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DISCORDANCE OF EROSIONAL TEMPOS: A NON-LINEAR AND SCALE DEPENDENT EVOLUTION IN THE ORANGE RIVER BASIN (SOUTHERN AFRICA)

ABSTRACT: LAGEAT Y., *Discordance of erosional tempos: a non-linear and scale dependent evolution in the Orange River basi (Southern Africa)*. (IT ISSN 0391-9838, 2013).

Various erosion rate monitors have been exploited to evaluate the denudational history of the Orange River basin from the Mesozoic to the present. Extrapolation back in time from contemporary sediment loads is hazardous, even throughout the Holocene, and low temperature thermochronometry is unable to provide constraints on the recent cooling history, so that a gap cannot be bridged between a Cretaceous period of significant post-rift denudation and a recent acceleration of the human impact. In the first period the role of tectonics may be viewed as the driving force whereas cropland soil losses have dramatically increased through gully erosion during the last century. This paper highlights the limitations of comparing denudations rates over a long period of time as measurements of current processes appear irrelevant for interpreting long-term landscape evolution. Two ways of acquiring an understanding of landforms are to be considered as they bear evidence of an irreducible disparity between the tectonic and anthropic imprints.

KEY WORDS: Erosion rates, Southern Africa, Passive margin, Fluvial discharge, Soil erosion.

INTRODUCTION

Over the past decades Earth scientists have expanded the range of methods used to infer erosion rates the application of which has spurred significant advances in quantifying the rhythms of geomorphic processes, a familiar approach for our colleague and yet friend Monique Fort. A quarter of century after a defence of a thesis about South Africa (Lageat, 1989), the following contribution to her «*Mélanges*» intends to assess the advancement of researches dealing with erosion tempos at various spatial and temporal scales, a fundamental topic in geomorphology. This paper reports average rates of denudation for various time intervals from the Mesozoic to the present, determined for the Orange drainage system, at areal scales ranging from the headwaters of the major tributaries to the whole catchment and over time scales varying from decades to millions of years. An opportunity is thus offered to aim at characterising and quantifying the terrigenous supply eroded in the drainage area, this set of data providing a rich information about erosion rates which extends along various time spans. The relevance of this knowledge for the modelling of landscape evolution remains a topic of considerable debate, allowing a critical approach of the methods applied to address this cardinal question.

THE ORANGE RIVER: A POWERFUL CONVEYOR

The Orange River originates in the north-eastern corner of Lesotho, in the Maloti Highlands, at Thaba Putsoa, 3350 m above sea level, at a distance of only 193 km from the Indian Ocean (fig. 1). After leaving Lesotho, where it is named Sanqu, the river flows westwards for 2092 km through regions of steadily increasing aridity, finally discharging its water and sediment in the South Atlantic Ocean at Alexander Bay.

The natural discharge

The Orange River Basin encompasses all of Lesotho, 48% of South Africa, 27% of Namibia and 12% of Botswana. It is by far the largest catchment in southern Africa, but the effective catchment area is difficult to de-

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FIG. 1 - The Maloti Highlands in Lesotho (photo: Y. Lageat).

termine since it includes several large ephemeral tributaries (Molopo, Hartebeest, Fish). Their contributions are erratic, particularly the Molopo which has not flowed into the Orange River in living memory (Blanchon, 2009). The most recent figures state that it covers 943,577 km^2 which are shared by four riparian countries, with 4% of the basin area lying in Lesotho, 62% in South Africa, 9% in Botswana, and 25% in Namibia (fig. 2).

Three sections can be characterised along its long journey to the Atlantic Ocean: (i) upper Orange above the Vaal confluence; (ii) middle Orange between the confluence and the Augrabies Falls; (iii) lower between the Falls and the sea. This segmentation has a double meaning: hydrological in the first instance, morphological in the second.

First of all, the main tributaries of the Orange River are the Caledon River, which forms part of the western border of Lesotho, the Kraal River, and the Vaal River (although «vaal» means «grey» in Afrikaans, «orange» does not refer to the colour of the waters but the river was named in honour of William V, King of Holland at the time of the East India Company), and most of the runoff (93%) is derived from a high-rainfall area of 294,000 km^2, 31% of the potential catchment area. Occupying only 3% of this basin, Lesotho itself provides 35% of the water filling the river. By contrast, downstream of its confluence with the Vaal River, the Orange River receives negligible runoff (7%) from the vast semi-arid surroundings (some 70%).

Secondly, a major discontinuity in the longitudinal profile of the river is represented by the Augrabies Falls, a

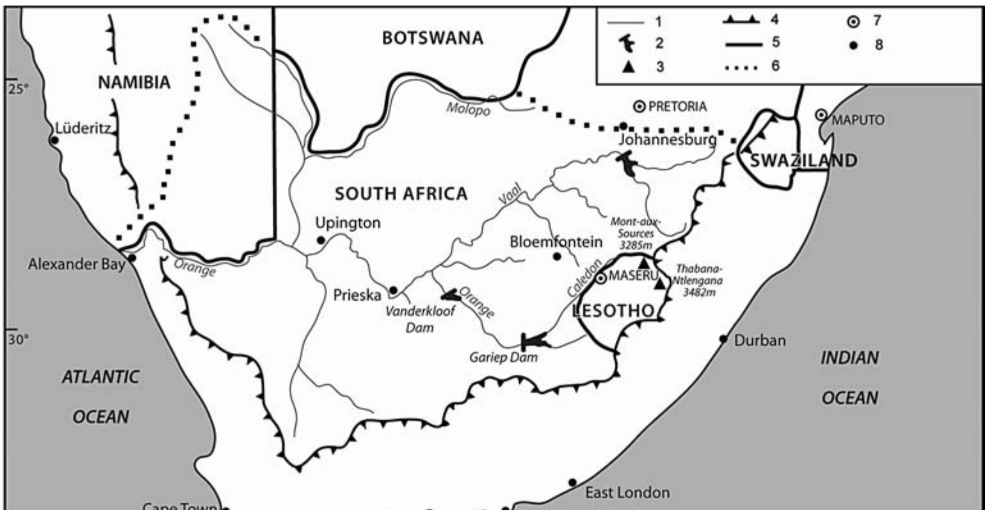

FIG. 2 - Map of the Orange River drainage basin (1. Main rivers; 2. Major dams; 3. Highest peaks; 4. Great Escarpment; 5. International borders; 6. Other watersheds; 7. Political capitals; 8. Main towns).

knickpoint, which separates its low gradient braided middle course from its deeply incised lower course (fig. 3): about 480 km from its mouth, the river plunges over the Falls, 120 m in height, into a 9 km long granitic gorge, through a series of spectacular cataracts, before crossing the coastal plain.

