Journal of Palaeosciences 70(2021): 267–288
0031–0174/2021

A river flowing through a desert: late Quaternary environments in the Nile Basin–current understanding and unresolved questions

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ABSTRACT

Williams MAJ 2021. A river flowing through a desert: late Quaternary environments in the Nile Basin–current understanding and unresolved questions. Journal of Palaeosciences 70(2021): 267–288.

Late Quaternary environments in the Nile Basin reflect the influence of the African summer monsoon upon plant cover, sediment yield and flood discharge in the Ethiopian and Ugandan headwaters of the Nile. Intervals of prolonged and very high Nile flow coincide with times of stronger summer monsoon and have been dated using a combination of ^{14}C, OSL and ^{10}Be methods. Periods of high Nile flow into the eastern Mediterranean coincide with the formation of highly organic sedimentary layers termed sapropels. Ages obtained so far for these times of sustained middle to late Pleistocene high flow in the Blue and White Nile are broadly coeval with sapropel beds S8 (ca 217 ka), S7 (ca 195 ka), S6 (ca 172 ka), S5 (ca 124 ka), S4 (ca 102 ka) S3 (ca 81 ka), S2 (ca 55–50 ka) and S1 (10–6.5 ka). Sapropel 5 (ca 124 ka) was synchronous with extreme Blue Nile floods and the formation of the 386 m lake in the lower White Nile Valley, as well as with a prolonged wet phase in the eastern Sahara. Fluctuations in Nile flow and sapropel formation reflect the influence of the precessional cycle upon the East African monsoon.

Between 75 ka and 19 ka the climate in the Nile headwaters region became progressively colder and drier. During the Last Glacial Maximum, Lake Tana in Ethiopia and Lake Victoria in Uganda became dry, flow in the White Nile was reduced to a trickle, and the Blue Nile and Atbara became highly seasonal bed–load rivers. The return of the summer monsoon at 14.5 ka ushered in extreme Blue Nile floods, widespread flooding across the Nile Basin and the formation of the 382 m lake in the lower White Nile Valley. There was a brief return to aridity during the Younger Dryas (12.8–11.5 ka), after which the climate again became wetter and widespread flooding in the Nile Valley resumed. The early Holocene floods were later followed by incision and creation of the modern relatively narrow flood plain.

Key–words—Nile, Floods, Sapropels, Blue Nile, White Nile, Atbara.

INTRODUCTION

IN his 1976 novella *A River Runs through it*, written when he was 73, Norman Maclean describes his boyhood experience of growing up close to a river in the backwoods of Missoula, Montana, and of the indelible memories it left him throughout his life (Maclean, 1976). The Egyptian people who lived alongside the Nile during dynastic and pre–dynastic times have also recorded their appreciation of the river to which they owed their very existence–but in the form of monuments and hieroglyphs (and the occasional recipe for making home–brewed beer) (Butzer, 1976; Clark, 1980; James, 2005; Wengrow, 2006; Romer, 2007; Welsby and Phillipson, 2008; Snape, 2014). The very word 'Nile' is a corruption of the original Egyptian word, which simply meant 'the river' (Griffiths, 1966). In a land flanked to east and west by desert, there was but one river. If the summer floods were too high, lives were lost; if the flood level was too low, famine stalked the land. Understanding the seasonal behaviour of the river was vital to survival and led to the construction of a variety of river gauges, or 'nilometers', over the last few thousand years. One of the most impressive nilometers still in existence is the Rhoda nilometer. It was built in 861AD on Rhoda Island, now in the heart of Cairo, in order to provide accurate information about variations in flood level from year to year.

Unlike the river that flows through Missoula, which is not much older than Holocene, the Nile is a venerable river, and has been flowing from its Ethiopian headwaters to the eastern Mediterranean for at least thirty million years (30 Ma) (Fielding *et al.*, 2016, 2018). In contrast to earlier views claiming that the Nile was no older than Pleistocene (Butzer & Hansen, 1968; Hassan, 1976), Laura Fielding and her colleagues have demonstrated that mineralogical, geochemical, and geochronological data offer strong support for a connection between the Nile Cone and the volcanic uplands of Ethiopia that extends back to at least 30 Ma (Fielding *et al.*, 2016, 2018). The term 'Nile Cone' includes the submerged sediments deposited by the Nile in the eastern Mediterranean (Bartolini *et al.*, 1975; McDougall *et al.*, 1975; Williams & Williams, 1980; Said, 1981, Macgregor, 2012) and refers to what later authors refer to as the 'Nile Deep Sea Fan' (e.g., Mologni *et al.*, 2020).

During that time the African tectonic plate has moved northwards by up to 1500 km, at an annual rate of about 5 cm/year (Habicht, 1979; Smith *et al.*, 1981; Owen, 1983; Williams, 2016). One consequence of this northward migration for North Africa was a shift from hot wet equatorial latitudes to tropical latitudes characterised by dry descending air masses and a prolonged dry season (Maley, 1980; Williams, 2019, p. 26). Global cooling and desiccation between 8 and 7 Ma (Herbert *et al.*, 2016) saw the inception of the Sahara as a desert, but one initially less arid than today. Drying out of the vast Tethys Sea during the Miocene deprived North Africa of a major source of moisture and accentuated aridity along the northern margins of Africa (Zhang *et al.*, 2014). The Sahara is thus far older than the earliest humans. However, the Saharan climate has fluctuated between hyper–arid and semi–arid. During wetter climatic intervals the desert has been sufficiently vegetated to provide adequate plant and animal resources for human occupation (Haynes *et al.*, 1989; Kröpelin, 1993; Kuper & Kröpelin, 2006; Bubenzer & Riemer, 2007). During drier times prehistoric human groups moved out of the desert to the Nile Valley or closer to the coastal regions of North Africa (Williams, 2019, pp. 322–333). To understand how humans in northeast Africa have reacted to environmental change, we need to consider the history of the Nile and that of the desert through which it flows. In this contribution I will only focus on the Nile and leave a comprehensive discussion of the Sahara for a later publication.

The aim of this paper is to show that we need to adopt the source–to–sink approach espoused by Woodward *et al.* (2015a) and Williams (2019) if we are to make sense of the environmental history of the Nile. I will argue later in this paper that a single–minded focus on the admittedly impressive sedimentary record preserved in the Nile Cone (the subaerial Nile Delta contains a very much shorter and more fragmentary sedimentary record) can lead to misinterpretation. To make full sense of the distal record preserved in marine sediment cores from the eastern Mediterranean we need to understand the often quite subtle response of major Nile tributaries to climatic and environmental changes in their headwaters (Williams, 2019). My purpose is to justify this contention using examples drawn from the three major tributaries of the Nile–the White Nile, the Blue Nile and the Atbara (Fig. 1). The data discussed in this paper are all in the public domain, having been previously published, but the interpretation of the data reflects my own experience of mapping soils and sediments in the Nile Basin since 1962 and working with teams of archaeologists in Sudan, Ethiopia and Egypt since 1973 (see Williams, 2016, chapters 5, 8, 9 and 21).

REGIONAL SETTING OF THE NILE BASIN

The Nile is 6,853 km long, making it the longest river in the world. It extends across 35° of latitude from 3°S to 32°N. As a result, the Nile Basin is subject to a wide variety of climates ranging from tropical climates, characterised by seasonal monsoonal regimes in the south to semi–arid, arid, hyper–arid and Mediterranean climates in the north. The Nile Basin covers an area of 3.3 million km^2, third in area only to the Amazon (7.2 million km^2) and the Orinoco (3.8 million km^2). In spite of its vast catchment area, the Nile has a paltry sediment yield which amounted to 230 ± 20 Mt before river regulation and sediment trapping in reservoirs (Garzanti *et al.*, 2006, 2015; Woodward *et al.*, 2015b, Table 1). We therefore need to ask why the modern sediment yield is so low and whether this has always been the case.

According to the most recent estimates, the volume of the Nile Cone amounts to about 580,000 km^3 (Macgregor, 2012). This sediment came from erosion in the Nile Basin during the past 30 Ma. According to Macgregor (2012), who based his argument on a combination of Apatite Fission Track Analysis, planation surface analysis and Red Sea sink volumes to deduce that denudation in the Red Sea Hills amounted to at least 1300 m and a maximum of 2000 m (Macgregor, p. 420), a significant amount of Nile Cone sediment (220,000 km^3) came from erosion of the Red Sea Hills via rivers draining westwards into the main Nile. A further substantial amount of sediment in the Nile Cone (150,000 ± 50,000 km^3) came from erosion in the Ethiopian Highlands, notably from the upper Atbara and the upper Blue Nile (McDougall *et al.*, 1975). The Ethiopian headwaters of the Nile date back to the early Oligocene, some 30 Ma ago (McDougall *et al.*, 1975; Pik *et al.*, 2003, 2008; Gani *et al.*, 2007) and appear to have been contributing sediment to the main Nile for at least 30 Ma. Compared to the Blue Nile and Atbara, the White Nile is a relatively young river and is probably no older than about 0.5 Ma (Williams & Talbot, 2009). Obtaining a precise and accurate age for when the White Nile first flowed into the main Nile remains an important and as yet unresolved research question.

