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# River sediments

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River history is reflected in the nature of the sediments carried and deposited over time. Using examples drawn from around the world, this account illustrates how river sediments have been used to reconstruct past environmental changes at a variety of scales in time and space. Problems arising from a patchy alluvial record and from influences external to the river basin can make interpretation difficult. The Nile is treated in some detail because its history is further complicated by tectonic, volcanic and climatic events in its headwaters and by enduring human impacts. It arose soon after 30 Ma. Since that time approximately 100 000 km$^3$ of rock have been eroded from its Ethiopian sources and deposited in the eastern Mediterranean, with minor amounts of sediment laid down along its former flood plains in Egypt and Sudan. From these fragmentary alluvial remains, a detailed history of Nile floods and droughts has been reconstructed for the last 15 kyr, and, with less detail, for the past 150 kyr, which shows strong accordance with global fluctuations in the strength of the summer monsoon, which are in turn perhaps modulated by changes in solar insolation caused by changes in the Earth's orbit and by variations in solar irradiance.

**Keywords: alluvial history; Nile basin; river metamorphosis; suspension load; traction load**

## 1. Introduction

> For it is clear to any intelligent observer … that Egypt … is, as it were, the gift of the river and has come only recently into the possession of its inhabitants … the greater part of the country I have described has been built up by silt from the Nile … the soil of Egypt does not resemble that of the neighbouring country of Arabia, or of Libya, or even of Syria … but is black and friable as one would expect of an alluvial soil formed of the silt brought down by the river from Ethiopia.
>
> Herodotus (*ca* 485–425 BC) [1, pp. 104–106]

Most rivers are efficient conveyors of water and sediment from their headwaters to the sea, as Herodotus discovered nearly 2500 years ago [1]. The Amazon presently contributes 20 per cent of all fresh water brought by rivers to the sea and carries on average approximately 1200 million tonnes of sediment to the Atlantic each year [2]. In a pioneering study, Gibbs [3] estimated that roughly 90 per cent of both the dissolved and suspended loads transported by the Amazon to the sea came from only about 10 per cent of the total catchment area, namely the

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One contribution of 10 to a Theme Issue 'River history'.

 

Andean headwaters. The amount of fluvial erosion in other mountainous regions is equally impressive. The Arun River in Asia has maintained its course through the Himalayas, keeping pace with Cainozoic uplift, and creating one of the deepest valleys on the Earth [4]. In Africa, the Blue Nile and Tekazze rivers have eroded approximately 100 000 $km^3$ of rock from the Ethiopian Highlands during the last 30 Myr, carving gorges to rival the Grand Canyon. Much of the resultant sediment has been deposited in the eastern Mediterranean, with minor amounts of sediment laid down as flood plain deposits in Egypt and Sudan [5].

Given the importance of the upper reaches of river basins as sources of water and sediment, it is little wonder that recent deforestation and land-use change in the headwaters of many tropical rivers has led to a change in flow regime and, in some instances, to an influx of sediment onto previously unaffected farmland [6,7]. These present-day changes in river behaviour and sediment transport mimic the changes associated with the climatic fluctuations of the last few million years, albeit taking place at a rate far faster than the environmental changes evident in the Quaternary alluvial record.

The following account considers how river sediments may be used to reconstruct alluvial history and the environmental fluctuations responsible for changes in fluvial deposition over time. A perennial problem inherent in using river sediments to reconstruct river history is the fragmentary nature of the alluvial record. Indeed, it is often necessary to seek a more complete archive in the correlative marine sediments. Another issue is recognizing the influence of factors external to the river basin, such as changes in river basin boundaries caused by tectonic and hydrological events. It is therefore important to recognize that problems arising from a patchy alluvial record, from human activities and from influences external to the river basin can make interpretation difficult.

The Nile is treated in some detail in this account not only because it is a big and very well-studied river, spanning a wide climatic and altitudinal range, but also because its history is further complicated by tectonic, volcanic and climatic events in its headwaters and by prolonged and varied human activities. In addition, a focus on the Nile allows the reader to follow in some detail the different methods employed to decipher river history from river sediments, with some salutary but far from obvious lessons for natural resource management.

## 2. River sediment inputs to the ocean

Thirty years ago, the best estimates for the total amount of river sediment transported each year to the oceans amounted to about 20 billion tonnes (approx. $2 \times 10^{10}\,\mathrm{t\,yr^{-1}}$) [8]. That estimate was based on data from about 100 rivers and has since been revised upwards, partly because better data, from over 500 rivers by 1997, revealed that the huge input of sediment from many small and hitherto poorly monitored rivers such as the Fly in Papua New Guinea ($0.085 \times 10^6\,\mathrm{t\,yr^{-1}}$) had not been adequately considered, but chiefly because of accelerating changes in land use within major tropical rivers [9]. However, with the proliferation of large dams in the last few decades, such as the Aswan High Dam in Egypt and the Three Gorges Dam on the Yangste in southern China, some of the sediment that would normally reach the ocean is now retained in large reservoirs, thereby reducing their water storage capacity and effective life. Another consequence of recent dam construction has been increased coastal erosion along the deltas downstream, because the former equilibrium between sediment input from rivers flowing to the sea and sediment removal by wave action and longshore drift has been altered.

Drilling into ocean sediments during the search for oil in the 1960s and 1970s allowed rates of oceanic sediment accumulation to be determined for the past 65 Myr of Cainozoic time (figure 1) [10]. Interpreting the results has not been straightforward, with some workers arguing that low rates of accumulation reflected low rates of sediment input during episodes of global aridity, citing the relatively low sediment yields from present-day Australian rivers as a possible modern analogue. Another explanation for the high rates of non-carbonate sedimentation in the latter part of the Cainozoic (figure 1) is that they reflect accelerated tectonic uplift and mountain building at this time, most notably in the Himalayas, Andes and Rockies. Indeed, Ruddiman *et al.* [11] have argued that the increased rates of erosion triggered by tectonic uplift in the latter half of the Cainozoic have led to enhanced silicate weathering and a draw-down of atmospheric carbon dioxide, accompanied by Cainozoic cooling and desiccation. It is also no coincidence that over half of all river sediments supplied to the ocean today come from a few big rivers, each originating in regions of recent tectonic uplift. These rivers include the Amazon ($1.2 \times 10^9\,\mathrm{t\,yr^{-1}}$), the Yangste ($0.48 \times 10^9\,\mathrm{t\,yr^{-1}}$) [12] and the combined Brahmaputra–Ganges rivers in Bangladesh ($1.7 \times 10^9\,\mathrm{t\,yr^{-1}}$) [13]. Such figures do not take account of sediment stored in flood plains or of the contributions from dissolved and traction load. (*Traction load* or *bed load* is that part of total sediment load moved on or just above the channel bed, with particles being mostly boulders, cobbles, pebbles and gravel. *Suspension load* refers to mainly silt- and clay-sized particles of low settling velocity held in suspension and kept aloft as a result of upward currents in eddies of turbulent flow. Once the current slackens, they will settle to the bottom of the fluid in accordance with Stokes' Law.)

Both the Indus and the Hwang Ho (Yellow River) have very large offshore deltas but little sediment now reaches the ocean today from those highly regulated and flow-deprived rivers [13,14]. The distal reaches of many river valleys have become more and more intensively cultivated, so that the Loess Plateau of central China, for example, has some of the highest rates of erosion on the Earth [15]. However, rapid rates of rock weathering, high relief, steep slopes and intense summer monsoon rains are the decisive factors in the high rates of sediment yield from the big rivers.

Much effort has gone into using changes in the rates of fluvial sedimentation offshore, together with changes in the terrestrial pollen content, to reconstruct past changes in climate in Africa and elsewhere [16]. One difficulty with this approach relates to how fluvial inputs are to be distinguished from silts derived from dust plumes, as in the case of the Harmattan dust blown from West Africa and studied by Darwin in the Cape Verde Islands in January 1832 [17]. Does an abundance of clay- and silt-sized sediment offshore denote an aeolian provenance or does it simply mean that the rivers of the day were transporting abundant fine sediment in suspension? More subtle interpretation is needed if the valley-fill deposits now being eroded originated as loess deposits many thousands of years earlier, as is the case with the Matmata Hills of Tunisia, the piedmont valleys of Namibia, the valleys of Sinai and the Flinders Ranges of South Australia.