The evolution of the discharge

The Orange River system, which drains half of the total surface area of South Africa accounts for only 22,5% of the total annual run-off to the sea. It has been estimated that its natural mean discharge was in the order of 11 km^3/a, or 350 m^3/s, with a miserable specific yield (0,4 $1.s^{-1}.km^2$) and a high inter-annual variability of 50% (6 to 15 km^3). The largest documented flood, which lasted three months (March, April, and May) in 1988, amounted to 24,3 km^3 at the mouth, with a peak of 8200 m^3/s.

These figures refer to the natural runoff which would have occurred had there been no developments in the catchment. The actual runoff reaching the river mouth is estimated to be in the order of 5.7 km^3, considerably less than the virgin value. The difference is due mainly to the extensive water utilisation in the Vaal River basin, which supplies the large industrial and domestic demands of the Gauteng urban area.

The Orange River system has become highly regulated by virtue of 31 major dams with a storage capacity of some 20 km^3 and numerous weirs within its catchment. The two major storage dams are: (i) the Gariep Dam (previously known as the Hendrik Verwoerd Dam), the South Africa's largest reservoir, with a gross storage of 5.3 km^3, which was completed in 1971, and (ii) the 3.2 km^3 Vanderkloof Dam (ex-P.K. Le Roux Dam), which, since 1977, currently supplies the downstream demands, but its water must be released well in advance since it takes 2 to 6 weeks to reach the mouth some 1400 km away.

Whereas the lower Orange River happened to dry out due to severe droughts before the dams were built, the regulation has resulted in changing flow patterns downstream of the Vanderkloof Dam, from a pronounced seasonal flow (a 82:18 summer to winter flow) to a nearly even flow distribution (a 59:41 summer to winter flow), throughout the year.

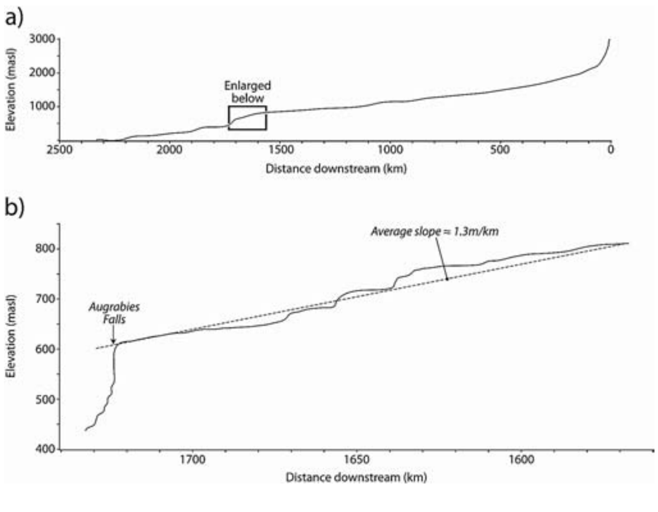

FIG. 3 - Longitudinal profiles of the Orange River.

The sediment delivery to the Ocean

The Orange River catchment area is underlain by geological formations that span a wide age range: from the sources to the sea, the outcrops belong successively to the basaltic lava flows of the Liasic Drakensberg Group, to the sedimentary rocks of the Karoo succession (Permian to Triassic shales and sandstones) and to the basement (Kalahari craton) consisting of granitoid-gneisses and greenstone belts. All these lithologies have been intruded by kimberlite pipes in Cretaceous times.

At the border between South Africa and Namibia, the Orange River forms a large estuarine and a vast submarine delta. Once they arrived in the sea before the rivers were dammed, sediments were dispersed north and southwards of the river mouth according to their granulometry: (i) gravel and sand fractions were transported equatorward by a vigorous longshore drift driven by high-energy waves and strong southerly winds, whereas (ii) the mud fraction (silt and clay < 63 µm) was moved south by a weaker poleward undercurrent and deposited in a mudbelt (Bluck & *alii*, 2007).

The coarse Orange River outfall, originated from the inland kimberlite areas, has been reworked into diamantiferous gravel beach deposits. These Pleistocene littoral sequences extend over 300 km along the Namibia's southwestern coast. This *Sperrgebiet*, which hosts one of the world's largest gem diamond placer deposits, has been mined intensively since the discovery of the first diamond near Lüderitz in 1908.

A strict composition resemblance between Orange sand and desert dunes, as documented by heavy-mineral suites, percentages of feldspar species and even volcanic lithic fragments, proves that most of the Namib sand is ultimately derived from the Orange River (Garzanti & *alii*, 2012). Sand has been blown off the beaches onto the land where it accumulated in the Namib Sand Sea (34,000 km^2) that stretches for nearly 600 km from Lüderitz (27°S) to the Kuiseb River (23°S), and for 100-150 km inland. The present rate of sand input has been estimated by Lancaster (1989) to be 400,000 m^3/a, and at least 1 Ma would have been required to form the modern erg. Recent cosmogenic-nuclide measurements on sand grains have established the consistence of such a residence time (Vermeesch & *alii*, 2010).