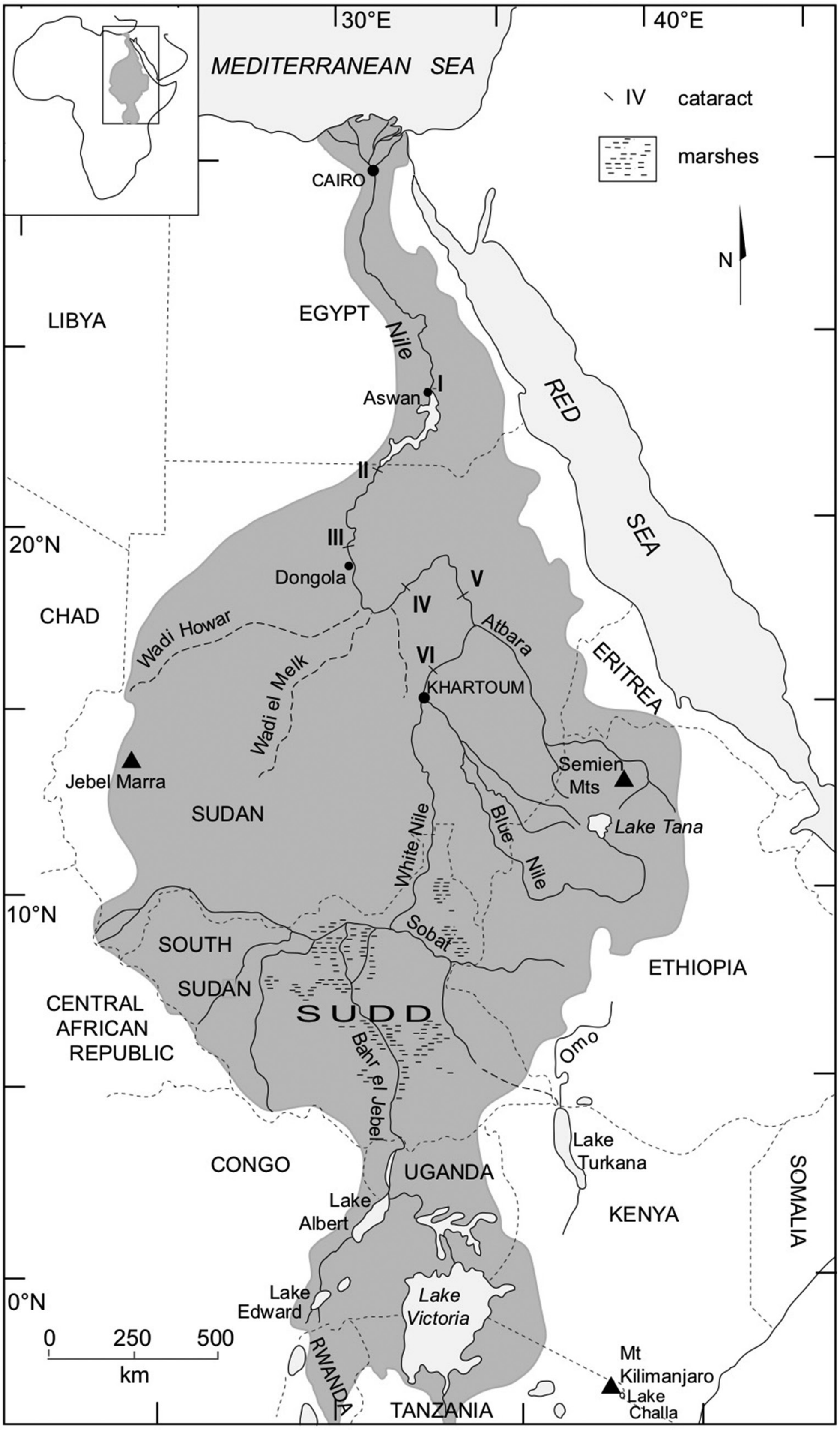

Fig. 1—The Nile Basin (after Williams, 2019, Fig. 1.1).

CENOZOIC EVOLUTION OF THE NILE BASIN

The Cenozoic evolution of the Nile Basin reflects the interaction of tectonic, volcanic and climatic events operating at a variety of different spatial and temporal scales (Talbot & Williams, 2009; Abdelsalam, 2018). Uplift of the Ethiopian–Arabian dome during the Oligocene triggered widespread regional erosion (Avni *et al.*, 2012) and culminated in extrusion of the Trap Series basalts in the Ethiopian Highlands in late Oligocene times, with much of the volcanism apparently taking place during a million or so years centred on 30 Ma (Hofman *et al.*, 1997). Uplift seems to have occurred in stages with long intervals of tectonic stability interspersed with a few phases of relatively rapid uplift (Minucci, 1938; Merla, 1963; Gani *et al.* 2007; Abdelsalam, 2018). The inception of the ancestral Atbara/Tekezze and Blue Nile/Abbai dates from about 30 Ma, subsequent to initial uplift and widespread extrusion of the highly fluid Trap Series basalts (Williams, 2016). According to Fielding *et al.* (2016, 2018), there has been a more or less continuous connection between rivers flowing from the Ethiopian Highlands and the eastern Mediterranean since that time.

The connection between the Ugandan headwaters of the White Nile and the main Nile (or Desert Nile) is far younger–probably no more than 0.5 Ma, which, on present evidence, seems to be about the maximum age for Lake Victoria (Fig. 1) on the equatorial Ugandan Lake Plateau (Williams & Talbot, 2009). Furthermore, the Ugandan lakes have had a chequered history, with both Lake Victoria and Lake Albert drying out at intervals during the late Quaternary, and most recently during the Last Glacial Maximum (Livingstone, 1980; Stager *et al.*, 1986; Johnson *et al.*, 1996; Stager & Johnson, 2008), when soils developed on the exposed beds of both lakes (Williams *et al.*, 2006). These events would have had serious consequences for flow along the main Nile, as discussed below.

NILE HYDROLOGY AND SEDIMENT SOURCES

Before large modern dams were built to regulate the flow of the Nile and its major tributaries, the unregulated White Nile used to provide a steady flow to the main Nile while the two main tributaries from the Ethiopian Highlands–the Blue Nile and the Atbara–provided a large but highly seasonal contribution (Hurst & Phillips, 1931; Hurst, 1950; Hurst, 1952). Of the peak monthly flow, the White Nile provided 10%, the Blue Nile 68% and the Atbara 22% (Hurst, 1950; Hurst, 1952; Williams *et al.*, 1982, p., 119, Table 7.1). Offsetting its meagre contribution to peak flow, the White Nile provided 83% of the low monthly flow (Williams *et al.*, 1982). The White Nile contribution to flow during the dry season was crucial to the survival of those dwelling alongside the main or Desert Nile in northern Sudan and Egypt, because the Blue Nile only furnished 17% of the low season flow while the Atbara dried up altogether during the winter months (Williams *et al.*, 1982). Several of the major tributaries to the Blue Nile in Sudan, such as the Dinder and the Rahad which rise in the western highlands of Ethiopia, also dried up during the long dry season, just as they do today (Hurst, 1952; Williams *et al.*, 1982).

Most of the sediment presently brought down from the Ethiopian Highlands via the Atbara and Blue Nile is trapped in the reservoirs downstream of such dams as the Aswan High Dam on the main Nile, the Sennar and Roseires dams on the Blue Nile–soon to be followed by the Grand Renaissance Dam in the Ethiopian Blue Nile or Abbai–and the Khashm el Girba dam on the Atbara (Woodward *et al.*, 2007, 2021). However, in historic times when the river was largely unregulated, the Blue Nile provided 72% of the total annual sediment load, the Atbara a further 24.5% and the White Nile a meagre 3.5% (Williams *et al.*, 1982). There are four main reasons why the historic White Nile sediment load is so trivial relative to that of the Atbara and Blue Nile. In Uganda the headwaters of the White Nile emerge from lakes and swamps, where locally derived sediment is trapped. Further downstream in South Sudan the vast Sudd swamps (Fig. 1) act as a gigantic biogeochemical filter trapping much of the sediment load (Williams & Adamson, 1973; Williams, 2019, pp.108–110). About half of the incoming water is lost through seepage and evapotranspiration in the Sudd, further reducing the sediment transporting capacity of the river. Finally, before completion of the Jebel Aulia dam south of Khartoum, water in the White Nile was ponded up by the Blue Nile floods (see section 3.1) and so any coarse sediment was quickly deposited along the Holocene flood plain of the White Nile.

At intervals during the Quaternary there have been two other important sources of sediment to the main Nile: desert dust and sediment brought in by local wadis flowing west from the Red Sea Hills and on a more local scale ephemeral streams draining the Nubian sandstone plateaux on both sides of the Nile (Butzer & Hansen, 1968; Butzer, 1980; Honegger & Williams, 2015; Woodward *et al.*, 2015b). Butzer and Hansen (1968) inferred that the wadis draining the Red Sea Hills in late Pleistocene times probably flowed during the winter–a view confirmed recently by uranium series ages obtained on tufa deposits in the northern Red Sea Hills (Hamdan & Brook, 2015).

Woodward *et al.* (2015b) have provided the first quantitative estimates of the comparative roles played by desert dust and local wadis in Holocene Nile sediment transport. They used strontium and neodymium isotope ratios of dated Holocene floodplain sediments to distinguish between Nile parent alluvium, local wadi deposits and Saharan desert dust. In northern Sudan up to about 40–50% of the sediment brought into the Nile came from actively flowing local wadis during the early to mid–Holocene (Woodward *et al.*, 2015b). Thereafter, from about 4.5 ka onwards, desert dust supplanted local wadis as the main secondary source of sediment to the Nile.

They found no evidence of any significant input of sediment from the Holocene White Nile in the floodplain sediments of the main or Desert Nile. One reason for this may be that the White Nile sediments were transported northwards during the low water season, remaining within the Nile channel and so not contributing to sediment accumulation along the Nile floodplain during the months of maximum flow from the Blue Nile and Atbara. This could explain why some authors have identified a significant input of sediments apparently derived from the White Nile within Holocene sediments of the Nile Delta (Foucault & Stanley, 1989) and further offshore (Blanchet *et al.*, 2013, 2015) when such White Nile sediments are invisible along the Desert Nile floodplain.

THE ENIGMATIC WHITE NILE MEGA–LAKES

The vast Sudd swamps (Fig. 1) have long been a subject of speculation, as has the very gentle flood gradient of the lower White Nile, amounting to 1: 100,000 or 1 cm/km (Willcocks, 1904; Williams *et al.*, 1982). The brilliant Italian hydrologist and engineer Elia Lombardini (1794–1878) was the first to suggest that the White Nile owed its gentle gradient to the fact that it flowed across the bed of a former vast lake (Lombardini, 1864, 1865). Lyons (1906) rejected this idea on the grounds of a lack of geological evidence in support of such a lake. In contrast to Lyons, the American geologist Lawson popularised Lombardini's views (Lawson, 1927) and just over a decade later Dr John Ball, the highly experienced Director of Desert Surveys in Egypt, proposed the name 'Lake Sudd' for this hypothetical lake (Ball, 1939). Ball argued that the lake reached as far north as Sabaloka–the site of the sixth and southernmost cataract on the Nile (Fig. 1)–which he maintained had acted as the dam for the lake. According to Ball, this lake would have reached an elevation of about 400 metres. Three key questions arise from this debate. Are there lake sediments in the valley of the White Nile? If so, how and when did the lake (or lakes) originate?