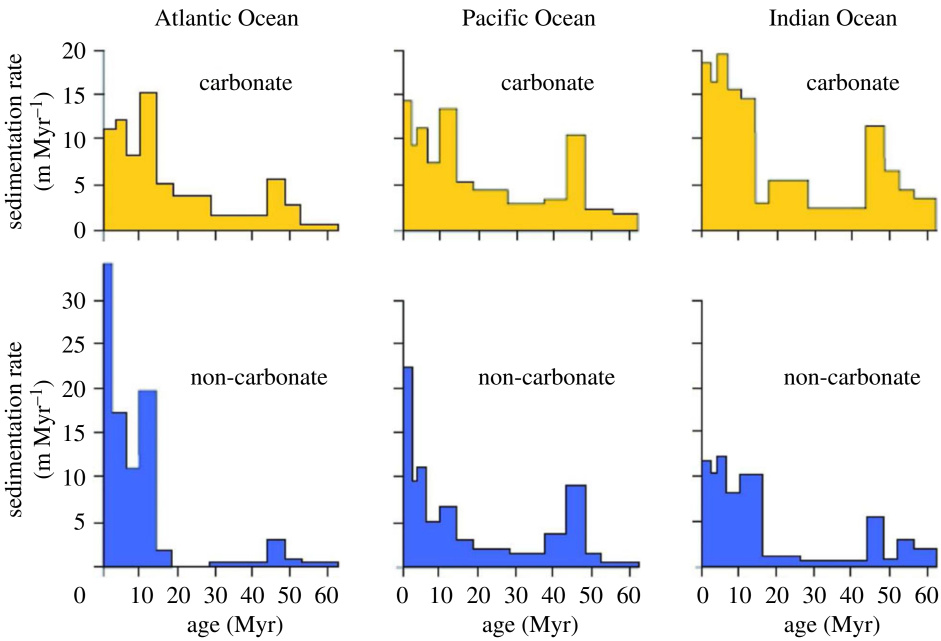


Figure 1. Cainozoic sedimentation rates in the Atlantic, Pacific and Indian Oceans. Rates were calculated from the graphs in Davies *et al.* [10] and are displayed here as histograms. (Online version in colour.)

## 3. Factors controlling river sediment yield

Figure 2*a* illustrates very schematically the dominant roles of tectonic history and climate in controlling sediment load in rivers, with the former primarily controlling relief, rock type and basin form, and the latter rainfall and run-off. Soil and plant cover reflect the interaction between climate and geology. The reality is of course more complex, with soil development determined by the ‘passive’ factors of rock type and topography and the ‘active’ factors of soil climate and biological activity, operating in combination over time. Figure 2*b* provides a schematic overview of the factors controlling dissolved load in rivers, with climate and rock type as primary agents once again modulated by plant cover and soil. In arid areas, sediment production is weathering-limited and, in humid areas where weathering is rapid, sediment production is erosion-limited.

This simple schema ignores sediment inputs to rivers from outside the immediate catchment in the form of wind-blown dust or advancing sand dunes, as well as inputs from more extreme events such as the sudden release of water from a glacially dammed lake [18,19], resulting for example in the ‘Channeled Scablands’ of Dakota [20]. Climate operates primarily through rainfall and run-off, with good evidence that highest sediment yields coincide with areas of highest seasonality of rainfall, such as the seasonally wet tropics and summer monsoon domains of Asia, Africa, South America and Australia [3,9,21,22]. In areas of long dry season and frequent fires, for the same unit momentum of raindrop impact, soil loss from bare burnt surfaces at the start of the wet season can be an order

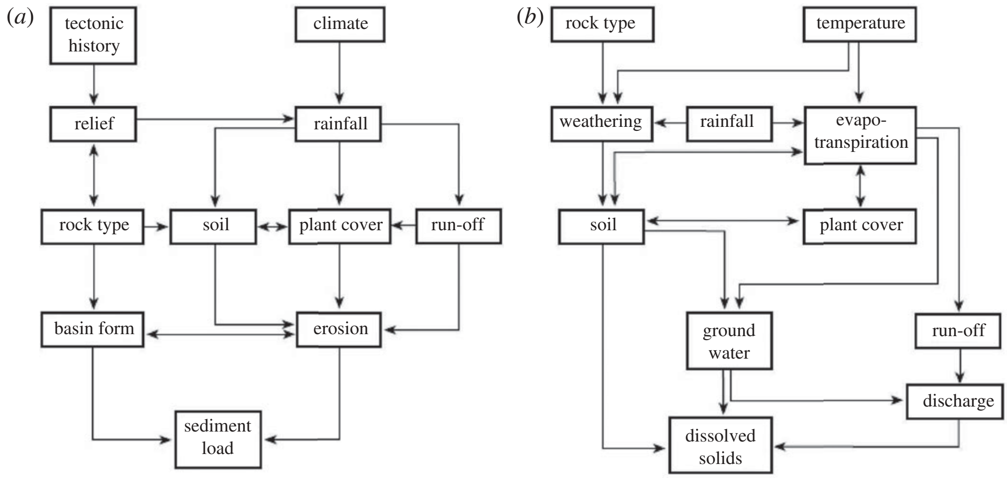


Figure 2. Factors that control (*a*) suspended load and (*b*) dissolved load in rivers.

of magnitude higher than at the height of the wet season, when plant cover is at a maximum [23]. If extensive areas of ground are left bare during the wet season, as a result of either forest clearing or cultivation, the ensuing rates of erosion can be several orders of magnitude higher than would otherwise be the case, with concomitant choking of wetlands and river channels with excess sediment.

An interesting set of questions for river managers is how much of the sediment contributed from the valley slopes to river channels is carried to the sea and how much is stored within the river terraces, active flood plains, levees and back swamps, and for how long? Detailed mapping and dating of alluvial deposits in the lower reaches of the Alligator Rivers in the Northern Territory of Australia demonstrated that the river channels had been actively shifting course and building up their flood plains for several thousand years at least, in response to the attainment of a reasonably stable base level once the sea in this region had reached its present level some 6 kyr ago [24].

## 4. Suspension load, bed load and river metamorphosis

Nearly half a century has passed since Leopold *et al.* [25] published their ground-breaking monograph *Fluvial processes in geomorphology*. This far-reaching synthesis inspired later workers to undertake more detailed studies of river channel hydraulics, channel patterns and stream power, flood plain processes, palaeohydrology, human influences on rivers and river management [26–32].

One outcome of these studies was confirmation of the empirical insights of Leopold & Wolman [33] and Leopold *et al.* [25] that river channel pattern and form are intimately related to sediment type, changing in response to changes in sediment supply, sediment size and precipitation regime. These latter changes are in turn determined by events upstream (such as changes in vegetation type and cover) that influence the ratio of load to discharge as well as the particle size range contributed from hill slope to valley bottom. For example, rivers in arid areas and in recently deglaciated regions such as much of New Zealand, Alaska and northern Europe often have a braided channel pattern, carry a large traction load of generally coarse and non-cohesive pebbles and boulders, shift channel course frequently and have steeply sloping, wide and shallow channels. However, no single channel morphology can be considered as typical of the periglacial zone [34], and the same is true of arid areas.

At the other extreme are rivers that flow in wide meanders across low-gradient floodplains composed mainly of clay. Such winding rivers tend to have relatively deep and narrow channels, a classic example (and origin of the noun *meander*) being the Büyükmenderes in what is now Turkey, known to the ancient Greeks as the River Maiandros in Phrygia. As a general rule, such meandering rivers have 'suspension-load' channels and braided rivers have mainly 'bed-load' channels [28,35]. An intermediate category of 'mixed load' channels shares some of the attributes of both ends of the spectrum, depending upon the ratio of load in suspension to load in traction. River channels can change during the course of several seasons, decades or centuries from one type of channel to another, depending upon changes in sediment and water influx. Schumm [36] termed such a change 'river metamorphosis'.

The physical processes involved in such changes depend upon often small changes in stream power ($W$), defined by Bagnold [37] as the rate of energy loss per unit length of stream, expressed per unit width of channel as the product of tractive force ($r$) and velocity ($V$),

$$W = rV. \tag{4.1}$$

Tractive force is the product of hydraulic radius ($R$) (i.e. the channel cross-sectional area divided by the wetted perimeter), slope ($S$) and the specific weight of the fluid ($y$),

$$r = yRS. \tag{4.2}$$

Both stream power and sediment transport rate are proportional to stream velocity cubed [28]. Once stream power falls below a limiting threshold value, bank erosion and sediment transport will diminish, leading to a change in channel pattern from braided to meandering.

In his investigation of the former channels that now criss-cross the floodplain of the Murrumbidgee (figure 3) in the aptly named Riverine Plain of southeast Australia, Schumm [35] noted that the sediments filling two types of former channel differed in lithology. The 'ancestral stream' channels were sinuous with meander wavelengths several times those of the present meandering channel, and were filled with mainly fine sediment, consistent with their sinuous channel pattern. The 'prior stream' channels on the other hand contained a coarser channel fill and were linear in plan, with wide relatively straight channels. Schumm concluded that the sinuous ancestral channels were suspension-load channels formed at a time when the overall climate was wetter than that of today and bankfull discharge several times greater than at present, in contrast to what he considered the more seasonal flow regime of the prior channels, which experienced episodically very high discharge from more sparsely vegetated headwaters. Bowler [38] built upon the pioneering work of Schumm and provided