Whereas gravels and sand are transported northwards, the separated mud fraction is carried out onto the «Namaqualand mud-belt», an elongate belt-shaped deposit occurring along the middle-inner shelf that parallels the western shore, from the Orange River prodelta (28°S) to the Cape Canyon (34°S), where most mud transported is assumed to be lost from the margin. As the continental shelf has been progressively submerged by the last transgression, this terrigenous sediment supply, deposited in water depths less than 115 m is of Holocene age (Meadows & *alii*, 2002). The mass of this offshore deposit has been estimated from marine geophysical and coring surveys to derive from the mean Holocene mud flux of the Orange River (Compton & *alii*, 2010). With the amount of terrigenous clay which is supposed to be lost from the margin, the total Holocene mud sedimentation is estimated to be 59 (37-85) Gt (10^9 t), *i.e.* some 5 Mt/a and 5 $t.km^{-2}.a^{-1}$ (or 4 m/Ma). Over the last 11,500 years, the flux of mud to the coast is assumed to be more or less in steady state with catchment erosion in spite of sediment storage on the long-term and extreme paleofloods as evidenced respectively from colluvial deposits and slackwater sediments. This suspended mud load provides a useful measure of the subsequent accelerated soil erosion as revealed by the longest record available starting in 1929.

THE HUMAN-INDUCED FORCING

It has been established that the measured loads at Upington and Prieska represent practically the total pre-dam yield, as very little sediment enters the river downstream of the latter stations.

Pre- and post-dam load

The pre-dam mean annual runoff looks modest in comparison with most other major rivers, but the Orange River carries such a large suspended load that it ranks as the most turbid river in Africa and the fourth most turbid in the world. It delivers each year 60 Mt of sediment, 50 of which is mud, to the western margin of South Africa, as much as the Amazon River the water discharge of which is 600 times bigger (fig. 4). The mean annual erosion value for the whole of the basin was thus *ca.* 60 t/km^2 or 45 Bubnoff (mm/ka or m/Ma).

Global estimates of the annual flux of sediment under pre-human and modern conditions suggest that human activities seem to have raised the sediment yield by a factor of 10 in reasonable agreement with the difference between the pre-dam (1930-1969) mud flux of 50 Mt/a and the mean Holocene of 5 Mt/a.

The mean annual sediment discharge measured (amounting to *ca.* 60 $t.km^{-2}.a^{-1}$ or 45 m/Ma) showed a steady downward trend, as values of 80-90.10^6 t/a in the 1930's declined to 30-40.10^6 t/a in the 1960's before the construction of dams. The peak values have been attributed to agricultural malpractices in the early period and Rooseboom & Von Harmse (1979) invoke a decrease in availability of fine topsoil for erosion and transport to explain the decline in sediment discharges in the past few decades. This explanation is not convincing as most of Orange River suspended sediment is produced by gullying and not by sheetwash, and it may be found in the effective sediment trapped by numerous farm ponds.

The bulk of this sediment carried to the sea was reportedly derived from the upper portion of the catchment, upstream of the Caledon-Orange River confluence, and even from the sub-catchment of the Gariep Dam, an area of 70,000 km^2 which comprises only 8.5% of the potential catchment (Rooseboom, 1992). The sediment load has been greatly reduced since 1971 by sediment trapping behind this dam, which was expecting to trap an average of 47.10^6 t/a. Actually present day discharges of sediment are considerably lower than those recorded prior to the early 1970's and are estimated to be around 17.10^6 t/a (Bremner & *alii*, 1990). So the main sources of the large suspended sediment load carried by the 1988 flood, mainly in suspension, which amounted to some 80.10^6 t, were bank erosion and riverbed scoured downstream of the Vanderkloof Dam.

FIG. 4 - The Orange River mouth before the building of large storage dams (photo: M. Petit).

Since the 20^{th} century, much has changed in terms of sediment delivery, as the transport in suspension has increased through soil erosion, and, simultaneously, the flux of sediment reaching the coast has been reduced because of the retention within artificial reservoirs.

Reservoir sedimentation

Some excessive estimates say that it will take the silt load of the Orange River only 50 years to fill Gariep Dam. Nevertheless, in South Africa, a by-product of accelerated soil erosion over the last few decades is the reduction of reservoir storage capacity through sedimentation, and consequently of their lifetime, since the reservoirs trap about 95% of the sediment yield. As a consequence data obtained downstream from this major dam suggest higher rates of chemical than physical weathering: 1.6 and 0.7 $t.km^{-2}.a^{-1}$ respectively, according to Keulder (1979).

A special case of reservoir sedimentation is the Welbedacht Dam on the Caledon River, which was commissioned in 1973 to supply drinkable water to Bloemfontein Municipality via a 115 km long pipeline. Due to siltation the capacity of the dam reduced rapidly from the original 114 Mm^3 to approximately 30 Mm^3 in 9 years and to 17 Mm^3 in 18 years, that is only 15% of the initial storage (Rooseboom, 1992). The reduction in storage created problems in meeting the city of Bloemfontein's demand at an acceptable level of reliability. As a result, a new dam, the Knellpoort Dam, with a gross storage capacity of 137 Mm^3, was built in 1988 to supplement the supply, and in order to prevent similar siltation problems, it was designed as an off-channel storage dam that is fed by water from the Caledon River through a 2 km long canal which is equipped with a silt trap to reduce sedimentation in the reservoir. Nevertheless, it was supposed to lose 50% of its capacity within approximately 20-25 years (fig. 5a).