THE WHITE NILE LAKES WERE CAUSED BY BLUE NILE FLOODS

Although the origin of the White Nile lakes has puzzled many observers and provoked some ingenious speculation, the actual cause of such lakes was very clearly explained by the distinguished hydraulic engineer Sir William Willcocks early last century. He provided the solution to the enigma in his book *The Nile in 1904*. His observations deserve to be quoted in some detail because they explain a good deal:

'…the flood of the Blue Nile in July, August and September travels up the White Nile, holds back its waters and converts the valley of the White Nile into a flood reservoir' (Willcocks, 1904, p. 46).

'We are in a lake rather than in a river, and in flood when the waters of the Blue Nile travel 300 kilometres up the White Nile, and wait for a fall in the Blue Nile to discharge themselves into the Nile, we are indeed in a pulsating lake and not in a river' (op. cit., p. 42).

'… the White Nile which, with its slope of 1/100000 in flood, is not a river but a flood reservoir' (op. cit., p. 43).

'… in September, when the discharge of the Blue Nile had fallen from some 11,000 to 6,000 cubic metres per second, and the stored–up waters in the valley of the White Nile were forcing themselves down to take the place of those cut off from the Blue Nile…' (op. cit., p. 42).

The process described by Willcocks in 1904 can be observed today throughout the world wherever two streams join during times of flood (Fig. 2). The stream with the greatest

Fig. 2—Photograph showing localised ponding of water at confluence of two small streams in the hills south of Adelaide, Australia, as a modern analogue of White Nile ponding by the Blue Nile during flood (Photo courtesy of Frances M.Williams).

discharge will operate as a temporary dam to impound the waters of the stream of lesser discharge to form a body of sometimes turbulent water which is released downstream once flow slackens in the previously dominant stream (Willcocks, 1904). With the Blue Nile in spate damming the lower White Nile, there is no need to invoke a bedrock dam such as Sabaloka, or a dam formed by dunes or vegetation. Furthermore, the lake thus formed would have been a seasonal and not a permanent lake, depending for its ephemeral existence on the duration of high flow in the Blue Nile. The Blue Nile paleochannels that radiate from southeast to northwest across the surface of the northern Gezira alluvial plains (Fig. 3) have migrated progressively northwards during the last 120 ka (Williams *et al.*, 2010, 2015; Williams, 2019, pp.142–53 and 160–1). The northern boundary of the former lakes would have likewise shifted progressively to the north, probably reaching the present confluence of the Blue and White Nile by the early Holocene (Arkell,1949, 1953; Williams, 2019, pp. 311–3).

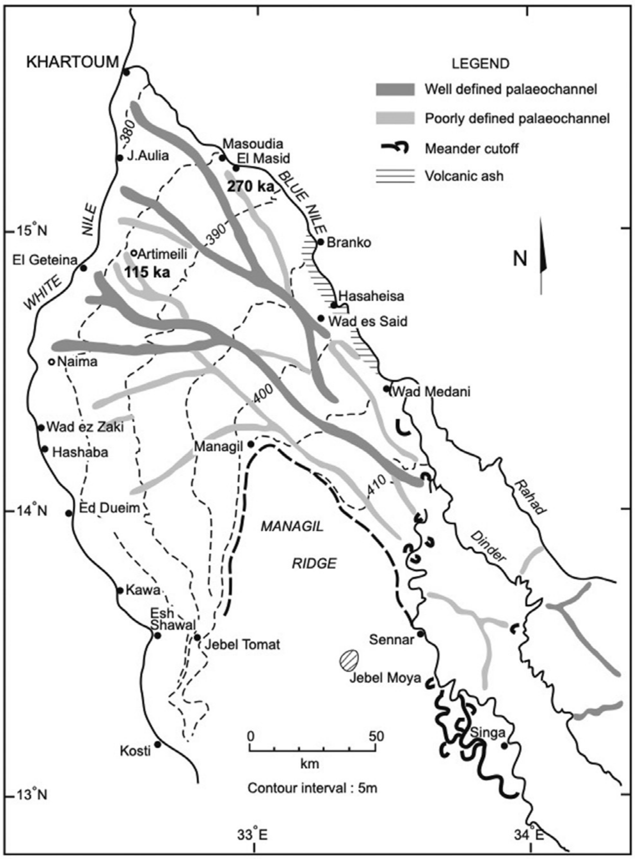

Fig. 3—Progressive northward migration of late Pleistocene Blue Nile channels in the northern Gezira, central Sudan, culminating in the Holocene junction of the Blue and White Nile at Khartoum. The northern channels are generally younger and more clearly defined than the older southern channels. (modified from Williams, 2009, Fig. 3a; Williams *et al.*, 2015, Fig. 1a; Williams, 2019, Fig. 11.1).

GEOMORPHOLOGICAL EVIDENCE OF THE WHITE NILE LAKES

Unless tilted by later tectonic movements, horizontal shorelines are good evidence of the presence of former lakes and have the advantage that they may contain aquatic fossils and dateable sediments (Williams, 2014, pp. 192–3). During 1952–54, while working for Sir Alexander Gibb and Partners, Mr R.H. Gunn mapped the soils and vegetation in south–central Sudan and found evidence of two major terraces cut into White Nile right bank alluvium between Rabak and Renk (Fig. 4) (Sir Alexander Gibb & Partners, 1954; Gunn, 1982). The elevation of these terraces was, respectively, 382 m and 386 m relative to the Alexandria zero height datum used by topographic surveyors in Sudan and Egypt before the advent of satellite altimetry.

Using satellite imagery, air photos and topographic surveys, Williams *et al.* (2003) mapped a series of large arcuate shorelines located at an elevation of about 386 m that ran roughly parallel to and above the eastern margin of the White Nile flood plain between Rabak and Renk (see Figs 4 and 5). Working further to the north, at Esh Shawal (Fig. 4) on the right bank of the White Nile, they also obtained undisturbed sediment samples and sub–fossil gastropod shells from soil pits dug to a depth of up to 6 m. The basal sediments were finely laminated green lacustrine clays. These clays were overlain by a bed of fine sand up to one metre thick which had OSL ages between 170 ka and 210 ka near the base of the bed, becoming older with increasing depth. The upper few centimetres of the green laminated clays had a single OSL age of >240 ka. Extrapolating likely sedimentation rates would give an age of about 400 ka at the base and may possibly represent when Lake Victoria first came into existence and began to supply water to the White Nile headwaters. The uppermost shell–bearing clays at 1.4 m depth had five calibrated ^{14}C ages between 14.3 ka and 13.2 ka and three OSL sediment ages in the bed immediately beneath that were between 21 ka and 16 ka, becoming older with increasing depth (Williams *et al.*, 2003).

THE LAST INTERGLACIAL 386 M WHITE NILE LAKE

The 386 m White Nile mega–lake has proven hard to date although its cuspate shoreline is often remarkably well preserved, especially in the sector east of the present river between Jebelein and Renk (Williams *et al.*, 2003). Barrows *et al.* (2013) were successful in obtaining a ^{10}Be age of ca. 110 ka for the 386 m beach ridge east of Jebelein (Fig. 4). This age indicates the last time that the White Nile lake attained a level of about 386 m, after which it receded progressively, as shown by scattered sand splays located between the upper limit of the early Holocene flood plain and the 386 m beach ridges. Overflow of Lake Turkana (Johnson & Malala, 2009;

Williams, 2019, p. 130–1) into the White Nile via the Pibor and Sobat (Fig. 6) during the last interglacial and again towards 14 ka would have helped to augment the volume of the White Nile lakes at these times. Even now, the Sobat contributes an extra 10 km^3 each year to White Nile flow (Hurst & Phillips, 1931; Hurst, 1952; Williams *et al.*, 1982). This was also a time of very wet conditions in the eastern Sahara, including at Kharga Oasis in the Western Desert of Egypt (Smith *et al.*, 2004) and Bir Tarfawi and Bir Sahara in the far south of the Western Desert (Wendorf *et al.*, 1993). The Last interglacial 386 m White Nile Lake occupied an area of at least 45,000 km^2, with a north to south length of more than 650 km and an east to west width up to 80 km (Williams, 2019, pp. 112–3).

One striking result of the exposure of the very gently sloping late Pleistocene lakes in the lower White Nile Valley was a change in the nature of the Blue Nile channels flowing across the northern Gezira (see Fig. 3). For most of their length these channels were relatively straight but on reaching the exposed lake floors they began to meander. In the area around El Geteina (for location, see Fig. 3) on the White Nile right bank several generations of meandering channels are clearly visible on aerial photographs and satellite images although they are less obvious on the ground. The channels contain fine sand at depth overlain by several metres of heavy clay. The channel sands have optically stimulated luminescence (OSL) ages dating between 100 ka and 70 ka (Williams *et al.*, 2015). Some of the meandering channels are overlain by sand dunes that originated as source–bordering dunes. The optical ages obtained from one of the largest north–south aligned dunes to the north of El Geteina indicate that it was active at 115–105 ka, 60 ka and 12–7 ka, which were times of extreme Blue Nile floods (Williams *et al.*, 2015).