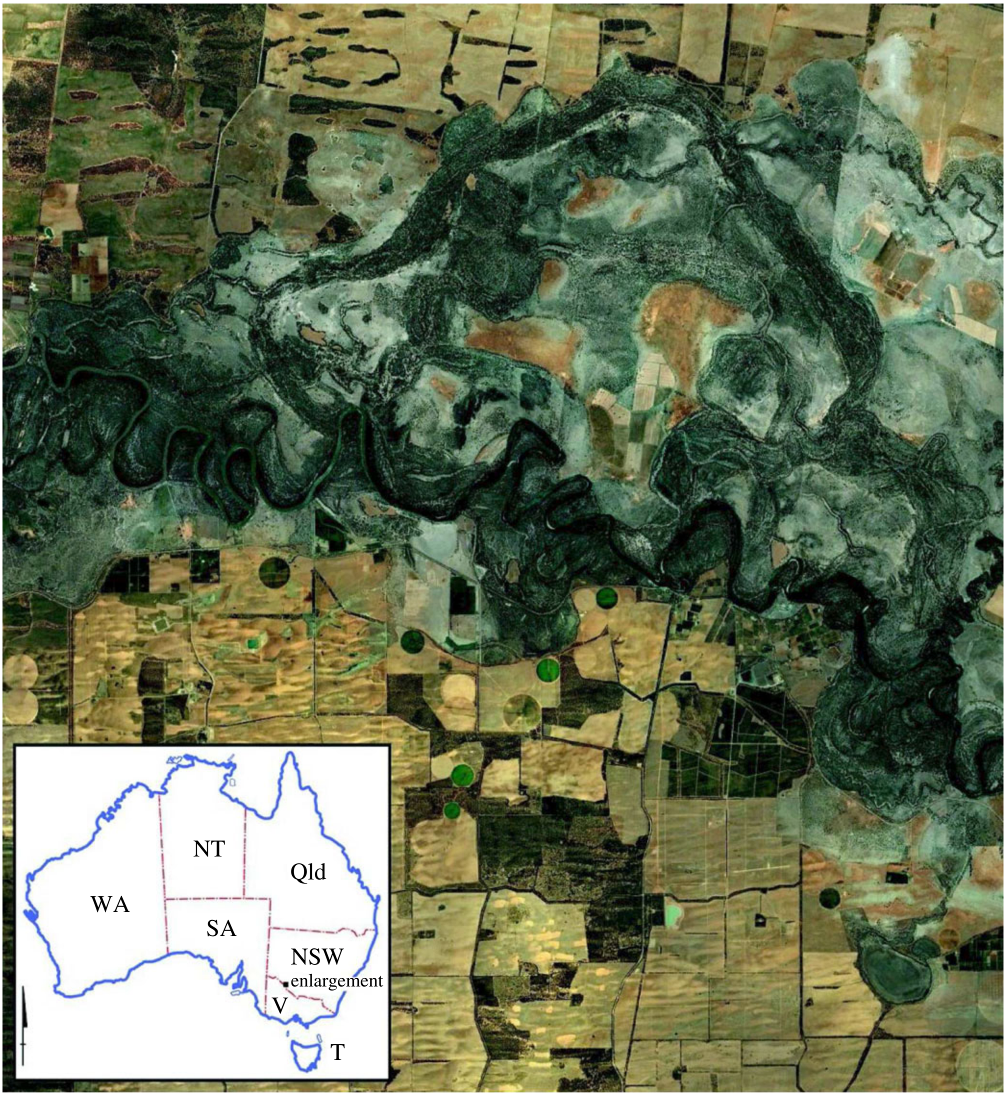


Figure 3. Wide Late Pleistocene palaeochannels and narrow modern river channel, lower Murrumbidgee valley, southeast Australia, 27 December 2006. (Google 2010, Image 2011 Digital Globe, 2011 Cnes/Spot Image, 34°40′40.03″ S 143°10′31.30″ E, elevation approx. 60 m). (Online version in colour.)

the first coherent radiocarbon chronology of the Late Pleistocene palaeochannels of the Riverine Plain, while noting that some of the alluvial sediment exposures reflected neotectonic activity within parts of the catchment.

Later workers used a combination of radiocarbon and luminescence techniques to extend the alluvial chronology of the Murray–Darling basin of Australia and of the rivers draining westwards into Lake Eyre, finding that times of high inferred river flow coincided with independently dated times when lake levels were also high [39–42]. One interesting aspect of this research was the recognition of a process of sediment recycling, with source-bordering dunes being generated from river channel sands and later becoming reincorporated into the fluvial sediments. For source-bordering dunes to form, three conditions are necessary: a regularly replenished supply of channel sands, sparse riparian vegetation and strong unidirectional winds, at least seasonally.

In presently semi-arid west-central New South Wales, the alluvial landscape consists of highly seasonal sandy alluvial channels, source-bordering dunes, Late Quaternary alluvial fans derived in part from the reworking of such dunes and small ephemeral lakes with Late Pleistocene sandy, gypseous and clay lunettes on their downwind margins. Strong similarities between the Late Quaternary sedimentary facies of western New South Wales and the Late Triassic of Somerset in southwest England, together with similar fossils (charophyte oogonia) and evaporite minerals (carbonate, gypsum), led Talbot *et al.* [43] to postulate that the Late Quaternary landscape of east-central Australia provided a modern analogue for the Triassic environment of Somerset.

## 5. Dating alluvial sediments

Efforts to reconstruct river history usually start with a careful analysis of the physical properties of these sediments, followed by a detailed investigation of their geochemistry and fossil content [44–46]. Inevitably, the study will require a precise and reliable chronology, i.e. one in which the age of the sample obtained actually relates to its time of deposition. However, very recent flood sediments may contain reworked charcoal fragments several thousand years older than the actual sediment [47]. Caution is also required when interpreting ages because many alluvial formations are time-transgressive [48–50].

Earlier workers had to rely upon a few flecks of charcoal, sporadic prehistoric stone artefacts, occasional sherds of pottery and, if lucky, ancient coins, in order to devise an alluvial chronology for coastal valleys around the Mediterranean [51] and tropical floodplains in Asia and Africa. In Australia, the 'post-European' alluvial horizon is demarcated at the base by fragments of bottle glass and rusting lengths of fencing wire. In peninsular India, widespread volcanic ash deposits were laid down across the sub-continent as a result of the eruption of Toba volcano in northern Sumatra 73 kyr ago. The jury is still out on whether or not they constitute an isochronous marker bed, with some claiming that reworking of the primary air-fall mantle nullifies this claim and others arguing for rapid remobilization of the original ash, with deposition in back-swamps and depressions and preservation beneath younger alluvium [52].

Useful as relative dating methods are as a first approximation in the field, there is now a veritable battery of techniques (table 1) for dating alluvial sediments over and above the long-established and generally reliable radiocarbon dating with its effective upper limit of 40–50 kyr. These methods include magneto-stratigraphy, potassium–argon ($^{40}K/^{40}Ar$) and argon–argon ($^{40}Ar/^{39}Ar$) dating methods, and dating using luminescence techniques, cosmogenic nuclides and uranium series. Each method is useful at different temporal scales and for particular types of sediment [53–55].

Luminescence dating is a widely used and versatile technique for dating when grains of quartz or feldspar were last exposed to daylight [56,57]. Under ideal conditions, it can provide a million year record, but its more usual upper limit is

Table 1. Methods commonly used in direct and indirect dating of river sediments, modified from Williams *et al.* [53, table A1]. AMS, accelerated mass spectrometry; TIMS, thermal ionization mass spectrometry.

| method | range | precision |
|---|---|---|
| I. correlation methods | | |
| tephrochronology | 0–5 × $10^9$ years | 0.5% |
| geomagnetic reversals | 0–5 × $10^9$ years | 0.5% |
| orbital variations | 0–5 × $10^9$ years | 0.5% |
| II. radioisotope parent–stable daughter | | |
| potassium–argon | approximately 50 kyr to 5 × $10^9$ years | 0.5% |
| argon–argon | approximately 10 kyr to 5 × $10^9$ years | 0.5% |
| radiocarbon (conventional) | 0–40 kyr | 0–6 kyr, 60 years; 6–30 kyr, 1%; greater than 30 kyr, greater than 10% |
| radiocarbon (AMS) | 0–50 kyr | as above |
| III. disequilibrium between parent and daughter radioisotopes | | |
| U-series: I, zero initial $^{230}$Th | 0–250 kyr ($\alpha$-spec)<br>0–500 kyr (TIMS) | 1% ($\alpha$-spec)<br>less than 0.5% (TIMS) |
| U-series: II, excess initial $^{230}$Th | 0–250 kyr | 10–20% |
| IV. trapped electrons | | |
| thermoluminescence | 0 to 100–500 kyr | 10% |
| OSL (optical dating) | 0–100 kyr | 7–10% |
| ESR (electron spin resonance) | 0–1 × $10^6$ years | 10–20% |
| V. cosmogenic isotopes | | |
| Beryllium-10 | $10^1$–$10^6$ years | 15% |
| VI. chemical methods | | |
| amino acid racemization | 0 to 100–500 kyr | more than 15% |

approximately $4 \times 10^4$ yr [58]. Magneto-stratigraphy can operate at scales from $10^1$ to $10^{10}$ yr, but can only provide a relative chronology and so needs independent calibration by other methods. Cosmogenic nuclide dating is becoming an increasingly useful method of dating soils, sediments and rocks back to $10^6$ yr, but has not so far been much used for dating alluvial sediments, although initial results appear promising. Uranium-series dating is useful for dating tufa deposits within alluvial sequences but has mainly been used to date flowstone deposits in limestone caves. Both $^{40}K/^{40}Ar$ and $^{40}Ar/^{39}Ar$ dating methods are widely used to date tephra and lavas sandwiched within Quaternary and older river sediments, and are generally used for material older than about $5 \times 10^4$ yr.