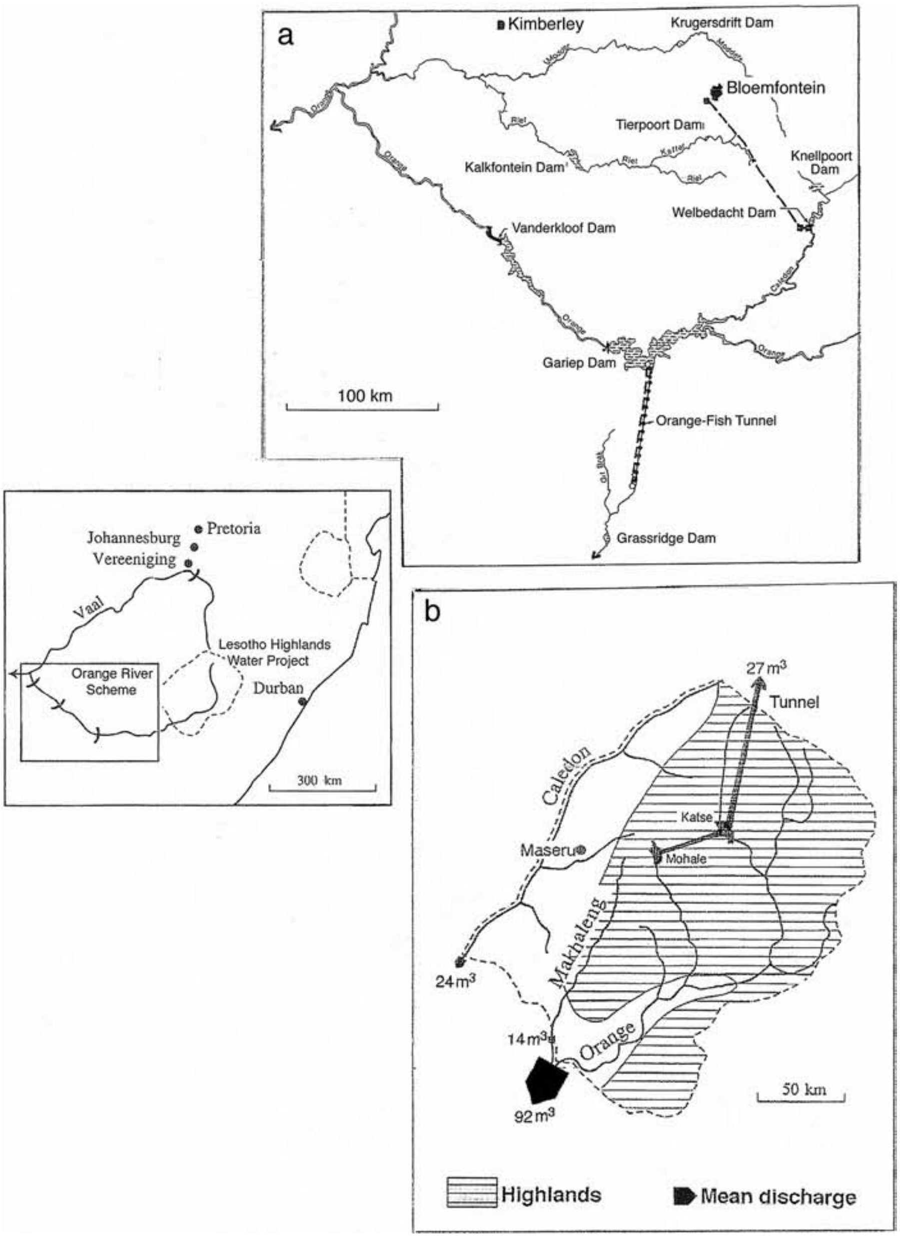

FIG. 5 - A. Implementation of water schemes in Southern Africa; B. The Orange River Project; C. The Lesotho Highlands Water Project.

The high siltation rate is due to the suspended load carried by the Caledon River which, given a drainage area of *ca.* 15,000 km^2, was found to be 360 $m^3.km^{-2}.a^{-1}$ (360 m/Ma), or some 500 $t.km^{-2}.a^{-1}$, mostly made up of fine sand (37% between 80 and 250 μm) and very fine sand (53% < 80 μm) according to Rooseboom (1992). Moreover Chakela (1981) refers to a sediment yield of 1979 $t.km^{-2}.a^{-1}$ for a 945 km^2 subcatchment of the same river which exhibits the highest sediment yields in southern Africa and has a significant part of its upper reaches in Lesotho (43% in surface).

The Lesothan upper reaches

Soil erosion for the Lesotho is estimated to be 5.4 Mt/a based on 12% arable land (3600 km^2) and 18,000 km^2 grazing land having average erosion rates of 750 and 150 $t.km^2.a^{-1}$, respectively. So the mean sediment yield within the country could amount to 400 $t.km^2.a^{-1}$ (270 m/Ma). The highest values are measured in the Lowlands areas where the severity of erosion is aggravated by intensive cultivation of footslope colluvial sediments (fig. 6).

Although weak bedrocks, particularly mudstones, can support badlands (named *dongas*), most of the suspended load is not derived primarily from sedimentary rocks of the upper portion of the Karoo Supergroup (sandstones, silstones and mudstones of the Beaufort Group), as quoted by numerous authors (*e.g.*, Compton & Maake, 2007), but mostly from the dissection of accumulation glacis which are severely gullied. Unconsolidated deposits, named «pedisediments» by South African pedologists, cover large areas of the Lowlands to depths over 20 m, above a basal veneer of gravels, the «Peeble Marker», which contains reworked Acheulian stone artefacts (fig. 7). This observation confirms the importance of near-channel storage close to the sites of erosion, and suggests a significant imbalance in the sediment budget.

Although their presence was mentioned by the first missionaries at the beginning of the 19^{th} century, badlands probably developed as a result of overgrazing and poorly suited cultivation practices between 1850 and 1950. Once established they persist, acting as efficient conduits of water and sediments, and presently small tributaries with drainage area of approximately 5 km^2 are characterised by intense gully erosion, which generates specific sediment yield amounting to some 500 $t.ha^{-1}.a^{-1}$.

The first natural resource of this land-locked country is its abundance of water that has become its greatest source of foreign exchange. As the Gariep and Vanderkfloof dams were no longer sufficient to supply the growing demands, a treaty was signed between Lesotho and South Africa in 1986 in order to divert water from the headwaters of the Orange-Senqu River in the Lesoto Highlands to the area of the Vaal River catchment where South Africa's economic heartland, Gauteng, is situated (fig. 5b). The first of five possible phases has been completed: it consists of two dams: the Katse Dam (the highest dam in Africa at 185 m high) and the 145 m Mohale Dam, respectively completed in 1998 and 2003. Water is gravitated through 82 km of subterranean tunnels from Katse down into the Ash River and eventually into the Vaal Dam (±29 m^3/s). The volume amounts to some 0.9 km^3, but the treaty between the two countries allows for four further feasible phases with a maximum diversion capacity of 2.2 km^3 after the construction of three other dams on the Senqu River. They all will be located in the Maloti Mountains as waters draining basalt rocks are clear; they become charged with suspended sediment only once they have crossed the Lowlands area where 73% of Lesotho's 2.3 million people are living and where the annual sediment production rate may

FIG. 6 - Dongas in the Lesotho Lowlands (photo M. Lageat).