THE TERMINAL PLEISTOCENE 382 M WHITE NILE MEGA–LAKE

The first recorded evidence of this lake seems to be that of Tothill (1948) who described a wave–cut shoreline at an elevation of 382.14 m etched into the north–south aligned sand dune situated immediately east of the White Nile and on which the village of Hashaba is located (for location see Fig. 3). He commented that this dune was not a true desert dune but the remains of a former lakeside dune. The strandline contained water worn aquatic snail shells of *Cleopatra bulimoides* and *Melanoides tuberculata* 'representing dead adults washed up from the river or lake bed by wave action'. (Tothill, 1948, p. 134) When I examined this locality in February 1963, the evidence of this shoreline bench had been obliterated by the villagers and their animals. However, a few weeks later I found freshwater shells at a depth of a metre in one of my soil pits near Esh Shawal (for location see Fig. 4). A shell sample (I–1486) yielded an uncalibrated radiocarbon age of 11,250 ± 220 BP (Williams, 1966). Work a decade later revealed that a buried shell–bed at a depth of 1.0–1.4 m sloped gently upwards from west to east for nearly 11 km from Esh Shawal to just west of the two small granite hills of Jebel Tomat (for location see Fig. 3). The shell–bed extended nearly to the edge of the early Holocene White Nile floodplain. The calibrated ^{14}C ages of the shells collected along an east–west transect of the buried shell–bed ranged between 14.7 ka and 13.1 ka (Williams, 2009, 2019, p. 117). Detailed levelling showed that

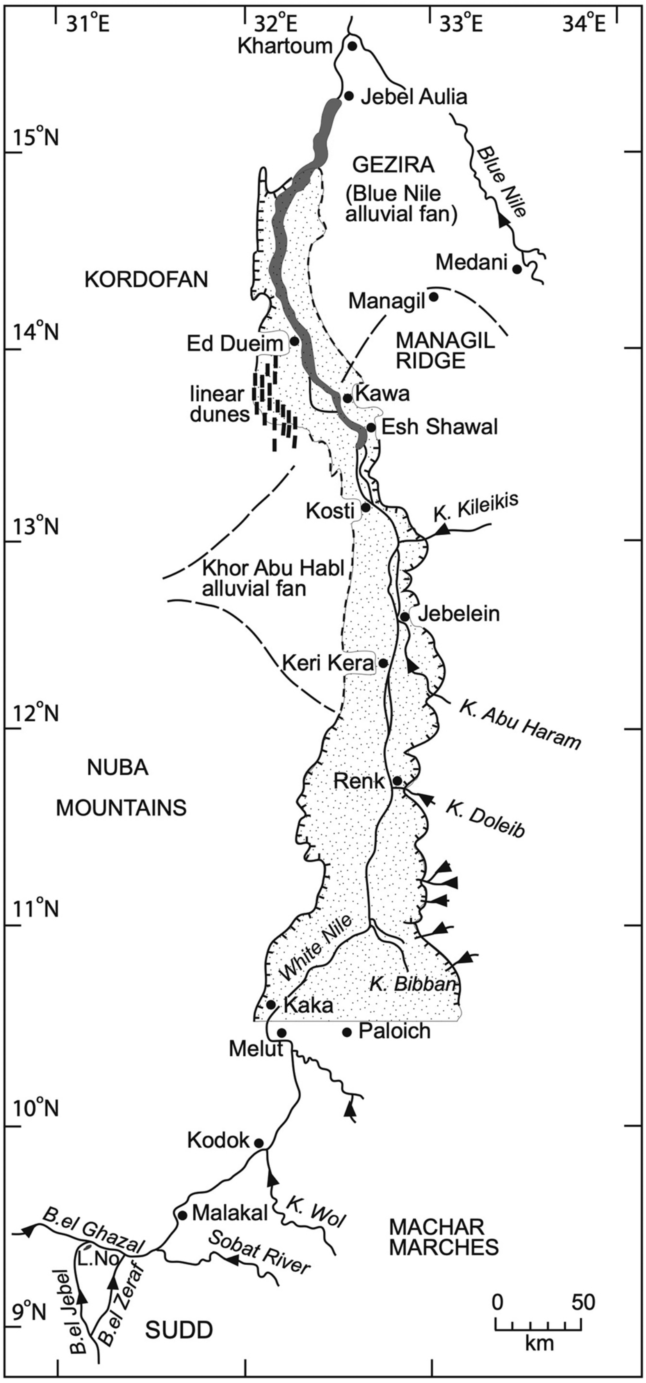

Fig. 4—The 386 m White Nile mega–lake (after Williams *et al.*, 2003, Fig. 1; Williams, 2009, Fig. 2; Williams, 2019, Fig. 8.2).

Fig. 5—False colour satellite image showing a 50 km long stretch of the arcuate White Nile last interglacial 386 m lake shoreline extending from 10 km south of Renk northwards. The modern White Nile is on the left of the image. For location see Fig. 4.

the highest elevation attained by the shell–bed was close to 382 m. The buried shell–bed at 1.0–1.4 m depth indicates a progressive recession by the White Nile seasonal lake from its maximum flood level of 382 m, with the shells representing a former beach facies. The most common species of freshwater gastropods and bivalves were *Melanoides tuberculata*, *Lymnaea natalensis*, *Cleopatra bulimoides*, *Biomphalaria pfeifferi*, *Bulinus truncatus* and *Corbicula fluminensis* and Nile oyster (*Etheria elliptica*). At its maximum, this White Nile lake was at least 400 km long from north to south and up to 25 km wide from east to west. The flooding to this level reflects the resumption of flow from the equatorial headwaters of the White Nile at this time (Talbot *et al.*, 2000; Williams, 2009, 2019). This event is also evident in the initial accumulation of the youngest sapropel layer in the eastern Mediterranean, discussed in the next section.

NILE FLOODS AND MEDITERRANEAN SAPROPELS

A sapropel is 'a mud or ooze composed predominantly of anaerobically decomposing organic material, usually in aquatic environments' (Goudie *et al.*, 1985, p. 376). Sapropel layers are common in Miocene, Pliocene and Quaternary sediments on the floor of the Mediterranean (Lourens *et al.*, 1996; Kroon *et al.*, 1998; Cramp & O'Sullivan, 1999; Larrasoaňa *et al.*, 2003). The anaerobic conditions that enable sapropels to form are widely considered to result from a change in seawater stratification caused by a substantial influx of freshwater from major rivers such as the Nile (Rossignol–Strick *et al.*, 1982; Rossignol–Strick, 1985, 1999). Rossignol–Strick *et al.* (1982) were the first to suggest that changes in the strength of the African summer monsoon were controlled by the earth's precessional cycle, which is related to the season of the year in which the earth is closest to the sun and is controlled by the direction in which the spin axis of the earth points in space, with an average cycle duration of about 20,000 years.

In order to test the hypothesis that these beds of highly organic mud found in marine sediment cores collected from the Eastern Mediterranean did indeed accumulate during times of sustained very high discharge from the Nile two things are necessary. One is an accurate and precise chronology of times of sapropel accumulation in the eastern Mediterranean (Lourens *et al.*, 1996; Ducassou *et al.*, 2008, 2009); the other is an equally accurate and precise chronology for extreme flood episodes from the Nile.

Zhao *et al.* (2011) obtained a long record (1.75 Ma) from a single deep–sea sediment core located on the distal Nile Cone. They identified 21 sapropel layers, defined as sediments with at least 1% organic carbon. These sapropels were enriched in barium. A further 21 dark layers enriched in barium, but from which the carbon had been oxidised and removed they termed 'ghost sapropels' and 'hidden sapropels.' Only the later part of the sediment record has a chronology one can consider precise, but the question arises as to how representative a single marine sediment core might be for deducing changes in Nile water and sediment discharge.

Ducassou *et al.* (2009) analysed changes in sediment input within forty deep–sea sediment cores located across the Nile Cone and spanning the last 200 ka. They found that times of sapropel formation were associated with times of very high inferred Nile discharge. The sapropel chronology appears robust, but it is difficult to see how changes in sediment input can be used as a precise and accurate measure of changes in Nile flood discharge. What is needed to test the hypothesis that sapropel accumulation does indeed coincide with phases of prolonged high Nile discharge is an equally robust terrestrial recorded from well–dated alluvial deposits laid down during times of extreme floods on land. A useful contribution to this

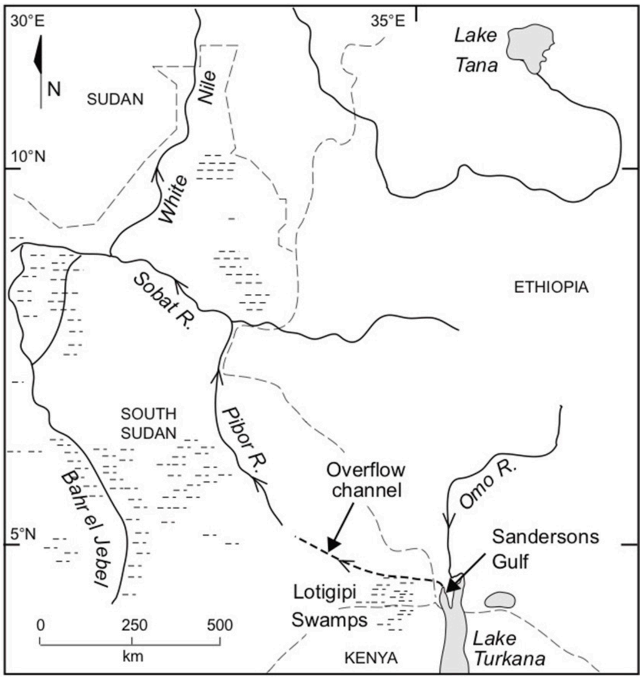

Fig. 6—Map showing approximate location of the early Holocene overflow channel between Lake Turkana and the White Nile via the Pibor and the Sobat. (after Williams, 2019, Fig.9.2, compiled from data in Butzer, 1971, Fig.1.5; Adamson & F Williams, 1980, Fig. 10.5; Harvey & Grove, 1982, Fig. 2; Johnson & Malala, 2009).

debate is that of Mologni *et al.* (2020) who compare their high–resolution Holocene marine sedimentary record with proxy terrestrial evidence for wetter episodes in the Ethiopian Highlands during the Holocene.

