Landscapes and Landforms of the Chobe Enclave, Northern Botswana

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Abstract

The northern part of the Chobe Enclave (an administrative district of northern Botswana) is an agricultural area situated between relatively pristine national parks situated in the Middle Kalahari Basin. It belongs to the Linyanti-Chobe structural basin and constitutes a syntectonic depocenter formed within a large structural depression, known as the Okavango Graben, a tectonic structure of a likely trans-tensional nature. The landscape includes fossil landforms, such as sand dunes, pans, sand ridges, and carbonate islands resulting from palaeo-environmental and palaeo-drainage changes through the Quaternary and associated to (neo)tectonic processes. In addition to river- and wind-reworked Kalahari sands, the sediments include diatomites and carbonate deposits, forming inverted reliefs and originating from palustrine palaeo-environments. The Linyanti-Chobe basin is at the convergence of several ecoregions from tropical and subtropical grasslands to savannas and shrubland biomes. The hydrological cycle in the northern Chobe Enclave is governed by a complex interplay between the Okavango, Kwando, and Upper Zambezi drainage basins, which originate from tropical watersheds of the Angolan highlands. Finally, the widespread development of termite mounds impacts the diversity of soils and sediments of the northern Chobe Enclave, which is also reflected in the vegetation.

Keywords

Trans-tensional basin • Palustrine and floodplain environments • Calcrete • Savana vegetation and soils • Termites • Linyanti-Chobe basin • Environmental change • Quaternary

6.1 Introduction

6.1.1 Where Is the Chobe Enclave and Why Is It so Interesting?

Landscape evolution in the Chobe-Linyanti basin is closely associated with active tectonics and surface processes. Sedimentary basins forming under extensional tectonic regimes are important records of geological history, among other types of depositional environments. The architectures of these basins and the basin-fill are influenced by the rheological structure of the lithosphere, the availability of crustal discontinuities that can be tensionally reactivated, the mode and amount of extension, and the lithological composition of pre- and syn-rift sediments (Ziegler and Cloetingh 2004). An important factor in controlling geomorphic processes during landscape evolution can be the topography of a region affected by recent and/or active processes such as faulting (Mayer 1986; Cox 1994; Bishop 2007; Giaconia et al. 2012a, 2012b). Therefore, it is crucial to report the underlying geology in order to understand the way landforms evolve.

The study site, as a part of the Middle Kalahari Basin of northern Botswana, is situated in the northern Chobe Enclave, an administrative district, close to the Chobe-Linyanti depocenter (Fig. 6.1). It is bounded by the Linyanti

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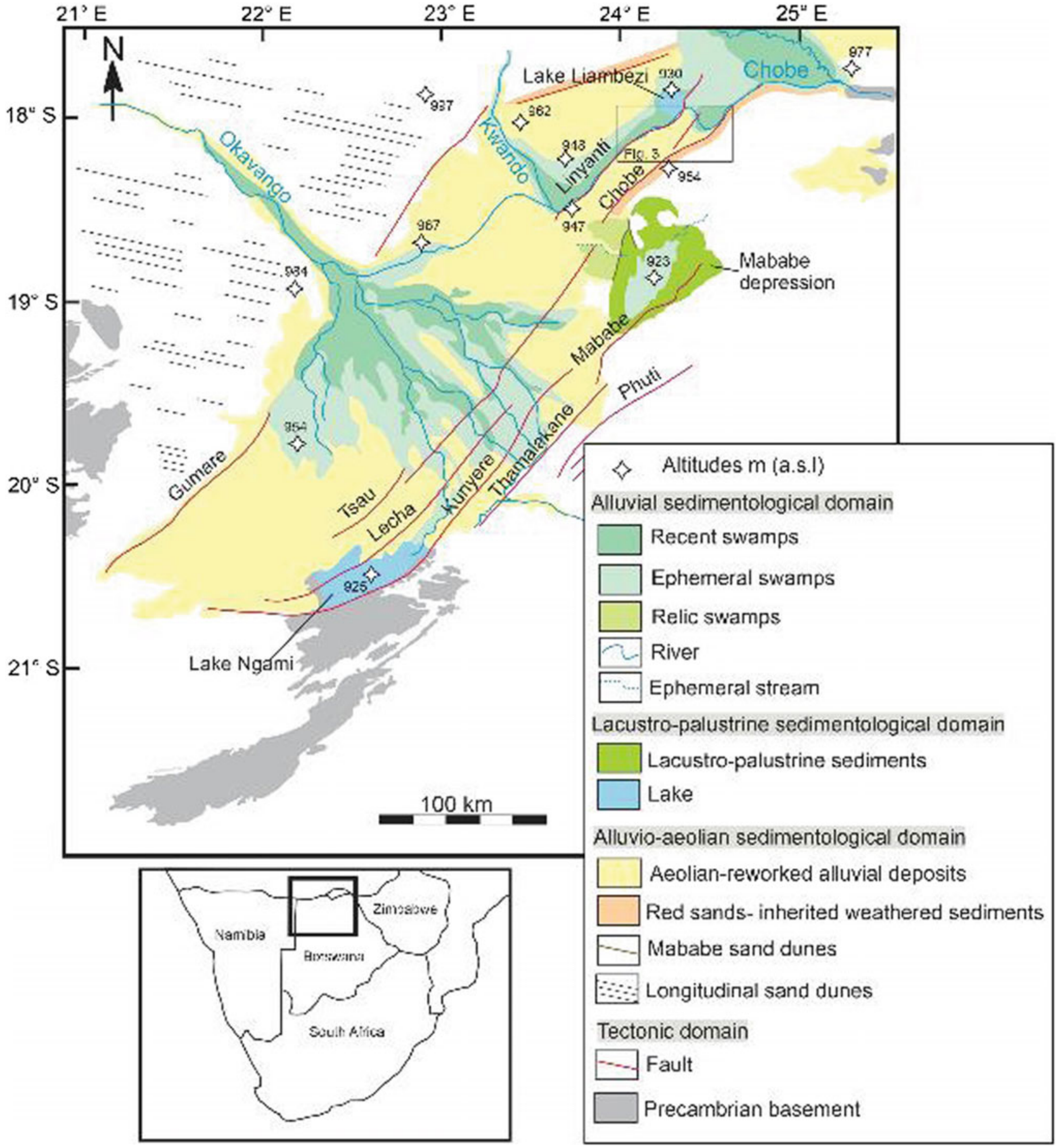

