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Human impact on the environment in the Ethiopian and Eritrean highlands—a state of the art

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Abstract

This review analyses the environmental evolution of the Ethiopian highlands in the late Quaternary. The late Pleistocene (20,000–12,000 ^{14}C years BP) was cold and dry, with (1) low lake levels in the Rift Valley, (2) large debris fans on the flanks of Lake Abhé basin, and (3) the Blue Nile transporting coarse bedload. Then, a period with abundant and less seasonal rains existed between 11,500 and 4800 ^{14}C years BP, as suggested by increased arboreal pollen, high river and lake levels, low river turbidities and soil formation. Around 5000–4800 ^{14}C years BP, there was a shift to more arid conditions and more soil erosion. Many phenomena that were previously interpreted as climate-driven might, however, well be of anthropic origin. Thick sediment deposits on pediments as well as an increase of secondary forest, scrub and ruderal species in pollen diagrams are witnesses of this human impact.

One important aspect of the late Quaternary palaeoenvironment is unclear: changes in Nile flow discharges and Rift Valley lake levels have been linked to changes in precipitation depth. Most authors do not take into account changes in land use in the highlands, nor changes in the seasonality of rain, both of which can lead to a change in runoff coefficients. Tufa and speleothem deposition around 14,000 years ago tend to indicate that at the end of the Last Glacial Maximum (LGM), conditions might have been wetter than generally accepted.

The most important present-day geomorphic processes are sheet and rill erosion throughout the country, gullying in the highlands, and wind erosion in the Rift Valley and the peripheral lowlands. Based on existing sediment yield data for catchments draining the central and northern Ethiopian highlands, an equation was developed allowing to assess area-specific sediment yields:

$$\mathrm{SY} = 2595A^{-0.29} \quad (n = 20;\ r^2 = 0.59)$$

where SY = area-specific sediment yield (t km^{-2} year^{-1}), and A = drainage area (km^2).

With respect to recent environmental changes, temporal rain patterns, apart from the catastrophic impact of dry years on the degraded environment, cannot explain the current desertification in the driest parts of the country and the accompanying land

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degradation elsewhere. Causes are changing land use and land cover, which are expressions of human impact on the environment. Deforestation over the last 2000–3000 years was probably not a linear process in Ethiopia. Studies on land use and land cover change show, however, a tendency over the last decades of increasing removal of remnant vegetation, which is slowed down or reversed in northern Ethiopia by a set-aside policy.

Ongoing land degradation requires urgent action at different levels of society. Soil and water conservation (SWC) structures are now widely implemented. Local knowledge and farmer's initiatives are integrated with introduced SWC techniques at various degrees. Impact assessments show clear benefits of the soil conservation measures in controlling runoff and soil erosion. In high rain areas, runoff management requires greater emphasis during the design of soil conservation structures. In such areas, investment in SWC might not be profitable at farm level, although benefits for society are positive. This pleads in favour of public support.

The present land degradation in the Ethiopian highlands has a particular origin, which includes poverty and lack of agricultural intensification. Causes of these are to be found in the nature of past and present regional social relations as well as in international unequal development. This review strengthens our belief that, under improved socio-economic conditions, land husbandry can be made sustainable, leading to a reversal of the present desertification and land degradation of the Ethiopian highlands.

Keywords: Deforestation; Desertification; Ethiopia; Human impact; Land degradation; Late Quaternary; Sediment deposition; Soil erosion

1. Introduction

This paper presents the state of the art regarding past and present erosion processes as well as environmental changes in the Ethiopian highlands (Fig. 1).

If many changes may have occurred in the environment, one factor has remained constant: the geological background. Peneplained Mesozoic sedimentary rocks have been concealed by basaltic flows (Mohr, 1963; Merla et al., 1979). The dome-like uplift of the Arabo-Ethiopian region started during the Oligocene and had two periods of intense tectonic activity: in the Miocene, about 25 million years ago and in the Plio-Pleistocene (Williams and Williams, 1980). The elevation above sea level of the base of the basalt shows the importance of the uplift: about 500 m in 25 million years to the west of Lake Turkana and more than 2000 m in Eritrea. (Physiographic continuity led us to include the Eritrean highlands in this review.) All the rivers are deeply incised (Adamson and Williams, 1980), and various lithologies are exposed. Steep slopes appear to be a major factor controlling the intensity of erosion processes. This setting strongly determines the impact of environmental changes, as will be discussed throughout this paper.

First, the late Quaternary environmental conditions are discussed through the analysis of a certain number of proxies. Then, the present-day rain, runoff and erosion processes are reviewed, and an attempt is made to quantify soil loss rates. Environmental change will be analysed, including rain and land use changes, which induce land degradation and desertification (UNEP, 1994). Human impact is important, not only as a factor of degradation, but also as a factor of environmental recovery, as will be discussed in Section 6.2.

2. Palaeoenvironmental evolution since late Pleistocene times

2.1. Introduction

Hurni (1989) and Messerli and Rognon (1980) already discussed palaeoclimatic conditions of the region during the Quaternary. However, many data on late Quaternary palaeoenvironmental evolution and erosion processes are dispersed over disciplines as diverse as geology, glaciology, geography, palaeoecology, -climatology, -botany, soil sciences, palaeontology, hydrology and palynology.

Some ways of exploring the past environmental conditions have never been exploited in Ethiopia. With respect to dendrochronology, only *Juniperus procera* has been found to be promising, especially in Northern Ethiopia, which has a well-defined longer dry season (Conway et al., 1998). The presence of tephra in depositional areas in the Rift Valley has been

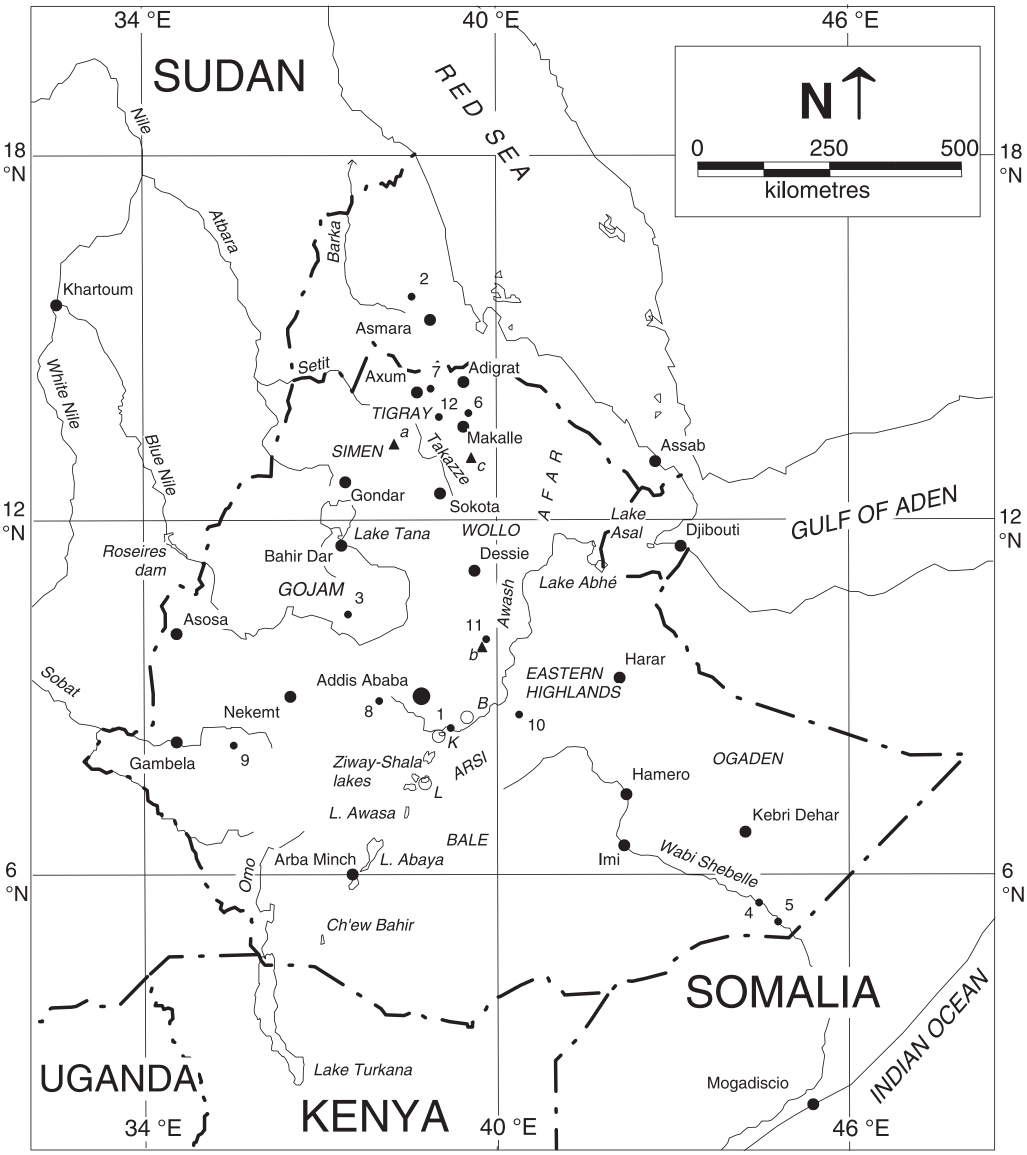

Fig. 1. Map of Ethiopia and Eritrea, with indication of localities mentioned in the text. (1, Nazret/Adama + Debre Zeit; 2, Afdeyu; 3, Anjeni; 4, Kelafo; 5, Mustahil; 6, May Makden; 7, Adwa; 8, Ambo; 9, Dizi; 10, Hunde Lafto; 11, Debre Sina; 12, Dogu'a Tembien; ▲, mountain summits: a, Ras Dejen; b, Ankober; c, Amba Alage; ○, lakes: B, Lake Besaka; K, Koka reservoir; L, Lake Langano).

recognised, but until now, only the early Pleistocene tephra layers of the southern Rift Valley and the Omo/Turkana basin have been correlated at a regional scale (Pyle, 1999; Katoh et al., 2000).

This section concerns the palaeoenvironmental evolution since late Pleistocene times through a number of its proxies. After a discussion of the implications of the late Oxygen Isotope Stage (OIS) 3 glacial—a benchmark in this review—lake and river levels and deposits, pollen cores and palaeosols will be examined. Throughout this paper, '^{14}C years BP' will be used for uncalibrated radiocarbon ages only.

2.2. Quaternary glaciations in the Ethiopian mountains

The presence of glacier cirques, moraines and periglacial solifluction deposits is clear evidence of glaciations in the highest mountain areas of Ethiopia (Grab, 2002), i.e., Simen (4620 m a.s.l.), Arsi (4180 m a.s.l.) and Bale (4357 m a.s.l.) (Fig. 1). Early researchers correlate traces of glaciation with temperature lowering and increased precipitation (Nilsson, 1940, 1964; Büdel, 1954; Abul Haggag, 1961). Pluvial periods in East Africa were expected to correspond to glaciations at higher latitudes. Coetzee and Van Zinderen (1989) review the literature regarding the shift from the "pluvial theory," in use until the 1960s, to the present-day interpretation of glaciations corresponding to dry periods in tropical Africa.

Messerli and Rognon (1980, p. 108) provide an extensive overview of the late Pleistocene glaciations. The lowest terminal moraines would have been situated around 3650 m a.s.l. (Potter, 1976) (3400 m a.s.l. according to Hastenrath, 1977), and the snow limit around 4000 m a.s.l., as materialised by the upper limit of the lateral moraines. Potter (1976) estimates that in the late Pleistocene, 580 km^2 (most in Bale) were glaciated in Ethiopia, out of 1900 km^2 of known Pleistocene glaciation in Africa. The periglacial limit was situated around 3000 m a.s.l. in Bale (Potter, 1976), at 3500 ± 100 m in Simen (Hurni, 1981) and at 3600–3700 m a.s.l. in other Ethiopian mountains north of Addis Ababa (Hurni, 1989). Nilsson (1940) and Hövermann (1954a,b) argue for the existence of lower limits of the snow belt, witnessing earlier glaciations in Ethiopia. This is questioned by later workers (Semmel, 1963; Hastenrath, 1974; Messerli and Rognon, 1980). Kuls and Semmel (1962) discuss how moraine- and kar-like formations, found down to 1800 m a.s.l. result from differential weathering of basalt, rather than from glacial activity.

Hurni (1981, 1982), in the first 'overall' study of glaciations in Simen, distinguishes the following landforms due to the last cold period: (a) >3000 m a.s.l.: fluvio-solifluvial deposits in valley bottoms, (b) 3400–4200 m a.s.l.: up to 15-m-thick periglacial solifluction deposits, (c) >3780 m a.s.l.: moraines (and nivation deposits), (d) >4250 m a.s.l.: glacier cirques.

A reconstruction of the climate during the Last Glacial Maximum (LGM) can be made by analysing the altitude of the periglacial belt nowadays (4250 m a.s.l.) (Hastenrath, 1974; Messerli and Rognon, 1980; Hurni, 1981, 1982) and during the last cold period (3500 m a.s.l.). The present-day snow limit, however, is estimated at 4800 m (Büdel, 1954; Messerli and Rognon, 1980), which is higher than Ras Dejen (4620 m a.s.l.), the culminating point of the Simen Mountains and of Ethiopia. Since this snow limit is only estimated and not materialised, it is less useful for the estimation of the shift of the altitudinal belts. As valley floors down to 3000 m are completely filled with solifluction deposits without bedding, it is deduced that there was less runoff than nowadays and thus, lower precipitation. Under such dry conditions, Hurni (1982) accepts a temperature gradient of 1 °C per 100 m lowering of the periglacial belt: at the LGM, the temperature in Simen was thus some 7 °C less than the present one. This is in accordance with temperature values found in other East African mountain areas (Coetzee and Van Zinderen, 1989).

Hurni (1982) proves that this glacial period is not sub-recent, as it was sometimes argued, taking into account reports by 17th to 19th century travellers (Hövermann, 1954a). Postglacial organic horizons were found to be more than 4000 years old. Gasse and Descourthieux (1979) found a palaeosol at 4040 m a.s.l. and dated it at 11 500 ^{14}C years BP. Dating glacial events usually involves pollen analysis and radiocarbon dating of organic matter deposited behind or on top of moraines (Messerli and Rognon, 1980). Based on ^{14}C dated peat deposits, the beginning of deglaciation in the Bale Mountains is estimated at 14,000–13,000 ^{14}C years BP (Mohammed and Bonnefille, 1998).

The last cold period is thus estimated to have been a dry period, corresponding to the late OIS 3 (20,000–12,000 ^{14}C years BP) (Hurni, 1981, 1989).

2.3. Lake levels

Many authors observed and dated previous levels of endoreic lakes in the main and northern Ethiopian Rift Valley, using various methods: analysis of shells in lake terraces, of littoral molluscs in life position and of in situ charcoal; interpretation of dune activity, palaeosols and the presence of marginal stromatolites (Gasse et al., 1974). Lake bottom sediments were analysed for magnetic minerals, diatoms, geochemistry and pollen (Gillespie et al., 1983). Synchronous change of levels of these endoreic lakes draining the highlands are believed to provide information about previous climatic changes because they often confirm other independent evidence of dryer or moister conditions during phases of low/high lake levels (Williams et al., 1977). Even in the tectonically active Afar basin, deformations due to Holocene tectonic activity are minor, compared to climatic lake level fluctuations (Rognon, 1975). Levels of endoreic lakes in Ethiopia are among the best documented in the world (Street and Grove, 1979). Whereas groundwater-fed lakes in the Rift Valley bottom such as Lake Besaka (Williams et al., 1981) are not sensitive indicators of climatic changes in the highlands, most Rift Valley lakes fall in the 'amplifier' category (Street, 1980), i.e., they strongly reflect the runoff conditions in their basin.

The Ziway-Shala basin is located in the highest part of the main Ethiopian Rift floor. The present-day four lakes merged into one during high stands (Grove and Goudie, 1971; Grove et al., 1975). Analysis of the lake levels was systematised by Gasse and Fontes (1989) and later confirmed in the study of Le Turdu et al. (1999). They found high lake levels before the LGM (30,000–27,000 ^{14}C years BP), low levels during late OIS 3 (22,000–12,000 ^{14}C years BP), after which the lake levels rose again (Fig. 2) with wet pulses around 11,500, 9500–8000, 7000–6000 and around 5500–5000 ^{14}C years BP. The general trend was interrupted by frequent, short, arid intervals marked by dramatic lake level lowering (Sagri et al., 1999), interpreted by Gillespie et al. (1983) as the influence of forcing mechanisms overriding the general trend (volcanism, solar variability, glacial surges, etc.). Benvenuti et al. (2002) suggest possible nonclimatic controls, such as faulting activity, on the high terminal Pleistocene lake levels. A flight of shorelines indicates that from 4800 ^{14}C years BP, the lakes retreated quickly to the present-day low level (Le Turdu et al., 1999). Based on carbon isotope records, Lamb et al. (2000) also found low lake levels at the nearby Lake Tilo from 4200 ^{14}C years BP on. Low lake levels existed during the whole middle and late Holocene, with the exception of a short high stand around 1500 ^{14}C years BP (Williams et al., 1977; Gasse and Street, 1978; Gasse and Fontes, 1989; Le Turdu et al., 1999). High lake stands in the late Pleistocene are also confirmed by the fact that Middle Stone Age sites (100,000–35,000 ^{14}C years BP) in that area are all located at more than 150 m above Lake

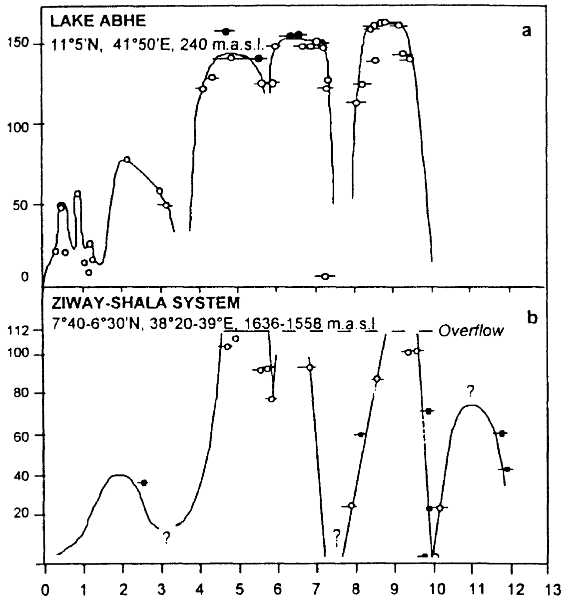

Fig. 2. Levels of Ziway-Shala and Abhé lakes during the Holocene. The x-axis represents the age (in 10^3 ^{14}C years BP); the y-axis is the elevation of the lake level (in meters above the present level). Reprinted from Gasse (1998), with permission from IAHS Press.