Revel *et al.* (2010, 2015) analysed the strontium and neodymium isotopic composition of land–derived sediment in two deep sea cores located about 100 km WNW of the outlet of the Rosetta distributary channel on the Nile Delta. They identified significant desert dust influx during the two main arid intervals MIS 4 and MIS 2 (Last Glacial Maximum). Changes in major elements, notably Fe, which were analysed with a 10–year resolution, coincided with wetter phases across North Africa during 98–69 ka, 60–50 ka, 38–30 ka and 14–5 ka. The 98–69 ka wet phase spans the time when sapropel S4 (102 ka) and S3 (81 ka) accumulated in the eastern Mediterranean; the 60–50 ka wet phase is coeval with sapropel S2 (55 ka), and the 14–5 ka wet phase with sapropel S1 (13.5–9 ka). Using high values of Fe as a proxy for sediments derived from an Ethiopian volcanic source, they concluded that the Ethiopian headwaters of the Atbara and Blue Nile were much wetter between 14.8 and 8.4 ka, consistent with a stronger East African Monsoon. The gradual reduction in Blue Nile sediment deposition and flood discharge between 8.4 ka and 3.7 ka reflects the onset of climatic desiccation culminating in extreme aridity between 3.7 and 2.6 ka (Revel *et al.*, 2015). These paleoclimatic inferences are supported by the results obtained from a 12 m sediment core retrieved from the Dendi eastern crater lake near the Blue Nile headwaters in Ethiopia, which indicates peak wetness–and presumably high Nile flow–between 10.0 ka and 8.7 ka (Wagner *et al.*, 2018).

An independently dated chronology of high Nile flood events obtained directly from non–marine sediments is therefore needed to improve our understanding and interpretation of distal Nile deltaic and marine deposits. Williams *et al.* (2015) dated extreme flood episodes in the Blue, White and main Nile valleys using OSL, ^{10}Be and ^{14}C and demonstrated, for the first time, that dated high Nile flood deposits coincide with sapropel units S9 (240 ka), S8 (217 ka), S7 (195 ka), S6 (172 ka), S5 (124 ka), S4 (102 ka), S3 (81 ka), S2 (55 ka) and S1 (13.5–9 ka). The younger the deposit, the more precise the age. Given the inevitably large error terms on ages older than about 125 ka, any correlations older than that age should be viewed as being plausible but not definitive.

CHANGES IN SEDIMENT FLUX IN THE NILE VALLEY

Some 2,500 years have elapsed since Herodotus (c. 485–425 BC) wrote that 'Egypt is, as it were, the gift of the river ……the soil of Egypt is black and friable as one would expect of an alluvial soil formed of the silt brought down by the river from Ethiopia' (Herodotus, trans. 1954). However, there were many times during the Pleistocene when the Nile had a very different sediment load to the suspension load silts and clays described by Herodotus. On a number of occasions during the Pleistocene, both the Blue Nile and the Desert Nile were transporting and depositing alluvial gravels rather than clays. For example, near Masoudia in the northern Gezira (for location see Fig. 3), the Blue Nile was depositing gravels at 270 ± 30 ka (roughly coeval with sapropel S10) and 190 ± 20 ka (roughly coeval with sapropel S7). These gravel beds are at least 10 m thick in total and are overlain unconformably by dark alluvial clays up to 5 m thick. Nile oyster (*Etheria elliptica*) shells at the base of the clay unit have calibrated AMS ^{14}C ages of 40,949 ± 477 cal BP and 43, 411 ± 545 cal BP. The clay bed has an OSL age of 52 ± 7 ka near its base. The clay unit is therefore coeval with sapropel S2 (Williams *et al.*, 2015). The oysters appear to have died as a result of a sudden influx of fine sediment. Their preferred habitat is a rocky or stony channel bed.

GRAVEL ACCUMULATION IN NUBIA

In the Nubian Desert of northern Sudan and southern Egypt, the Desert Nile was actively aggrading and depositing sand and gravel at intervals during 75–50 ka and 30–20 ka (Williams, 2019, p. 205). Butzer and Hansen (1968) considered that these late Pleistocene gravels were laid down at a time of increased Nile flow and wetter conditions in the Ethiopian headwaters of the Nile but that the Nile lacked the necessary competence to transport them as far as the Nile Delta. Fairbridge (1962, 1963), in contrast, considered that the gravels represented a time of diminished Nile flow at a time of glacial aridity in the headwaters. Adamson *et al.* (1980) and Williams and Adamson (1980) offer a more nuanced interpretation, elaborated below, and one that takes into account the decisive role of the White Nile.

DESICCATION OF THE UGANDAN LAKES FEEDING THE WHITE NILE

During the Last Glacial Maximum (LGM: ca. 24–18 ka) flow into the Nile from its equatorial headwaters ceased when Lakes Victoria and Albert became closed basins. Soils developed on the exposed floor of Lake Albert during 20.75–17.75 ka and during 16.73–15.1 ka (Williams *et al.*, 2006: Fig. 9b). At the same time that the late glacial White Nile was reduced to a seasonal trickle, the Desert Nile was depositing sand and gravel in northern Sudan and southern Egypt. The highlands of Ethiopia were glaciated at this time (Williams, 2019, pp. 164–71). These three phenomena–glaciation in the Ethiopian headwaters of the Blue Nile and Atbara, severely diminished White Nile flow, and gravel deposition in Nubia–are all interlinked and are discussed in greater detail in the following sections. Once the Desert Nile was deprived of water from the White Nile it probably ceased to flow during the winter months. The Atbara and Blue Nile would still have contributed to flood flow during the summer months, allowing coarse sand and gravel to be transported as far north as southern Egypt, but they most likely dried out during the winter months, so that the Desert Nile would have been deprived of water at these times. The coarse nature of the sediment reflects events in the Ethiopian headwaters of the Nile.

GLACIATION IN THE ETHIOPIAN HIGHLANDS

The Ethiopian Highlands (Figs 7, 8) were glaciated during and before the LGM and the snowline was lowered on several occasions by up to about 1,000 m ((Hövermann, 1954; Hastenrath, 1974, 1977; Williams *et al.*, 1978; Messerli *et al.*, 1980; Hurni, 1982; Osmaston *et al.*, 2005; Williams *et al.*, 2015; Williams, 2019, pp. 164–175). Small cirque glaciers were common above about 4,000 m in the Semien Highlands which is where the headwaters of the Tekezze/Atbara rise (Fig. 9). The Abbai/Blue Nile headwaters rise just south of the Semien Highlands. Williams *et al.* (2015) obtained seven ^{36}Cl exposure ages from the glacial moraines of two former cirque glaciers in the Semien Highlands (Fig. 10). The ages from Mount Bwahit cirque were 68.5 ± 3.5 ka, 32.7 ±1.4 ka and 15.3 ± 0.7 ka. The moraines on Mount Mesarerya cirque had ^{36}Cl exposure ages of 46.2 ± 2.1 ka, 36.4 ± 1.6 ka, 27.4 ± 1.0 ka and 18.1 ± 0.9 ka. Some of the older ages may denote inheritance of ^{36}Cl from older erosional surfaces

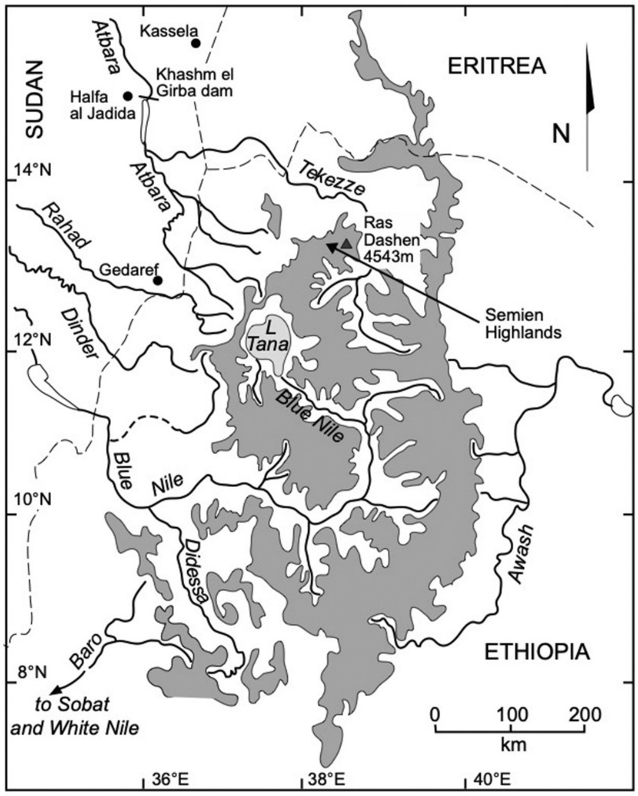

Fig. 7—Semien Highlands of Ethiopia showing the upper catchment of the Tekezze/Atbara River (after Williams, 2019, Fig. 12.1).

due to shallow glacial erosion in these small cirque glaciers rather than a prolonged period of glaciation. However, we can say with some confidence that ice was present on Mount Mesarerya 18.1 ka ago and on Mount Bwahit 15.3 ka ago (Williams *et al.*, 2015).

Periglacial deposits (Figs 11, 12) exposed in one of the tributary valleys of the Tekezze and attributed to the LGM extend about 1000 m below the present–day lower limit of periglacial activity, indicating that the upper tree–line would likewise have been up to 1,000 m lower at that time (Williams *et al.*, 1978; Williams, 2019, p. 166). Bare, unstable hill slopes and active physical weathering in the headwaters of both the Atbara and the Blue Nile would mean a substantial influx of coarse debris to these rivers during the summer meltwater months when the ground was no longer frozen.

RIVER RESPONSE TO CLIMATE CHANGE IN THE NILE HEADWATERS

The Last Glacial Maximum was a time of reduced intertropical precipitation (Gasse *et al.*, 2008), with much diminished summer rainfall in the Ethiopian headwaters of the Nile. Lake Tana, near the headwaters of the Blue Nile, dried out during the LGM, thereby depriving the Blue Nile of a perennial supply of water (Lamb *et al.*, 2007; Marshall *et* *al.*, 2011). The large tributaries below Lake Tana would have provided the Blue Nile with water for part of the year. The outcome was high seasonal runoff, with high peak flows, but a much–reduced total annual discharge (Fig. 13a), with the Atbara drying out for many months and the Blue Nile most likely for several months, much like the pre–regulated Atbara (Adamson *et al.*, 1980). In the Semien Highlands, winters were between 4°C and 8°C cooler than today, with small cirque glaciers present above 4200 m elevation (Williams, 2019, pp. 166–8). Periglacial solifluction was active, the upper tree line was 1000 m lower, and slopes were unstable down to 3000 m, with abundant coarse debris being supplied to the Atbara and Blue Nile headwaters (Williams, 2019, pp. 168–71). Down in the lowlands of central Sudan on the Gezira alluvial fan, deflation of the Blue Nile distributary channel sands resulted in the formation of source–bordering dunes (Williams, 2019, pp.158–9).