Fig. 6.1 Map of the sedimentary domains, Okavango region

river and Lake Liambezi in the north, bordering the eastern part of the Caprivi Strip, and the Savuti and Chobe rivers in the west and south, respectively. It is an agricultural area between relatively pristine national parks covered mostly by the Kalahari sands with unexpected occurrences of carbonate beds (Diaz et al. 2019). Equally so, the various types of sedimentary deposits (Fig. 6.1), including diatomites, fluvial, and aeolian sands, span the site. Palaeo-shorelines within the Middle Kalahari have been investigated as Quaternary landform relicts, which formed under different hydrological conditions (e.g., Burrough and Thomas 2008; Burrough et al. 2009a, 2009b; Shaw et al. 1988). Some investigations focused on understanding geomorphological features, such as sand ridges associated with lacustrine systems, namely, the Mababe Depression, Lake Ngami, and the Makgadikgadi Pan (Grey and Cooke 1977; Grove 1969; Shaw 1985). Understanding the timing of formation of these sand ridges has been helped by thermoluminescence (TL) and optically stimulated luminescence (OSL) dating techniques (e.g., Cooke and Verstappen 1984; Shaw 1985; Burrough et al. 2007; Burrough and Thomas 2008; Ringrose et al. 2008; Burrough et al. 2009a, 2009b). Burrough and Thomas (2008) reported OSL ages of sand ridges from the Mababe Depression and Lake Ngami consistent in ages with sand ridges in the northern Chobe Enclave. However, given the nature of emplacement of these sand ridges and associated sedimentary deposits, such as carbonate beds (Diaz et al. 2019), and the complexity of interpreting the emplacement of geomorphological features, some questions remain unanswered, such as:

- What is the origin of the sediments forming the sandy landforms?
- What is the origin of the carbonate deposits?
- What depositional environments and landscapes are recognized?
- What could be the main processes at work in this landscape?

The area is also characterized by a large diversity of soils and vegetation (Romanens et al. 2019; Vittoz et al. 2020). The region covered in this study is unique as it mirrors some aspects of the inland alluvial fans of the Okavango and Kwando rivers, as geologic and tectonic processes extend into this arid to semiarid region, hence establishing a regional-scale and even local-scale zoning of the same large-scale vegetation zones adjacent to the subtropical swamps/flooded areas. The Linyanti-Chobe basin, in particular, is at the convergence of several ecoregions that are all part of the "tropical and subtropical grasslands, savannas and shrubland" biome (www.worldwildlife.org/biomes/tropical-and-subtropical-grasslands-savannas-and-shrublands). It includes the Zambezian flooded grasslands, the Zambezian and mopane woodlands, the Kalahari *Acacia-Baikiaea* woodlands, and the Kalahari xeric savannah. The Linyanti-Chobe basin flora thus consists of a mix of species from these contrasting, neighboring ecosystems. Finally, the hydrological cycle in the Chobe Enclave is governed by a complex interplay between the Okavango, Kwando, and Upper Zambezi drainage basins. In addition to local rainfall, it receives waters from the Okavango, Kwando-Linyanti, and Zambezi-Chobe River systems, which drain tropical watersheds in the Angolan highlands.

6.1.2 Topography and Climate of the Chobe Enclave

Tectonics influenced both the topography and hydrologic conditions of the study region, possibly altering palustrine/shallow lacustrine and sedimentary budgets. The Middle Kalahari is a semiarid environment with a record of wetter climate in the past (Cooke 1975; Shaw and Cooke 1986), and depositional environments that preserve a complex history of Quaternary environmental change (Thomas and Shaw 1991, 2002). Today, northern Botswana is classified as a hot semiarid (steppe) climate (BSh) (Peel et al. 2007). This type of climate is characterized by a mean annual temperature above 22 °C, low mean annual precipitation of which 90% falls between October and April (Skarpe et al. 2014). The northern Chobe Enclave district (average precipitation of 650 mm; Jones 2002) has a wet season from November to March with average minimum and maximum temperatures of 20 °C and 32 °C, respectively, and a dry season from April to October with average minimum and maximum temperatures of 13 °C and 30 °C, respectively.

6.1.3 Hydrological Setting of the Chobe-Linyanti Basin

As noted above, this region is bounded by the Linyanti river and Lake Liambezi in the north and the Savuti and Chobe rivers in the west and south, respectively. These bounding rivers are part of a complex drainage system that includes all three of the regionally important rivers of the Okavango, the Kwando, and the Zambezi (McCarthy et al. 2000).

The headwaters of the catchments to the Okavango, Kwando, and Zambezi are characterized by a humid, subtropical climate with a mean annual precipitation of up to 1300 mm, decreasing down to about 570 mm in the region of the Chobe Enclave district. The rainy season in the headwater region, as well as the study region, is commonly between November and March (Mizlow et al. 2009) and the intensity depends critically on fluctuations in the positioning of the Intertropical Convergence Zone (e.g., McCarthy et al. 2000). The discharge of the three river systems, along their course which is individually >1800 km, provides a maximum flooding to their respective distal wetlands, towards the south, with a delay of about 6 months. The Okavango's and Kwando's maximum floods occur from June to August (e.g., Pricope 2013), while the Zambezi has an earlier annual flood pulse that may spill over into the Linyanti-Chobe wetlands through the Bukalo Channel, as well as the Chobe River. The flooding in the Linyanti-Chobe basin thus peaks between March and May, a few weeks after the Zambezi reaches its maximum discharge (Tweddle and Hay 2011; Pricope 2013). The Okavango waters penetrate the Linyanti-Chobe basin only during exceptional floods via the Selinda spillway. The relative exchange from the Zambezi and/or the Okavango/Kwando into the Linyanti-Chobe area may well have been more important during the past humid periods, and the existence of a larger palaeo-lake during the Quaternary is postulated (Burrough and Thomas 2008), as supported by inherited geomorphological features in the landscape.