Ziway. Late Stone Age artefacts (11,000–10,000 ^{14}C years BP) were only found on hilltops more than 34 m above lake Ziway (Wendorf and Schild, 1974; Laury and Albritton, 1975). Using lake levels, Street and Grove (1979) modelled palaeoclimatic data for the Ziway-Shala basin, as compared to present conditions (Table 1).

Although Lake Abhé is located in the central Afar, it depends chiefly on runoff from the highlands (Gasse, 1977; Gasse and Street, 1978). Lake Abhé registered a dry phase at the early OIS 3 (around 70,000–60,000 years). The recorded lake levels indicate a wet period from 50,000 to 17,000 years, interrupted by a minor regression around 30,000 years. A major regression started at 20,000 ^{14}C years BP with a hyperarid period from 17,000 to 10,000 ^{14}C years BP (Fig. 3). During the LGM, Lake Abhé was dry (Gasse, 1974, 1990). From 9500 to 4500 ^{14}C years BP, the major lacustrine phase of the Holocene period (Gasse, 1974, 1990), lake levels were high, with the exception of a short regression around 8000–7500 ^{14}C years BP, which can be correlated with a fall in the levels of Lakes Ziway-Shala, Turkana and Nakuru as well as with a change in

Table 1
Palaeoclimatic estimates for the Ziway-Shala basin (after Street and Grove, 1979)

Time period	Years BP	Temperature (°C, departure from present)	% of present precipitation	Comparable data were found for other lakes
Late Holocene maximum	2000	0	122	
Early Holocene maximum	9400–8500	0 to −2	128–147	Nakuru, Naivasha, Manyara[a]
Terminal Pleistocene minimum	13,500	−3 to −6	68–91	Nakuru, Kivu, Victoria

[a] See also Hastenrath and Kutzbach (1983).

the flow regime of the Blue Nile. From 4000 ^{14}C years BP on, Lake Abhé dried out rapidly and during the late Holocene, the lake had only minor fluctuations (a small transgression between 2700 and 1000 ^{14}C years BP) (Fig. 3) (Gasse et al., 1974; Gasse, 1977, 1980; Gasse and Street, 1978).

Located more to the east, Lake Asal (−155 m a.s.l.) had high stands (+160 m a.s.l.) between 9000 and 6000 ^{14}C years BP, when it drained the Awash basin through groundwater circulations (Fontes et al., 1973). During dry periods, it is fed by infiltration of marine water. It differs thus from lake Abhé by its specific geological and hydrological setting, as illustrated in an eloquent way by Gasse and Fontes (1989, Fig. 9). This lake should therefore not be used as a proxy for palaeoclimatic conditions on the highlands.

Littmann (1989) reviews the late Pleistocene water budget tendencies in Africa. Five generalised maps of the main water budget tendencies at continental scale fully confirm, with respect to the Ethiopian highlands, the above-presented fluctuations. This author also finds that in the Ethiopian parts of the East African Rift system, there is a general coincidence of cool and dry conditions on the one hand and of warm and wet conditions on the other. It was already stated earlier that Ethiopia followed "the general pattern of Holocene climatic events of the southern margin of the Sahara" (Gasse et al., 1974).

There are, however, present-day lake level variations, which cannot be explained by variations in annual precipitation. The Lake Awassa rise between 1954 and 1972 was not accompanied by increasing precipitation but has been explained by a reduction in evaporation: the number of wet days increased, and land was cleared for cultivation, increasing runoff (Makin et al., 1974; Grove et al., 1975; Lamb,

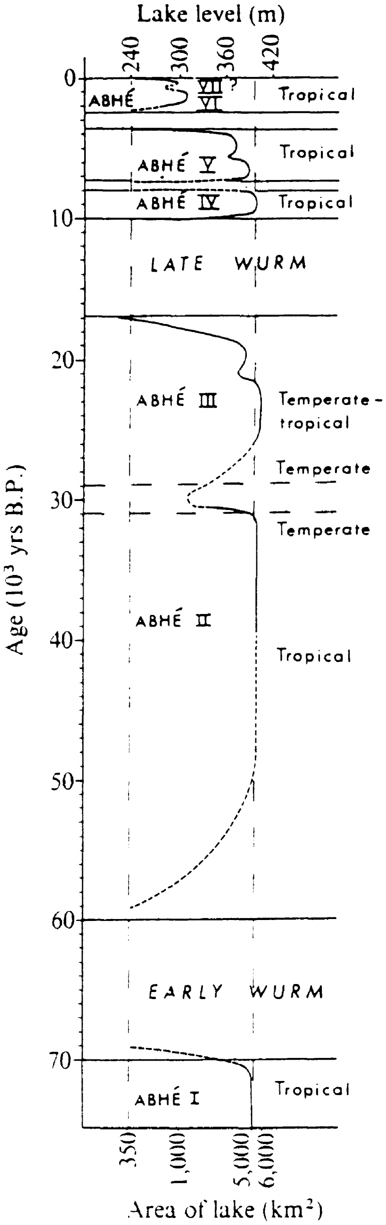

Fig. 3. Variations of Lake Abhé level and area during the late Pleistocene. Reprinted from Gasse (1977), with permission from *Nature*.

2001). The studied endoreic lake levels in the Rift Valley are a proxy for runoff from the highlands (Street, 1980). Such runoff volumes cannot, however, be interpreted as a straightforward indicator for total precipitation. Runoff coefficients should also be considered, which depend on rain seasonality and vegetation cover. Especially, man-induced land use change is a variable, which is rarely taken into account when explaining late Holocene lake level changes.

2.4. Fluvial and lacustrine deposits in Ethiopia and along the Nile

Sediment successions deposited in Rift Valley lakes (Ziway-Shala group, Turkana, Abhé) as well as in the Nile valley and delta inform on the palaeoclimate as well as on past erosion processes and rates in the highlands. Sediment production, especially under natural conditions, in a specific area, is dependent on annual and seasonal rain, as, e.g., expressed by Fournier's (1962) degradation coefficient (C) for large catchments (>2000 km^2):

$$C = p^2/P \tag{1}$$

where: p = monthly precipitation (mm) in the catchment during the wettest month; P = yearly precipitation (mm) in the catchment.

For short periods (in geological terms), variations in river load are associated with climatic fluctuations, rather than with oscillations in the volcano–tectonic processes (Williams and Adamson, 1980).

The Awash River has its origin in the highlands. After a long trajectory through the Rift Valley, it reaches the endoreic Lake Abhé where essentially chemical precipitation takes place. Between 50,000 and 17,000 years, only material in solution entered the lake (there was, however, a short period of detritic deposits around 30,000 years ago), signature of a wet phase with well-distributed rains and dense vegetation cover in the highlands. From 17,000 to 12,000 ^{14}C years BP, during the arid episode, large debris fans formed on the basin flanks (Gasse, 1977). As already mentioned, there was no inflow (and no lake) during that period. In the beginning of the Holocene, from 10,000 to 8000 ^{14}C years BP, sedimentary facies reflect very low turbidities, indicating that precipitation was again abundant and evenly distributed throughout the year (Gasse, 1977). Increased detritic deposits (silt, clay) during the middle Holocene, indicate more irregular and intense rains. At the upper Holocene, there is enrichment in clay and silt, corresponding to more arid conditions (Gasse, 1977) (Fig. 4). Stable isotope analyses of inorganic calcite in Lake Turkana, which drains Southern Ethiopia by the Omo River, show an increase in (heavier) ^{18}O isotopes between 4000 and 2000 ^{14}C years BP. This is explained by a change in climatic conditions in the Ethiopian plateau (Ricketts and Johnson, 1996). The change in isotopic composition could have resulted from a drop in water temperature, a fall in humidity, a drop in the percentage of advected air, an increase in wind speed, an increased influence of the Atlantic air mass on the Ethiopian highlands, or a combination of these. However, there is clearly a "shift in the climate between approximately 4000 and 2000 BP, resulting

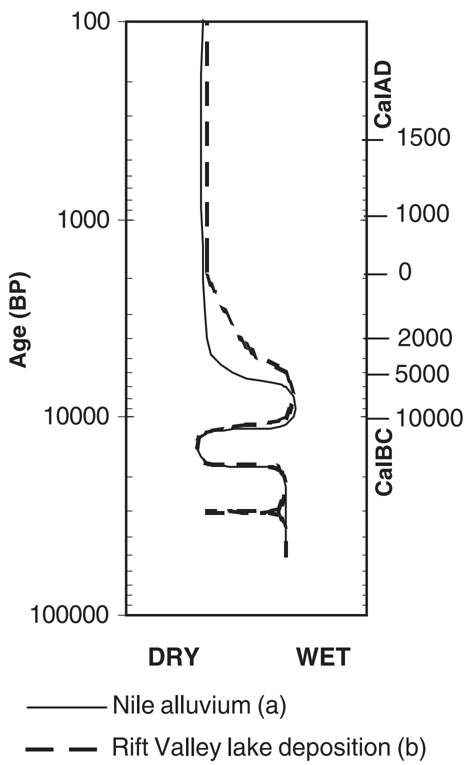

Fig. 4. Variations in annual rain, as derived from lake sediment deposits and Nile alluvium. Sources for (a) are Williams and Adamson (1980), Butzer (1980) and Williams et al. (2000); for (b) Laury and Albritton (1975), Gasse (1974, 1977), Gasse and Street (1978) and Ricketts and Johnson (1996).

in a new state unlike what existed prior to 4000 BP" (Ricketts and Johnson, 1996).

More studies have been carried out on Nile sediments. Although only 13% of the drainage area of the Nile is situated inside Ethiopia, it nevertheless contributes 80% of the annual discharge of the Main Nile (Misganaw, 1989). The variability of the discharges of the Main Nile thus largely reflects runoff from the Ethiopian highlands. Discharges during the late Quaternary were reconstructed among others through the analysis of river terraces and fluvial sediment composition.

Given the different regimes and origins of the affluents composing the main Nile, it must not surprise that their contribution to the sediment load differs greatly: the Atbara contributes actually 14×10^6 t yearly, the Abay or Blue Nile 41×10^6 t and the White Nile 2×10^6 t, most of which is brought down by the Sobat from the southwestern Ethiopian highlands (Williams and Adamson, 1980). With regard to the depositional history of the Nile, which cannot be discussed at length here, two review papers (Butzer, 1980; Williams and Adamson, 1980) are summarised below. From 25,000 to 18,000 ^{14}C years BP, there was deposition of 33 m Ethiopian silt in Nubia, 8 m near Cairo, with numerous sand channels (Butzer, 1980). During the dry and cold late Pleistocene, the Blue Nile transported an abundant coarse bedload and deposited sands in Central Sudan. Precipitation was less but more seasonal, runoff rates from barren surfaces were high (Williams and Adamson, 1980). For the main Nile, Butzer (1980) finds important summer inundations witnessed by deposits of gravel and braided river systems ("wild Nile" between 12,000 and 11,500 ^{14}C years BP).

Extensive dark clays from the Ethiopian highlands were deposited in Sudan during the early Holocene (11,200–6000 ^{14}C years BP), when there were higher temperatures and increased rain (Williams and Adamson, 1980; Butzer, 1980). This massive alluviation of allogenous material is also found in Nubia and the Nile delta during the same period. Around 6000 ^{14}C years BP, the Nile incises by 15 m, and since then, deposition of allogenous alluvial material occurred (Butzer, 1980).

Research has been carried out on correlations between Nile floods and climate in Europe (Hassan, 1981), El Niño Southern Oscillation (Quinn, 1992; Whetton and Rutherford, 1996; Eltahir, 1996; Amarasekera et al., 1997) and lunar and solar cycles (Fairbridge, 1984). The influence of vegetation cover changes and deforestation of the Ethiopian highlands on long-term variations in Nile discharges and sediment deposition is generally omitted in these studies. It might, however, be a useful study object, given the changes in vegetation cover throughout the late Quaternary.

2.5. Pollen analysis

Pollen diagrams provide useful indications on previous vegetation and climate as well as their evolution. Studies in Ethiopia started with late Tertiary data (Yemane et al., 1985; Bonnefille and Hamilton, 1986), but in this review only the upper Quaternary will be considered. Unfortunately, very few pollen data are available for the northern half of Ethiopia, neither of ancient deposits nor of modern pollen rain.

Knowledge of modern pollen rain, as it exists for SW Ethiopia (Bonnefille et al., 1993), is essential for the interpretation of fossil pollen floras. Bonnefille et al. (1993) found that in the three main types of forest (lowland woodlands, intermediate forests, afromontane forests), there is a reasonably good agreement between the floristic component and pollen rain, taking into account, however, that insect-pollinated species such as *Acacia* are underrepresented. Since sampling did not concern areas with secondary, degraded or artificial vegetation, there is only a sporadic occurrence of introduced species, such as Cupressaceae (other than *Juniperus procera*) and *Eucalyptus* in the pollen diagrams.

During the LGM, East African mountains had grassy/shrubby vegetation, including some forest refuges, as proven by the presence of tree pollen in low abundance (Jolly et al., 1998).

Most Quaternary pollen studies deal, however, with sediments deposited since the last glacial–interglacial transition period. In a highland peat (3000 m a.s.l.) in Bale, Mohammed and Bonnefille (1998) found between 13,000 and 10,500 ^{14}C years BP a pioneer treeless vegetation, consisting mainly of Umbelliferae, which colonised the deglaciated terrain (Fig. 5, pollen zone 3). The absence of arboreal pollen indicates that during that period, the forest limit was well below 3000 m. From 10,500 ^{14}C years BP on,

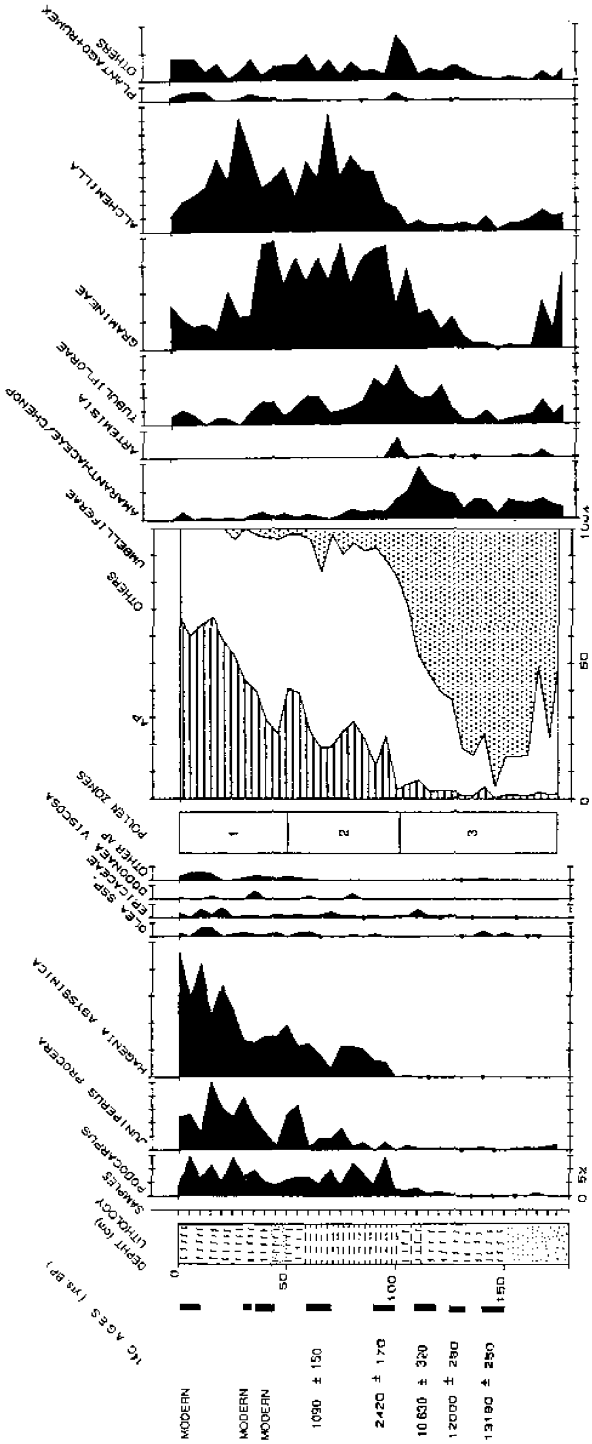

Fig. 5. Pollen diagram, Tamsaa (3000 m a.s.l.), Cyperaceae are excluded from the pollen sum; AP = arboreal pollen; vertical tildes = fibrous peat; broken lines = clayey peat; dots = silty sand; small and big dots = sandy gravel. Reprinted from Mohammed and Bonnefille (1998), with permission from Elsevier.

there is an increase of Chenopodiaceae and Amaranthaceae (Fig. 5, pollen zone 2). Bonnefille and Hamilton (1986) found such an increase in Chenopodiaceae around 11,500 ^{14}C years BP in a peat deposit in Arsi (3800 m a.s.l.) and interpreted it as an evidence for late Pleistocene aridity, the pollen being transported from the nearby beds of shrunken lakes in the Rift Valley. In this highland peat, there is an increase in montane tree pollen around 7500 ^{14}C years BP. Similar increase in tree pollen (and especially *Podocarpus*) in Lake Abiyata between 7000 and 5000 ^{14}C years BP is interpreted to the presence of a *Podocarpus* forest at 1600 m a.s.l. As the present-day lower limit is situated at 1800 m a.s.l., the presence of the forest at 1600 m a.s.l. would have required 20–40% more precipitation at that altitude than it is the case now (Lezine and Bonnefille, 1982). Another short increase of *Podocarpus* was registered around 3500–2000 ^{14}C years BP at 3000 m a.s.l. (Mohammed and Bonnefille, 1998). At these altitudes, there is a small pollen production, and most pollen is transported from lower altitudes. Bonnefille et al. (1993) observed, e.g., that contemporary *Podocarpus* pollen is transported by wind over 50–100 km. There is a meager knowledge of the ecology of *Podocarpus* pollen; palynologists do not agree if the rise of different *Podocarpus* species from 4000 ^{14}C years BP on, observed everywhere in East Africa, is due to a dryer climate (Hamilton, 1982) or to the wet pulse around 2500 ^{14}C years BP (Bonnefille and Hamilton, 1986; Mohammed and Bonnefille, 1991).