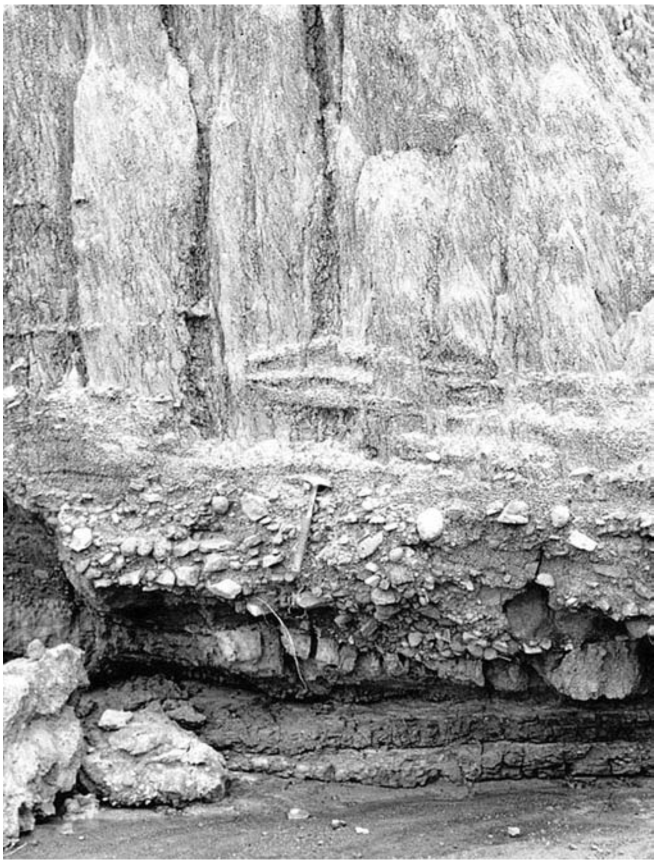

FIG. 7 - The «Pebble Marker» underlying colluvial deposits in the Lesotho Lowlands (photo: Y. Lageat).

exceed 2000 t/km^2, whereas yields are less than 10 t/km^2 in the Highlands (Makhoalibe, 1984).

Soil erosion processes have been studied intensively throughout the last decades in Lesotho (references *in* Lageat, 1999). It is generally assumed a decreasing area-specific sediment yield with increasing drainage basin area. However this scale effect fails to explain the modest production of sediment measured at the monitored agricultural plot scale (< 1 t.ha^{-1}.a^{-1}): soil loss addresses sheet and rill erosion at a given site, although gullies are by far the most important sediment source (> 20 t.ha^{-1}.a^{-1}). Nevertheless this simple law proves to be suitable while comparing erosion rates in badlands areas with the amount of sediment transported to reservoirs. It expresses the importance of sediment storage, which may increase the total amount of soil erosion considerably. Some figures infer a «sedimentary delivery ratio» of 20-30%. On a century timescale, the total amount of erosion from the land surface can be considerably greater than the amount transported out of the catchment because of this temporary storage.

All the former figures have been calculated, assuming a sediment density, which has been found to be about 1,35 t/m^3 in reservoirs. We must return to the density of the solid rocks (*ca.* 2.7) while examining the long-term evolution.

CONTRASTED GEOLOGICAL RECORDS

Following the eruption of the Karoo flood basalts at 183 ± 1 Ma, the geological history of southern Africa has been dominated by the break-up of Gondwana which occurred in the South Atlantic at about 120 Ma. It was preceded by a period of continental rifting starting at about 160 Ma (Brown & *alii*, 2002). Immediately following break-up, the rifted continental margin was subjected to a phase of major denudation.

The epeirogenic uplift

The southern Africa's topography is obviously strongly «bimodal» (Doucouré & de Wit, 2003) in spite of some oversimplification: there is a vast inland plateau of low relief and high average elevation, the so-called «Kalahari Plateau», which is separated from a narrow coastal plain of high relief and low average elevation by an horseshoe-shaped Great Escarpment encircling much of the sub-continent (fig. 8).

A wide range of mechanisms have been put forward by geophysicists in an attempt to model the morphological evolution of the rifted passive margin and to explain this apparently anomalous topography, either episodes of plate-boundary reorganisation or plume-sustained model. The causes and timing of the «Kalahari uplift» remain elusive.

Thermochronology, using mostly apatite fission track analysis (AFTA) across the escarpment flanking the Atlantic coast shows that accelerated cooling, and therefore denudation, occurred in two main episodes with most results clustering at *ca.* 100 Ma: (i) early-Cretaceous (140-120 Ma), and (ii) mid-Cretaceous (100-80 Ma).These peaks complement the measured offshore sedimentation rates as the episodicity of vertical motions at the end Mesozoic is not only established from the exhumation history of southern Africa, but it also seems confirmed by the stratigraphy of sediment accumulated on its western margin where offshore rates indicate also two peaks at 130 to 115 Ma and 86 to 78 Ma (Dingle & *alii*, 1983). That a considerable denudation occurred is demonstrated by the elevations in excess of 1.5 km of the Damaraland Complexes (137-125 Ma) above the surrounding Namib plains (Goudie & Eckardt, 1999).