The complexity of the drainage system in the Chobe Enclave is related to uplift along the Okavango-Kalahari-Zimbabwe axis and extension of the East African Rift system during the Cretaceous (Moore and Larkin 2001; Tooth et al. this book). The Kwando River has been deviated from the Zambezi due to faulting along the Linyanti fault, creating the endoreic systems and resultant wetlands of the Linyanti river and Lake Liambezi. The southern bounding fault of the Chobe, in turn, separated the Kwando from the Chobe River, together with movements along the Linyanti fault, also limiting the Linyanti overflow into the Chobe River. Instead, the Chobe River and its wetlands may now receive backflow waters from the Zambezi (Moore and Larkin 2001; Tooth et al. this book). However, depending on the relative amounts of precipitation in the Okavango, the Kwando, and finally also the Zambezi catchments, additional water in the Linyanti-Liambezi basin may enter via the Selinda spillway from the Okavango drainage basin and/or via the Bukalo Channel from the Zambezi drainage basin (McCarthy et al. 2000). Depending on the water levels in the Linyanti and Lake Liambezi drainage system, the Chobe River serves as the major surface outflow (Seaman et al. 1978; Burrough and Thomas 2008; Kurugundla et al. 2010; Peel et al. 2015) and finally returns the waters through a confluence with the Zambezi River further south.

6.2 Geological Settings

6.2.1 Structural Context

The northern Chobe Enclave district forms part of the Linyanti-Chobe sub-basin, bounded by the Linyanti and Chobe faults (Fig. 6.1). The Linyanti-Chobe sub-basin together with the Ngami and Mababe basins are the three characteristic syntectonic depocenters (Kinabo et al. 2007) within the large structural depression known as the Okavango Graben. This structural depression is made up of a series of NE-SW trending normal to dextral strike-slip faults (Modisi et al. 2000; Campbell et al. 2006; Bufford et al. 2012). The bounding faults of the Okavango Graben in the southern part form an *en-echelon* pattern with a direction following the strike of the Precambrian basement structures (Mallick et al. 1981; Moorkamp et al. 2019). The Okavango Graben has been associated with the formation of the southwestern branch of the East African Rift system (Modisi et al. 2000; Alvarez and Hogan 2013); however, Pastier et al. (2017) argued that the tectonic structure of the Okavango Graben better fits a trans-tensional basin model. Trans-tensional or pull-apart basins are topographic depressions that form at releasing bends or steps in a basement strike-slip fault system, with basin margin characterized by the developments of *en-echelon* oblique—extensional faults that soft- or hard-link with increasing displacement in the principal displacement zones (Wu et al. 2009). Kinabo et al. (2008) coupled analysis of Shuttle Radar Topography Mission (SRTM), Digital Elevation Model (DEM), and aeromagnetic data in the Chobe region and revealed the development of soft linkage on segments of the Linyanti fault and evidence of a hard linkage forming between two *en-echelon* right-stepping segments of the Chobe fault.

6.2.2 Bedrock Geology

The Ghanzi-Chobe basin in northwestern Botswana consists of a linear belt of volcano-sedimentary sequences that were deposited during Meso-Neoproterozoic times following extension tectonics associated with the Namaqua orogeny (Modisi et al. 2000). This basin was subsequently deformed as part of the continent-wide tectonic event, the Pan-African Damara orogeny, resulting in the inversion of the Proterozoic volcano-sedimentary basins and formation of a fold belt along the northern margin of the Kalahari Craton (Modie 1996). Sediments of the Mesozoic Karoo Supergroup then became part of this fault-bounded graben system (Bordy et al. 2010). The giant Okavango dyke swarm cut across the basin during this time (Le Gall et al. 2002). Significantly, this sequence in northwestern Botswana was further affected by the reactivation of ancient faults in Pan-African belts, resulting in the Proterozoic NE-SW strike (Pastier et al. 2017). Thus, the Ghanzi-Chobe basin in northwestern Botswana preserves an important record of subsequent tectonic and depositional events that brought together two prominent southern African cratons—the Kalahari and Congo cratons.

6.2.3 Surficial Sedimentary Geology

The Kalahari sands mainly dominate the sediments observed at the ground surface of the area. They outcrop as the upper part of a sedimentary body of more than 200 m in thickness

(Thomas and Shaw 1991). Two other sedimentary units, not documented in the region, have also been identified: continental carbonate landmasses (described as ``islands" in this chapter) and diatomites.

First, X-ray diffraction (XRD) analyses confirm the almost pure quartz nature of the sands in the region. They are usually extremely well sorted and are present in all types of sediments and soils, whatever their nature (Fig. 6.2). They obviously belong to the large body of the Kalahari sands (Table 6.1). They are also found associated with finer fractions, likely originating from (i) a fine loess fraction and/or dust (Crouvi et al. 2010), (ii) diatoms, or (iii) pedogenic redistribution and neoformation, the latter including the role of termites (see below; Jouquet and Lepage 2002; Jouquet et al. 2011). Their ubiquity, and their mostly aeolian and fluvial reworked nature at the surface of the landscape (Thomas and Shaw 1991, 2002), make these sands suitable for OSL dating (see below).