From 2500 ^{14}C years BP on, there is an important peat accumulation in Bale (Mohammed and Bonnefille, 1998), which corresponds to a moist phase during the generally dry late Holocene. There is a general rise in tree pollen, especially *Juniperus procera* and *Hagenia* (Fig. 5, pollen zone 2). During that period, there is also a greater content of arboreal pollen in Lake Turkana, which drains the southern Ethiopian highlands (Mohammed et al., 1995). After 1850 ^{14}C years BP, there are major pollen changes associated with disturbance by mankind (Bonnefille and Hamilton, 1986). Although the earliest agriculture is older, major forest destruction starts in that period, marked by a decrease of *Podocarpus* and an increase of secondary forest and scrub (*Dodonaea*, *Hagenia*).

Relative increase in *Juniperus procera* is thought to be due to selective felling. There is also a sharp increase in Chenopodiaceae, *Plantago* and *Rumex*, which commonly accompany human disturbance (Bonnefille and Hamilton, 1986). Near Dessie, Lamb (2001) found pollen evidence for forest clearing at 2500 ^{14}C years BP. Pollen associated with cultural debris dating from the same period also show striking dominance of non-arboreal taxa, including many ruderal species (Bard et al., 2000). In a core from the bottom of Lake Langano (Ziway-Shala basin), which is fed by three rivers draining the nearby Arsi highlands, the same tendencies are observed, with an increase in *J. procera* and *Podocarpus* between 1000 and 800 ^{14}C years BP. Increase in *Podocarpus* is considered as a sign of a slight humid trend. With respect to *J. procera*, it is not clear whether its increase is due to climate change or to selective deforestation (Mohammed and Bonnefille, 1991). Finally, a pollen record from a site in Arsi, situated near the upper tree limit, allows to recognise colder (predominance of Ericaceae) and warmer periods (predominance of arboreal pollen) since 3000 ^{14}C years BP (Fig. 6) (Bonnefille and Mohammed, 1994).

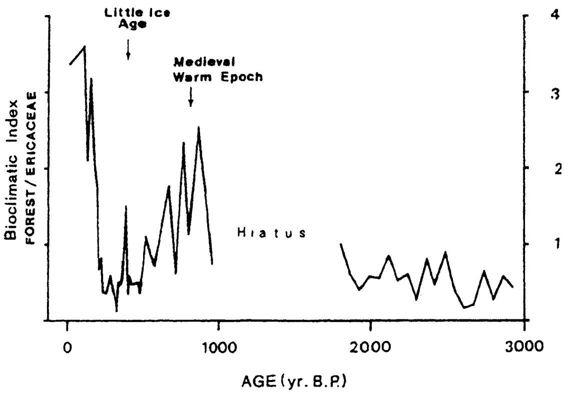

Fig. 6. The ratio between forest tree pollen and Ericaceae expresses the movement of the forest limit and is used as a bioclimatic index. During cold periods, the forest limit is lower and Ericaceae dominate. Reprinted from Bonnefille and Mohammed (1994), with permission from Elsevier.

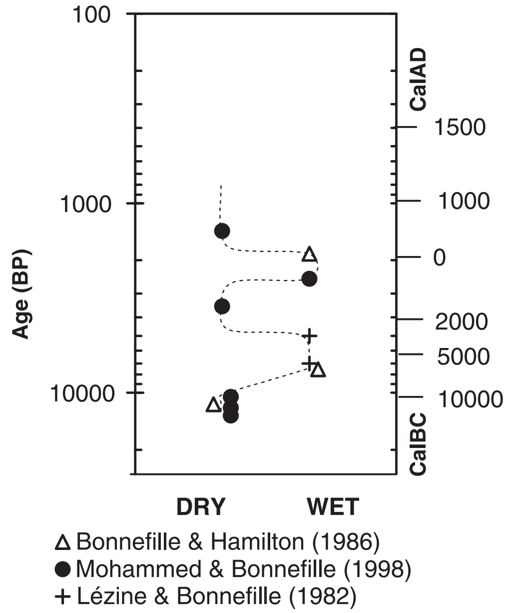

Fig. 7. Variations in annual rain in the Ethiopian highlands, as derived from pollen analysis.

Most cores sampled from lake sediments and peat bogs do not give data about the most recent deposits; on lake bottoms, for technical reasons, the upper few centimeters are often lost (Mohammed and Bonnefille, 1991). One should thus not be surprised that especially the introduced species do not appear in the pollen diagrams. Analysis of active freshwater tufa deposits, which fix pollen without allowing any disturbance by pedoturbation, might be a tool for the analysis of the most recent changes in vegetation (Nyssen et al., 2002b).

Overall, climatic conditions indicated by pollen analysis (Fig. 7) seem to follow the general tendency.

2.6. Soil formation

Soils and palaeosols are the result of a complex interaction between different factors and processes of the ecosystem. Rather than precise climatic statements, only general indications of pedogenetic conditions such as temperature and rain can be deduced from it. A well-developed soil, for instance, can either be the result of a short intensive soil formation process and/or of a long period of slow pedogenesis.

Interpretation of palaeosols necessarily includes existing knowledge about palaeo-temperature and -precipitation. It must be stressed that most reviewed studies use the term 'soil formation' to indicate strong rock weathering and mineralisation of soil organic matter, resulting generally in deep and genetically mature soils. Soil formation, however, also exists under dry conditions, e.g., by redistribution of soluble salts.

Except for Brancaccio et al. (1997) (see below), ages obtained for soil formation are either ages of overburdening sediment or ages of charcoal or pottery included in the soil, which are expected to be older than pedogenesis. Furthermore, often, samples were taken from vertic soils of which the churning properties are well known, and hence, there is a great risk that reworked samples have been dated (Nyssen et al., 2000a). Obtained ages allow illustrating general tendencies but should not be used for high-resolution analysis (Aitken, 1990; Lang et al., 1999).

Messerli and Frei (1985) analysed a palaeosol transect through the Saharan and East African mountains and compare their findings to Maley (1983). Palaeo-Vertisols, together with the deposition of fine clay-rich materials on footslopes and in lakes, were thought to correspond to a soil formation period between 12,000–10,000 and 7000 years ago, characterised by well-distributed rains. No chronological evidence, however, is available to support this theory. From 7000 to 4000 years, there would have been a seasonal distribution of rains with intensive summer rains and important runoff. This resulted in the formation of ferric Luvisols dominated by kaolinites, and transport of very coarse material by the rivers. Slope deposits were incised, which resulted in terrace formation.

As we will discuss later, recent work provided more details on this pattern for the Ethiopian highlands. Furthermore, from a theoretical point of view, this issue raises two questions:

(a) It is generally known that Vertisols need an ustic moisture regime to be formed (Mermut et al., 1996). This does not fit with the 'well-balanced rain distribution' between 10,000 and 7000 years ago. The vast and thick Vertisols of the Ethiopian highlands would rather not have been developed during this period. This leads to the question if it is justified to correlate soil formation periods in the Eastern African highlands to those found by Maley (1983) around Lake Chad.
(b) It is unclear why the formation of Luvisols and Vertisols, on the tertiary basalt of the Ethiopian highlands, is thought to be mutually exclusive. Their position on the slope is very different, in the upper part of the catena for the Luvisols, on toeslopes and valley bottoms for the Vertisols (Driesen and Dudal, 1991; Nyssen et al., 2000a). According to climatic conditions, Luvisol formation can coincide with the formation of coarse colluvial deposits on steeper slopes or of Vertisols and other vertic soils in bottomlands.

Verheye (1978) analysed a soil chronosequence on Lake Ziway terraces: Haplustalfs (USDA, 1975) or Haplic Luvisols (FAO et al., 1998) on the higher terrace level formed around 9000 ^{14}C years BP are witnesses of a wetter environment. Soils on the lower terraces are poorly developed and illustrate that pedogenesis is low under the present-day climatic conditions. More developed soils on the intermediate terraces can be correlated with a climate somewhat moister than the present.

Research on palaeosols in Ethiopia is mainly carried out in the northern highlands. Brancaccio et al. (1997) dated many palaeosols by ^{14}C dating on humic samples extracted by NaOH leaching. Returned ages are between 8300 and 4480 ^{14}C years BP, "a period with thick vegetation cover." They insist that all the analysed palaeosols are strongly truncated, which means that most probably, soil formation took place even beyond that last date. In the Afar depression and near Dessie, Semmel (1971) dated charcoal in 'Latosols' (Petric Plinthosols, FAO et al., 1998) and 'Tirs' or Vertisols (Deckers et al., 2001). The humid soil formation period returned dates between 8380 and 3670 ^{14}C years BP. Hurni (1982) dates palaeosols at 3710 and 4570 ^{14}C years BP in Bale, as well as 4120 in Simen. Sagri et al. (1999) see the moist climate of 10,000–5000 ^{14}C years BP with geomorphic stability as ideal conditions for active pedogenetic activity resulting in widespread well-developed soils. Mohammed (2001) finds several gaps in the early Holocene soil formation period, corresponding to the abrupt arid intervals, which are also deduced from lake levels.

Growing aridity, from 5000 to 4000 ^{14}C years BP on, provokes soil erosion and deposition, which covers existing soils, as witnessed by analysed palaeosols (Messerli and Frei, 1985; Ogbaghebriel et al., 1997; Sagri et al., 1999). Brancaccio et al. (1997) relate burial of soils under alluvial and colluvial sediments to the period of deforestation of the primary forest. The ages of the observed buried soils range between 5160 ± 80 (May Makden, Tigray) and 300 ± 60 ^{14}C years BP (Adi Kolen, Tigray) (Fig. 8). Charcoal sampled from similar colluvial deposits in Tsigaba (Tigray) returned a ^{14}C age of 3090 (Nyssen, 2001). In Adi Kolen, a strong degradation only occurred after 300 ^{14}C years BP, whereas degradation near May Makden started early, what can be related to the finding of remnants of dwellings under 5 m of colluvial sediments (Ogbaghebriel et al., 1998). Soil formation, by weathering and mineralisation of soil organic matter, continues in those areas

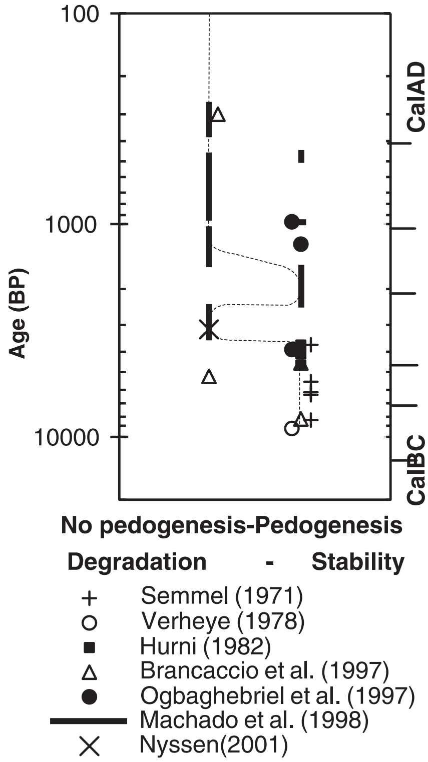

Fig. 8. Periods of soil formation and degradation as derived from the stratigraphy of colluvial deposits.

covered by forests and afroalpine grass–steppe (Messerli and Frei, 1985). All this seems to confirm that present-day environmental conditions began during the second millennium BC (Brancaccio et al., 1997) (Fig. 8).

Machado et al. (1998), using detailed profile descriptions of infilled valley bottoms in the Axum-Adwa area, present a high-resolution sequence of the main degradation and stability phases in Northern Ethiopia over the last 4000 years. More humid soil formation periods existed between 2500 and 1500 ^{14}C years BP, as witnessed by two levels of Vertisol and vertic soils, expected to correspond to the above documented wet pulse with high lake levels and increased arboreal pollen. The more arid degradation periods with coarse sediment transport from slopes into valleys (3500–2500 and 1500–1000 ^{14}C years BP) (Machado et al., 1998) were interrupted by the aforementioned period of soil formation around 2000 ^{14}C years BP, which was also recognised in Central Ethiopia by Sagri et al. (1999). Machado et al. (1998) inferred increasing aridity after 1000 ^{14}C years BP from an anomalous increase of aggradation phenomena and interpret it as a result of intensive land use. Within this arid period, Ogbaghebriel et al. (1997) and Machado et al. (1998) interpret palaeosols as witnesses of short wetter intervals. However, besides the already mentioned uncertainties on soil formation dating, the calibration of radiocarbon ages obtained for shallow palaeosols at different places in the Adwa area (480 ± 30, 380 ± 30, 290 ± 30, 140 ± 70 ^{14}C years BP) (Machado et al., 1998) yields important overlaps, covering the period 1400–1960 AD, which makes it impossible to point to one or two ‘humid phases.’ To our understanding, such short, localised soil formation periods can as well be seen as the expression of local evolution of forest cover, including temporary reforestations. Recent observations show that within only 15 years after shrub regrowth, a 40-cm-thick OM-rich layer can be deposited (Nyssen, 2001). Similarly, Brancaccio et al. (1997) see a correspondence between soil formation around 1250 and 970 ^{14}C years BP and decreased population at the end of the Axumite kingdom around 1000 AD (Michels, 1990). Taking into account the uncertainties mentioned, Fig. 8 shows the major periods with presence or absence of pedogenesis.

2.7. Past geomorphic processes

In a number of studies dealing with Ethiopia, geomorphic processes active in the past have been recognised and explained, although few events have been dated. We shall proceed chronologically through the late Quaternary, but poor dating of the palaeomorphological processes, with the notable exception of lake terraces, prevents us from making a high-resolution analysis.

It is generally accepted that, except at the margins of the Rift Valley, the present-day stepped morphology (Fig. 9) of the Ethiopian highlands is due to differential weathering of rocks with variable hardness. Abul Haggag (1961) demonstrated that the different flat surfaces are part of a structural system, and not a witness of successive erosional cycles (Davis' cycles), as was for example defended by Desio (1940).

Selective weathering explains also the existence of landforms within one geological formation, such as the intra-montane plateaux in Godjam, eroded in basalt formations (Semmel, 1963).

With respect to the omnipresent escarpment-debris slope–footslope–toeslope morphology, Riché and Ségalen (1971), in the Wabi Shebele headwaters, see the redistribution of material, originating from retreating escarpments, under the form of coarse deposits on the steeper slopes and fine material on the footslopes, as an actual process. In the same sense, Ogbaghebriel and Brancaccio (1993) think that the pediments between hills/mountains and highland plains in Central Ethiopia are the result of very recent sheetflood processes: the pediments are not much dissected and sheetflood can only occur if the catchment is devoid of vegetation. These pediments are 'commonly veneered with rock waste in transit and deposited during the waning stage of floods' (Ogbaghebriel and Brancaccio, 1993).

It is on such footslopes, south of Harar, that Riché and Ségalen (1973) observe decimetre-thick stonelines in many profiles, which are recovered by soft material, among others by Vertisols. This is explained by the present-day ustic climate: erosion occurs on steep slopes, fine material is transported further downslope covering previously eroded coarse material. No variation in climate or land use is invoked. From the analysis of profiles in the Tigray highlands, we concluded (Nyssen et al., 2000a) that these concentrations of stony material, overlain by fine material on footslopes are witnesses of a chronological succession of processes. Slope deposits by mass movements underlie fine material.

Fig. 9. The subhorizontal structural topography induced by the presence of alternating hard and soft layers is a common view when flying over the Ethiopian Highlands. Amba Raesat, Dogu'a Tembien. Photo B. Muys.

Recent research showed furthermore that the main processes resulting in high rock fragment covers in the Ethiopian highlands are tillage-induced kinetic sieving, argillipedoturbation in Vertisol areas and differential water erosion on sloping areas (Nyssen et al., 2000a, 2001b).

Abul Haggag (1961) analyses river terraces in the northwestern highlands. Especially along the lower reaches of the rivers draining the highlands to the west (Barka, Setit and Takazze), extensive accumulations of alluvia cover the plains and have buried the bases of the hills. Nowadays, the rivers in these valleys do not carry any water during most of the year. The alluvium can only have been deposited during a wetter period when the rivers had a greater ability to transport materials eroded from the highlands over longer distances, at least during a certain period of the year. From more recent research, it is clear that this pluvial period did not correspond to the late OIS 3 (as thought by Abul Haggag, 1961). The fact that this alluvium is not yet incised over its whole thickness also demonstrates that the shift to the present-day runoff conditions is quite recent. Eastbound runoff to the Rift Valley shows the same image (own observations). No chronology for these deposits could be found in literature.

Messerli and Frei (1985), Ogbaghebriel et al. (1997), Machado et al. (1998) and Sagri et al. (1999) see, with the coming of the arid/degradation phase around 5000 ^{14}C years BP, soil erosion accelerated in the Central highlands. Toposequences analysed in a remnant forest and its surroundings in Hechi (Dogu'a Tembien) show the impact of deforestation and subsequent land use changes on soils. Nyssen (2001) measured an average soil surface lowering between 30 (on a structural flat area) and 90 cm (on slopes >0.3 m m^{-1}) since deforestation.

The arid phase was interrupted by short wet or drier periods. Around 400–250 ^{14}C years BP, the Little Ice Age was arid in Ethiopia: there was aggradation of ephemeral streams, a sign of increased slope erosion together with decreasing stream power.