Since the break-up of Gondwana, the Orange River has delivered large amounts of sediment to the western continental margin, and the Orange Basin has been the main depocentre for Mesozoic and Tertiary sedimentation. Post-rift sedimentary wedge reaches a maximum thickness (*ca.* 6 km) above the transitional crust. Dingle & *alii* (1983) used at least 3 km thick Mesozoic sediments, overlain by up to 600 m of Palaeogene sediments and almost 400 m of Neogene sediments, in order to calculate a sedimentation rate of about 64 m/Ma for the time interval Aptian-Maestrichtian, a thickness of 600 m to estimate the rate of 22 m/Ma for the Palaeogene sedimentation, and a rate of 18 m/Ma during Neogene times.

Rouby & *alii* (2009) have determined the long-term signal of sedimentary supply (at the 10 Ma scale) by com-

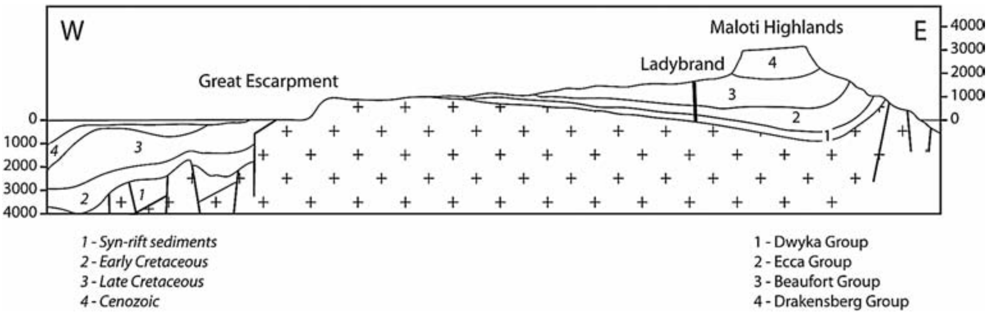

FIG. 8 - Simplified geological cross-section through the South African subcontinent.

paring offshore isopach maps and denudation volumes deduced from thermochronology. Offshore sedimentation studies and results of fission-track indicate that the bulk of the denudation of the south-western margin occurred during the early post-rift phase. The total amount of denudation apparently prompted by the break-up event, decreases from a total of *ca.* 4 km of bedrock between the Great Escarpment and the coast (some 60-70 m/Ma), while inland of this escarpment, vertical denudation removed less than 2 km (30-35 m/Ma).

The kimberlite message

Most of the diamond-bearing kimberlite pipes were emplaced within the main Karoo Basin and therefore within the present drainage network of the Orange basin. Two distinct intrusion peaks have been reported at 143-117 Ma and 95-78 Ma, corresponding to kimberlite Group II and I respectively (Basson & Viola, 2003). Are these two distinct «spikes» of kimberlite intrusion synchronous with the two main regional episodes of accelerated offshore sediment accumulation and onshore denudation? This correspondence remains a question to be discussed.

Kimberlite diatremes contain a wealth of information concerning the composition of the mantle and the lower crust, but they also make possible to reconstruct the erosion history of central South Africa since the level to which kimberlites of different ages are eroded may provide estimates of rates of denudation with time. According to a model which was suggested by Hawthorne (1975) and applied by Lageat (1989) to the morphological evolution of the «interior plateau», upper-crustal xenoliths provide a record of the host sequences into which the kimberlite intrusions were emplaced: certain country rocks may not outcrop in the vicinity of the pipes as a result of post-emplacement erosion, and xenoliths that occur in older diatremes may be absent in younger ones, due to an intervening period of denudation.

Thus a greater former extent of the flood basalts, covering much of central South Africa, is inferred by the presence of basalt xenoliths in many of the kimberlite diatremes, an evidence that clearly indicates the removal of some 950 m of Drakensberg Group basalts subsequent to the effusion of lavas, but prior to the extrusion of the Group II kimberlites. It indicates a surface lowering of *ca.* 30 m/Ma (*ca.* 80 t.km^{-2}.a^{-1}). Such a statement refutes the ultimate claim by King (1978) of the preservation of a pre-break-up Gondwana surface.

The presence or absence of xenoliths of Karoo Supergroup sedimentary rocks in the kimberlite diatremes reflects respectively the presence or absence of specific stratigraphic units within the main Karoo Basin at the time of eruption. Rock material sunken into volcanic pipes enable the reconstruction of the land surfaces: thus, in the Kimberley area, approximately 500 m of erosion is estimated to have occurred from 120 to 85 Ma and 850 m from 85 Ma to the present day at average rates of approximately 15 m/Ma and 10 m/Ma respectively (Hanson & *alii*, 2009). But the cross-sections which illustrate the paper suggest that most of the denudation was over at 65 Ma and could have therefore exceeded 40 m/Ma during late Cretaceous, even in the continental interior (fig. 9). On the basis of thermal history modelling of apatite fission track data for the Ladybrand borehole west of the Lesotho

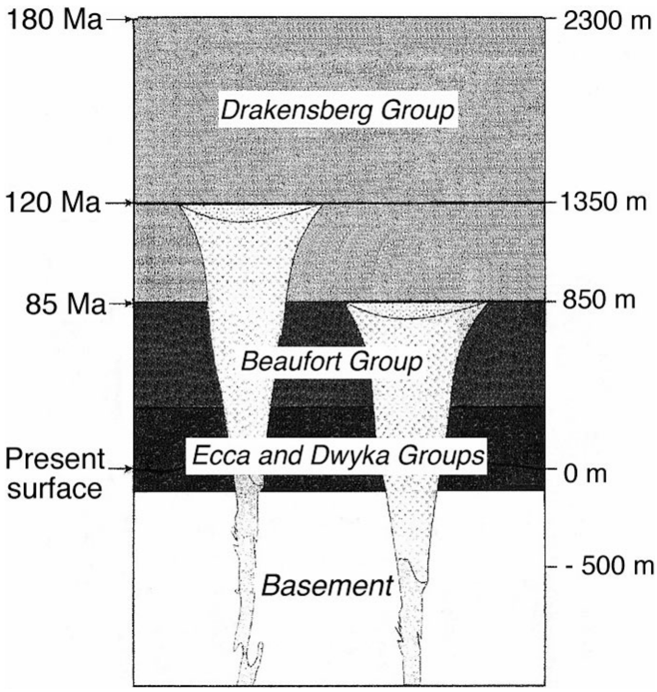

FIG. 9 - Emplacement positions and post-emplacement erosion of kimberlite pipes in relation to the stratigraphy of the host-rock sequences of the Karoo Supergroup (after Hanson & *alii*, 2009).