Second, carbonate formations in the region appear as limited landmasses identified as "islands" in the sand sea (Fig. 6.3). They have variable thicknesses but are frequently thicker than 1 m (Figs. 6.4 and 6.5e–g). In the literature, these carbonate layers are usually defined as "calcrete". Although this term refers to a large variety of processes explaining their formation (and consequently palaeo-environments; Wright, 2007), their genesis has been attributed in Botswana to a combination of pedogenic, groundwater, and geochemical-diagenetic processes (Nash et al. 1994; Nash and Shaw 1998; Ringrose et al. 2002, 2014). The beds, hardened to various degrees, include all the features observed in palustrine limestones at the macroscale (e.g., traces of roots, burrows and/or galleries, shells,

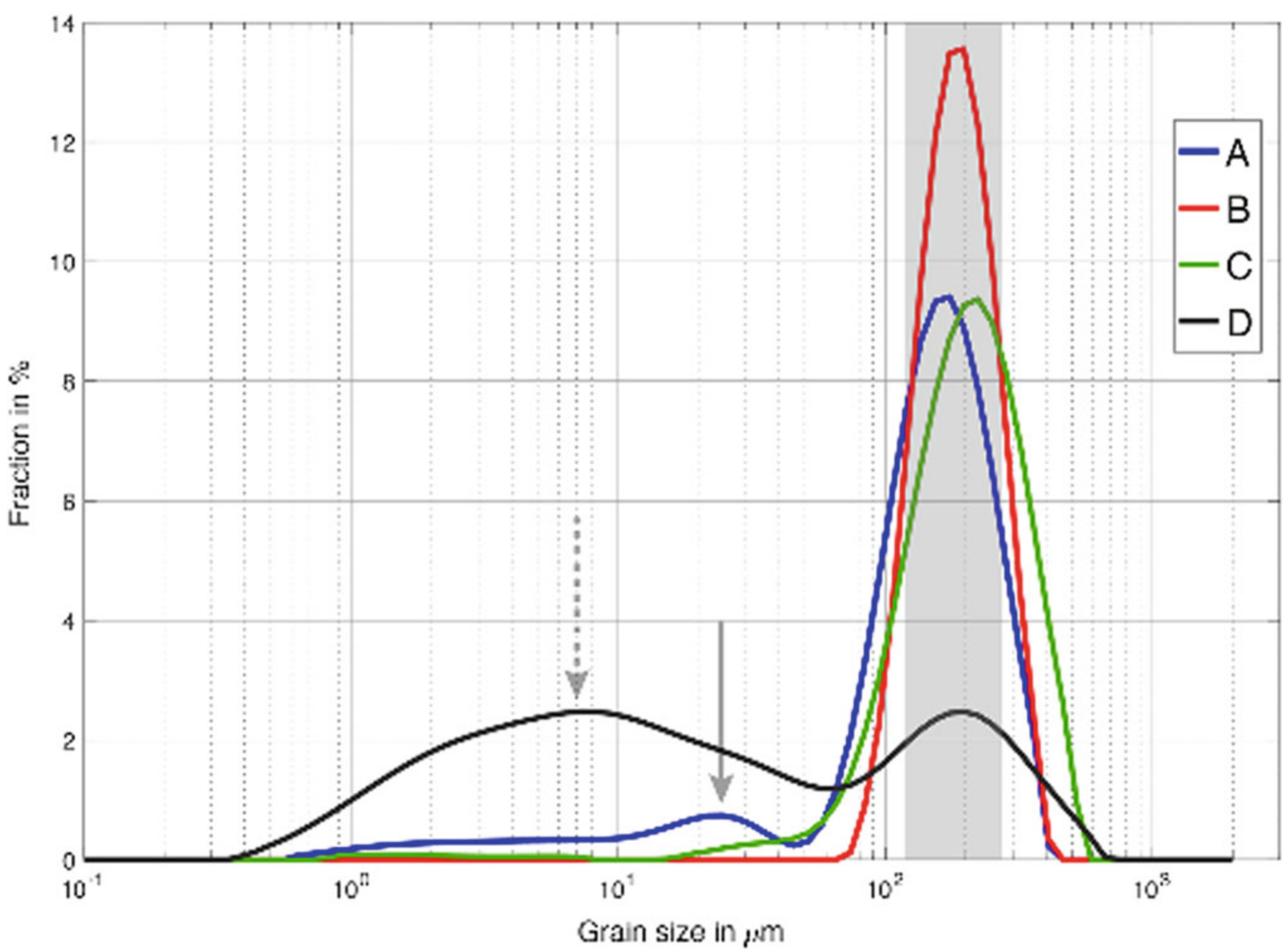

Fig. 6.2 Four examples of grain-size distributions from the Chobe Enclave: the Kalahari sand mode is always displayed (shaded area centered on 200 µm). In addition, two modes of fine particles are detected and likely correspond to fine peri-desertic loess (solid arrow at 30 µm) and/or dust/very fine silt due to termite activity and/or diatomite (dashed arrow at 8 µm; see Crouvi et al. 2010 and Sect. 4.3). **A** Sandy and homogeneous layer (5.40 m deep), Baobab pit (18°06′10″ S, 24°18′30″ E, 935 m asl). **B** Recent fluvial sand from the Linyanti (18°04′30″ S, 24°08′43″ E, 937 m asl). **C** Recent aeolian sand, east plateau of the Linyanti river (18°10′16″ S, 24°14′54″ E, 936 m asl). **D** Baobab pit, upper layer (soil)