Finally, some observations about karstic phenomena are reported. Though many of the Ethiopian caves have been described (Catlin et al., 1973), little is known about their speleogenesis. The same authors also associate the large tufa dams and tufa deposits along spring lines in Tigray with a different climatic regime, "possibly such as existed in the pluvials of the Pleistocene" (Catlin et al., 1973). Voigt et al. (1990) in Northern Somalia date the most recent tufa deposits at 9105 ^{14}C years BP. Ogbaghebriel et al. (1998) analyse a tufa dam some 10 km north of Mekelle. Sediment deposited behind the dam was dated 7310 and 5160 ^{14}C years BP; no evidence is given that these deposits are contemporaneous with the dam. Brancaccio et al. (1997) estimate that the period of freshwater tufa buildup started during the more humid and mild climatic conditions at the end of OIS 3. Ogbaghebriel et al. (1998) expect that the end of the tufa deposition is due to a reduction of the vegetation cover, which provoked a decrease of CO_2 in runoff water. Dramis et al. (1999) link the end of tufa precipitation in Ethiopia, Northern Africa and Europe to delayed groundwater temperature increases after the LGM. The analysis of late Quaternary tufa dams in Dogu'a Tembien shows that the final stages of their buildup coincides with and is probably due to deforestation (in a period starting before around 1430–1260 BC), which led to increased sediment transport by rivers (Nyssen, 2001). The shift from perennial to seasonal discharge would also have decreased the length of the yearly period of buildup and increased the erosive power of flash floods on tufa (Virgo and Munro, 1978).

Soils developed on the sediment deposits behind the dam; the same climatic conditions are favourable for both soil development and tufa deposition. The tufa deposits are then partially covered by alluvio-colluvial material, which is thought to be a witness of the shift to dryer conditions, but also of progressive clearing of natural vegetation induced by the growth of Axum from 1500 BC on (Brancaccio et al., 1997; Ogbaghebriel et al., 1998). These authors thus conclude that the effectiveness of human impact increases with drier conditions.

2.8. Synthesis

(Peri-)glacial processes occurred in the mountain areas above 3000 m a.s.l. during the late Pleistocene (before 12,000 ^{14}C years BP). It was a cold and arid period, with low lake levels in the Rift Valley; large debris fans were formed on the flanks of Lake Abhé basin and the Blue Nile transported coarse bedload. At the end of the glaciation (shift from arid to moist phase) in the Ethiopian highlands, large volumes of sediment were produced (Hamilton, 1982; Hurni, 1989).

A long period with abundant and evenly distributed rains existed between 11,500 and around 4800 ^{14}C years BP (see Fig. 10), as witnessed by increase in arboreal pollen, high lake levels (with, however, wet pulses and more arid intervals). There were low turbidities in the rivers, large Nile discharges with deposition of dark clays in Central Sudan, Nubia and the Nile delta. There were thus low erosion rates du-ring the early Holocene. It was also a period of palaeosol formation.

Around 5000–4800 ^{14}C years BP, Ricketts' and Johnson's (1996) "shift in climate" got into motion. It must be stressed here that from now onwards, many phenomena which are interpreted as 'climatic' might well be of anthropic origin. Low lake levels, variable Nile levels and silt transport by the lower Awash point in the direction of arid conditions. There are slope instability and soil erosion. Coarse sediments are transported from slopes into valleys. Terraces of rivers draining the highlands are formed during the beginning of this period.

Short high lake levels between 2500 and 1500 ^{14}C years BP, peat accumulation in intra-montane depressions, rise in tree pollen in Southern Ethiopia indicate probably a wet pulse. There were exceptionally high Nile stands and a soil formation period was recognised in Northern and Central Ethiopia (Machado et al., 1998; Sagri et al., 1999). This and other minor changes during the middle and late Holocene are caused both by climatic and land use changes.

The data and profile descriptions obtained by Brancaccio et al. (1997) and Ogbaghebriel et al. (1998) strongly suggest a primary relation between deforestation and land degradation on one hand and human occupation on the other. Machado et al. (1998) reached, however, the conclusion that "climate is the main long-term driving factor of environmental change in the region." In earlier work (Machado et al., 1995, 1997), the degradation episodes were thought to correspond to large-scale human intervention (introduction of plough and cereals, decline of the Axumite kingdom, 18th–19th centuries). It is striking that both the Spanish (Machado et al.) and the Italian groups (Brancaccio, Billi, Dramis, Ogbaghebriel, Sagri, ...) who worked quite independently on palaeosols in North and Central Ethiopia seem to hesitate about the interpretation of their data: are recent oscillations in land degradation intensity mainly due to human impact or do they have a climatic origin? Possible links between ancient deforestation in the Ethiopian highlands, decreased evapotranspiration and dryer climatic conditions have never been explored.

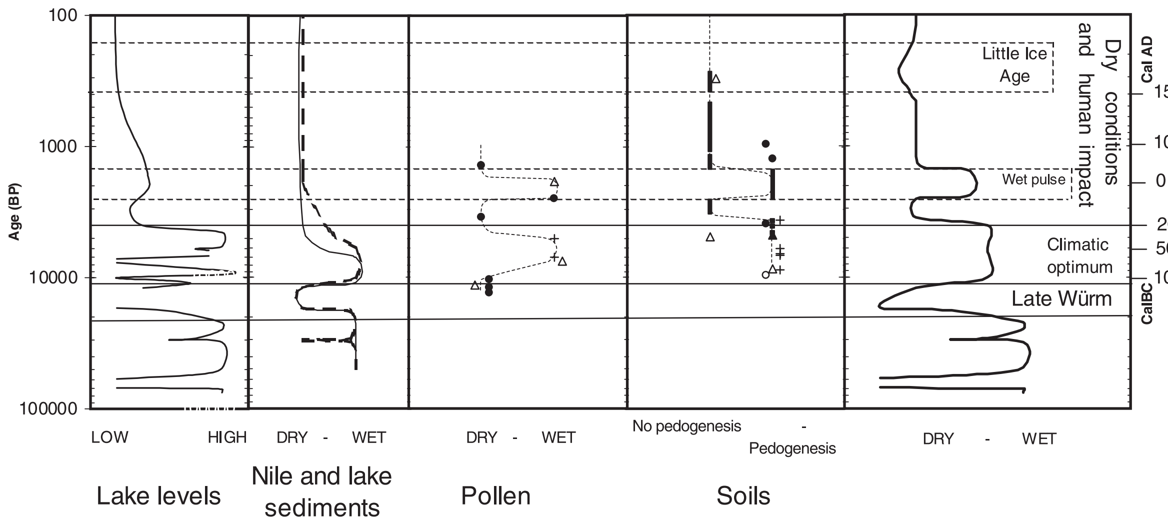

Fig. 10. Palaeoenvironmental indicators in the Ethiopian highlands: synthesis. All curves oscillate between arid, degradational (left) and more stable (right) situations.

From 2500 to 1850 ^{14}C years BP onwards, pollen diagrams reveal important human impact through an increase in secondary forest and scrub as well as in ruderal species such as Chenopodiaceae, *Plantago* and *Rumex* (Bonnefille and Hamilton, 1986; Bard et al., 2000). The conditions become even more arid after 1000 ^{14}C years BP, when there are low lake levels and an anomalous increase of aggradation phenomena, locally interrupted by short periods of soil formation (Machado et al., 1998). Since 4000 ^{14}C years BP, at least there is major impact of human activities, although it is unclear if soil formation by mineralisation of soil organic matter should be ascribed to climatic 'noise' (e.g., short wet pulses—Machado et al., 1998) or to a nonlinear deforestation process.

The Little Ice Age (15th–18th centuries), its permanent snow on Simen and its aridity witnessed by the aggradation of ephemeral streams, might be seen as one of the periods were there is a short climatic interference with what is increasingly considered as a man-induced land degradation process.

3. Contemporary erosion processes and their driving forces

The magnitude of erosion processes in the Ethiopian highlands finds its cause in the combination of erosive rains, steep slopes due to the quick tectonic uplift during Pliocene and Pleistocene and human impact by deforestation, an agricultural system where the openfield dominates, impoverishment of the farmers and stagnation of agricultural techniques (Ståhl, 1974, 1990; Girma and Jacob, 1988). Most reports from the first half of the 20th century (e.g., Giglioli, 1938a,b; Joyce, 1943) recognise the soil erosion problem but do not estimate that it is a major problem for agricultural development. More recently, in their controversial paper, Bojö and Cassells (1995) stress that nutrient loss is more important than soil loss. It should be underlined that despite the aforementioned unfavourable conditions, a thick soil cover is still present on most of the Ethiopian highlands, which can be attributed to overall low soil erodibility, high rock fragment cover and awareness of the soil erosion problem by the farmers (Nyssen, 2001). Fig. 11 (after Oldeman et al., 1991) gives an interpretation of the state of anthropogenic soil degradation in Ethiopia and Eritrea. According to this study, the most important soil degradation processes occurring in the highlands are loss of topsoil due to water erosion (Wt) and water-induced terrain deformation/mass movements (Wd). It should be stressed that Wd, besides landsliding, also includes gullying. The description of this last unit can lead to misinterpretations: i.e., an overestimation of the importance of mass movements (see Keyzer and Sonneveld, 1999). Sheet erosion occurs almost everywhere in Ethiopia and Eritrea, whereas linear erosion features are more restricted to the highlands. In the Rift Valley and the peripheral lowlands, wind erosion (Et) is also a problem. It is evident that such a map is based on partial observations and should not be considered final. Assefa (1986) published another soil degradation map for Ethiopia.

3.1. Rain and runoff as driving forces for erosion processes

The climates of Ethiopia are complex: "Within short horizontal distances, climates from tropical to subhumid, and subtropical to arctic can occur" (Krauer, 1988, p. 19). Rain and temperature vary mainly with elevation. Furthermore, at a given altitudinal level, precipitation decreases and seasonality increases with latitude.

3.1.1. Rain in relation to seasonal circulation patterns

During the winter in the northern hemisphere, the Intertropical Convergence Zone (ITCZ) is situated to the south of the equator in Eastern Africa. At this time of the year, the western highlands of Ethiopia receive hot and very dry winds from the Sahara. On the other hand, the Red Sea coast and the eastern part of the country are under the influence of east winds, with high moisture content after their journey over the Indian Ocean. These winds bring also spring rains in the southern part of Ethiopia. From March till May, intense rains, particularly in regions situated at high altitude, accompany the movement to the north of the ITCZ and the equatorial air masses. These rains are provoked by the convergence of humid equatorial air and colder extratropical air (Suzuki, 1967; Troll, 1970; Daniel, 1977; Messerli and Rognon, 1980; Goebel and Odenyo, 1984).

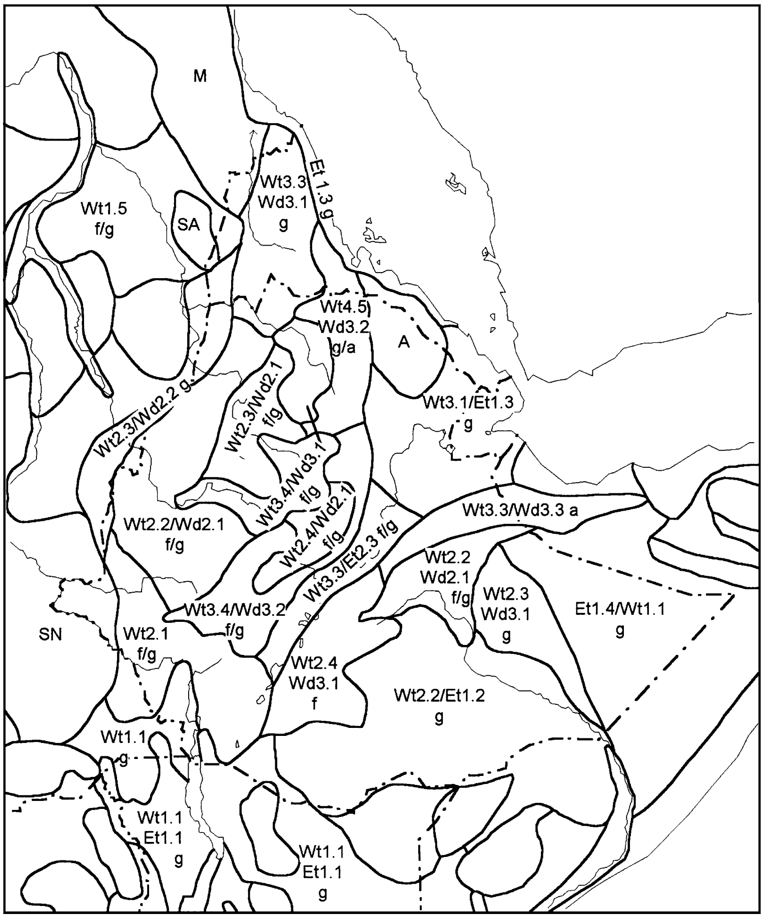

Fig. 11. Extract of the GLASOD map (Global Assessment of Soil Degradation, after Oldeman et al., 1991), which represents the main anthropogenic soil degradation processes for each major physiographic unit of Ethiopia. Wt = loss of topsoil due to water erosion; Wd = terrain deformation by water erosion, especially rill and gully formation, and mass movements; Et = loss of topsoil due to wind erosion; SN = stable terrain under natural conditions; SA = stable terrain with well-managed permanent agriculture; M = arid mountain regions; A = deserts. The first figure after the soil degradation process symbol refers to the degree of the process (1 = light; 4 = extreme), whereas the second figure refers to the area percentage of the physiographic unit that is affected by the process (1 = up to 5%; 5 = over 50%). The letter under the soil degradation symbol refers to the main cause (f = deforestation, g = overgrazing, a = agricultural activities). For a detailed key to the symbols and for more information regarding the methodology used we refer to Oldeman et al. (1991).

From the end of June onwards, the ITCZ is situated at its most northerly position (16–20°N). The southeast monsoons, limited to the lower layers of the atmosphere, bypass the highlands by the south and reach them from the west, provoking the rainy season (Goebel and Odenyo, 1984). Other authors attribute the origin of these humid air masses, coming from the west, to the Atlantic Ocean. They would pass over the equatorial forest where vapour is taken in (Hemming, 1961; Suzuki, 1967; Daniel, 1977; Rudloff, 1981). Generally, clouds are formed at the end of the morning, as a result of evaporation and convective cloud formation due to daytime heating of the soil, and it rains in the afternoon. In Afdeyu station, on the Eritrean highlands, 80% of daily precipitation takes place between 12 and 16 h (Krauer, 1988). Slopes exposed to the west, to the afternoon sun, receive less sunshine. This convective nature of rain also explains why individual showers have a very local distribution.

At the end of the summer, the ITCZ returns quickly to the south, preventing the arrival of monsoons. This is the end of the rainy season in the highlands.

3.1.2. Annual precipitation

Abebe and Apparao (1989) calculated from 241 stations in Ethiopia a mean annual precipitation of 938 ($\pm$83) mm year^{-1}. For the highlands, annual rain varies between 600 mm year^{-1} in Tigray and more than 2000 mm year^{-1} in the southwest (Krauer, 1988). The interaction of latitude and altitude in controlling total annual precipitation is described as follows (Troll, 1970): "The highlands of Northern Ethiopia represent, from the point of view of climate, the mountainous facies of the Sudanese zone; as in the Sudan plain, total rain and the length of the rainy season decrease from south to north, whereas the annual magnitude of the temperature increases. Moreover, because rain increases with altitude, and evapotranspiration decreases, due to lower temperatures, mountain areas are much moister and have a much longer rainy season than lowland areas at the same latitude." At the regional scale, one should, however, also take into account the facts that during the rainy season, winds essentially come from the west, as well as orographic effects: valleys are preferred flow paths for the penetration of humid air masses in the highlands (Nyssen et al., 2003a).

3.1.3. Rain erosivity in the Ethiopian highlands

High rain erosivity is an important factor of soil erosion in the highlands. Data of automatic rain gauges installed in Central Tigray during 1 year indicate that 65–77% of rainfall has an intensity >25 mm h^{-1} (HTS, 1976, p. 30). Krauer (1988) obtains (from the rain data of six Soil Conservation Research Programme [SCRP] stations) mean annual Universal Soil Loss Equation (USLE) erosivity indices R between 166.6 (Afdeyu, Eritrea) and 543.7 J cm m^{-2} h^{-1} year^{-1} (Anjeni, Godjam). Hurni (1979), in an analysis of rain erosivity in the Simen Mountains, insists on two other particularities of Ethiopian mountains: erosivity due to hailstorm (2.5 times more important than erosivity due to rain) and the influence of hillslope aspect. A surface unit exposed to the wind receives a greater quantity of water than a surface unit with an opposite exposure. Hurni (1988) presents a trigonometric equation allowing the calculation of the effective rain received on sloping soil surfaces. This equation assumes knowledge of aspect and slope, both for rain and topography.

Given that rain characteristics are very different, it is difficult to apply erosivity equations, such as those proposed in (R)USLE (Wischmeier and Smith, 1978; Renard et al., 1997), which have been developed for North America, to rain on the Ethiopian highlands. Based on drop size measurements, Nyssen et al. (2003a) showed that for the same rainfall intensity, rain erosivity is higher in the Ethiopian highlands compared to elsewhere in the world. Moreover, in the absence of a network of automatic rain gauges, maximum hourly rain intensity could only be measured in a small number of research stations in Ethiopia.

Rain erosivity is a function of the depths and intensities of the individual rainstorms, and these are not closely related to annual precipitation. However, the United States data indicate that for a given annual precipitation the range of likely erosivity values can be somewhat narrowed by knowledge of the general climatic conditions in the particular geographic area (Wischmeier and Smith, 1978). In East Africa (i.e., Tanzania, Kenya, Uganda), the relationship between total rain and erosivity index improves if rain stations are grouped by geographic area (Moore, 1979). For Ethiopia, Hurni (1985) and Krauer (1988) elaborated, from monthly data of six SCRP stations, correlations between USLE's R-factor and mean annual rain, and Krauer (1988) presents an isoerodent (rain erosivity) map of Ethiopia.