Against this general background, we now consider one river—the Nile—in some detail, because this will allow us to provide a more integrated account of how different techniques can be deployed to reconstruct river history using river sediments, with some unexpected lessons for natural resource managers.

## 6. Reading Nile alluvial history from sediments

### (*a*) *Nile hydrology and sediment load*

The Nile is the longest river in the world, with a total length of 6670 km from the headwaters of the Kagera River in Uganda to the Mediterranean Sea, and drains an area variably estimated at 2.96–3.25 million $km^2$, although this has varied in the past (figure 4). Two of the three most important contributors of water and sediment to the main Nile are the Blue Nile/Abbai and the Atbara/Tekezze, both of which rise close to one another in the volcanic highlands of Ethiopia, where they have eroded approximately 100 000 $km^3$ of rock and cut spectacular gorges nearly 2 km deep and 35 km wide [5]. These two rivers provide, respectively, 68 per cent and 22 per cent of the peak flow and 61 per cent ($140 \pm 20$ million t $yr^{-1}$) and 25 per cent ($82 \pm 10$ million t $yr^{-1}$) of the total Nile sediment load, which amounts at present to $230 \pm 20$ million t $yr^{-1}$ [59,60]. This is in strong contrast to the White Nile, which provides relatively little sediment to the main Nile but is important for quite a different reason. The White Nile flows from the lake plateau of Uganda and disappears into the extensive swamps of southern Sudan, where it emerges as a river of nearly constant flow throughout the year. It provides 83 per cent of Nile discharge during the month of lowest flow and is responsible for maintaining perennial flow in the main Nile during extreme drought years in Ethiopia [61]. The main Nile flows north through the eastern Sahara desert, receiving no further water or sediment once north of the Atbara confluence, reaching the Mediterranean Sea after a waterless journey of 2689 km.

Samuel Baker [62] aptly described the Blue Nile as 'a rapid mountain stream, rising and falling with great rapidity' in contrast to the White Nile that flowed through a land of 'malaria, marshes, mosquitoes, misery' but had a far more equable flow regime than the Blue Nile. The distinction is an important one. Owing to its highly seasonal flow and vigorous bank erosion, the alluvial record of the lower Blue Nile is fragmentary in the extreme, with only the more recent sediments well preserved, while that of the lower White Nile is remarkably complete over the last 250 kyr, despite very low rates of sedimentation. It is instructive to compare the Late Quaternary sedimentary record preserved in the upper few metres of the lower Blue and White Nile valleys, because they provide a useful record of environmental events in their respective headwaters.

### (*b*) *Cainozoic evolution of the Nile*

Nile history closely reflects the influence of tectonic, volcanic and climatic events in its Ethiopian and Ugandan headwaters. The hydrological differences between the Blue and White Nile rivers reflect their very different geological and geomorphic histories. The Blue Nile gorge is one of the most spectacular features in the Nile basin, and post-dates the Ethiopian flood basalts that were erupted within the space of a million years some $3 \times 10^7$ years ago [63,64]. The volume of rock eroded from the Abbai and Tekezze basins since then amounts to $100\,000 \pm 50\,000$ $km^3$ from an area of 275 000 $km^2$, which is comparable to the volume of the Nile cone in the eastern Mediterranean, estimated at $150\,000 \pm 50\,000$ $km^3$ [5]. The concordance between these two independent estimates is consistent with an Ethiopian source for the bulk of the Nile cone sediment. The major drainage

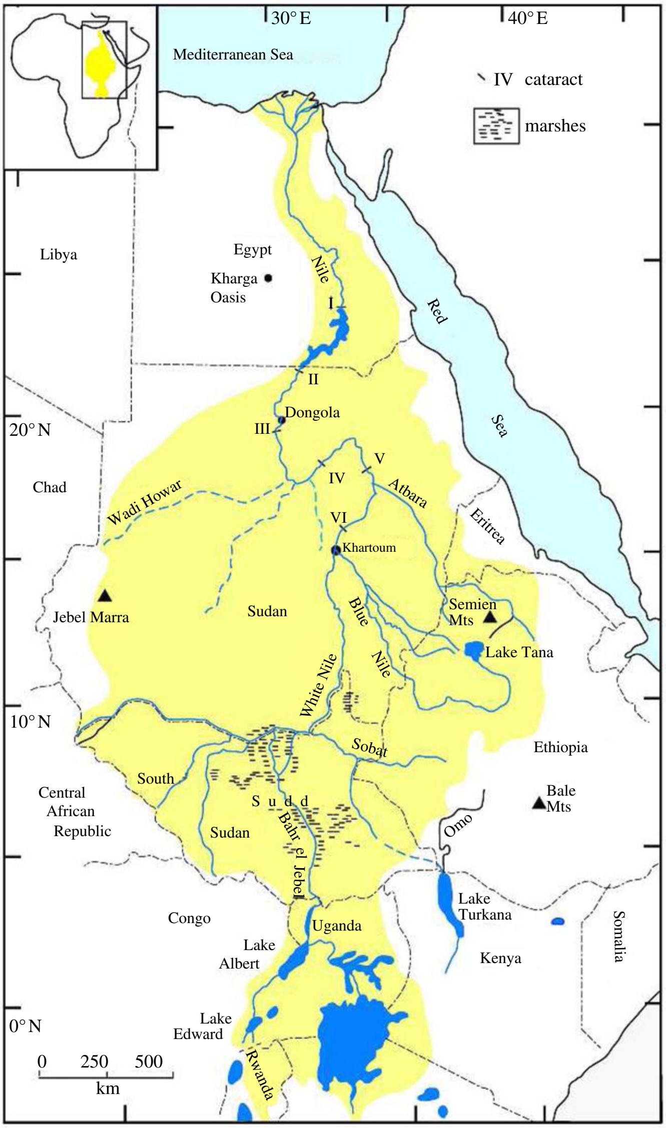


Figure 4. Nile basin. (Online version in colour.)

divides date back to 2–3 × $10^7$ years ago and pre-date the rifting and break-up of the original Ethiopian volcanic plateau, which did not begin until after 2 × $10^7$ years ago [63,64]. Uplift of the Ethiopian plateau was in three stages (29–10, 10–6 and 6–0 × $10^6$ years ago) with long-term erosion rates accelerating at approximately 10 and 6 × $10^6$ years ago [65,66].

Climatic cooling and progressive desiccation in the Ethiopian highlands at approximately 2.5 Ma [67,68] ushered in an era of glacial–interglacial cycles characterized by cold, dry conditions during glacial maxima and warm wet conditions during interglacial phases, when the summer monsoon was stronger than today. During the last glacial maximum (LGM) at $21 \pm 2$ kyr ago, the Semien Highlands (figure 4) were glaciated down to 4200 m, the lower limit of periglacial solifluction was 1000 m lower (3100 m), and temperatures were 4–8°C colder [69]. Lake Tana (figure 4) became a closed basin until 17–15 kyr ago [70]. The rivers also became more seasonal and carried sands and gravels to the Nile until 17–15 kyr ago, when they deposited silt and clay across their floodplain [61,71]. Deforestation in the headwaters over the past 100 years has increased erosion by an order of magnitude, leading to widespread silting up of reservoirs downstream [72]. The discussion that follows amplifies these statements. We begin with the Blue Nile.

### (*c*) *Blue Nile fining-upwards alluvial sequence*

Fifty kilometres south of Khartoum, a series of gravel pits have been excavated in recent years close to the present Blue Nile just south of Masoudia (figure 5) and afford a striking insight into the alluvial history of that river. The uppermost layer of sediment consists of a black cracking clay with a discontinuous band of Nile oyster shells (*Etheria elliptica*) near the base of the clay layer, above a thin band of rolled carbonate gravel (figure 5*a*). Such cracking clays (or *vertisols*) form a continuous mantle, 1–2 m thick, across the surface of the Gezira alluvial plain [73–75]. The Gezira is a low-angle alluvial fan built up during the Late Cainozoic by the Blue Nile and its tributaries. It forms a triangular plain bounded to the south by the Manaqil Ridge, to the east by the Blue Nile, to the west by the White Nile, with its apex at Khartoum, where these two great Nile tributaries meet (figure 5).

Exposed beneath the surface vertisol in the gravel quarries, there are two alluvial units, each several metres thick, composed of rolled quartz and ironstone gravel in a sandy clay matrix, with large cross-beds indicative of high-energy flow. Both gravel units contain up to 25 per cent of irregular and linear calcium carbonate concretions 10–30 cm long, reflecting replacement of former tree roots by carbonate and indicating prolonged intervals of soil development accompanied by carbonate precipitation following deposition of the river gravel units (figure 5*a*). The overall sedimentary sequence shows two phases of high-energy Blue Nile flow, each followed by a shift in the channel and prolonged soil formation under semi-arid conditions. The final phase of widespread clay deposition across the former Blue Nile floodplain indicates a major change in the type of sediment being carried by the river, followed by exposure of the alluvial clays and ensuing pedogenesis.