In a pioneering study using strontium isotope ratios in dated White and Blue Nile gastropod shells and in Lake Albert sediment cores, Talbot *et al.* (2000) were able to put more precise limits on the date at which flow resumed in the White Nile. Williams *et al.* (2006) followed up this work in more detail and demonstrated that there was a relatively abrupt return of the summer monsoon at about 14.5 ka. As a result, the wet season lasted longer and there was higher summer rainfall (Fig. 13b). In the lowlands of central and northern Sudan, savanna woodlands and grasslands expanded into regions that were previously semi–desert. In the Ethiopian Highlands the slopes became vegetated and stable above 3000 m, with chemical weathering of the volcanic rocks leading to soil formation and a supply of fine sediment to the headwaters of the Blue Nile and Atbara. Bastian *et al.* (2017) provide a useful overview of the importance of enhanced silicate rock weathering processes at this time. Both the Atbara and the Blue Nile had a higher annual discharge and an increased base flow, resulting in attenuated but still very significant flood peaks. Flow in the Blue Nile was now perennial and there was widespread wet season flooding by suspension load channels in the Blue Nile, its tributaries, and along the Desert Nile.

Both the Blue Nile and the Desert Nile responded to changes in discharge and sediment load by changing their channel planform. During those times when they were mainly transporting a traction load of sand and gravel, their longitudinal channel forms had a relatively straight or braided pattern and channel cross–sections were wide and shallow with abundant mid–channel bars and point–bars. Conversely,

Fig. 8—Dissected northern escarpment of the Semien Highlands, Ethiopia, near the headwaters of the Tekezze River. The horizontal basalt flows exposed in the cliffs were erupted about 30 million years ago.

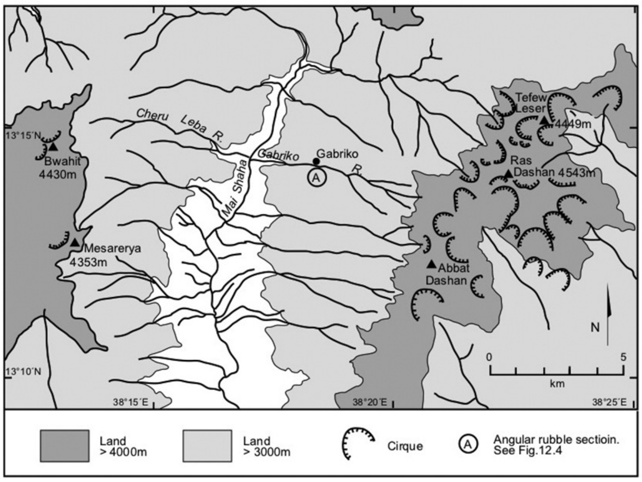

Fig. 9—Geomorphic map of the Ras Dashan summit region, Semien Highlands, Ethiopia (after Williams, 2019, Fig.12.2, based on data in Werdecker, 1967; Hastenrath, 1974; Williams *et al.,* 1978; Hurni, 1982; field observations in 1971, 1974, 2009; and air photo interpretation. The location of the two glaciated summits where the moraines were sampled for ^{36}Cl exposure dating are shown as B (Mount Bwahit) and M (Mount Mesarerya).

during times of mostly fine suspension load transport, the channels became more sinuous and developed U–shaped cross–sections. As time progressed and the channel bed and banks became lined with clay, less kinetic energy was lost in friction, and the river began to incise its channel bed. Although the precise time is not well established, incision by the Desert Nile and Blue Nile probably began soon after 9–8 ka, leading to a progressively narrower zone flooded each year. Blue Nile incision also led to incision by the White Nile, which propagated slowly upstream, causing drying out of previously perennial swamps and wetlands.

The White Nile derives its sediments from highly weathered rocks in its equatorial headwaters (Garzanti *et al.*, 2006, 2015; Padoan *et al.*, 2011). Under present climatic conditions it transports a fine suspension load of clay–sized particles. In South Sudan the alkaline soils of the alluvial and swampy lowlands are smectite–rich with subordinate kaolinite, chlorite and illite (Grazanti *et al.*, 2015, p. 29). This has not always been the case, (although an important unresolved research question is whether smectites were present in the oldest unequivocal White Nile sediments). At some time between about 30–25 ka, possibly for only a few centuries, the White Nile seems to have been able to carry

Fig. 10—Late Pleistocene glacial moraine at about 4000 m elevation in the Semien Highlands of Ethiopia, close to the headwaters of the Tekezze River.

a substantial traction load of medium to coarse quartz sand under conditions of very high energy flow. White Nile bedload sand with large cross–beds indicative of high energy flow and exposed in a quarry near Ed Dueim (Fig. 4) west of the modern White Nile had an OSL age of 27.8 ± 3 .2 ka (Williams *et al.*, 2010). At about the same time as this phase of high energy flow in the White Nile, alluvial fans were being deposited along the foot slopes of granite inselbergs near Jebelein (Fig. 4) close to the 386 m White Nile shoreline, which they overlie and partially obscure (Williams *et al.*, 2010). The OSL age of one such fan was 27.5 ± 2.7 ka (Williams *et al.*, 2010). There was renewed fan deposition at this site during the terminal Pleistocene at 14.5 ± 3.2 ka, which also coincides with the initial overflow from Lakes Victoria and Albert into what later developed into the Sudd swamps.

The interval between 14.5 ka and 12,800 was one of enhanced East African Monsoon and of wetter conditions across the Sahara and Nile Basin (Gasse & Roberts, 2004; Drake *et al.*, 2011). However, the very warming which had strengthened the summer monsoon now produced the reverse effect. A massive influx of glacial meltwater into the North Atlantic caused a major change in global ocean circulation and brought cold conditions to North America and Europe, where it is known as the Younger Dryas event. It also brought prolonged cold and drought to North Africa (Roberts *et al.*, 1993). The Younger Dryas event lasted for more than a thousand years, from 12.8 ka to 11.5 ka, and was associated with reactivation of previously vegetated and stable desert dunes across much of the Sahara (Swezey, 2001) and a significant reduction in Nile flow.

The Younger Dryas drought finally ended about 11,500 years ago and the climate became wetter once more across North Africa and the Nile Basin (Grove, 1993; Claussen *et al.*, 1999; Garcin *et al.*, 2007). The Sahara was home to numerous small lakes (Faure *et al.*, 1963; Faure, 1966, 1969). Lake Chad deepened and expanded (Servant & Servant–Vildary, 1980) until it started to overflow into the Niger River and into the Atlantic. At its Holocene maximum, about 9,000 years ago, it covered an area of nearly 400,000 km^2 (Drake & Bristow, 2006). Pollen grains from former lake sites in northern Sudan indicate a northward displacement of vegetation zones (and of the associated summer rainfall zones) by about 400–450 km between 10 ka and 6 ka (Ritchie *et al.*, 1985; Ritchie & Haynes, 1987).

The Sudd swamps would not exist without the sustained input of water from the rivers that now flow into and through them. When the rivers ceased to flow, the swamps dried out. Once flow resumed, there would have been a time lag

Fig. 11—Late Pleistocene periglacial blockfield at an elevation of about 3500 m in the Semien Highlands of Ethiopia.

of several centuries, at least, before the swamps became fully established. During this hiatus between no flow and resumption of flow through the swamps, the hydrological regime of the White Nile would have been very different. With no water losses from seepage and evapotranspiration, the White Nile discharge would be greatly increased, possibly to nearly double its present amount. Furthermore, without the buffering influence exerted upon discharge, the river would respond more rapidly to precipitation and would have become far more seasonal than now. At present, the ratio of peak monthly flow to minimum monthly White Nile flow is only about 5: 2. With no swamps to filter out any coarse sediment, the White Nile would have been capable of carrying a substantial bedload of sand as opposed to its present limited suspension load of clay–sized particles.

Once the Sudd swamps were re–established, which on present evidence seems to have been at about 14.7 ka, the modern flow regime would have begun, but not for long. The cold dry Younger Dryas climatic event (12.7–11.5 ka) brought renewed aridity to the Nile Basin, so that the White Nile would again have experienced a change in flow regime, from much reduced during peak aridity to much increased immediately after, for a few short centuries. Garcin *et al.* (2007) have documented this second abrupt return of the African monsoon at the Younger Dryas–Holocene climatic transition, just as Williams *et al.* (2006) have documented the initial abrupt return of the African summer monsoon between about 15 ka and 14.5 ka.

INTERPRETING THE ENVIRONMENTAL RECORD PRESERVED IN THE NILE DELTA AND THE NILE CONE

In a classic study of the heavy mineral composition of Holocene sediments in the Nile Delta, Foucault and Stanley (1989) postulated that a high pyroxene content denoted a volcanic provenance and hence a major input of sediment from the Blue Nile and Atbara. They further postulated that high counts of amphibole relative to pyroxene indicated a significant influx of sediment from the White Nile. Times of inferred low Blue Nile and Atbara sediment input were coeval with times when Lake Chad and the Ethiopian Rift lakes were high (Gasse & Street, 1978; Servant & Servant–Vildary, 1980; Gasse, 2000; Gasse & Roberts, 2004). They concluded that during times when the climate in the Ethiopian headwaters of the Nile was warm and humid, a dense vegetation cover protected the hill slopes from erosion, so that sediment loads were low although flood discharge was high. According to

Fig. 12—Detail of weathered basalt blocks in the blockfield shown on Fig. 11.