3.1.4. Runoff and infiltration

Active agent of soil erosion, runoff has been measured in Ethiopia at various temporal and spatial scales (from runoff plot to catchment). Runoff has been monitored in the SCRP catchments and analyses of up to 12 years long series are available (SCRP, 2000). Generally, runoff coefficients (RC) from small (< 1000 m^2) runoff plots are very variable (0–50%) (Table 2), which is attributed to the variable experimental conditions. Besides different slope gradients, local differences in soil texture, land use, vegetation cover, organic matter content or rock fragment cover result in a wide range of infiltration rates obtained from runoff plots (Feleke, 1987; Mwendera and Mohamed, 1997). Results from runoff plots may

Table 2
Runoff data for selected experimental plots in the Ethiopian highlands

Location	Slope gradient (%)	Area (ha)	Period	Mean annual precipitation P (mm)	Mean annual runoff R (mm)	Runoff coefficient ($100 \times RP^{-1}$)	Land use	Source
Melkassa	10–11	0.008	2 years	806		45.5	bare fallow	Feleke, 1987
Afdeyu	31	0.018	2 years	475	240	50.5	arable	Bosshart, 1997[a]
Afdeyu	31	0.018	2 years	475	115–155	24.2–32.6	arable with SWC	Bosshart, 1997[a]
Debre Zeit	4–8	0.002	74 days	350[b]	80	22.9	very heavy grazing	Mwendera and Mohamed, 1997[a]
Debre Zeit	4–8	0.002	74 days	350[b]	22	6.3	no grazing	Mwendera and Mohamed, 1997[a]
Debre Zeit	0–4	0.002	74 days	350[b]	34.5	9.9	very heavy grazing	Mwendera and Mohamed, 1997[a]
Debre Zeit	0–4	0.002	74 days	350[b]	7.3	2.1	no grazing	Mwendera and Mohamed, 1997[a]
Maybar[c]	28	0.018	2 years	1049	16.3	1.6	tradit. cultivation	Mulugeta, 1988
Maybar[c]	28	0.018	2 years	1049	4.5	0.4	id, with grass strips	Mulugeta, 1988

[a] RC calculated from data presented by the author.
[b] In 74 days.
[c] West of Dessie.

certainly not be extrapolated to catchments. At the scale of small catchments, some relatively low runoff coefficients are explained by specific physical conditions (Table 3). In the case of the Dombe catchments, W of Lake Shala: dense vegetation between the fields facilitates infiltration (Bosshart, 1998a). Moreover, this author finds a significant difference in runoff volume between these two similar catchments, of which one has been treated by physical conservation measures and the other not. Small catchments are sensitive to human intervention. In Afdeyu (Eritrea), there is low precipitation and a large grass-covered floodplain, which decreases runoff velocity (Bosshart, 1997). However, in the openfield of the northern highlands, especially at higher, rain-rich elevations, such as in Anjeni (Bosshart, 1998b), runoff depth is important, both in absolute and in relative (RC) terms.

Large catchments ($A \geq 100$ km^2) show decreasing runoff coefficients with increasing catchment area (Fig. 12). The already mentioned conditions for high

Table 3
Runoff data for selected small watersheds in the Ethiopian Highlands

Catchment	Altitude range (m)	Area (km^2)	Period (years)	Mean annual precipitation[a] P (mm)	Mean annual runoff[a] R (mm)	Runoff coefficient ($100 \times RP^{-1}$)	Source
Anjeni (Godjam)	100	1.13	10	1615.8 (± 238.4)	731.0	45.2	Bosshart, 1998b
Dombe[b] (without SWC)	125	0.94	12	1308.0 (± 249.4)	246.3 (± 158.2)	18.8	Bosshart, 1998a
Dombe[b] (with SWC)	105	0.73	12	1308.0 (± 249.4)	148.7 (± 111.1)	11.4	Bosshart, 1998a
Hunde Lafto	352	2.37	11	935	80	9	Herweg and Stillhardt, 1999
Maybar (W of Dessie)	328	1.13	12	1211	324	27	Herweg and Stillhardt, 1999
Andit Tid (Ankober)	504	4.77	10	1379	754	55	Herweg and Stillhardt, 1999
Dizi (Western Ethiopia)	224	6.73	4	1512	73	5	Herweg and Stillhardt, 1999
Afdeyu (Eritrea)	210	1.61	7	382.5 (± 123.7)	19.6 (± 16.5)	5.1	Bosshart, 1997

[a] Standard deviation between brackets.
[b] West of Lake Shala.

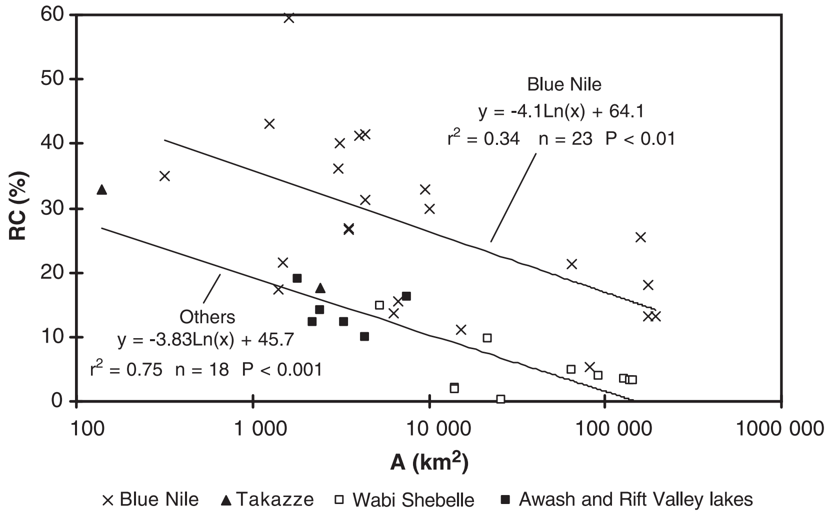

Fig. 12. Runoff coefficients (RC) vs. drainage area (*A*) for catchments of the basins of the Blue Nile (USBR, 1964; Daniel, 1977; Conway, 2000a), Takazze (HTS, 1976), Wabi Shebele (Bauduin and Dubreuil, 1973), Awash and Rift Valley lakes (Daniel, 1977; Fekadu and Bauwens, 1998; Vallet-Coulomb et al., 2001). See Fig. 1 for location. Observation periods were from 3 to 30 years, with the exception of USBR (1964) data, which cover 1 or 2 years (Conway, 1997).

RC (presence of openfield and much rain) are mainly found in the Blue Nile basin. For this reason, two data series can be considered. The Takazze, Awash and Wabi Shebele basins are mainly situated in dry sub-humid to arid regions (Engida, 2000). In the Wabi Shebele basin, Bauduin and Dubreuil (1973) explain decreasing RC with increasing catchment size by the fact that the small catchments are mostly situated in the headwaters where nearly impervious, basalt-derived soils dominate, and also by a smaller mean annual basin precipitation in the larger catchments which include (semi)arid lowlands. The rain and runoff data for catchments of the Blue Nile basin suffer, according to its authors (USBR, 1964), from lack of precision in the drainage areas (*A*) for smaller basins. Representative catchment precipitation data are also difficult to obtain given poor station density and great spatial variability of rain (Conway and Hulme, 1993). Conway (1997) stresses also the short average recording period (1.5 years) and possible errors in rain data. Despite the wide scatter for the Blue Nile basin, it can be observed that RC are larger than in the other basins but that they follow a parallel trend (Fig. 12). Decreasing RC values with increasing *A* values in the Blue Nile basin are thought to be a result of (a) runoff transmission losses due to evaporation and possibly lithological changes and (b) less rain and larger potential evapotranspiration in the western areas of the Blue Nile catchment along the border with Sudan. These areas reduce the overall runoff depth for the whole catchment (Conway, pers. comm., 1999). Here, we touch the problem of scale dependency, which will be discussed when reviewing sediment yield estimations (Section 4.1).

3.2. Weathering and soil formation

Few studies have been made on weathering of parent material in the highlands. Hövermann (discussion in Bakker, 1967) studied the basal Precambrian granites in Northern Ethiopia where weathering is up to 120 m deep. No studies exist for Mesozoic sedimentary rocks or for Tertiary volcanics, but depth of weathering mantle is obviously much less.

Hurni (1983a,b), through the study of soil developed on periglacial slope deposits, extrapolated soil formation rates for the different agroclimatological zones of Ethiopia (Table 4). Zonation in Ethiopia is based on altitude and, more specifically, on the climate gradient, which it provokes. These soil formation rates are mean rates, taking into account rain and temperature conditions, but not lithology. They are intended to

Table 4
Average soil formation rates in Ethiopia (after Hurni, 1983a,b)

Agroclimatological zone[a]	Altitudinal limits (m a.s.l.)	Annual rain (mm $year^{-1}$)	Soil formation rate (t ha^{-1} $year^{-1}$)
High Wurch	>3700		2
Wet Wurch	3200–3700	>1400	4
Moist Wurch	3200–3700	900–1400	3
Wet Dega	2300–3200	>1400	10
Moist Dega	2300–3200	900–1400	8
Wet Woina Dega	1500–2300	>1400	16
Moist Woina Dega	1500–2300	900–1400	12
Dry Woina Dega	1500–2300	< 900	6
Moist Kolla	500–1500	900–1400	6
Dry Kolla	500–1500	< 900	3
Berha "desert"	< 500		1

[a] Delimitation of agroclimatological belts after Daniel (1977) and Hurni (1986).

be used in comparison with soil loss rates, but cannot, to our understanding, be applied to the vast areas where the soil mantle results from sediment deposition rather than from pedogenesis.

3.3. Sheet and rill erosion

Most research on soil erosion in Ethiopia deals with sheet and rill erosion (Fig. 13). Hurni (1975, 1978, 1979) studied thoroughly the Jinbar valley (3200–4000 m) in the Simen Mountains. Andosols occupy the whole valley, which is partially under rangeland and degraded forest and partially under barley. The depth of the A-horizon was measured at some 300 sites in cropland and compared with A-horizon depth in uncultivated areas for slopes with similar gradient. Mean total soil profile truncation depth from cropland, occurring between the beginning of permanent human occupation (500–200 years ago) and 1974, was estimated to be 14.5 ± 2.1 cm, or 950 ± 200 t ha^{-1}, or 2–5 t ha^{-1} $year^{-1}$. Hurni (1979) seems to have accounted for a low soil bulk density. Due to elevation and to the proximity of the climatic limit of barley cultivation, deforestation here has started much later than in most other parts of the highlands (Hurni, 1982). The variability in soil loss depth is correlated with slope aspect and probably with the age of deforestation (Hurni, 1975, 1978).

Soil loss is thought to occur mainly in the beginning of the main summer rainy season (kremt). In those regions where spring (belg) rains are sufficient for cultivation, these crops have been harvested and the land ploughed again before kremt (Mulugeta, 1988). In the northern highlands, spring rain is unreliable; the land is only sown in the beginning of the kremt season,

Fig. 13. Rills are especially discernible on long and steep slopes without soil conservation structures (about 50 km south of Sokota, February 1999).

when rains are intensive. The fields have then undergone at least two tillage operations, are bare and offer less resistance to splash and runoff erosion (Virgo and Munro, 1978). With the advance of the rainy season, soil loss decreases, as it is negatively correlated to crop cover (Mulugeta, 1988). We observed, however, in Tigray that there was only substantial runoff more than 1 month after the beginning of the kremt rains. In the beginning of the rainy season, most rain infiltrated quickly in the dry, tilled fields. Marque and Rosenwald (1997) stress the importance of tillage for soil moisture conservation in the northern Ethiopian highlands. Furthermore, on Vertisols, which are well represented in Ethiopia (Kanwar and Virmani, 1986; Deckers, 1993), the first rains are well absorbed, because of the deep shrinkage cracks. After some time, with the closing of the cracks, these soils become completely impervious and favour important runoff (Bauduin and Dubreuil, 1973). Fekadu and Bauwens (1998) find the most significant correlation between monthly rain and summer flow in the Awash headwaters to be situated in August, at the beginning of the second half of the rainy season, when "there is greater opportunity for flow generation (even for smaller storms) since the catchment is already moist." Sutcliffe and Parks (1999) estimate that "early rainfall is required to replenish the soil moisture storage after the dry season." More monitoring is necessary on precipitation, runoff and sheet and rill erosion throughout the rainy season.

Soil loss on experimental plots is positively correlated to runoff volume (Feleke, 1987). Based on the Soil Conservation Research Programme (SCRP) data set, Sonneveld et al. (1999) find that annual soil loss (supposedly through sheet and rill erosion) is an almost linear function of total annual runoff. Mulugeta (1988) finds a good correlation between runoff volume and soil loss for individual storms on a runoff plot without conservation measures and concludes that soil erosion on steeper slopes can be controlled to a greater extent if the runoff volume is reduced by means of soil conservation measures.

On large mechanised farms, despite the fact that very few steep slopes are cultivated, high rates of soil loss by sheet and rill erosion have been measured (Alemayehu, 1992). These high rates are attributed to monoculture cropping, large field sizes ('extremely long slopes up to 200–500 m') and mechanisation. Joyce (1943) observed already "that on the large Italian tractor-ploughing schemes, erosion and gullying are more marked than elsewhere."

From his research and from data collected in the SCRP stations, Hurni (1985) adapted the Universal Soil Loss Equation (Wischmeier and Smith, 1978) to Ethiopian conditions (Table 5) for use by development agents in the field of soil and water conservation. Due to the large difference between its minimum and maximum values, the cover-management (C) factor has outstanding importance and slight mistakes in the analysis of land cover can easily result in largely over- or underestimated soil loss assessments. For an appropriate use of this equation, we might suggest for several factors to possibly go back to equations developed for the (R)USLE. The slope gradient factor S is the one presented by Wischmeier and Smith (1978) for slope gradients up to 0.22 m m^{-1}. Above that limit, data collected in Simen (Hurni, 1979) show that this equation cannot be extrapolated and that the curve of soil loss vs. slope gradient shows a flattening. Such conclusions were also reached in RUSLE (Renard et al., 1997; Nearing, 1997). The soil erodibility factor K can be assessed from soil textural data and organic matter content. The inclusion of the often-occurring rock fragment cover (Nyssen et al., 2001b, 2002c) in the management factor P is surprising, and the assumed percentage of soil erosion reduction seems underestimated and not supported by data. Nyssen et al. (2001a) found that reducing rock fragment cover from 20% to 0% in a 12.5 m m^{-1} plot in the Tigray highlands results in a threefold increase of soil flux due to water erosion. Relationships between rock fragment cover and rill and interrill soil loss, established from numerous measurements on experimental plots (Poesen et al., 1994), result in exponential equations where an 80% rock fragment cover yields some 90% reduction of the erodibility of these soils (Wischmeier and Smith, 1978; Renard et al., 1997).

The adapted USLE has been used for the assessment of soil loss through GIS models (estimation of all the factors, representation in different layers and multiplication) by Helldén (1987), Eweg and Van Lammeren (1996) and Eweg et al. (1998). These studies conclude that the USLE model is not applicable as a quantitative model at a regional level. Errors are mainly due to the important locational inaccuracy for each USLE factor and the lack of reliability of the analysis of topography and land cover, as also dis-

Table 5
The Universal Soil Loss Equation (USLE) adapted for Ethiopia (Hurni, 1985) (R in J cm m^{-2} h^{-1} $year^{-1}$, K also in SI units, following Wischmeier and Smith's (1978) conversion coefficient)

THE UNIVERSAL SOIL LOSS EQUATION (USLE) ADAPTED FOR ETHIOPIA

SOURCE: WISCHMEIER AND SMITH, 1978

ADAPTIONS: R CORRELATION: HURNI, 1985
K VALUES: FROM BONO AND SEILER, 1983, 1984; AND WEIGEL, 1985
S EXTRAPOLATION: HURNI, 1982

EQUATION: A = R * K * L * S * C * P (TONS PER HA PER YR)

1. R: RAINFALL EROSIVITY

ANNUAL RAINFALL (MM)	100	200	400	800	1200	1600	2000	2400
ANNUAL FACTOR R	48	104	217	441	666	890	1115	1340

2. K: SOIL ERODIBILITY

SOIL COLOUR	BLACK	BROWN	RED	YELLOW
FACTOR K	0.15	0.20	0.25	0.30

3. L: SLOPE LENGTH

LENGTH (M)	5	10	20	40	80	160	240	320
FACTOR L	0.5	0.7	1.0	1.4	1.9	2.7	3.2	3.8

4. S: SLOPE GRADIENT

SLOPE (%)	5	10	15	20	30	40	50	60
FACTOR S	0.4	1.0	1.6	2.2	3.0	3.8	4.3	4.8

5. C: LAND COVER

DENSE FOREST:	0.001	DENSE GRASS:	0.01
OTHER FOREST:	SEE GRASS	DEGRADED GRASS:	0.05
BADLANDS HARD:	0.05	FALLOW HARD:	0.05
BADLANDS SOFT:	0.40	FALLOW PLOUGHED:	0.60
SORGHUM, MAIZE:	0.10	ETHIOPIAN TEF:	0.25
CEREALS, PULSES:	0.15	CONTINUOUS FALLOW:	1.00

6. P: MANAGEMENT FACTOR

PLOUGHING UP AND DOWN:	1.00	PLOUGHING ON CONTOUR:	0.90
STRIP CROPPING:	0.80	INTERCROPPING:	0.80
APPLYING MULCH:	0.60	DENSE INTERCROPPING:	0.70
STONE COVER 80%:	0.50		
STONE COVER 40%:	0.80		

Note: We recommend to use the equations presented in RUSLE (Renard et al., 1997) for S and K factors as well as for an assessment of the impact of stone cover on soil erodibility.

cussed in this paper. Eweg et al. (1998) stress that "this does not rule out that the USLE model can be used to indicate the erosion risk of the area, notably in a qualitative way, to support strategic planning for rehabilitation of degraded areas." Gete (1999) tested the WEPP model (Water Erosion Prediction Project) on data obtained from Anjeni (Godjam) SCRP station and found that WEPP overpredicts soil loss by a mean value of 48.8%.