### (*d*) *Gezira alluvial fan*

The youngest portion of the Gezira alluvial fan (figures 5 and 6) consists of a veneer of dark clay (generally, 1–2 m thick) that mantles alluvial sands and gravels with very large cross-beds indicative of very high-energy flow. The clays contain freshwater mollusc shells with maximum calibrated radiocarbon ages of 15 kyr near the base of the clays [61]. (Since the strength of the Earth's magnetic field has varied over time, with concomitant changes in the cosmic ray flux in the atmosphere and hence the conversion of $^{14}N$ to $^{14}C$, radiocarbon years are not calendar years and so need to be calibrated against other less variable chronologies, such as that devised by counting tree rings, or the marine coral

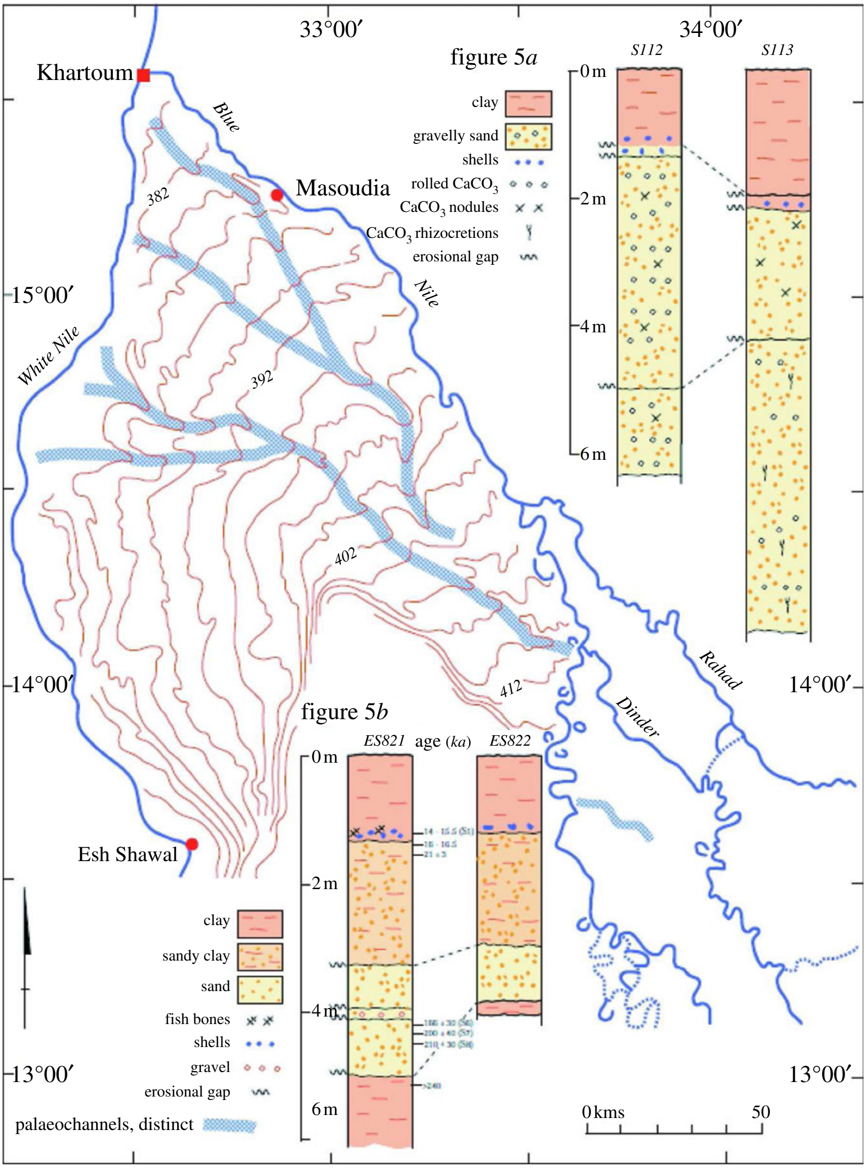


Figure 5. Gezira alluvial fan. Stratigraphic logs show (*a*) Pleistocene Blue Nile gravels and Holocene clay mantle, and (*b*) Pleistocene White Nile clays, sandy clays and sands and Holocene clays. (Online version in colour.)

record dated by very high-resolution uranium series measurements.) The aquatic gastropod species (*Melanoides tuberculata*, *Biomphalaria pfeifferi*, *Bulinus truncatus*, *Corbicula fluminensis* and, appropriately, *Cleopatra bulimoides*) become less frequent towards the surface, and are replaced progressively by semi-aquatic snails such as *Pila wernei* and *Lanistes carinatus*, until finally replaced towards 5 kyr ago by the land snail *Limicolaria flammata*. This snail is common today in the acacia tall grass savannah region of south-central Sudan with a

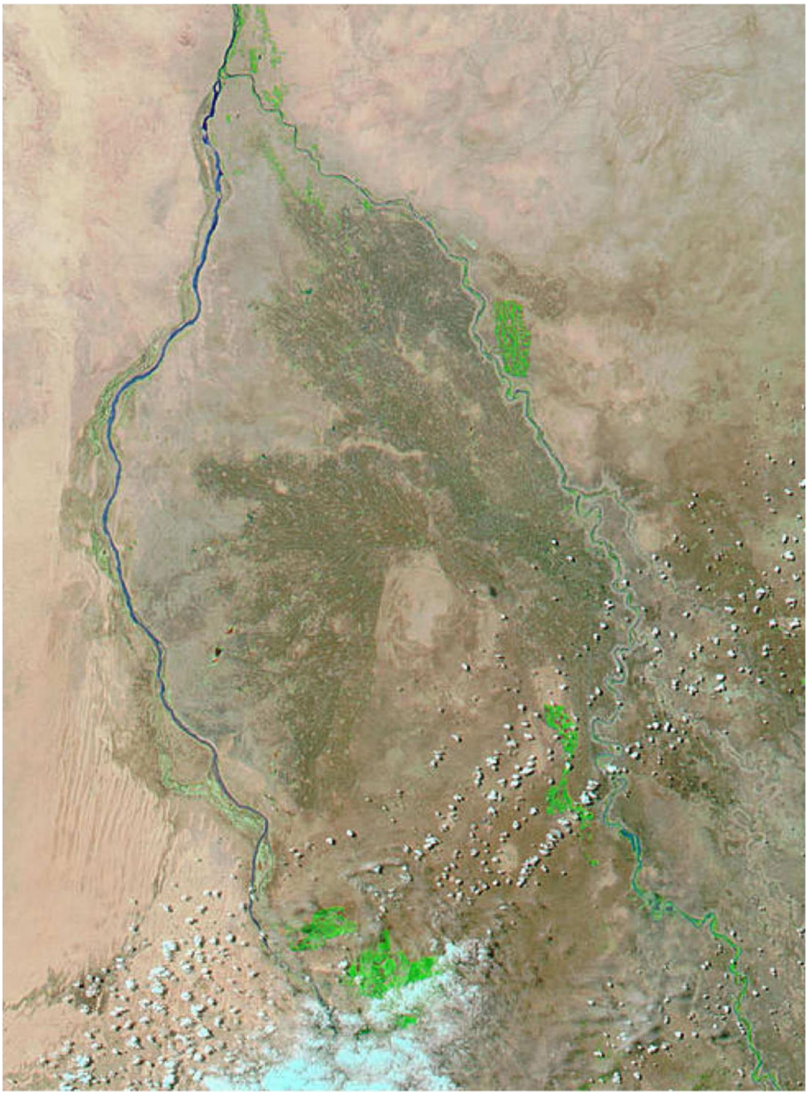

Figure 6. Gezira alluvial fan. Aqua satellite image (19 June 2003, at the start of the rainy season) showing clay-mantled irrigated land traversed by sandy palaeochannels. (Image courtesy of Jacques Descloitres, MODIS Rapid Response Team at NASA, GSFC). (Online version in colour.)

minimum rainfall of 500 mm per annum. The changes in snail species through time indicate prolonged flooding along the Blue and White Nile rivers, followed by a much more seasonal flood regime, culminating in drying out of the flood plains some 5 kyr ago, as a result of either progressive Blue Nile incision or climatic desiccation or both, for both of which there is good independent evidence [61,71,76]. The sands and gravels beneath the clay surface mantle are less well dated but belong to the Late Pleistocene, with radiocarbon and luminescence ages from at least 40 to 17 kyr ago [61,77].

