a rapid response. Slope failure in colluvial deposits, and erosion of hillwash and aeolian deposits indicate the important role of forms and deposits which are inherited from the past. Long-term processes, such as deep weathering, can contribute to the humanly-induced instability of hillslopes, once intrinsic thresholds are exceeded due to a continuous lowering of the shear strength or the increase in soil thickness (Shroder, 1976).

Studies on the impact of climatic changes on erosion processes in the late Pleistocene and the early Holocene indicate that a complete adjustment to the changed conditions requires a simultaneous response of all landscape components throughout a period of time that is long enough, to overcome the inertia of the geomorphic system (Thomas, 2004). With respect to the time frame of change in the vegetation-soil systems, these adjustments are considered to have been accomplished within a period of 10^3 to 10^4 years (Thomas, 2004, 2006). The expected rates of response point to the temporal and spatial differences between natural changes and humanly-induced change. Human interference is capable of changing the vegetation cover and the hydrologic regimes of extensive areas within a relatively short time. Repeated biomass burning in the savanna and rain forest zones coupled with intensified land-utilisation activity resulted in a degradation of the vegetation-soil system in several areas and often initiated an array of self-reinforcing processes.

Predictions of IPCC (2001) on the climatic development in Africa suggest that the climate is likely to get warmer, while the total amount of rainfall will not change significantly. However, a higher number of days with heavy rainfall is likely. These changes may affect the biota, the land use pattern and the hydrologic regimes. In the alpine Usambara Mountain area of East Africa (Tanzania) the lower replacement of montane forest trees seems to have been accompanied by general global warming over the last 100 years (Binggeli, 1989, Hamilton and Macfadyen, 1989). As a result of global warming, a general decline in the extent of the Afroalpine areas is likely (Taylor, 1999). The predicted increase in heavy rains may promote the increase of runoff, whilst the decrease of soil moisture is likely to bring with it edaphic aridity and an increase in erodibility (Beckedahl, 2002). This may result in a reinforcement of soil erosion. The decline in the number of rain days, on the other hand, may promote vegetation decay and leave more areas unprotected from heavy rainfalls. However, land use changes seem to have a much greater impact on susceptibility to soil erosion (Beckedahl, 2002, Valentin et al., 2005).

The increase in population in Africa is expected to result in an extension of the area cultivated land, even in steeply sloping mountainous regions. The impact of change in the climate and the intensified land use are likely to cause a reinforcement of degradation processes in the landscapes and may result in a lowering of the carrying capacity of land. However, predictions on future rates of change also depend on socioeconomic processes and political decisions. The devastating impact of desertification in the Sahel was not only a result of drought but was also associated with one of the highest population growth rates in the world.

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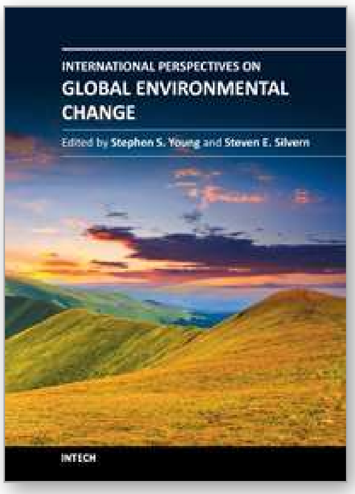

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Environmental change is increasingly considered a critical topic for researchers across multiple disciplines, as well as policy makers throughout the world. Mounting evidence shows that environments in every part of the globe are undergoing tremendous human-induced change. Population growth, urbanization and the expansion of the global economy are putting increasing pressure on ecosystems around the planet. To understand the causes and consequences of environmental change, the contributors to this book employ spatial and non-spatial data, diverse theoretical perspectives and cutting edge research tools such as GIS, remote sensing and other relevant technologies. International Perspectives on Global Environmental Change brings together research from around the world to explore the complexities of contemporary, and historical environmental change. As an InTech open source publication current and cutting edge research methodologies and research results are quickly published for the academic policy-making communities. Dimensions of environmental change explored in this volume include: Climate change Historical environmental change Biological responses to environmental change Land use and land cover change Policy and management for environmental change

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