3.4. Gullying

Gullying is not restricted to the highlands of Ethiopia but is widespread at subcontinental scale in Africa (Moeyersons, 2000). In Tigray, the change in hydrological behaviour has been attributed to an overall lowering of the infiltration capacity of the soils due to removal of the natural vegetation (Hunting, 1974; Virgo and Munro, 1978; Machado et al., 1998). Buried soils witness deforestation, which might have started around 5000 ^{14}C years BP in the Ethiopian highlands (Fig. 14). Since the 20th century, vegetation removal, however, concerns also shrubs and small trees, as well as grass strips in between the fields and on steep slopes. This lowers the infiltration capacity of the soil, favours the occurrence of flash floods and is estimated to be the major cause of rapid gullying in many areas. One should also stress the importance of the abandonment of cropland for gully initiation, especially if it is converted into grazing land: the overgrazed soil surface has a higher runoff coefficient than regularly ploughed fields; soil and water conservation structures are no longer maintained, and bank gullying (Poesen et al., 2003) often starts at places where these structures collapse.

Brancaccio et al. (1997) explain the present-day processes of channel incision by the increased erosional power enhanced by lower sediment load, associated with the advanced phase of soil erosion on the hillslopes where bedrock is outcropping.

Gully systems in Ethiopia can often be considered as discontinuous ephemeral streams (Bull, 1997) comprising a hillslope gully, an alluvial–colluvial cone at the foot of the hill and renewed incision with gully head formation further downslope in the valley Vertisol. Pediments dissected by gullies are a common feature in many areas (Riché and Ségalen, 1971; Ogbaghebriel and Brancaccio, 1993; Ogbaghebriel et al., 1997; Billi and Dramis, 2003). In valley bottoms, initial gully heads often coincide with sinking polygonal structures in Vertisols (Nyssen et al., 2000a).

Active gullying induced by road building on pediments is described in Ogbaghebriel et al. (1993). In a case study along the road Mekelle-Adwa, built in 1994, Nyssen et al. (2002a) analyse how road building, through the important enlargement of drainage areas and the concentration of runoff, induces an artificial passing of the critical catchment size at which gully heads are formed for a given slope gradient.

Recent tectonically induced gullying was analysed in the Rift Valley (Williams, 1981; Belluomini et al., 2000). It should be remembered that ongoing rift widening at rates of 0.5 mm year^{-1} (Mohr, 1974, quoted by Williams, 1981) causes cracks to develop. Though not a primary cause for gullying in Ethiopia, it can certainly be important at a regional scale. Near Dessie, a 300-m-deep gully was created after river capture (Asfawossen et al., 1997).

Fig. 14. In some highland areas, trees in between the fields are well managed. Hechi, Dogua Tembien, Tigray, August 2001.

Assessments of gully erosion volumes in Ethiopia are rare. Using photogrammetric techniques, Shibru et al. (2003) estimate that between 1966 and 1996, 700,000 t of soil were lost by gully erosion from a 9-km^2 catchment in the eastern highlands (26 t ha^{-1} year^{-1}). Using monitoring and interview techniques to establish average long-term soil loss rates by gully erosion in Central Tigray, Nyssen (2001) obtained 5 t ha^{-1} year^{-1} and De Wit (2003) 6.2 t ha^{-1} year^{-1} (average of 65 year in a 18-km^2 catchment).

So far, no study has been made on the regional distribution of gullies in Ethiopia, which would allow to recognise the relative importance of main controlling factors, comprising (1) the relationship drainage area-slope gradient for incipient gullying (Patton and Schumm, 1975; Vandaele et al., 1996), (2) soil characteristics, (3) land use, (4) sediment load, (5) intensity of the causative rain, (6) tufa dam incision, (7) direct human intervention (road building, settlement, digging of small canals to intercept runoff in fields) and (8) tectonics.

3.5. Tillage erosion

Soil translocation due to tillage by the ox-drawn ard plough appears to be an important erosion process in the Ethiopian highlands. Calculated tillage erosion (Govers et al., 1994) rates indicate that this process contributes on average to half of the sediment deposited behind stone bunds (Nyssen et al., 2000b). Colluviation occurs in the lower part of the field and soil profiles are truncated in the upper part (Herweg and Ludi, 1999; Nyssen et al., 2000b). A soil sequence on a progressive terrace in Regosol overlying strongly weathered rock in the central highlands of Ethiopia (Ankober area) is analysed by Bono and Seiler (1986). At the upper part of the terrace, soils are shallow and water and nutrient storage capacity low. However, in Dogua Tembien, intraparcel variability of soil fertility parameters is low (Nyssen, 2001). A larger content of soil moisture and of soil organic matter was even observed at the foot of the stone bunds, at the very place where the soil profile has been truncated after stone bund building (Vancampenhout, 2003). Possible effects of soil profile truncation on the values of these two parameters are outbalanced by increased infiltration rates, induced by stone bund building. The most common soils in Dogua Tembien (Regosols, Vertic Cambisols and Vertisols) have a quite homogenous composition with depth, which explains the absence of soil fertility gradients in terraced land (Nyssen, 2001).

3.6. Wind erosion

In the Ethiopian highlands, wind erosion has not been measured nor mentioned. To our knowledge, wind erosion only occurs under the form of dust devils in areas with important trampling by humans or cattle, such as market places, footpaths, unmetalled roads and around cattle drinking places. Nedeco (1997) also rejects the hypothesis of wind erosion in the highlands. A known exception to this are the numerous isolated mountains or ‘inselberge.’ Here, important wind erosion, including the formation of dunes, occurs due to the aerodynamic situation (Uhlig and Uhlig, 1989).

Wind erosion is especially important in low-lying, dry and hot regions, adjacent to the highlands, such as many places in the Rift Valley. Desert pavements, created by wind erosion, exist around Lake Turkana (Hemming and Trapnell, 1957). Wind erosion and deposition contribute to the formation of dunes in the alluvial plains of the Wabi Shebele and to overall deposition of aeolian sediments in that region (Riché and Ségalen, 1973). The Eritrean coastal plain is in many places covered by stone mantles produced by deflation as well as by loose sand occurring either as a mantle of variable depth or in the form of mobile dunes (Hemming, 1961). Aeolian sediments in the coastal plains can be composed of eroded materials from nearby rocks or brought in by dust storms, which are quite common (Horowitz, 1967).

3.7. Soil creep

Soil creep occurs in many areas and is quite visible on steeper slopes where it creates terracettes (irregular micro-terraces—Vincent and Clarke, 1976) and grass-free strips (Fig. 15). Soil creep has rarely been described in Ethiopia (Büdel, 1954; Nedeco, 1997), let alone analysed. Büdel (1954) observed creep mainly in the more humid Dega zone, on slopes with gradients >7° on agricultural land. In mountain forests, creep occurs on slopes with gradients >30°. Nyssen et al. (2002b) measured creep rates of 3–6

Fig. 15. Soil creep is evidenced where it creates terracettes and in the extreme case where the movement is faster than vegetation regrowth, which results in grass free strips. Ksad Adawro, Dogua Tembien, Central Tigray.

cm $year^{-1}$ on ancient mass movement deposits with a slope of 0.43–0.47 m m^{-1}. Shear resistance measurements in this particular case indicate the danger for failure.

3.8. Landslides and rockfall

Landsliding is common in the highlands with its deeply incised valleys and high precipitation. Landslides and other mass movements have mostly been studied when interfering with human activity, such as road damage (Parkman, 1996; Almaz, 1998; Ayalew, 1999) and loss of human life (Berhanu et al., 1999b), some landslides having killed 300 people (Ayalew, 1999). Ayalew analysed information on 64 landslides in several parts of the Ethiopian highlands and observed that they increase in size and number over the last three decades and that they often concern reactivated old landslides and related mass wastings.

Most studied landslides can be related to three types, depending on the nature of the bedrock lithology and slope morphology, defined by Berhanu et al. (1999b): (1) rock fall and landsliding from escarpments (Almaz, 1998), described as active processes in the lower Wabi Shebele valley, leading to slope retreats parallel to the escarpment (Riché and Ségalen, 1971); (2) rapid landslides in the colluvial cover of steep slopes, the main triggering factor being rain (Almaz, 1998; Berhanu et al., 1999b); (3) slow landslides in clayey material, sometimes passing to earthflow and mudflow (Büdel, 1954; Canuti et al., 1986).

Almaz (1998) and Ayalew (2000) stress the importance of rock and soil type in the Blue Nile gorge. Important mass movements can of course be expected on the slopes of a 1000-m-deep gorge, but Almaz (1998) indicates the substrates which are most sensitive: i.e., intercalation of soft beds within hard, competent rock, unfavourable joint systems, thick unconsolidated material on steep slopes. Mass movements occur in rainy years (Nyssen et al., 2002b) or after extreme events (Ayalew, 1999). This last author finds a good correlation between landslide occurrence and significantly above-average rain.

On the one hand, there are natural conditions such as substrate, slope gradient, precipitation and river captures (Asfawossen et al., 1997). On the other, particular human interventions on steep slopes may trigger landslides: i.e., irrigation ditches (Berhanu et al., 1999b); cultivation of shallow rooted permanent crops such as ‘chat’ (*Catha edulis*) and ‘enset’ (*Ensete ventricosum*) (Berhanu et al., 1999b), road building (Nedeco, 1997; Ayalew, 2000), deforestation (Nedeco, 1997; Almaz, 1998), gullying (Canuti et al., 1986; Asfawossen et al., 1997) and exclosures on steep slopes, resulting in important grass regrowth, and hence increased infiltration (Nyssen et al., 2002b).

3.9. Sediment accumulation

On the back- and footslopes of the numerous cliffs, a 'classic' sorting of deposited sediment often occurs, the coarse sediments (rock fragments) being deposited on the debris slope, finer material on the footslope, as shown by Riché and Ségalen (1973) in the Wabi Shebele basin. Belay (1998) stresses the importance of continuous deposition of colluvium on convergent footslopes preventing the development of mature soil profiles.

Hurni (1985) shows, for a 116-ha catchment in Wollo, that the rate of sediment accumulation (17 t ha^{-1} $year^{-1}$) is more important than the rate of sediment export through the drainage system (7 t ha^{-1} $year^{-1}$). In a vegetation-rich catchment in southwestern Ethiopia, sediment accumulation rates are 30 t ha^{-1} $year^{-1}$, and sediment export rates through the river only 1.1 t ha^{-1} $year^{-1}$. Here, most of the sediment deposition occurs in densely vegetated areas along riverbanks. A sediment budget for a 200-ha catchment in the Tigray highlands indicates that 59% of sediment produced by water erosion is deposited within the catchment (Nyssen, 2001). Reuter (1991) stresses the importance of the amount of organic matter stored in colluvium on footslopes in Embatkala (Eritrea). Naudts et al. (2003) stress the importance of the exclosures in Tigray as traps for sediment and associated organic carbon. Sediment deposition in floodplains and natural lakes is important, but its rates seem to have never been studied in Ethiopia.

4. Assessment of soil loss and sediment yield

Most quantitative research on soil erosion has dealt with soil loss rates due to rill and interrill erosion, mostly from runoff plots (Hurni, 1985; Kefeni, 1992; Herweg and Ludi, 1999). Other erosion processes were less quantified. Tillage erosion, despite being assessed as generating soil fluxes of the same order of magnitude as rill and interrill erosion, was only studied once (Nyssen et al., 2000b). Few data on soil loss through gully erosion in the Ethiopian highlands are available.

Sediment budgets, i.e., the detailed account of the sources and deposition of sediment as it travels from its point of origin to its eventual exit from a drainage basin (Reid and Dunne, 1996), have rarely been established for the Ethiopian highlands (Nyssen, 2001). Among the better quantified processes of such a budget are water erosion at plot scale and, at the outlet of the system, suspended and dissolved sediment loss through the rivers draining the highlands.

4.1. The importance of spatial scale in sediment yield assessment

In June 2000, frequent electric power cuts in Ethiopia were attributed to the silting up of the 41-year-old Koka reservoir, which reduced hydroelectric power production at the end of the dry season (Ethiopian News Agency, 2000). Yearly, 18×10^6 t of sediment is estimated to enter into the reservoir from its approximately 4050 km^2 catchment, which is maybe overestimated (Shahin, 1993, 2001). The Ethiopian highlands are also the main source of sediment deposited in the Aswan High Dam. Fahmy (1998) reviews the well-known siltation problems in that reservoir, where some 134×10^6 t $year^{-1}$ is deposited, corresponding to a specific sediment yield of 40.5 t km^{-2} $year^{-1}$ for a basin of around 3×10^6 km^2. This includes the Atbara, Blue and White Nile subbasins. As compared to the White Nile, the Blue Nile and Atbara-Takazze have seasonal flows and carry a significant bedload during the flood period (Sutcliffe and Parks, 1999), and vegetation is relatively sparse in their basins. Specific sediment yield is largest in small headwater basins, where reservoirs can be rapidly silted up. For one of the Takazze's tributaries, the Midmar river near Axum, Machado et al. (1995) estimated an area-specific sediment yield of 2100 t km^{-2} $year^{-1}$ for a watershed of 6.7 km^2, based on the filling of a reservoir.

Data on sediment yield for different basins in the highlands have been compiled. Such data are not only difficult to find but sometimes also difficult to compare. Measurements of sediment deposition in reservoirs with high trap efficiency, such as Aswan or Koka, give relatively accurate data over longer periods. Most sediment transport measurements are, however, done by regular sampling at gauging stations. The major problem here is that bedload transport is not always taken into account. HTS

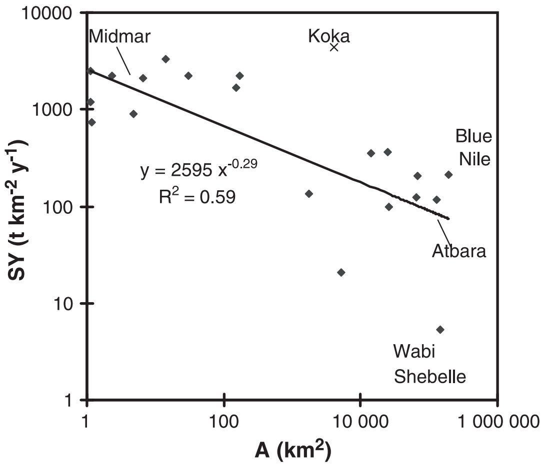

Fig. 16. Area-specific sediment yield (SY) vs. drainage area (A) for different catchments draining the Ethiopian Highlands. Koka data are probably overestimated (Shahin, 2001) and were not taken into account in the regression analysis. Data are from Bauduin and Dubreuil (1973), Virgo and Munro (1978), Williams and Williams (1980), Hurni (1982, 1985), Shahin (1993), Machado et al. (1995), El Swaify and Hurni (1996), Bosshart (1997, 1998a,b), Sutcliffe and Parks (1999), Herweg and Stillhardt (1999), Billi (2000, 2002), MOIWR (2001) and Abbas (2001).

(1976), for a river in the Takazze basin, includes an assumed bedload mass of 10% of the suspended sediment mass. Bauduin and Dubreuil (1973) in the Wabi Shebele area estimate that in the steeper river sections, bedload and suspended sediment transport would be approximately of the same order of magnitude, but do not include bedload transport in their sediment yield estimations.

Again, as for runoff coefficients (Section 3.1.4), data from small catchments are very variable due to particular conditions. Despite these shortcomings, a power relation of calculated sediment yield on drainage area outside the less-erodible areas in southern Ethiopia gives a fairly good correlation for catchments >1 km^2 (Fig. 16):

$$SY = 2595A^{-0.29} \quad (n = 20;\ r^2 = 0.59) \qquad (2)$$

where: SY = area-specific sediment yield, in t km^{-2} $year^{-1}$, and A = drainage area, in km^2.

Walling (1984) stresses that the highest suspended sediment yields in Africa are in the mountain and upland areas of east and southern Africa, as well as the fact that there are considerable depositional losses within the river systems. In the Wabi Shebele, 95% of solid transport (14×10^6 t $year^{-1}$) is deposited in the alluvial plains around Kelafo and Mustahil. Sediment concentrations in the Wabi Shebele near the Somali border are low (maximum 400 g m^{-3}) (Bauduin and Dubreuil, 1973). The downriver increase in clay/silt ratio in the Nile (Walling, 1984) is also an indication of sediment deposition. The inverse relationship between area-specific sediment yield and basin area, which we found for the Ethiopian highlands, shows clear similarities with relationships for comparable environments in Tanzania (Rapp, 1977) and more generally for world mountain areas (1000–3000 m a.s.l.) (Milliman and Syvitsky, 1992).

4.2. Soil loss from runoff plots and sediment deposition

Hurni (1985) estimates mean soil loss rates for different types of land use (Table 6) and stresses that sediment export via rivers and soil loss from fields are two different things; differences between the two can be explained by the vast amount of sediment deposited on footslopes. Stocking (1996) and Hänggi (1997) review the reasons why runoff plot rates cannot be extrapolated towards catchment scale: (1) runoff plots exemplify a ridge situation, they are neither affected by runoff or sediment deposition from upslope; (2) erosion is a redistribution of soil particles: only a small part of the soil lost from a plot is lost from the catchment; (3) due to the restricted size

Table 6
Estimated rates of soil loss by sheet and rill erosion on slopes, for various land uses in Ethiopia (after Hurni, 1990; updated in Bojö and Cassells, 1995)

Land cover	Area (%)	Soil loss (t ha^{-1} $year^{-1}$)
Grazing	47	5
Uncultivable	19	5
Cropland	13	42
Woodland/bushland	8	5
Swampy land	4	0
Former cropland	4	70
Forests	4	1
Perennial crops	2	8
Total for the highlands	100	12

of runoff plots, gully erosion and often rill erosion are not taken into account.

Nevertheless, there is a tendency in reports towards extrapolation of runoff plot data to regions or even to the whole country, as if it were crop yield data, neglecting thus the scale effect. In a reaction to such simplifications, Bojö and Cassells (1995) insist on the necessity not to omit deposition when calculating soil loss rates. Their controversial paper sometimes sounds somewhat confusing: on the one hand, the Ethiopian Highlands Reclamation Study and the SCRP are blamed for neglecting deposition; on the other, the authors acknowledge these studies for assuming that "only 10% of the gross soil loss represents a net annual loss to the system, but the 90% that is redeposited is spread proportionally on several categories of land (57% on grazing land, 20% on cropland, 14% on forest land)" (1995: 3–4). The confusion arises from the fact that the importance of spatial scale is not sufficiently taken into account.