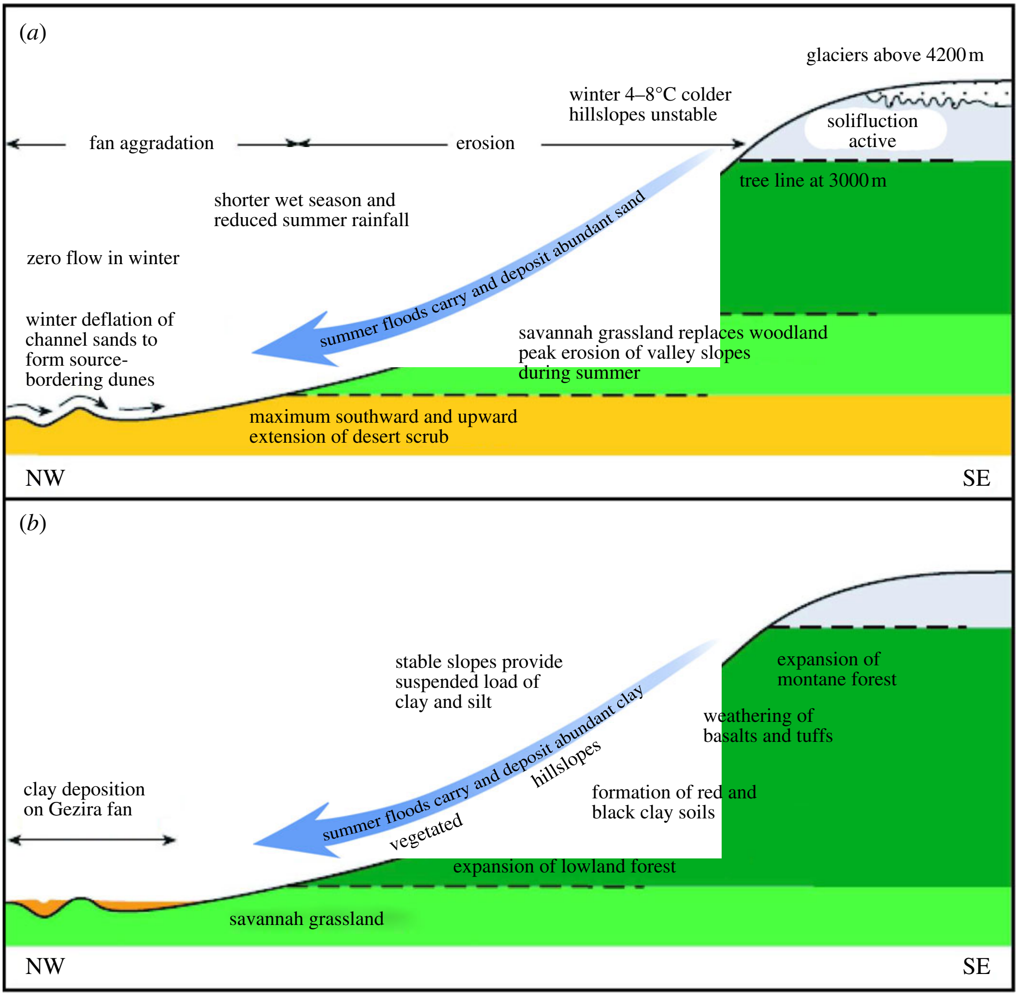


Figure 7. Depositional models for the Blue Nile: (*a*) during the last glacial maximum and (*b*) during the Early Holocene. (Online version in colour.)

## (*e*) *Depositional models for the Late Pleistocene and Holocene Blue Nile*

Thirty years earlier, when far fewer ages were available for this region, Adamson *et al.* [78] and Williams & Adamson [79] proposed a simple depositional model linked to climate to account for these changes. During cold, dry glacial intervals, the headwaters of major Ethiopian rivers would be sparsely vegetated, hill slope erosion would be accelerated and rivers would become highly seasonal, low-sinuosity, bed-load streams which carried and deposited large volumes of poorly sorted gravels and sands (figure 7*a*). Conversely, with a return to warm, wet conditions and re-establishment of a dense plant cover in the headwaters, we should see a change to high-sinuosity, suspended-load streams that carried and deposited silts and clays (figure 7*b*). A fining-upwards alluvial sequence [80] from coarse basal gravels through sands to horizontally bedded silts and clays is thus a predictable outcome of a change from a bed-load to a suspended-load regime, related to a change from glacial aridity to interglacial and postglacial climatic amelioration. The precise timing of the last glaciation in the Ethiopian Highlands is still being investigated. Osmaston *et al.* [81] considered that up to 180 km$^2$ of the Bale Mountains of Ethiopia could have been glaciated at this time, with a central icecap of at least 30 km$^2$. Glacial moraines and periglacial deposits in the Semien Mountains near the sources of the Tekezze and Blue Nile/Abbai rivers (figure 4) are presently being dated using cosmogenic nuclides. The few available radiocarbon ages point to colder LGM conditions (4–8°C cooler), with a lowering of the upper tree line by approximately 1000 m during the LGM [69,82].

### (*f*) *Glacial aridity in the Blue Nile basin*

It can be argued that this depositional model is based upon an unproven assumption, namely glacial aridity. After all, other workers had used the evidence afforded by Late Pleistocene Nile sands and gravels in northern Sudan and southern Egypt to argue for greater fluvial competence and so higher discharge and more pluvial glacial conditions [83]. The inference by Adamson *et al.* [78] that the Late Pleistocene was a time of greater aridity was based on the fact that, during the last glacial, maximum lake levels were low in Ethiopia, Kenya and Uganda [84]. In addition, Saharan desert dunes were active up to 800 km south of their present limits, to within at least 12° N of the equator [85]. In addition, there are strong theoretical reasons why glacial aridity prevailed in the intertropical zone, not least being a cooler sea surface and reduced evaporation from the oceans, thereby curtailing the supply of moist air derived from tropical convection.

Later work has confirmed that much of equatorial Africa was drier and colder than today during the LGM [86,87], with Lakes Victoria, Albert and Edward no longer flowing into the upper White Nile [88]. Two Late Pleistocene cores from Lake Albert contain palaeosols dated to 20.7–17.7 kyr ago and 16.6–15.1 kyr ago, indicating lake desiccation at those times, with a high lake level before 20.7 kyr ago and after 15.1 kyr ago evident in the $^{87}Sr/^{86}Sr$ values [89]. The older palaeosol coincides with the LGM.

The White Nile, deprived of the run-off from its headwaters by the closure of the Ugandan lakes, dried out in the winter months, during which sand dunes migrated across its former bed [61,71]. To sum up, the Late Pleistocene Blue Nile and Atbara rivers were highly seasonal bed-load streams that, together with their tributaries, ferried and deposited vast quantities of poorly sorted sands and gravels in central Sudan and southern Egypt [77]. With the return of the summer monsoon towards 17 kyr ago, strengthening at 15 kyr ago [89], run-off increased in the Ethiopian headwaters, and Lake Tana overflowed once more [70]. From approximately 15 until 7.5 kyr ago and perhaps slightly later [61], the Holocene Blue Nile was depositing clays across the low-angle Gezira alluvial fan in the central Sudan. Thereafter, it began to incise, terminating its fining-upwards depositional cycle.

### (*g*) *Relevance of the Blue Nile depositional models to India and Australia*

A similar pattern of widespread deposition of Late Pleistocene sand and gravel, followed by terminal Pleistocene to Early Holocene fine-grained alluviation culminating in vertical river entrenchment has been documented for the Son and

Belan rivers in semi-arid north-central India [90,91] as well as in the sub-humid to semi-arid Murray and Murrumbidgee river basin in southeastern Australia [38–42]. It thus appears that rivers in semi-arid catchments are sensitive to changes in plant cover, whether once glaciated or not. A substantial reduction in vegetation cover in their headwaters is conducive to a bed-load regime, reverting to a suspension-load regime once the plant cover has been restored and a soil cover widely established in the headwaters. In essence, in the absence of any eustatic, isostatic or other tectonic causes of changes in base level, a river will tend to aggrade its valley when the load-to-discharge ratio is high and to degrade its valley when the load-to-discharge ratio is low. However, care is needed to avoid falling into the trap of circular argument in which a given type of climate (wetter, drier, transitional from wet to dry or dry to wet) is inferred from the presence of a river terrace, and the inferred climate is then used to account for the existence of the same terrace. Some independent check upon the purely fluvial evidence is therefore necessary when seeking to reconstruct hydrological and climatic changes in river basins [92].

### (*h*) *Strontium isotope evidence for Late Pleistocene integrated Nile drainage network*

Careful studies of the lake sediments in Uganda and Ethiopia and the alluvial sediments along the lower Blue and White Nile valleys demonstrated that, after a long dry interval during the LGM, rainfall increased and the Ugandan lakes overflowed once more into the upper White Nile in southern Sudan, causing widespread flooding in the lower White Nile valley, from approximately 14.5 kyr ago onwards. However, Beuning *et al.* [93] put forward a contrary view, averring that Lake Victoria remained a closed basin until approximately 7.2 kyr ago. This conclusion ran counter to three decades of research by many workers and, if correct, would have required a complete re-thinking of Nile prehistoric archaeology.

In order to test this hypothesis and ascertain more precisely the time at which Lake Victoria overflowed, Talbot *et al.* [94] used strontium isotopes as tracers to determine just when Lakes Victoria and Albert overflowed into the upper White Nile. They analysed the strontium isotope ratio ($^{87}Sr/^{86}Sr$) preserved in freshwater gastropod shells from Blue and White Nile sediments ranging in age from terminal Pleistocene to present day, and compared these with the strontium isotope ratios obtained from the Ugandan lakes. Since these ratios are not changed by weathering and hydrological cycles, the strontium ratios of river and lake waters give a weighted average for the type of rocks within the various basins making up the overall river system. All of the shells analysed had been tested for carbonate re-crystallization using X-ray diffraction. This work demonstrated the overflow from Lake Victoria by approximately 14.5 kyr ago and provided an independent confirmation that the present-day integrated Nile drainage network became re-established at that time, a conclusion vindicated by later studies [89].