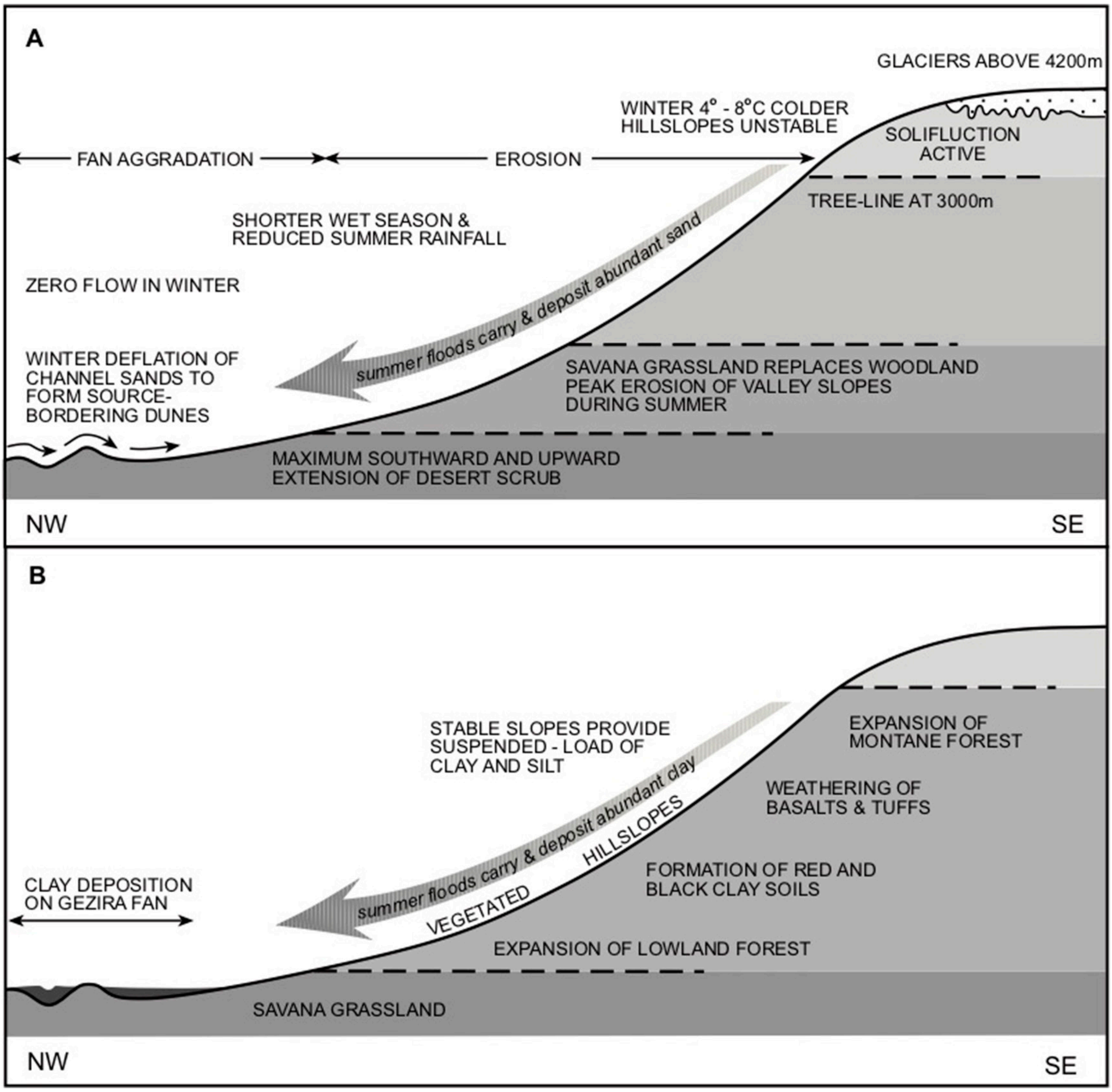

Fig. 13—(A) Depositional model for the late Pleistocene Blue Nile (after Williams & Adamson, 1980, Fig. 12.4a; Williams, 2019, Fig. 6.7a). (B) Depositional model for the early Holocene Blue Nile (after Williams & Adamson, 1980, Fig. 12.4b; Williams, 2019, Fig. 6.7b).

their model, during cold dry climatic phases, the Ethiopian Nile tributaries would have brought more sediment down to the Delta. They also proposed that during times of wetter intertropical climate, when the Intertropical Convergence Zone (ITCZ) extended further north into southern Egypt and the eastern Sahara, wadis that are dry at present would have flowed into the Nile, bringing sediments derived from Basement Complex rocks rich in amphibole. However, Garzanti *et al.* (2015) conducted a comprehensive analysis of modern Nile sediments along the length of the Nile and have demonstrated that the amphibole/pyroxene ratio is not an appropriate indicator of White Nile *versus* Ethiopian provenance.

Hamroush and Stanley (1990) investigated changes in trace elements and rare earth elements from three Nile Delta sediment cores spanning the last 30 ka. They used peaks in the ratio of chromium to scandium (Cr/Sc) as an index of high inputs of sediment from Ethiopia and peaks in the ratio of lanthanum to lutetium (La/Lu) to infer a substantial input of sediment from the Ugandan headwaters of the White

Nile. During the humid climatic phase in Ethiopia revealed by high lake levels between 7 ka and 4 ka (Williams *et al.*, 1977; Gasse & Street, 1978), the Cr/Sc ratio was at first high, indicating a substantial input of Ethiopian sediment to the Nile. Over time, the plant cover became dense enough to minimise hill slope erosion. Flood flow remained high but sediment input to the Blue Nile and Atbara fell. Here again, caution is advisable. Both Cr and Sc are present in unknown proportions in pyroxene and in volcanic rock fragments and the relative abundance of chemical elements is more likely to be controlled by hydraulic sorting processes than by provenance (personal communication, Professor Eduardo Garzanti, 12 October 2020).

A much longer record extending back to 1.75 Ma came from marine sediment core MD90–964 on the far edge of the Nile Cone (Zhao *et al.*, 2011). Variations in the ratio of titanium to vanadium (Ti/V) at different depths were used as a proxy for fluctuations of sediment derived from the volcanic highlands of Ethiopia and showed a strong precessional influence with a periodicity of about 23 ka. The precessional cycle denotes the changing season of the year when the Earth is closest to the sun. It is controlled by the direction in which the spin axis of the Earth points in space. When the Earth is closest to the sun during the northern summer, the tropical monsoon will usually be stronger and the summer precipitation greater in amount and more intense across the intertropical zone, which includes the Nile headwaters. The interpretation proposed by Zhao *et al.* (2011) is similar to that put forward by Foucault and Stanley (1989) and posits that there was less erosion and less sediment input to the Atbara and Blue Nile during times of stronger monsoon and denser, more widespread vegetation cover.

In a subsequent study of the same marine sediment core, Zhao *et al.* (2012) used variations in the ratio of iron to aluminium (Fe/Al) as an index of sediment input to the Nile from its Ethiopian headwaters, with Fe–bearing heavy minerals diagnostic of a volcanic source. They concluded that times when the Fe/Al index were highest were also times of sapropel formation, once again with a spectral peak of about 23 ka denoting a precessional signal. They proposed that sapropel formation occurred during times of increased Nile flood discharge but reduced fine sediment load.

Box *et al.* (2011) analysed the strontium isotopic composition and major element geochemistry of two fine–grained sediment cores from the edge of the Nile Cone, one south of Cyprus and the other off the Southern Coast of Israel. They deduced that far less suspended sediment was delivered to the Nile from its Ethiopian headwaters during the 11–5 ka wetter climatic phase than during the drier phases before and after this humid episode. They further deduced that sediment from the White Nile increased by a factor of three from 13 ka to 5 ka, while the influx of Saharan dust decreased during this time. In order to account for these inferred observations, they postulated that an increase in precipitation in the equatorial headwaters of the White Nile would generate an increase in hill slope erosion and an increase in sediment transport through the Sudd swamps. They further postulated that in the seasonally wet Ethiopian Highlands an expansion of the vegetation cover would lead to less erosion and a lower influx of sediment from the Ethiopian headwaters of the Nile. Neither conclusion seems convincing.

In fact, there are a number of unproven assumptions implicit in all of these studies that require careful scrutiny. It is curious that the authors rely on very speculative inferences and ignore the vast literature on erosion and sediment transport in present–day tropical rivers, reviewed in great detail in Gupta (2007) and summarised more recently by Gupta *et al.* (2020). It is worth examining the links between monsoon precipitation, plant cover and erosion in the context of modern process studies, because it illustrates the dangers of relying solely upon analysis of distal sediments stored in river deltas and their offshore marine fans to deduce climatic events in the proximal headwaters of a large river basin like the Nile. Peak sediment yields in large present–day rivers occur in the monsoonal tropics and in semi–arid regions and are relatively low in densely vegetated catchments in equatorial climatic zones (Douglas, 1967, 1969; Milliman & Meade, 1983; Milliman, 1997; Gupta, 2007; Williams, 2012; Gupta *et al.*, 2020). It is hardly likely that the Sudd swamps would trap less sediment during times of increased regional precipitation. The opposite is far more likely because the area covered by swamps would increase, enhancing its sediment trapping efficiency.

Furthermore, it is hard to distinguish unequivocally between sediment transported by the White Nile and sediment transported by the Blue Nile and Atbara. Overflow from Lake Turkana via the Pibor into the Sobat and then into the White Nile (see Fig. 6) (Johnson & Malala, 2009; Williams, 2019, p. 130–1) means that during wetter climatic episodes the White Nile would be preferentially enriched in sediments from a volcanic source. Each of the studies summarised earlier neglects this factor. So, for example, during the very Late Pleistocene and Early Holocene, until about 5 ka, White Nile sediment would show an enhanced volcanic influence, and be difficult to separate out from sediments derived directly from the volcanic uplands of the Blue Nile and Atbara.

Finally, during times of wetter climate in the valley of the Desert Nile, whether from increased summer or increased winter rainfall, local wadis flowing from the Basement Complex granites and metasediments of the Red Sea Hills would transport sediments to the Nile that are geochemically very similar to sediments derived from similar rocks in the White Nile headwaters. Once again, care is needed to identify such inputs.