Table 6.1 Average grain sizes of the Kalahari sands from various locations

Locations	Average sizes (μm)	Authors
Kalahari dune field	125–220	Cook (1980)
Namib sand sea	200	Lancaster (1981)
South West Kalahari	190–240	Lancaster (1986)
Kalahari dune crest	175	Watson (1986)
Northern Botswana	200	Thomas and Shaw (1991)
Central Kalahari (Ghanzi)	170–225	Wang et al. (2007)
Central Kalahari (Tshane)	210–230	Wang et al. (2007)
Namib sand sea	170–340	Crouvi et al. (2010)

paleosol imprints) and at the microscale (e.g., desiccation features, redistributions in the micromass and nodulization, multiple vadose and phreatic phases in cements; see Freytet and Verrecchia 2002; Verrecchia 2007). Consequently, although designated as "calcretes" in the literature (sensu Wright and Tucker 1991), the carbonate beds from the northern Chobe Enclave district could also be described as palustrine limestones. Their position as low relief features in the landscape are due to an inverted relief, emphasizing their inherited nature (Diaz et al. 2019). This raises the question "Is the palustrine origin of these limestones likely?". Fine carbonate deposits have been observed in the Okavango Delta (McCarthy et al. 2012) in island-mounds forming within the Delta. However, these tree-covered islands do not accumulate significant amounts of calcium carbonate, only traces as subsurface crystals. Their formation is related to the chemistry at the interfaces between fresh groundwater flow and trapped saline groundwater through tree uptake and transpiration (McCarthy et al. 2012), with dust contributing substantially to the material found on the surface of islands (Humphries et al. 2014). If such a process cannot generate large calcium carbonate accumulations, it is for the simple reason that the Okavango waters are presently extremely poor in Ca^{2+}, i.e., 3–7 mg/L on average, with a conductivity of 33–40 μS cm^{-1} (Dr. Mogobe, Okavango Research Institute, Maun; personal communication, spring 2016). But one can imagine a palaeo-river system significantly enriched in Ca^{2+}, in which the same processes at work during large floods could have led to some of the fossil carbonate deposits outcropping today. They could correspond to large ponds and marshes and would constitute precious palaeo-climate and environmental indicators (Diaz et al. 2019; see below) to reconstruct the relationships between the Zambezi-Chobe system and the southern palaeo-lakes observed by Burrough and Thomas (2008). Presently, most of these outcropping carbonate deposits are karstified and eroded under the present-day climate and weathering conditions.

Third, diatomites are easily recognized in the sediments: they form a fine, white material, with a very low density and are sometimes so soft that they can be crushed to a powder in the field. X-ray fluorescence (XRF) analyses confirmed their pure siliceous nature (SiO_2). The presence of quartz, probably of aeolian origin, is also observed using scanning electron microscope (SEM). However, in the study region, diatomites can also be associated to calcite as vein, crack, fracture, and tubular infillings (Fig. 6.6). Such pure diatomites can only form in ponds or shallow lakes, in a range of water salinities, from freshwater to slightly brackish. They must correspond to periods during which wetter conditions prevailed. Their presence can correspond to periods similar to what has been suggested by Burrough and Thomas (2008), i.e., "that increased flow in the Chobe and Zambezi system significantly contributed to the Middle Kalahari lake phases" for the palaeo-lakes observed in the south and east of the Chobe. It can be hypothesized that their alternation with the carbonate deposits indicates that the Chobe underwent fluctuations in the hydrochemistry of water sources, as well as in climatic conditions. But this theory needs to be properly assessed and documented by further studies.

6.2.4 Insights into the Quaternary in the Chobe-Lynianti Region

6.2.4.1 Quaternary Palaeo-Lakes and the Chobe-Lynianti Region

The endorheic nature of the basin has led to aeolian, fluvial, and/or lacustrine sedimentary accumulations since the Cretaceous (Grove 1969). Regarding the Quaternary palaeodrainage history, it has mostly been reconstructed through geological and geomorphological archive studies (e.g., Thomas and Shaw 2002; Burrough et al. 2007; Moore et al. 2012). Fossil landforms, such as sand dunes, pans, sand ridges, and carbonate islands (Diaz et al. 2019; Fig. 6.3), testify to changes in hydrological conditions. However, the present-day semiarid climate leads to the deflation of sedimentary archival deposits and does not favor the preservation of organic proxies (Thomas and Burrough 2012). Regional palaeo-environmental records are thus fragmentary