Highlands, Brown & *alii* (2002) confirm a phase of accelerated denudation from 78 to 64 Ma.

Thanks to these results, it has been suggested that the Cretaceous is the time when most of the elevated topography of southern Africa was generated, ignoring the Cenozoic contribution.

The Cenozoic enigma

As previously noted, low-temperature geochronological methods suggest that a major epeirogenic uplift occurred during Cretaceous times, although Burke (1996) argues for a low-lying African continent prior to 30 Ma and even Partridge & Maud (1987) postulate a late Pliocene (*ca.* 2.5 Ma) vertical movements of more than 100 m along the western section of the Great Escarpment.

A suite of new techniques is supposed to have revolutionised researches in long-term landscape evolution, notably low-temperature thermochronometers: because of the poor resolution of apatite fission track thermochronology at very shallow crustal depths, attempts have been made to use the lower temperature sensivity of the (U-Th)/He analysis in apatite to constraint lower rates during the post-Cretaceous thermal history. Flowers & Schoene (2010) restricted Cenozoic denudation in eastern Transvaal to less than 850 m, which is far from being negligible in spite of their assertions.

If we turn to the western margin, mean annual sediment yield of the Orange River system would show a progressive decrease from 10.10^6 m^3 in the late Cretaceous to 3.86 in the Palaeogene and 0.3 in Neogene times (Dingle & Hendey, 1984). However, it is difficult to convert accumulated volumes to depths of denudation because it requires knowing the location of long-vanished divides and the limits of the contributing drainage area (Dollar, 1998).

Do these results entirely preclude a phase of uplift during Cenozoic times? Persuasive evidence for post-Cretaceous warpings has been advanced to account for a history of radical drainage arrangements, which have reconfigured the geometry of main rivers networks. There has been a major shift in the location of the mouth of the ancestral Orange River since studies of the sediment within the Orange Basin indicate that the river entered the South Atlantic through the Cape Canyon, some 300 km south of its present mouth, during the Oligocene (Dingle & Hendey, 1984). The subsequent abandonment of the flow is thought to be associated with tectonic warping, and it is well established through the distribution of diamantiferous gravels that the headwaters of the Vaal River were also truncated by recent movements occurring along the Griqualand-Transvaal axis which forms its northern watershed (Moore, 1999).

As previously noted, the Orange River exhibits a dramatic contrast between its alluvial channel in its middle course and a bedrock gorge which is under the structural control of a fracture line. The Augrabies Falls act as regional base level for areas that are upstream. The persistence of such a knickpoint at the edge of the interior plateau throughout tens of millions of years of landscape evolution is puzzling (fig. 10).

Unlike most tropical rivers devoid of coarse abrading-tools, that hence retards downcutting into the solid rock, the Orange-Vaal drainage has transported significant quantities of clastic sediment, including diamonds, from hinterland sources to the Atlantic Ocean since at least the Mid-Eocene. Whereas the early Orange River supplied enormous volumes of sand and fine-grained material to the continental margin during the initial construction of the continental shelf, the bulk of Namibian diamonds had been introduced to Atlantic littoral environments since at least Middle Eocene times (*ca.* 42 Ma), according to Spaggiari & *alii* (2009): «*This shift from fine- (silt to clay) to coarse-grained (gravel to sand) deposition at the mouth of the ancestral Orange River reflected a regional subcontinental uplift in the late to end Cretaceous that initiate deep fluvial incision which has continued intermittently through much of the Cenozoic*».

If the main flux of coarse gravel is derived from the Neoproterozoic Nama Group in the nearby escarpment, gravel terrace remnants are preserved along the lower courses of the Vaal and, owing to a large number of exposures available in diamond diggings throughout the area, it

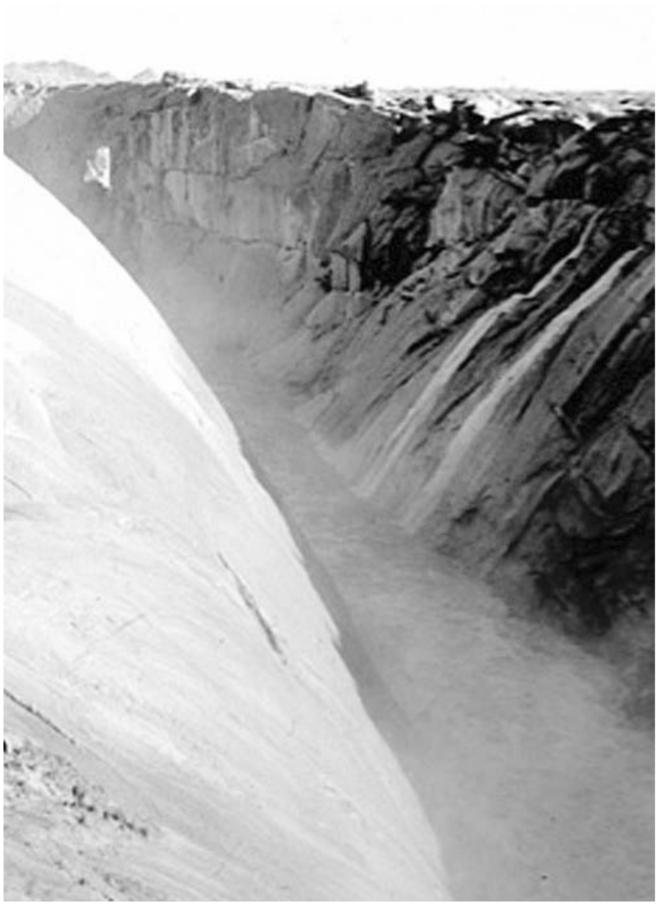

FIG. 10 - The knickpoint formed by the Aughrabies Falls (photo: Y. Lageat).

has been shown that the index of rounding of alluvial clasts argues for a transport over long distances (Helgren, 1979). So, even if traction loads (estimated to be only 5% of the suspended load) are presently difficult to obtain, the coarseness of the sediment the Orange River carried in the past should have been more effective as an agent of abrasion in the steepened reaches after the rejuvenation of the drainage network.