The abrupt return of the summer monsoon was also reflected in wet conditions along the southern margins of the Sahara, with dune stabilization [95], high lake levels in the mountains of Tibesti and Jebel Marra [61,96,97] and a sudden reduction in dust flux from the Chad basin into the Atlantic [98]. The approximately 14.5 kyr ago return (or strengthening) of the summer monsoon evident in the Nile basin was a global event, and has been identified elsewhere in Africa [87] as well as in India, China and Australasia [99,100].

### (i) *White Nile and main Nile floods and Mediterranean sapropels*

During phases of very high Nile flow, clastic muds rich in continental organic matter and highly organic sapropels accumulated on the floor of the eastern Mediterranean [101–107]. Flood deposits exposed in trenches dug east of the present White Nile near Esh Shawal village 250 km south of Khartoum [108] (figure 4) show episodes of middle to Late Pleistocene high flow (figure 5*b*), which, within the limits of the dating errors, coincide with sapropel units S8 (217 kyr ago), S7 (195 kyr ago) and S6 (172 kyr ago) [104]. Sapropel 5 (124 kyr ago) was synchronous with major flooding in the White Nile valley and with a prolonged wet phase at approximately 125 kyr ago at Kharga Oasis in the Western Desert of Egypt. Recently dated high flood deposits on the main Nile are roughly coeval with sapropel units S6 (172 kyr ago) and S3 (81 kyr ago) [77]. There are as yet no well-dated Nile sediments synchronous with sapropel unit 2 (55 kyr ago).

Thanks to a very gentle flood gradient (1:100 000), the post-LGM flood deposits in the lower White Nile valley are well preserved. Calibrated $^{14}$C ages obtained on freshwater gastropod and amphibious *Pila* shells and fish bones show high White Nile flood levels around 14.7–13.1, 9.7–9.0, 7.9–7.6, 6.3 and 3.2–2.8 kyr ago. The less complete Blue Nile record shows very high flood levels towards 13.9–13.2, 8.6, 7.7 and 6.3 kyr ago [61]. The Blue Nile has cut down at least 10 m since approximately 15 kyr ago and at least 4 m since 9 kyr ago, with concomitant incision by the White Nile amounting to 4 m since approximately 15 kyr ago and at least 2 m since 9 kyr ago. Such incision would help in draining previously swampy flood plains, freeing them for cultivation.

The most recent sapropel, S1, in the eastern Mediterranean is a composite unit, with ages of 13.7–12.4 kyr near the base and 9.9–8.9 kyr near the top [101–103]. The gap in the S1 record may coincide with the arid phase seen in other parts of Africa coeval with the Younger Dryas (approx. 12.5–11.5 kyr ago) [98]. Higgs *et al.* [109] considered that formation of sapropel S1 may have ended as recently as 5 kyr ago, which is also when the Nile deep-sea turbidite system became inactive as a result of reduced sediment discharge from that river [107]. The interval from approximately 13.7 to 8.9 kyr ago and locally up to 5 kyr ago also coincides with a time when freshwater lakes were widespread across the entire Sahara and when the White Nile attained flood levels up to 3 m above its modern unregulated flood level.

Where independently dated comparisons exist between sapropel formation and Nile floods, they point to synchronism between sapropel accumulation and times of higher Nile flow, indicative of a stronger summer monsoon at these times. Although the sapropel record in the eastern Mediterranean is incomplete, with some evidence of complete removal of sapropels by post-depositional oxidation [109], it is a longer and more complete record than that presently available on land, and so can serve as a useful surrogate record for Nile floods and phases of enhanced summer monsoon precipitation.

## 7. Nile floods, orbital and solar irradiance cycles, and El Niño Southern Oscillation events

The 'African Humid Period' (14.8–5.5 kyr ago) [98] was a time of wetter climate in the Blue and White Nile headwaters. This wet interval was linked to changes in the tilt of the Earth's axis, such that the Earth was closest to the sun during the northern summer, leading to a 7 per cent increase in summer (June–July–August) insolation and a corresponding 7 per cent decrease in winter (December–January–February) insolation relative to the present [110,111]. This asymmetrical heating of the Earth's surface increased temperature and pressure gradients on either side of the equator, strengthening the monsoon circulation and bringing more summer rain to tropical northern Africa, including the now arid southern and central Sahara [112–114]. The wet phase seems to have begun quite suddenly but to have ended gradually, with lakes drying out first in higher latitudes and later in lower latitudes, although this may simply indicate an initially slow response to the northward shift of the Intertropical Convergence Zone and a time-transgressive response to its final southward displacement.

An abiding puzzle is how slow changes in insolation can trigger abrupt climatic responses. A possible reason is that, once insolation levels had passed a certain threshold value, certain nonlinear 'biogeophysical' feedback processes start to operate [98]. Such factors include changes in sea surface temperatures, surface cover, albedo and sensible heat flux [115–121]. Such feedbacks would have accentuated both Early Holocene humidity and the Mid-Holocene desiccation of the Sahara and Nile valley evident in the strontium isotope records from the Nile delta and the demise of the Old Kingdom dynasty in Egypt caused by the severe drought at 4.2 kyr ago [122,123].

The intervals of high White Nile flow listed in §6*i*, if correct, seem to occur at millennial-scale frequency. Bond *et al.* [124] found that Holocene episodes of ice-rafted debris (IRD) into the North Atlantic occurred at intervals of approximately 1500 years. Comparison of $^{14}C$ and $^{10}Be$ concentrations (in large part controlled by galactic cosmic radiation modulated by the strength of the solar wind's magnetic field) in tree rings and Greenland ice cores suggested a similar periodicity, prompting some workers to hypothesize that cyclical fluctuations in solar irradiance might have been responsible for at least some of the cyclical fluctuations discernible in Holocene marine, ice core and speleothem climate proxy records [125–127]. Bard and Franck [128] wisely advise caution in this regard. A further suggestion that fluctuations in solar irradiance may have modulated the frequency of El Niño Southern Oscillation (ENSO) events at millennial to centennial scales remains speculative [129,130] but prompts us to consider the links between historic Nile floods and ENSO events [131–133]. In essence, times of strongly negative Southern Oscillation Index (SOI)—i.e. El Niño events—were almost always synchronous with very low flow in the Nile, as well as with years of drought in northeast China, peninsular India, Java and southeast Australia [131–133]. Conversely, times of strongly positive SOI (i.e. La Niña events) were usually years of exceptional floods in those same regions. By way of example, rivers were in spate across eastern Australia during the 2010 La Niña. More sombrely, during the great El Niño drought of 1877, six million people died in India and some 10 million in China [131]. Since there is substantial overlap between the domain of the monsoon and that of ENSO [90], years of weak monsoon will exacerbate the effect of El Niño-induced droughts, and conversely during La Niña events.

## 8. River sediments and river basin management

Regulation of river flow by dam building has a 5000 year pedigree in Egypt, but the effects are not always beneficial, with reservoir sedimentation, salt accumulation in irrigated soils, and downstream erosion as some of the more obvious consequences. For example, by 1996, the capacity of the Roseires reservoir (figure 4) on the Blue Nile had been reduced by almost 60 per cent through silt accumulation and that of the Khashm el Girba reservoir on the Atbara (figure 4) by nearly 40 per cent [72]. Forest clearance in the mountainous Ethiopian headwaters for cultivation is a prime cause of such rapid rates of sedimentation. Such clearance alters the hydrological balance through decreased infiltration and increased run-off. Hurni [6] recorded a decrease in the area under natural forest in the upper Blue Nile drainage basin from 27 per cent to 0.3 per cent between 1957 and 1995, with a corresponding increase from 40 to 77 per cent in the area cultivated. Annual soil loss amounted to $2\,\mathrm{mm\,yr^{-1}}$ on mountain slopes in this region, increasing to over $15\,\mathrm{mm\,yr^{-1}}$ in cultivation years. However, deforestation has varied over the last few thousand years in Ethiopia, with intervals of forest regrowth alternating with periods of forest removal [7].

More insidious is the slow build-up of salt within agricultural soils as a result of poor canal maintenance (Uzbekhistan), inadequate soil drainage (Egypt and Pakistan) and clearing of native vegetation (Australia). Another cause is geological inheritance. The high levels of subsoil salinity in Quaternary alluvial deposits flanking the lower White Nile have nothing to do with the present-day climate and cannot be understood without a detailed knowledge of the complex depositional history of that river [61,71,108,134]. This is doubtless equally true of other big rivers flowing through semi-arid regions, such as the Indus and the Tigris–Euphrates.