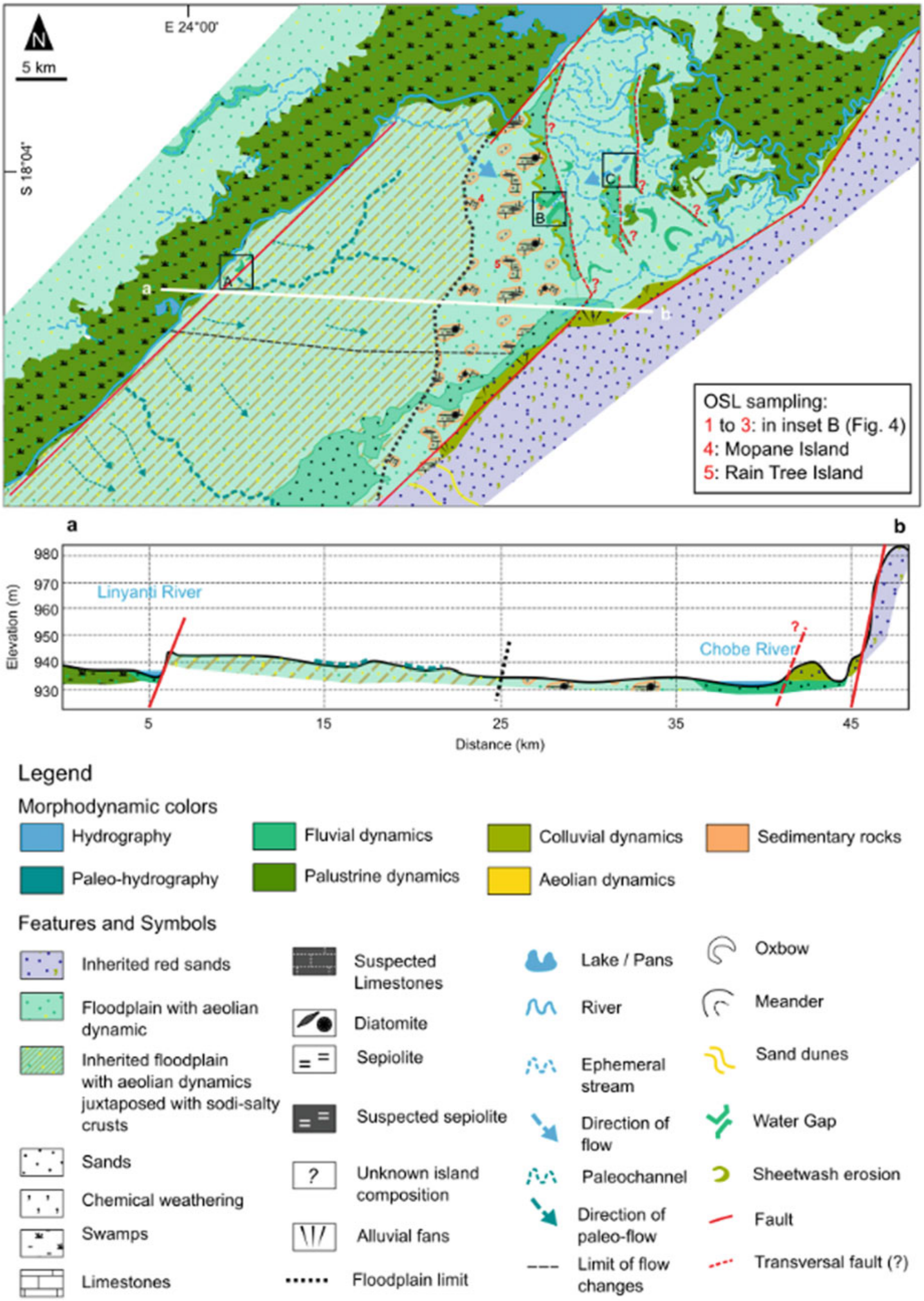

Fig. 6.3 Geomorphological map from the Chobe Enclave and elevation profile through the Chobe Enclave

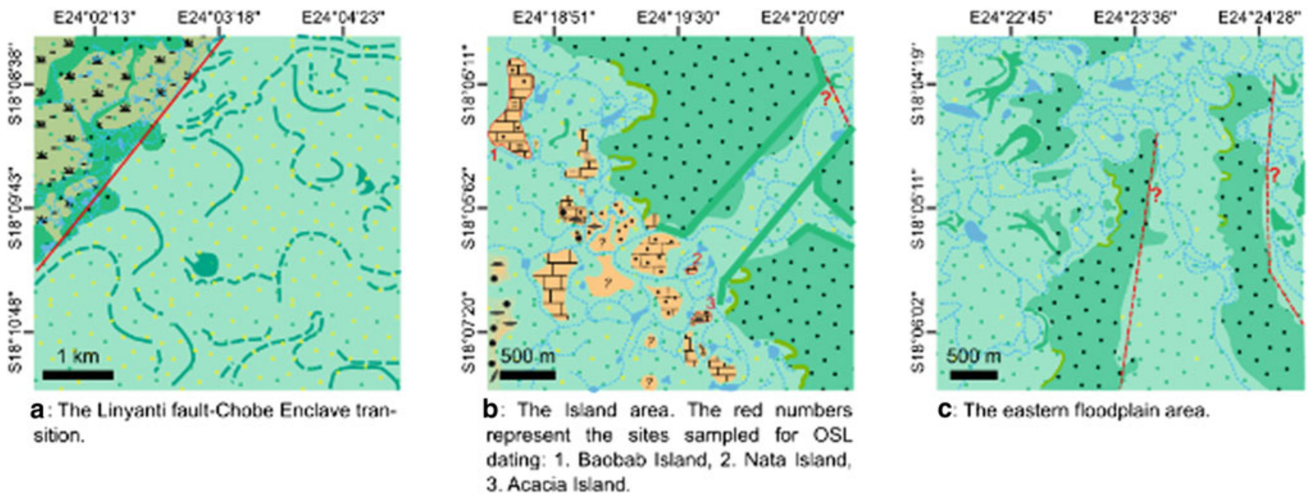

Fig. 6.4 Detailed geomorphological maps in three key areas from west to east (see Fig. 6.3 for location). **a** Palaeo-floodplain, **b** the island area, **c** the sand ridge area. The red numbers correspond to the sites where samples were taken for OSL measurements. Red question marks refer to potential transversal faults. See legend in Fig. 6.3

and difficult to assess (e.g., Cordova et al. 2017), making interpretations sometimes invalidated, particularly regarding influences of climate change and/or tectonic events on the palaeo-drainage (Moore et al. 2012; Bäumle and Himmelsbach 2018).

In the Okavango region, extensive geomorphological evidence for large palaeo-lake exist, referring to wetter climate periods (Grove 1969). Most previous research has focused on palaeo-lakes in this region, notably Lake Ngami (Huntsman-Mapila et al. 2006), Lake Mababe (Gamrod 2009), and Lake Makgadikgadi (Thomas and Shaw 1991). Burrough and Thomas (2008) proposed that these palaeo-lakes might have coalesced into a single large Mega-Kalahari Lake during previous humid periods. OSL dating of the main palaeo-lake shorelines (beach ridges) in the Okavango region suggested that high lake levels occurred during the Holocene and Late Pleistocene at 8.5 ± 0.2 ka, 17.1 ± 1.6 ka, 26.8 ± 1.2 ka, 38.7 ± 1.8 ka, 64.2 ± 2.0 ka, 92.2 ± 1.5 ka, and 104.6 ± 3.1 ka (Burrough and Thomas 2008; Burrough et al. 2009a, 2009b).