For this reason, and others (Lageat, 2000), we postulate that the western section of the Great Escarpement has experienced a discrete uplift in the recent past: the crustal unloading associated with its retreat would have lead to an isostatic uplift of the margin (Gilchrist & Summerfield, 1990), and the flexural effect of a marginal upward is substantially comforted by the percentage hypsometric curve for the Orange River drainage basin (fig. 11).

Unfortunately, no evidence may substantiate the reality and timing of these movements. If a general lack of Tertiary apatite ages suggests that major exhumation was over at the end of Cretaceous times, de Wit (2007) stated that «*subsequent Cenozoic motions are more subtle, ill-defined and poorly dated*».

The analysis and interpretation of cosmogenic nuclides (^{3}He, ^{10}Be, ^{10}Al, ^{21}Ne, ^{36}Cl) have given geomorphologists a new opportunity to measure rates of erosion as the window of time (10^3 to 10^5 years) falls between hard rock chronometers and methods used to assess contemporary sediment yield (Bierman et Caffee, 2001; Cockburn & *alii* 2000; Kounov & *alii*, 2007). Preliminary results have provided erosion rates for two sections of the escarpment and interior plateau, indicating they have been an order of magnitude lower during the last few hundred thousand years than the Cretaceous (less than 5 m/Ma). Conversely, this erosional atony revealed by cosmogenic isotope data allows to highlight modern, anthropogenically-enhanced erosion rates: increases in river fluxes estimated by quantifying sediment accumulation offshore indicate a 13-fold increase since the Neogene for the west coast (Dingle & Hendey, 1984), but it is generally admitted that a switch to a more arid climate has reduced the supply of sediment to the margin since the onset of Benguela upwelling at 15 Ma (Ségalen & *alii*, 2006).

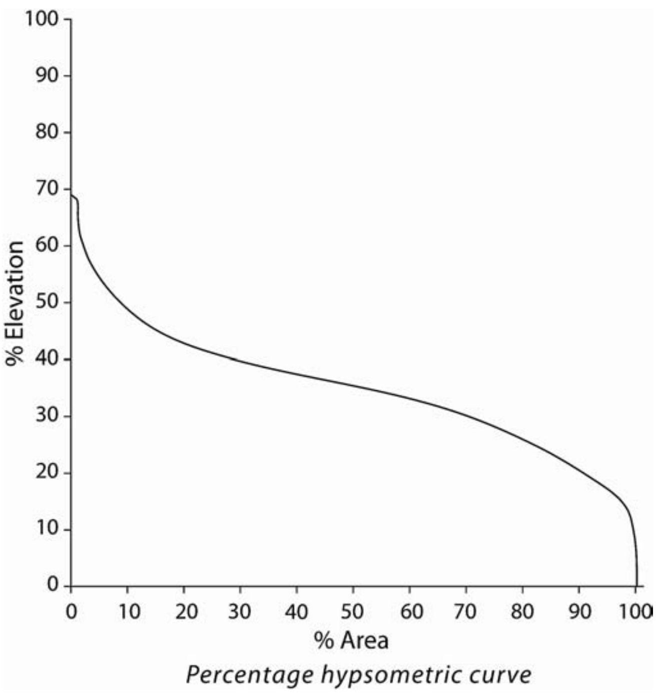

FIG. 11 - Hypsometric curve for the Orange drainage basin.

CONCLUSIONS

One of the biggest challenge faced by Earth scientists is the development of an ensemble of techniques that can be integrated to quantify geomorphic processes across a spectrum of timescales. If landforms are experienced recent and present dynamics, they are also the product of longer evolutions. Thereby they hold a central place in the geomorphological tradition. But what is the appropriate scale for studying these landforms? In the past five decades, two conflicting conceptions of the genesis of landforms have emerged: dynamic geomorphology or megageomorphology. This implies that there are two ways of acquiring an understanding of landforms through the studies of small-scale surface processes or large-scale endogenic processes. In the Orange River basin, the analysis of landforms evolution has been addressed at both contrasting spatial and temporal scales. The cross-checking of abundant available data demonstrates clearly such spatial and temporal heterogeneities in erosion rates (fig. 12).

Although the calculations are little more than rather crude approximations, measurements of current rates of processes appear irrelevant for interpreting long-term landscape evolution. Attempting to upscale short-term to long-term erosion rates appears to be a fruitless effort. It could indeed happen that some consistencies over extended periods of time may be observed, even in so contrasted morphostructural contexts as the Nanga Parbut area of

Period	Sediment discharge 10^6 t yr^{-1}	Erosion rate Bubnoff^{-1}
Late Cretaceous (78-64 Ma)		80
Palaeogene		<10
Neogene		<5
Holocene	5	4
Pre-1921	119	96
1929-1934	89	72
1934-1943	56	45
1943-1952	52	42
1952-1960	46	37
1960-1969	34	28
1980	<17	14

FIG. 12 - Variations in erosion rates at different time-scales (after Bremmer & *alii*, 1990, and various authors).

the western Himalayas (Burbank & *alii*, 1996) or the loess belt of central Belgium (Peeters & *alii*, 2008). However, in strong contrast with these attempts, linking modern process studies with the full compass of landscape evolution seems illusive since the nature of the forces in action and the landforms they are able to generate are irreducible to each other. A gully will not evolve into a valley, a valley-side slope into a regional scarp, a bench into a planation surface: they belong to two distinct orders with no filiation between them, each one having its own specific logic and significance. The detailed process understanding at the small-scale cannot be transferred to macroscale landscape development. This conspicuous and unsolvable disparity has been admirably demonstrated by Baulig (1959) before the emergence of surface process geomorphology and the advent of plate tectonics…

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