These *caveats* are not intended to depreciate the value of careful studies of the sediments preserved in the Nile Delta and Nile Cone, but to stress the need for greater caution when interpreting the isotopic and geochemical record preserved in

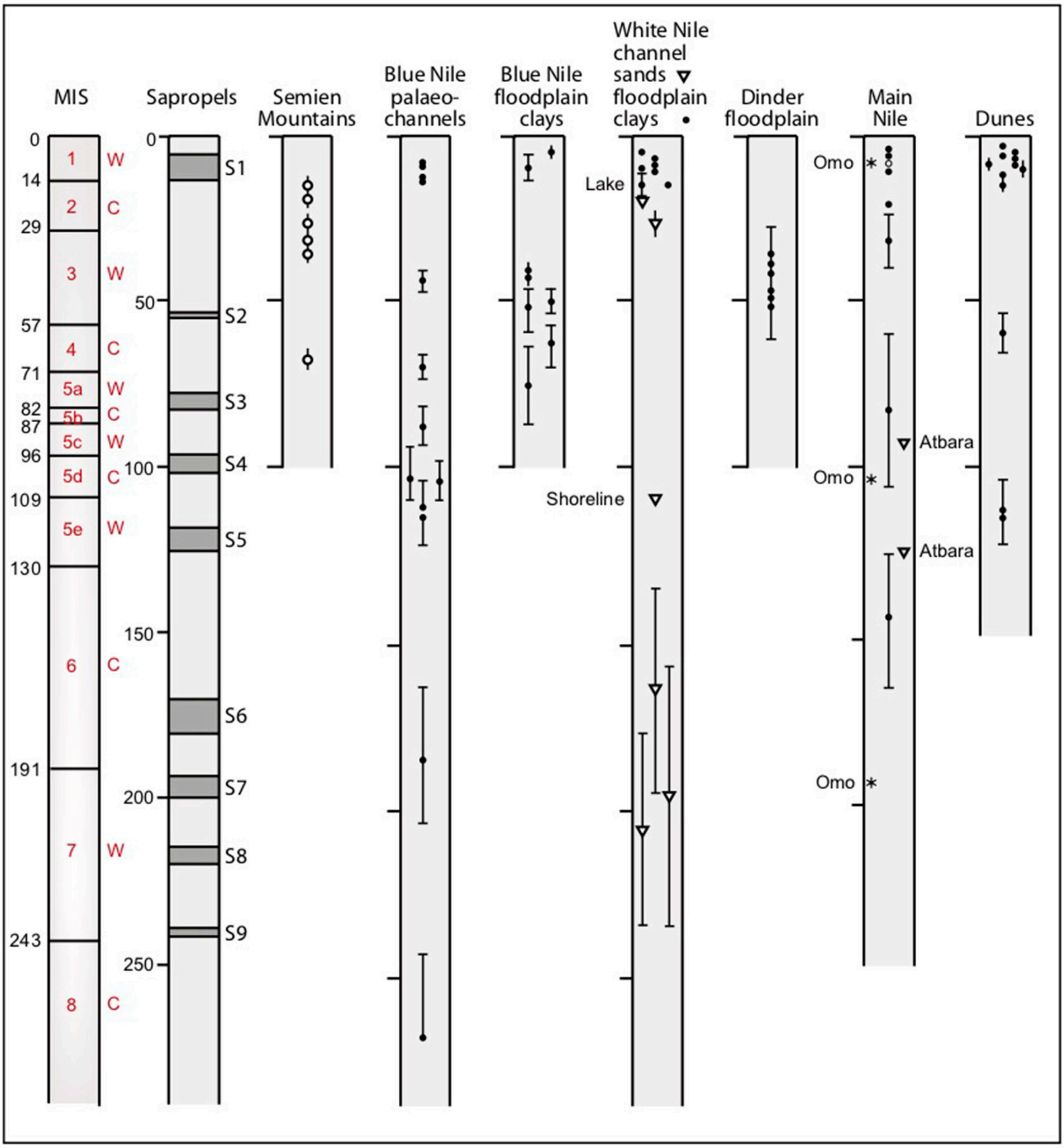

Fig. 14—Phases of sapropel formation in the Nile Cone in relation to episodes of high Nile flow during the last 250 ka. MIS is Marine Isotope Stage, W is warm, C is cold. Sapropel: S1 to S9 are sapropel units. Semien Mountains: open circles denote exposure ages for glacial moraines, with vertical error bars. Blue Nile: solid black dots with vertical error bars denote major flood episodes evident in once active palaeochannels and in widespread deposition of floodplain clays. White Nile: inverted triangles denotes deposition of high energy channel sands or sandy shoreline of Last interglacial lake in lower White Nile Valley; solid black dots denote episodes of very high White Nile floods. Dinder floodplain: solid black dots denote episodes of high flow. Main Nile: solid black dots denote phases of very high floods with vertical error bars; the three asterisks denote times of high flow in the Omo and overflow from Lake Turkana into the White Nile via the Sobat; the open inverted triangles denote phases of very high flow in the Atbara. Dunes: the solid black dots and vertical error bars denote optically dated phases of eolian activity and sand accumulation west and east of the lower White Nile and west and east of the main Nile (after Williams *et al.,* 2015, Fig. 8; Williams, 2019, Fig. 21.2).

such sediments. Analyses of the marine sedimentary record can generate useful hypotheses to be tested against the continental sedimentary record, and conversely. Only thus will real progress be possible.

One study that appears on chronological grounds to meet these criteria is that of Hennekam *et al.* (2014) who analysed the bulk sediment inorganic geochemistry and the stable oxygen isotopic composition of the planktonic foraminifer *Globigerinoides ruber* in a Holocene marine sediment core from the Nile Cone. They identified five periods of enhanced Nile flow for which they obtained calibrated ages of ca. 9.7 ka, ca. 9.1 ka, ca. 8.6 ka, 7.7 ka and 6.6 ka. These ages agree remarkably well with independent evidence of high White Nile flow at 9.7–9.0 ka, 7.9–7.6 ka and 6.3 ka, and of high Blue Nile flow at 8.6 ka, 7.7 ka and 6.3 ka (Williams, 2009). The recent study by Mologni *et al.* (2020) also takes into account Holocene terrestrial records of environmental change in the Ethiopian headwaters of the Blue Nile in its interpretation of a marine sedimentary core recovered from the Nile Cone.

POSSIBLE HUMAN RESPONSES TO LATE QUATERNARY NILE FLUCTUATIONS

An interesting question to raise for future work, but one beyond the present scope of this review to answer, is how prehistoric human groups may have responded to late Quaternary environmental changes in the Nile Basin. For instance, did the Nile act as a bridge or a barrier to human movement along the Nile Valley? This issue is discussed by Vermeersch and Van Neer (2015), Leplongeon *et al.* (2020) and Williams, (2020) and is relevant to wider issues of human migrations out of Africa, including when and where they might have taken place (Williams, 2019, pp. 322–33). The Nile Valley does not have a monopoly over prehistoric migration routes from Africa. Other possible prehistoric human migration routes out of Africa and into Eurasia are along the Coast of North Africa and across the Nile Delta through the Sinai Desert into the Levant and/or across the Red Sea to the Arabian Peninsula, possibly via the narrow, shallow Bab el Mandeb strait at the southern end of the Red Sea, using some form of watercraft. Any environmental impediment to movement northwards along the Nile Valley may have made these other routes more attractive. A key issue here is whether prehistoric humans became sufficiently well adapted to living in arid regions that crossing deserts like the Sahara or the Sinai was less of an impediment than we might imagine.

CONCLUSION

Late Quaternary environments in the Nile Basin reflect the influence of the African summer monsoon upon plant cover, sediment yield and flood discharge in the Ethiopian and Ugandan headwaters of the Nile. Intervals of prolonged and very high Nile flow coincide with times of stronger summer monsoon and have been dated using a combination of ^{14}C, OSL and ^{10}Be methods. Periods of high Nile flow into the eastern Mediterranean coincide with the formation of highly organic sedimentary layers termed sapropels (see Fig. 14). Ages obtained so far for these times of sustained middle to late Pleistocene high flow in the Blue and White Nile are broadly coeval with sapropel beds S8 (ca 217 ka), S7 (ca 195 ka), S6 (ca 172 ka), S5 (ca 124 ka), S4 (ca 102 ka) S3 (ca 81 ka), S2 (ca 55–50 ka) and S1 (10–6.5 ka). Sapropel 5 (124 ka) was synchronous with extreme Blue Nile floods and the formation of the 386 m lake in the lower White Nile Valley, as well as with a prolonged wet phase in the eastern Sahara. Fluctuations in Nile flow and sapropel formation reflect the influence of the precessional cycle upon the East African monsoon.

Between 75 ka and 19 ka the climate in the Nile headwaters region became progressively colder and drier. During the Last Glacial Maximum, Lake Tana in Ethiopia and Lake Victoria in Uganda became dry, flow in the White Nile was reduced to a trickle, and the Blue Nile and Atbara became highly seasonal bed–load rivers. The return of the summer monsoon at 14.5 ka ushered in extreme Blue Nile floods, widespread flooding across the Nile Basin and the formation of the 382 m lake in the lower White Nile Valley. There was a brief return to aridity during the Younger Dryas (12.8–11.5 ka), after which the climate again became wetter and widespread flooding in the Nile Valley resumed. The early Holocene floods were later followed by incision and creation of the modern relatively narrow flood plain.

Throughout the late Quaternary (and doubtless before then) the Nile has fluctuated between two extreme modes: a traction load river during times when the climate in its Ugandan and Ethiopian headwaters was cold and dry; and a suspension load river during warm wet intervals in these regions. These two contrasting scenarios represent either end of an environmental spectrum, with multiple variants in between.

Two as yet unresolved questions concern when the White Nile first joined the main Nile, thereby ensuring that the Nile became a perennial river, and whether smectite clays were present along the White Nile flood plain throughout its history. Answers to both questions are fundamental to correct interpretation of the marine sediment records preserved in the Nile Cone.

Acknowledgements*—For fieldwork help and advice, I thank Professor Osman et Tom (Khartoum), Neil Munro (Khartoum), Dr Yousif El Samani (Geological Research Authority of Sudan, Khartoum), Professor Yasin Abdl Salaam (University of Khartoum), Dr Donatella Usai and Dr Sandro Salvatori (Italian Archaeological Mission, Khartoum), Professor Matthieu Honegger (Swiss Archaeological Mission, University of Neuchâtel), Dr Derek Welsby (British Museum, London), Professor Geoff Duller (Aberystwyth University), Professor Jamie Woodward (University of Manchester),* *Professor Mark Macklin (Lincoln University), Professor Tim Barrows (University of Wollongong) and Dr David Haberlah (Sydney). I greatly appreciate the thorough, constructive and insightful comments on an early draft by Professor Eduardo Garzanti and Dr Carlo Mologni.*

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