However, it seems that tectonic processes also played a critical role in the drainage evolution. Moore et al. (2012) reviewed the Plio-Pleistocene history of the main rivers in the region belonging to the Congo Basin and the Upper Zambezi Basin. They proposed a model in which the general drainage conditions resulted in smaller lakes, whereas larger lakes were controlled by climate feedbacks, such as the rainfall and evaporation balance. The authors recognized different palaeo-lake systems: Palaeo-Lake Deception with a fossil sand ridge at ~995 m (McFarlane and Eckardt 2008), Palaeo-Lake Makgadikgadi with fossil sand ridges at 945 m (Thomas and Shaw 1991; Burrough and Thomas 2008), and Palaeo-Lake Thamalakane with fossil sand ridges at 936 m, 920 m, and 912 m (Grey and Cooke 1977). They proposed one mega-lake (Palaeo-Lake Deception and after Palaeo-Lake Makgadikgadi) during the Early Pleistocene (<500 ka) that progressively contracted, rather than oscillated, due to a reduction of inflow from the tributaries of the Upper Zambezi palaeo-river system that were cut off by tectonic subsidence and lifting, inducing headward erosion and river capturing (Bäumle and Himmelsbach 2018). The decrease of water inputs from the Boteti River controlled by the fault system at the foot of the Okavango Delta caused the Palaeo-Lake Thamalakane to shrink, indicated by a shoreline decreasing from 920 to 912 m between 300 and 100 ka. The 912 m lake would have desiccated in the last 100 ka (Moore et al. 2012).

6.2.4.2 The Carbonate Islands from the Chobe-Lynianti Region

Carbonate islands of various sizes and surrounded by ephemeral waterbodies are a common feature of the landscape (Figs. 6.3 and 6.4; Diaz et al. 2019). The precise origin and timing of such landforms remain unclear, but they are hypothesized to have been formed in response to regional palaeo-hydrological changes, making them interesting archives for the region. Different successions of sedimentary beds are exposed in quarries (Figs. 6.5e–g and 6.6). The Baobab Island (site 1, Fig. 6.4) is one of these quarries opened in an island. The other studied islands (sites 2, 3, 4, and 5, Figs. 6.3, 6.4 and 6.7) refer to outcrops; the sands below the carbonate layers are described from pits dug in the alluvial plain surrounding the carbonate island, as illustrated in the supplementary materials from Diaz et al. (2019).

The Baobab Island quarry is 5 m deep and the only place where the Kalahari sands can be reached directly below the carbonate beds. These sands have a thickness of 1.30 m

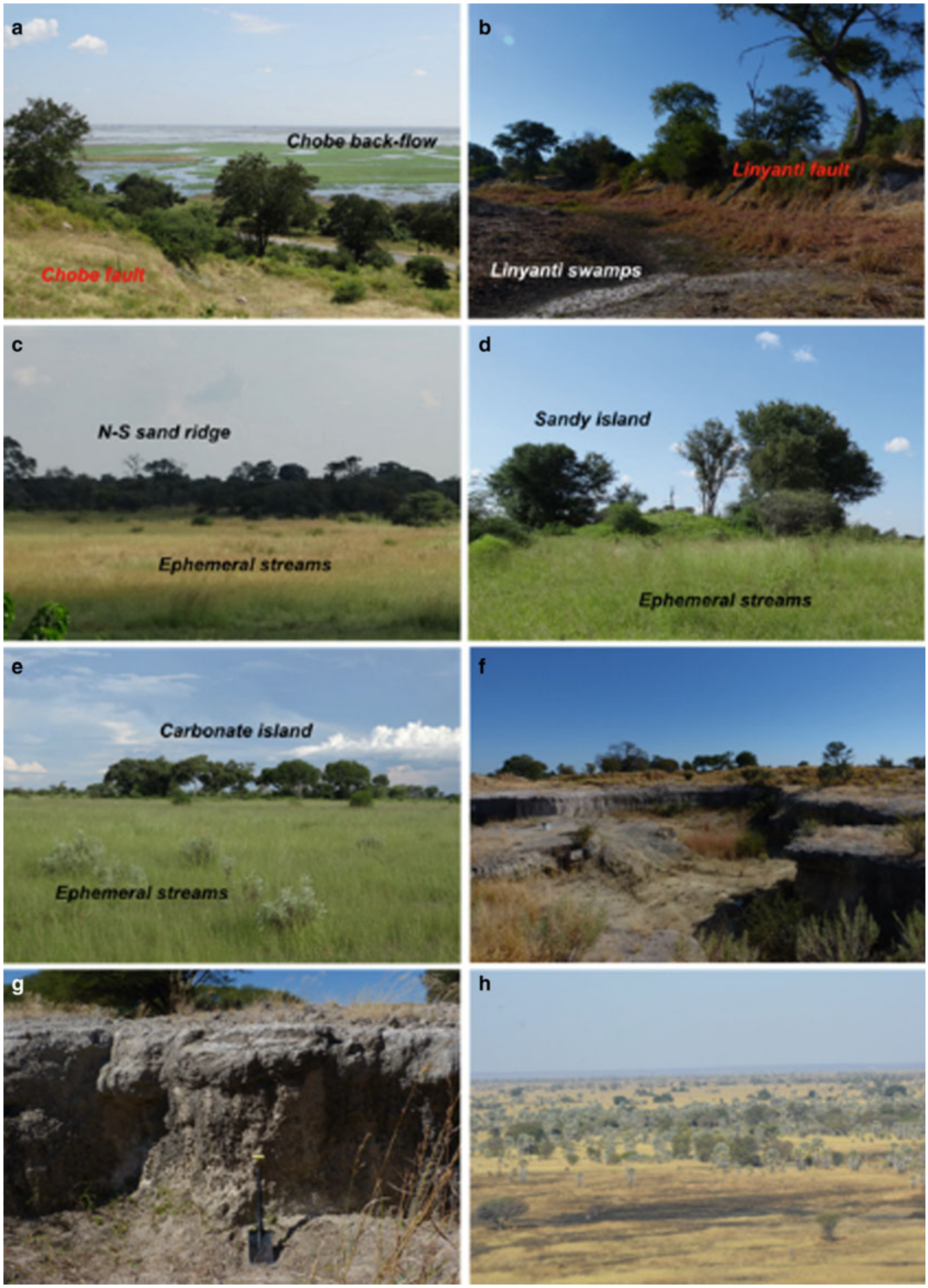

Fig. 6.5 Different landscapes and landforms from the Chobe Enclave. **a** View from the Chobe fault near Katchikawe on the Chobe backflow, **b** view from the Linyanti swamps toward the Linyanti fault during the dry season, **c** view on the western side of the N-S sand ridge, slightly elevated with tree vegetation and surrounded by ephemeral streams covered with grass, **d** Sandy island covered with trees and surrounded by ephemeral streams covered with grass, **e** view of a carbonate island (Diaz et al. 2019) slightly elevated, covered with trees and surrounded by ephemeral streams covered with grass, **f** Baobab quarry view with exposure of the carbonate beds (see location on Fig. 6.4b), **g** Carbonate layer from another quarry, Acacia pit (see location on Fig. 6.4b), where a diatomitic layer was also observed (Fig. 6.6), **h** view on the eastern floodplain with small N-S sand ridges covered by trees and surrounded by ephemeral streams covered with grass during the dry season