A final example will serve to illustrate the sometimes counterintuitive influence of river sediments on land use. The Jonglei Canal project aims to drain part of the Sudd swamps of South Sudan and so increase discharge downstream. The slogan ‘more water for the North, more land for the South’ conceals some less obvious pitfalls. About $40\,\mathrm{km^3}$ of water enters the Sudd and half that amount flows out, the balance lost in seepage and evapo-transpiration. However, the *concentration* of dissolved solids in water leaving the Sudd is the same, but with an altered *composition*. Moreover, the salinity levels in sediments beneath the Sudd are high. In fact, the swamps operate as a gigantic biogeochemical filter, and also buffer the flow regime [135]. Drainage resulting from the canal will probably expose saline subsoils, alter water chemistry downstream and increase flow variability, leading to bank erosion downstream and damage to the inlet pipes of pump schemes along the lower White Nile. The original project planners did not consider these possible consequences. The details are site specific; the principles are universal. Land use that does not take due account of river history and sediment characteristics is unlikely to endure.

Table 2. River responses to global and regional environmental changes discussed in this account (for sources, see text).

I. 250–25 kyr ago (interval includes two glacial–interglacial cycles, each approx. 100 kyr long)

*A. Nile basin, Sahara and Mediterranean Sea*

(1) high White Nile floods at more than 240 kyr ago, 210 ± 30 kyr ago, 200 ± 40 kyr ago, 166 ± 30 kyr ago and approximately 125 kyr ago coeval with intervals of stronger Northern Hemisphere summer monsoon
(2) intertropical convergence zone (ITCZ) displaced at least 500 km further north during the Northern Hemisphere summer
(3) prolonged wet phase at approximately 125 kyr ago at Kharga Oasis in the western Desert of Egypt. Sahara studded in lakes
(4) sapropel accumulation in the east Mediterranean Sea at 217 kyr ago (S8), 195 kyr ago (S7), 172 kyr ago (S6), 124 kyr ago (S5), 102 kyr ago (S4), 81 kyr ago (S3) and 55 kyr ago (S2), broadly coincident with phases of very high Nile discharge

*B. Rest of world*

(1) high interglacial sea levels at approximately 210 kyr ago and approximately 125 kyr ago
(2) warmer interglacial sea surface temperatures, high rates of evaporation from the oceans, and wetter climates in the intertropical zone
(3) high lake levels in Africa and Australia at approximately 130–110 kyr ago, 100–80 kyr ago, 65 kyr ago and 40 kyr ago
(4) super-eruption of Toba volcano approximately 73 kyr ago followed by several decades of intense global cooling and several centuries of sustained reduction in precipitation in Asia and possibly Africa. Low lakes levels in tropical Africa at this time may be related to the climatic impact of this eruption

II. 25–17 kyr ago (interval includes the last glacial maximum (LGM: 21 ± 2 kyr)

*A. Nile basin, Sahara and Mediterranean Sea*

(1) lakes drying out in headwaters of Blue and White Nile rivers coeval with southward displacement of the ITCZ relative to today during the Northern Hemisphere summer and weakening of the summer monsoon. Soils form on the floor of Lake Albert during 20.7–17.7 kyr ago and 16.5–15.1 kyr ago
(2) White Nile deprived of overflow from Ugandan lakes and reduced to a seasonal trickle and its lower reaches blocked by desert dunes
(2) lakes drying across the Sahara and East Africa
(3) temperatures 4–8°C cooler than today in the Ethiopian Highlands, with local glaciation in the Semien and Bale Mountains and intense periglacial action and mass movement on mountain slopes. Upper treeline lowered approximately 1000 m. Bare unstable slopes supply abundant coarse debris to highly seasonal Blue Nile and Atbara rivers
(4) widespread deposition of coarse sand and fine gravel across the Gezira alluvial fan in central Sudan and along the main Nile valley in Sudan and Egypt. Main Nile probably dries out during winter months. Humans mostly abandon the Main Nile valley and migrate south in search of more reliable supplies of water

(*Continued.*)

Table 2. (*Continued.*)

*B. Rest of world*

(1) maximum global ice volume and LGM sea levels approximately 120 m lower than today
(2) cooler sea surface temperatures and reduced evaporation from the oceans, with drier conditions in the intertropical zone
(3) lakes low or dry in Africa and Australia during the LGM
(4) desert dunes active in Africa, Asia and Australia, and widespread deposition of desert dust/loess in China, India, South America, Australia and Europe

III. 17–5 kyr ago
*A. Nile basin, Sahara and Mediterranean Sea*

(1) lakes in Blue and White Nile source regions begin to rise at approximately 17 kyr ago and overflow perennially at 15–14.5 kyr ago
(2) stronger summer monsoon and ITCZ extends approximately 500 km further north than today during the Northern Hemisphere summer
(3) upper catchments of Blue and White Nile densely vegetated and soil formation active
(3) perennial channel flow re-established in Blue and White Nile and Main Nile, which now carry a large seasonal suspension load of silt and clay
(4) Blue Nile incised more than 10 m into its former floodplain since 15 kyr ago and more than 4 m since 9 kyr ago, beheading its Gezira distributary channels, which dry out by approximately 5 kyr ago
(5) high White Nile flood levels at *ca* 14.7–13.1, 9.7–9.0, 7.9–7.6, 6.3 and 3.2–2.8 kyr ago. High Blue Nile flood levels at *ca* 13.9–13.2, 8.6, 7.7 and 6.3 kyr ago
(6) the Sahara is once more studded in sporadic lakes and supports a human population of Mesolithic hunter–fisher–gatherers and later Neolithic pastoralists
(7) a composite sapropel (S1) accumulates in the eastern Mediterranean Sea between approximately 13.7–12.4 kyr ago near the base and approximately 9.9–8.9 kyr ago near the top. The gap between the two sapropel sub-units may denote the influence of the Younger Dryas (YD) episode (approx. 12.5–11.5 kyr ago), which was marked by aridity in Lake Victoria and in the Sahara. Flow in the Nile was probably curtailed and more seasonal during the YD

*B. Rest of world*

(1) retreat of icecaps and mountain glaciers in both hemispheres and rapid sea-level rise from 17 kyr ago on
(2) sea surface temperatures higher once more, with greater evaporation from the oceans and wetter climates in the intertropical zone
(3) stronger summer monsoon in both hemispheres
(4) high postglacial lake levels in Africa, Asia and Australia
(5) desert dunes vegetated and stable along desert margins, and supply of desert dust/loess much reduced
(6) rivers of central India cut down approximately 30 m into their former floodplains at or before approximately 8 kyr ago
(7) ice-rafted debris (IRD) events in N Atlantic at 11.0, 10.3, 9.4, 8.1 and 5.9 kyr ago
(8) Donggwe Cave speleothem record (South China) shows strong Asia Monsoon (AM) at 9–7 kyr ago and weak AM events at 8.3, 7.2, 6.3 and 5.5 kyr ago
(9) Oman speleothems show reduced rainfall at 9.5, 9.0, 8.3, 7.4 and 6.3 kyr ago

(*Continued.*)

Table 2. (*Continued.*)

IV. 5–0 kyr ago

*A. Nile basin, Sahara and Mediterranean Sea*

(1) lakes in Blue and White Nile source regions continue to overflow
(2) weaker summer monsoon and ITCZ retreats approximately 500 km to the south during the Northern Hemisphere summer
(3) more seasonal flow in Blue and White Nile rivers and Main Nile. Blue Nile carries a mixed load of sand, silt and clay
(4) lakes dry out in the Sahara, which becomes abandoned by Neolithic pastoralists who move south into West Africa or east into the Nile valley
(5) deposition of sapropel S1 in the east Mediterranean Sea may have persisted until approximately 5 kyr ago, when the Nile deep-sea turbidite system also becomes inactive

*B. Rest of world*

(1) climates in the intertropical zone become less humid and more seasonal
(2) summer monsoon still strong but less vigorous than in the previous phase and its spatial domain probably somewhat reduced
(3) lakes dry out in arid and semi-arid areas
(4) sporadic glacier advances culminating in the most recent Little Ice Age
(5) increase in the frequency of El Niño Southern Oscillation (ENSO) events, leading to more frequent extreme flood and drought events in the Americas, Africa, Australia and Asia. Interaction between ENSO and summer monsoon leads to a more variable rainfall regime in both hemispheres in regions influenced by them, and to highly seasonal or very variable river flow regimes
(6) impact of human activities upon global climate (and on river basins) may have begun with the advent of agriculture and has increased ever since, not always to the advantage of either
(7) IRD events in North Atlantic at 4.2, 2.8 and 1.4 kyr ago
(8) Donggwe Cave speleothem record (South China) shows weak AM events at 4.5, 2.7, 1.6 and 0.5 kyr ago

## 9. Conclusions

River history is reflected in the nature of the sediments carried and deposited over time, allowing reconstruction of past environmental changes at a variety of scales in time and space, although a patchy alluvial record and influences external to the river basin can make interpretation difficult. The continental record of Late Quaternary Nile floods is consistent with the well-dated record of highly organic sediments (sapropels) in the eastern Mediterranean Sea. Times of high Nile flow accord with global fluctuations in summer monsoon strength (table 2), perhaps modulated by changes in solar insolation caused by changes in the Earth's orbit and by variations in solar irradiance. Understanding the alluvial record can assist catchment management and help avoid environmental damage. As Bacon [136] noted in 1620: *Natura non nisi parendo vincitur* (to command Nature, we must first obey her laws).

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