Although Hurni made a clear difference between soil loss by sheet and rill erosion from farmers' plots on one hand and sediment yield from catchments on the other (e.g., Hurni, 1979, 1982, 1985), the presentation of data in tables with estimated mean soil loss per type of land use and the popularisation of the adapted USLE (Hurni, 1985) may lead to the use of these tools for analyses for which they are not intended. It is an easy exercise to multiply mean yearly soil losses from field surfaces with total area of fields in a given area. Such a calculation (Hurni, 1990) ends up with a nationwide estimation of soil loss on slopes in Ethiopia of 1.5×10^9 t year^{-1}. Although Hurni (1990) states that this represents only the soil erosion component of a sediment budget, this figure has become an eye catcher in reports and some scientific papers (Gemeda and Fielder, 1995; Asgelil, 2001). USLE also fits well for use in GIS-type models. Again, generalisations are made, slope lengths are derived from DTMs, omitting any deposition area, and total derived soil loss from catchments is strongly exaggerated. Helldén (1987) thus calculated a mean soil loss for Ethiopia of 150 t ha^{-1} year^{-1}, and Reusing et al. (2000) found that the predominant part of a study area near Lake Tana is affected by soil loss rates in excess of 256 t ha^{-1} year^{-1}.

Bojö and Cassells (1995) estimate that "there is reason to adjust previous calculations of the rate of soil loss down." In our opinion, soil loss estimation methods depend on the objective of the estimation: if one wants to know soil loss estimates from cropland (from that part of cultivated fields where there is no deposition), a careful application of USLE (Renard et al., 1997) could be useful for estimating sheet and rill erosion rates, to which tillage erosion should be added. In order to estimate soil loss rates for small or large catchments, the best solution is a measurement campaign of suspended and bedload sediment. In the absence of such measurement, a rough estimation for an 'average' Ethiopian catchment can be made using Eq. (2), which needs to be refined with more data, possibly grouped by maximum headwater elevation (Milliman and Syvitsky, 1992) or by agro-ecological region.

With respect to reservoir siltation, El Swaify and Hurni (1996) estimate that minimising sediment production has a better technical feasibility, at lower economical cost, than removal of silt from major reservoirs.

Some conclusions presented by Milliman and Syvitsky (1992) are fully applicable to sediment yield estimations in the Ethiopian highlands: "topography and basin area have order of magnitude control over sediment discharge of most rivers. (...) The role of sediment erodibility (mainly a function of geology, vegetation cover and human activity) clearly cannot be discounted."

4.3. Dissolution

The specific dissolved solid transport rate of the Nile is 3.9 t km^{-2} year^{-1} (Martins and Probst, 1991). Blue Nile and Atbara deliver water poor in ions compared to the White Nile water, which is attributed to evaporation of the latter in the Sudd swamps (Kempe, 1983). The major dissolved cation transported is Na^+, closely followed by Ca^{2+} (Table 7) (Probst et al., 1994). The water chemistry of the Nile River is controlled by rock weathering (Abu El Ella, 1993), and high rates of dissolved calcium carbonates are expected in limestone areas (Fig. 17). Specific dissolved organic carbon output is the least of all major world rivers (0.089 t km^{-2} year^{-1}), main controlling factors being river discharge, slope gradi-

Table 7
Mean annual fluxes of dissolved major elements transported by the Nile (after Kempe, 1983; Probst et al., 1994)

Element	Dissolved flux (10^6 t year^{-1})
Ca^{2+}	2.57
Mg^{2+}	1.01
Na^{+}	3.03
K^{+}	0.59
Cl^{-}	2.84
SO_4^{-}	2.48
HCO_3^{-}	14.01
Organic C	0.17

ent and carbon content of the soils (Probst et al., 1994).

4.4. A sediment budget for the Ethiopian highlands

Data gained in quantitative studies allow the establishment of a crude sediment budget for "average" catchments in the Ethiopian highlands:

$$\text{Sources} - \text{Sinks} = \text{Yield} \tag{3}$$

The following components should be accounted for:

$$\text{Diff} + \text{ShR} + \text{Gull} + \text{Mass} + \text{Diss} - \text{Coll} = \text{SSL} + \text{DSL} \tag{4}$$

Diff = diffusion-type erosion processes (tillage erosion, creep, splash erosion); this concerns transfers inside the catchment. At the catchment outlet, Diff = 0 t km^{-2} year^{-1}.
ShR = soil loss from slopes by sheet and rill erosion. Here, a mean value for the Ethiopian highlands can be taken (1200 t km^{-2} year^{-1}; Hurni, 1990), or it can be computed taking into account land use in the catchment with measured soil loss rates or data from Table 6.
Gull = gully and river bank erosion. Average of the only three known gully erosion assessments (Shibru et al., 2002; Nyssen, 2001; De Wit, 2003) is 1240 t km^{-2} year^{-1}, but these assessments might not be representative of all the conditions in the Ethiopian highlands.
Mass = different mass movement processes which rarely carry sediment directly into the river system. If mass movement lobes are eroded, those masses will be accounted for in the "sheet" and "gully erosion" components. Hence, Mass ≈ 0 t km^{-2} year^{-1}.
Diss = dissolution. Specific dissolution rate in the Nile basin is 3.9 t km^{-2} year^{-1} (Martins and Probst, 1991). More refined data might be calculated taking into account lithology.
Coll = redeposited sediment inside the catchment. Deposition areas are diverse in type and locations; hence, volumes are difficult to account for. They can best be calculated by solving Eq. (5).
SSL = suspended sediment load which can be measured at the catchment outlet, or estimated from equations such as Eq. (2).
DSL = dissolved sediment load. This will be more or less equal to dissolution, if the circumstances in the catchment are such that only a very small part of the dissolved minerals are reprecipitated. Obviously, precipitation of dissolved material is more important in basins including large tracts of arid land.

Fig. 17. Lapiez resulting from $CaCO_3$ dissolution in Mesozoic Antalo limestone in the Tembien highlands.

Table 8
Tentative sediment budgets for average catchments of different sizes in the Ethiopian highlands

Catchment size	100 km^2		1000 km^2		10,000 km^2	
	Absolute (t year^{-1})	Specific (t km^{-2} year^{-1})	Absolute (t year^{-1})	Specific (t km^{-2} year^{-1})	Absolute (t year^{-1})	Specific (t km^{-2} year^{-1})
Sources						
Sheet and rill erosion[a]	120,000	1200 (49.1%)	1,200,000	1200 (49.1%)	12,000,000	1200 (49.1%)
Gully erosion[b]	124,000	1240 (50.7%)	1,240,000	1240 (50.7%)	12,400,000	1240 (50.7%)
Dissolution[c]	390	4 (0.2%)	3900	4 (0.2%)	39,000	4 (0.2%)
Total soil loss	244,390	2444 (100%)	2,443,900	2444 (100%)	24,439,000	2444 (100%)
Sinks						
Sediment deposition[d]	175,745	1757 (71.9%)	2,089,944	2090 (85.5%)	22,604,699	2260 (92.5%)
Sediment yield						
Solid sediment loss[e]	68,255	683 (27.9%)	350,056	350 (14.3%)	1,795,301	180 (7.3%)
Dissolved sediment loss[c]	390	4 (0.2%)	3900	4 (0.2%)	39,000	4 (0.2%)
Total catchment sediment output	68,645	686 (28.1%)	353,956	354 (14.5%)	1,834,301	183 (7.5%)

[a] Hurni (1990).
[b] Shibru et al. (2002), Nyssen (2001), De Wit (2003).
[c] Martins and Probst (1991).
[d] (Total soil loss) – (Total catchment sediment output).
[e] Calculated with Eq. (2).

Taking into account the above assumptions, sediment budgets in the highlands could be calculated using the following simplified equation:

$$\mathrm{ShR} + \mathrm{Gull} + \mathrm{Diss} - \mathrm{Coll} = \mathrm{SSL} + \mathrm{DSL} \quad (5)$$

Indicative sediment budgets, as shown in Table 8, suffer from lack of spatially distributed data on gully erosion rates. They are scale dependent; preliminary calculations show that with catchment size increasing from 100 to 10,000 km^2, the ratio between the mass of sediment deposited within the catchment and total sediment export increases from approximately 3:1 to more than 13:1.

5. Land degradation and desertification

Although climatic conditions [0.05 < (annual precipitation/potential evapotranspiration) < 0.065] in parts of the northern highlands and in the low-lying parts of the country would justify the use of the term 'desertification' (UNEP, 1994), the term 'land degradation' will be used to indicate environmental degradation throughout the country. Two major factors inducing land degradation and desertification in the Ethiopian highlands are generally considered: drought and land use changes.

5.1. Rain variations and drought

Attention in the media to famines in Ethiopia has created a popular view of a drought-stricken country, with a tendency towards decreasing annual rain. Analyses of time series till 1990 give contradictory results. For Yilma and Demarée (1995), a decline of the rainfall in the Sahel observed since about 1965 is also seen to a lesser extent in the north central Ethiopian highlands. Camberlin (1994) found a similar tendency. However, unlike the Sahel, a comparison between two reference periods (1931–1960 and 1961–1990) yields no significant changes in mean rain over Ethiopia, but an increased interannual variability (Hulme, 1992). Mattson and Rapp (1991) state "it is not clear whether this pattern signifies the beginning of a long-term reduction or is within the range of normal fluctuations." Analyses of time series of annual precipitation, reaching up to 2000 AD, both for Addis Ababa and the northern highlands (Fig. 18), show that although the succession of dry years between the late 1970s and late 1980s produced the driest decade of the last century in the Ethiopian

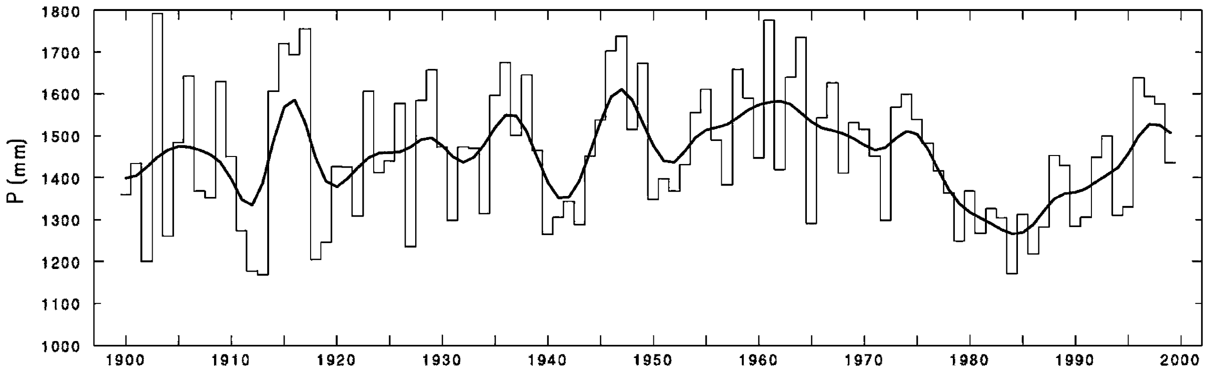

Fig. 18. Time series of annual precipitation (*P*): area-average (1–11 gauges) over the Blue Nile basin. The smooth curve is obtained using a 10-year Gaussian filter (Conway, 2001).

highlands, there is no evidence for a long-term trend or change in the region's annual rain regime (Conway, 2000b).

With respect to interannual rain variability, Conway (2000b) finds a coefficient of variation below 20% for the wetter areas, but far above that for drier areas to the north and at lower altitudes. Hoffmann (1987) also finds annual rain variability strongly dependent on climatic region: < 10% in the area around Jimma with a tropical rain climate and >45% in semi-desertic areas. Dry years were noted in 1913–1914, 1937, 1941, 1953, 1957, 1965–1966, 1969, 1973–1974, 1976, 1979, 1983–1984, 1987 (Camberlin, 1994). It is evident that, in an already degraded environment, a dry year has a very negative impact, not only on agricultural production, but also on the environment (i.e., overgrazing, cracking of Vertisols, groundwater depletion). Runoff coefficients in such a year are larger (Casenave and Valentin, 1992; Valentin, 1996) and result in increased soil erosion rates.

Besides yearly precipitation, its seasonal distribution must be considered as well. Unlike West Africa, according to Hulme (1992), rain seasonality over Ethiopia slightly decreased, between 1931–1960 and 1961–1990. Better spread rains mean, however, also that a larger percentage of rain falls outside of the agricultural season, or that there is a shift from one rainy season to another. Camberlin (1994) and Yilma and Demarée (1995, 1998), comparing 1939–1964 and 1965–1990, found decreased summer and increased spring rains in the northern Ethiopian and the Eritrean highlands. Differences between the temporal behaviour of spring and summer rains are expected to reflect different levels of influence from the Indian and Atlantic Oceans, respectively (Conway, 2000b).

5.2. Human settlement and changes in land use and land cover

Human settlement with concomitant agricultural exploitation induces changes in land use and land cover, which in turn alter infiltration and runoff conditions (Mengistu, 1999), as well as erosion processes (Olson, 1981; Bunney, 1990). Detailed studies show that settlement decisions were made on a clear ecological basis, especially from the beginning of the pre-Axumite era (700 BC) on. Preferred locations were at the margin between Vertisol areas and narrow alluvial valley bottoms which could be irrigated (Michels, 1988). Human activity expanded from such preferential places to the present-day occupation of steep slopes for agriculture through a number of stages, including forest clearing and removal of remnant trees and shrubs.

5.2.1. Deforestation

In many reports and even in scientific papers dealing with environmental degradation in Ethiopia, it is a commonplace to stress that 40% of the country was covered with forests "in 1900" (Allen-Rowlandson, 1989; Taddesse, 1995), "16%, two decades ago" (Assefa, 1986), or even in Eritrea "30–40% of total land area forested in 1900" (Robinson et al., 1995).

Comprehensive literature searches (Wøien, 1995; Gascon, 1998; McCann, 1998) show that no study ever produced such data. Pankhurst (1995) states that there are no reliable records of the extent of the country's forests prior to recent times. In 1946, Logan estimates that only 5% of the highlands were forested. Taking into account the climatic potential for forest cover in Ethiopia, 40% would have been the approximate cover at the beginning of agricultural exploitation (Von Breitenbach, 1961). For a comprehensive overview of forest, wood- and bushland types in Ethiopia, we refer to Fiedler and Belay (1988). Ritler (1997) has carried out an interesting study: based on a critical examination of numerous European travellers' accounts for the period 1699–1865, he made a reconstruction of the landscape of the central-northern Ethiopian highlands. Some short quotations from this work might be sufficient to invalidate speculations on widely forested Ethiopian highlands around 1900. In the 18th and 19th century, above 1500 m, "closed forests were rare, except for areas that were unfavourable for cultivation. The few forests mentioned usually ranged from several hectares to a few square kilometers in size. On the other hand, landscapes described as having isolated tall trees, groves of trees, or sections along streams and rivers with rows of trees and small forests, were mentioned frequently. (...) In certain locations the shortage of fuelwood and timber was a clearly recognisable problem. Dried dung mixed with earth and straw was used as a substitute for fuelwood" (Ritler, 1997, p. 13). Pankhurst (1995) gives evidence for widespread deforestation in the 16th century; locally reforestation occurred (Zewdu and Högberg, 2000a, 2000b). Boerma (1999, 2001) found that forest cover changes in Eritrea have been much less important than is generally assumed.

Changes in forest cover, though not as dramatic as often stated, are, however, real (Melaku, 1992), as it was found in several analyses using aerial photo interpretation. One study shows an increase in forest cover in a eucalyptus growing area, around the Ankober Pass, in 1986 as compared to 1957 (Wøien, 1995). Though no studies exist for Ethiopia, eucalyptus trees are generally expected to lead to increased soil loss rates due to reduced understorey cover. Pankhurst (1995) notes that therefore, in 1913, a decree was issued compelling the uprooting of eucalyptus trees; for economic reasons (Jagger and Pender, 2003), it was never enforced. One can expect at least a localised impact on soil erosion rates by the increased area planted with eucalyptus (Conway, 1997). For semi-natural forests, the trend is, however, the inverse. Kebrom and Hedlund (2000), comparing the 1986 situation to that of 1958 for a 110-km^2 area in southern Wollo, find a decrease of forest cover from 8% to 5% and shrubland from 28% to 14%. Near Ambo, "bush and woodland" decreased from 42% to 33% of the area between 1957 and 1994 (Van Muysen et al., 1998). Gete and Hurni (2001) found that the 27% natural forest cover in parts of Gojam had nearly totally disappeared after 1957.

For centuries, within deforested areas, both on arable and shrubland, many trees and shrubs were present (Fig. 14). Due to grazing, woodcutting and ploughing of vegetated areas in between fields, the vegetation cover by trees and shrubs gradually decreased. Wøien (1995) finds a decrease of scattered trees around Debre Sina. This deterioration of the woody vegetation cover is illustrated some 200 km more to the north by Kebrom and Hedlund (2000), who found that most of the 'lost' forest had become shrubland, and most of the 'lost' shrubland turned into 'open areas.'

It appears thus that deforestation is a very old phenomenon, as old as the basis of already discussed soil accumulations overlying burnt horizons (around 2450 ^{14}C years BP in parts of Wollo, e.g., Hurni, 1985), or as the signal of ruderal species in pollen diagrams in Bale and Arsi from 2000 ^{14}C years BP on (Bonnefille and Hamilton, 1986). What has triggered recent degradation is generally not rapid deforestation, but a progressive change in land cover, including removal of trees and shrubs in between fields, along rivers and in areas with secondary regrowth ('shrubland').

5.2.2. Grazing and changes in vegetation cover

Different types of climax grasslands exist in Ethiopia: i.e., at high elevations, on Vertisols and on dry places (Klötzli, 1977). Besides, cattle also graze vast deforested areas, commonly called rangeland. Much in the same way as in forests and woodlands, vegetation cover decreases in grass- and rangeland. Most of the above quoted studies of land use changes show, besides decreasing tree and shrub cover, an increase of the area occupied by 'bare land,' 'no vegetation,' 'open areas' and the like.

Unlike the situation in southern Ethiopia (Oba et al., 2000), overgrazing of rangeland is a particular problem in the cereal zones of the highlands, where current stocking rates are well in excess of estimated optimum stocking rates (Table 9). Livestock plays a key role in the agricultural system of the highlands, providing energy (traction, manure used as fuel), food, fertiliser, insurance and status (Bojö and Cassells, 1995; Kassa et al., 2002). If no measures, such as introducing alternative sources of fodder or restrictions to free grazing, are taken, Hurni (1993) predicts a severe livestock crisis to emerge in Ethiopia in the near future, which may seriously affect agricultural production in the ox–plough system.

Consequences of overgrazing on the environment are decreased surface roughness, compaction of fine-textured soils, increased bulk density, decreased soil organic matter content, soil structure decay and decreased hydraulic conductivity. All these factors contribute to decreased infiltration and increased runoff volumes. Mwendera et al. (1997) carried out experiments on grazing land with slope gradients < 0.08 m m^{-1} in an area between Ambo and Addis Ababa. Comparing ungrazed, moderately and heavily grazed land, they found significant differences in runoff volumes for slope gradients 0.05–0.08 m m^{-1}. Steady-state infiltration rates decrease significantly, even under light grazing intensity, and show the effect of animal compaction of the soil (Mwendera and Mohamed, 1997). On cropland, stubble grazing (a widespread practice) dramatically decreases the infiltration capacity. Field observations also indicate that topsoil degradation by cattle trampling significantly contributes to soil erosion and sediment delivery to water reservoirs.

Table 9
Current and optimum livestock densities (after EFAP, 1993; Bojö and Cassells, 1995)

Zone	Current (TLU[a] ha^{-1})	Optimum (TLU[a] ha^{-1})
High potential pasture zone (highlands)	0.67	0.69
High potential cereal zone (highlands)	0.78	0.66
Low potential cereal zone (highlands)	0.66	0.31
Lowlands	0.18	0.25

[a] TLU = Tropical Livestock Unit, equivalent to a standard zebu ox of 250 kg (FAO, 1988); for each other type of domestic animal there is a conversion equivalent.

5.3. Parameters for the assessment of the evolution of land degradation

Quantitative studies on the evolution of land degradation and desertification in Ethiopia are mostly not holistic, but concern only certain phenomena, such as drought, soil erosion, deforestation and other land use changes. Engida (2000) concludes that desertification is increasing, but his study only concerns mean annual precipitation and PET, and not any process induced by land use changes, such as infiltration or runoff. Furthermore, the measured areal extension of semiarid to hyperarid zones by 8% between 1992 and 2000 seems primarily due to different classification methods.

Out of the different phenomena indicative of land degradation and desertification, lowering of the water table can be observed in many areas in Ethiopia, but was, as far as we know, never discussed in literature. Qualitative studies and some assessments of soil fertility decrease exist (see, e.g., Bojö and Cassells, 1995). Quite a number of quantitative studies of soil erosion have been carried out, as discussed earlier in this paper, and a few of them give indications about the evolution of soil erosion rates through time. Though most observations on the evolution of soil erosion rates are only qualitative (e.g., Oldeman et al., 1991), they are supported by some quantitative data. In several soil profiles in Tigray, Machado et al. (1998) found an aggradational sequence for the last 1000 years of up to 13 m in thickness, in contrast to only 4 m of aggradation for the previous 3000 years. An important increase in soil erosion rate over the last decades could be inferred from aerial photo interpretation showing, e.g., that valley bottom entrenchment only started after the early 1960s (around Axum; Machado et al., 1998). Brancaccio et al. (1997) estimated that the process of downcutting, which characterises large parts of the streams in Tigray, should be associated with an advanced phase of removal of soil and sediments from slopes, the lower sediment load increasing the erosive power of runoff. Mean sediment load in the Atbara River was, however, reported to have increased from 5.0 g l^{-1} in 1969 to 8.0 g l^{-1} in 1989 (El Swaify and Hurni, 1996).

5.4. Social and historical impulses of land use and cover changes

It appears that rain, apart from the catastrophic impact of dry years on the degraded environment, cannot be invoked to explain the current land degradation. Causes are to be found in changing land use and land cover, which are expressions of human impact (Reid et al., 2000; Feoli et al., 2002). Though deforestation and other removal of vegetation cover over the last 2000–3000 years has probably been a cyclic, rather than a linear process, studies on land use and land cover change show that at present, there is a tendency of increasing removal of vegetation cover.

At this stage, it appears necessary to briefly outline the social and historical causes of this human impact. Under feudalism (until 1974), agricultural techniques stagnated for centuries (Crummey, 2000). Until the 1940s, the Agricultural Department's only real activity was collecting the agricultural tax (Joyce, 1943). Investment in agriculture only started in the 1950s and was in the beginning mostly oriented towards export crops such as coffee (*Coffea arabica* L.), grown in southern Ethiopia. Therefore, there was limited agricultural investment in the highlands, where subsistence production dominates (Ståhl, 1990; Mulugetta, 1992). Until the late 1970s, sharecropping prevented the farmers from investing in their fields. Impoverishment led them to prefer immediate returns, even if it induced environmental degradation (Taddesse, 1995). On the other hand, recent land redistributions in order to allocate land to landless households had a positive impact on land productivity (Benin and Pender, 2001). To increase agricultural production, most of the trees and shrubs between the fields and on steep slopes have been cleared during the 19th and 20th centuries, increasing runoff and soil erosion (Ståhl, 1974, 1990; Girma and Jacob, 1988). In short, in situations of poverty and social insecurity, short-term survival prevails over medium- and long-term conservation issues.

6. Human reaction to land degradation

6.1. Agricultural intensification and land rehabilitation

Faced with a deteriorating environment, society reacts in order to maintain/improve agricultural production, often leading to qualitative changes in the production system (Boserup, 1981). Blaikie and Brookfield (1987) and Ståhl (1990) insist on the necessity that modern science be involved in this innovative process. The present-day rise in food production in Ethiopia and Eritrea (Fig. 19) can, besides reestablished climatic conditions, also be attributed to a variety of human interventions at different levels (Nyssen et al., 2003b). Extension of cropped area and increased grazing pressure are still possible. However, not much space is left for it, and productivity decreases. Giglioli (1938a,b) and Joyce (1943) already reported widespread use of indigenous soil conservation technology. Krüger et al. (1997) compiled an overview of the various indigenous soil and water conservation techniques, such as lynchets, stone bunds, checkdams or vegetative control. According to Hurni (1998), such indigenous technologies can be used as a starting point, but need

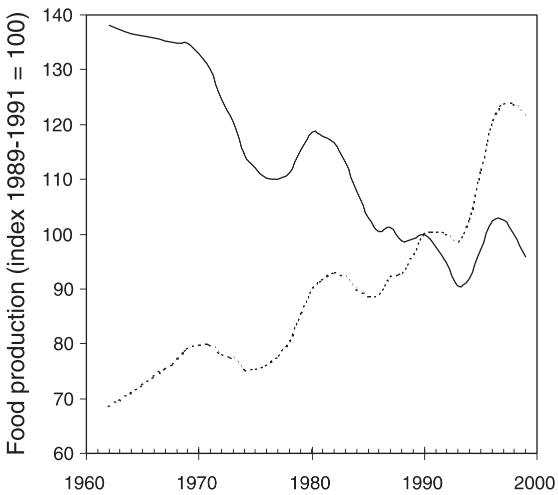

Fig. 19. Evolution of relative food production in Ethiopia and Eritrea (1962–2000) (3 years' moving average). Dotted line: total food production; solid line: food production per capita (index 1989–1991 = 100). Data from FAO (2001).

improvement in order to increase their ecological efficiency.

Nowadays, changes in the agricultural system appear, such as haymaking (‘cut and carry’) (Hurni, 1986), partially from exclosures (i.e., land under strict conservation management, often controlled by the community), which are increasingly being organised in the most affected northern highlands (Kebrom et al., 1997; Wisborg et al., 2000; Tenna et al., 2001; Nyssen, 2001, Aerts et al., 2001, 2002, 2003). Soil fertility is apparently increasing at certain sites (Eyasu and Scoones, 1999).

Different pathways to agricultural intensification are possible in Ethiopia. Mineral fertiliser should not be the overall option, given scarce capital resources. Due to decreased landholdings, a shift in the tillage system to gardening and minimum tillage (on self-mulching Vertisols) may be suggested (Abiye et al., 2002), as well as an extension of the cropping period on Vertisols (Abate et al., 1999). Asnakaw et al. (1994) obtain good maize yields with rock fragment mulching and no-tillage.

Inputs such as extension services, seed selection and credit services are increasingly made available. Adapted technology is being introduced, such as the International Livestock Research Centre’s (ILRI) Vertisol technology package. This includes the early planting of improved crops with fertiliser in order to utilise soil moisture during the main growing season, the planting of grass in gullies and of leguminous trees on steeper slopes to control erosion and provide quality feed for cattle. Most known is probably the broad bed maker, an adaptation of the traditional ard, which allows draining excess water during the rainy season (Gaspart et al., 1997; Ayele and Heidhues, 1999). The impact of this implement on soil loss is under discussion (Worku and Hailu, 1998). Such inputs are, however, limited due to budget problems. Subsidies for mineral fertiliser have been withdrawn.

6.2. Soil and water conservation

Soil and water conservation (SWC) activities are currently the most widespread form of agricultural intensification. Travelling around in the Ethiopian highlands, one will see that many stone and soil bunds dating from the 1980s are still in place. Apparently, their destruction is not as widespread as stated by Dessalegn (1994); often, the farmers accept these structures. Many, probably most of the soil bunds throughout Wollo have evolved into full-grown lynchets; even in the very rainy Ankober area, soil bunds have often been ‘opened’ to allow drainage, but are still in place over most of their length.

Local knowledge and farmer’s initiatives are integrated with these introduced technologies at various degrees (Hurni and Perich, 1992; Gaspart et al., 1997; Yibabe et al., 1999; Nyssen et al., 2000c, 2003b; Mitiku et al., 2000). The efficiency of particular techniques cannot be discussed by and large here; the reader is referred to the work conducted by the Soil Conservation Research Programme (SCRP) (Herweg and Stillhardt, 1999; SCRP, 2000). The results of this project show clear quantitative benefits of the structural (stone and soil bunds, checkdams) and biological (grass strips and exclosures) soil conservation measures in controlling runoff and soil erosion in a variety of agro-ecological zones and under various land uses (El Swaify and Hurni, 1996). In the SCRP, it was noted that quite a number of farmers accepted the introduced technologies and that they have maintained them ever since. Their judgement of the economic viability of those adopted technologies appears to be a key to success. Berhanu et al. (1999a) and Vagen et al. (1999) demonstrated that investment in stone terraces in Tigray is profitable at farm level. However, Bereket and Asafu-Adjaye (1999) found that investment in soil conservation technology may, from the farmer’s point of view, not be viable in the short term, and yet the net social benefits are positive. Hence, “there is a strong case for governments to provide incentives for soil conservation in view of the economic benefits.” Bekele and Holden (1999, 2000) also state the necessity for public intervention and incentives: “Where higher yields are lacking, society may have to look for other incentives (carrots and sticks) to persuade land users to install conservation practices.”

According to El Swaify and Hurni (1996), expected benefits of enhanced soil and water conservation are (1) control of upland soil erosion, (2) reduction in sediment loads of the region’s rivers, (3) improvement in the hydrologic regime, including reducing the peak flows, and (4) addressing food-security needs of Nile basin states. There is clearly a possibility of joint gains: rainfed agriculture in the Ethiopian highlands benefits from soil and water conservation, whereas irrigated agriculture in downstream Egypt, a country which fears reduced Nile flows due to Ethiopian conservation activities (Waterbury and Whittington, 1998) benefits nevertheless from decreased sediment load (Mengistu, 1997) and improved base flow.

Herweg and Ludi (1999) demonstrate that reduction of soil loss was considerable at all SCRP stations and with most conservation measures, although absolute erosion rates were still high in some cases. Runoff control requires, however, greater emphasis during the design of SWC structures. Especially in subhumid areas with secure high rain, structures must have a gradient to safely drain excess water. In this case, the drainage lines need to be protected from gully erosion. One of the major conclusions of the evaluation of the SCRP by Herweg and Ludi (1999) is that successful SWC is frequently connected with the following attributes: technical feasibility and adaptability, ecological soundness, economic viability and social acceptance. These authors, as well as Bewket and Sterk (2002), stress the need to involve the farmers in all stages of experimentation with SWC technology. This will certainly contribute to the development of more productive technologies, i.e., combinations of agronomic, biological and mechanical techniques, which are more attractive to the farmers compared to techniques, which are simply targeted at reducing erosion (El Swaify and Hurni, 1996).

6.3. Reservoir construction

Finally, a large irrigation programme for which scientific reports are few should be mentioned. Since the National Irrigation Commission decided in 1994 to construct 500 microdam reservoirs in Tigray for irrigation purposes (Nana-Sinkam, 1997; Mengistie, 1997), the landscape becomes studded with lakes, all a few tens of hectares in size. Up to 2002, only 50 of such microdams have been constructed and the initial project has been suspended. The main reasons are (1) the short life expectancy of these reservoirs due to a rapid capacity loss by sediment deposition, (2) the large number of leaking lakes, losing their waters into the Antalo Limestone substrate and (3) technical problems with the dams such as failure or seepage (Moeyersons et al., 2003). Only a few irrigation schemes operate in the way and place as originally foreseen (De Wit, 2003). Several reservoirs are left behind, unexploited and are considered as failures for the particular purpose they were established for. Nevertheless, their overall environmental impact is important. For instance, in many cases, these reservoirs significantly contribute to river regularisation (Waterbury and Whittington, 1998), as well as to a rapid recharge of aquifers, which, in some cases are already exploited (Moeyersons et al., 2003).

7. Conclusions

In the Ethiopian highlands, the late Pleistocene (20,000–12,000 ^{14}C years BP) was cold and dry (Fig. 20). At the end of the Last Glacial Maximum (LGM), large volumes of sediment were produced, due to the shift from arid to moist conditions. Then, a long period with abundant and evenly distributed rains existed between 11,500 and 4800 ^{14}C years BP, as witnessed by increase in arboreal pollen, high river and lake levels, low river turbidities and soil formation. Around 5000–4800 ^{14}C years BP, there was a shift to more arid conditions and slope instability. Increasing attention is given in literature to human interference with the environment in the mid-

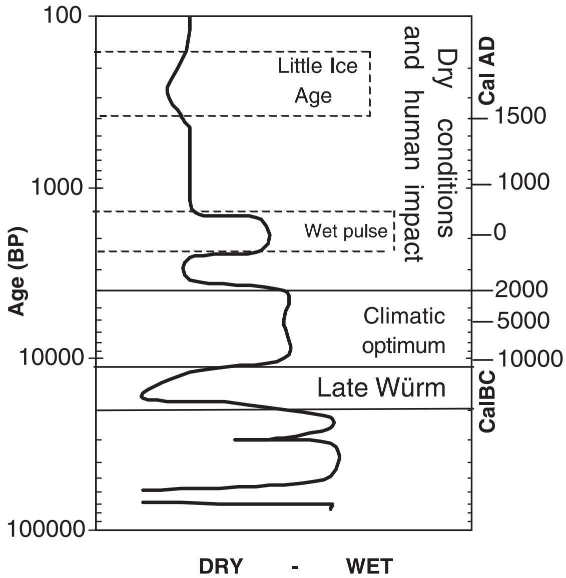

Fig. 20. Major environmental stages during the late Quaternary.

dle and late Holocene (Bonnefille and Hamilton, 1986; Brancaccio et al., 1997; Ogbaghebriel et al., 1998; Bard et al., 2000). Many phenomena that have been interpreted as climate-driven might well be of anthropic origin. Thick sediment deposits on pediments and increase of secondary forest, scrub and ruderal species in pollen diagrams are witnesses of this human impact. An important aspect remains obscure at the end of this review: high Nile runoff discharges and Rift Valley lake levels are automatically linked to great precipitation depths. Most authors do not take into account changes in land use and management, nor changes in rain seasonality, both of which can lead to a change in runoff coefficients.

The most important present-day degradation processes are sheet and rill erosion throughout the country, gullying especially in the highlands, and wind erosion in the Rift Valley and the peripheral lowlands. If many quantitative studies of sheet erosion have been made, few have dealt with gully erosion. Several studies of landslides have been made. On the other hand, tillage erosion was only recently recognised as a major process active in cultivated fields.

With respect to recent environmental changes, it appears that rain, apart from the catastrophic impact of dry years on the degraded environment, cannot be invoked to explain the current land degradation and desertification processes. Causes must be sought in changing land use and land cover, which are expressions of human impact. Though deforestation and removal of other vegetation cover over the last 2000–3000 years has probably been a cyclic, rather than a linear process, studies on land use and land cover change show that there is a tendency of increasing removal of vegetation cover, which is slowed down or reversed in Northern Ethiopia, since a decade or two, by a set-aside policy (exclosures).

Ongoing land degradation requires urgent action at different levels of society, and this is increasingly done in Ethiopia. Soil and water conservation activities are now widespread. Local knowledge and farmer's initiatives are integrated with these introduced technologies at various degrees. The results of the Soil Conservation Research Programme show clear benefits of the soil conservation measures in controlling runoff and soil erosion. In high rain areas, runoff evacuation requires greater emphasis during the design of soil conservation structures. In these areas, investment in SWC might not be profitable at farm level, although social benefits are positive. This pleas in favour of public support.

At the end of this literature review, the fundamental question remains of the underlying causes of land degradation in the Ethiopian highlands. The palaeoenvironmental evolution (Fig. 20) is not exceptional; it shows clear similarities to what happened at higher latitudes, and particularly in Europe. Furthermore, the present-day degradation cannot be attributed to drought, and deforestation also occurred in most other parts of the world. Why then the ongoing land degradation and desertification?

The present land degradation in the Ethiopian highlands has a particular origin, which includes the stagnation of agricultural technology and lack of agricultural intensification. Causes of these are to be found in the nature of past and present regional social relations (Ståhl, 1974; Blaikie, 1985), as well as in international unequal development (Amin, 1977). This has not been treated in this review and might be considered as one of its limitations.

Finally, it should be noted that the current geomorphic and hydrologic processes are of a lesser magnitude than what happened during some periods of the late Quaternary. This literature review thus strengthens our belief that, under improved socio-economic conditions, land husbandry can be made sustainable, leading to a reversal of the present land degradation from medium to low intensity.

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