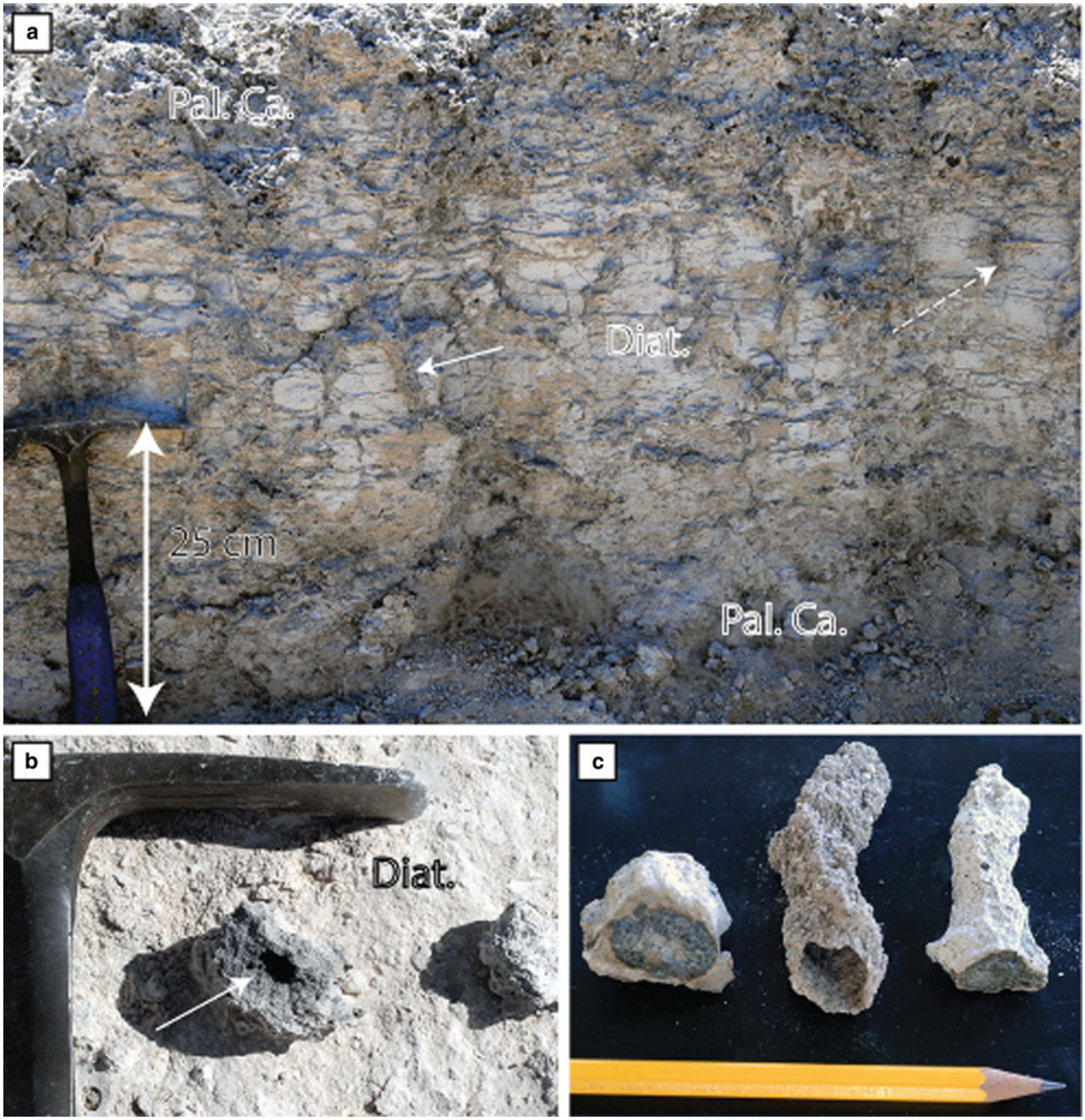

Fig. 6.6 Field photographs showing **a** the outcrop of a diatomite layer (Diat.) between two palustrine carbonate deposits (Pal.Ca.) at Acacia quarry (18°07′03″ S, 24°19′31″ E, 933 m asl; see Fig. 6.5g). The age of such deposits is still unknown. There are some diagenetic features, such as burrows or galleries (solid white arrow) and craze planes and fractures (dashed arrow), both infilled with $CaCO_3$. **b** Close-up of a carbonate-rich burrow or gallery inside a diatomite layer. **c** Comparison between present-day termite gallery (sample in the center) and fossil features found in diatomites. All these hand specimens are cemented by Ca-carbonate

(Fig. 6.5f). They are enriched in secondary silica concretions, as nodules or rhizoliths. The sands become carbonate-rich just below, and at the contact, with the overlying carbonate layer with a well-marked boundary between the sand-rich layers and the carbonate bed. The bottom part of the carbonate bed includes few small carbonate nodules, increasing in size and abundance toward the surface. Siliceous nodules display the same distribution as the carbonate ones. The uppermost 1.70 m of the pit comprises a hard carbonate bed without any apparent nodules.

The pit from site 2 (3 m deep) was observed next to Nata Island (Fig. 6.4b) located between the northern and the

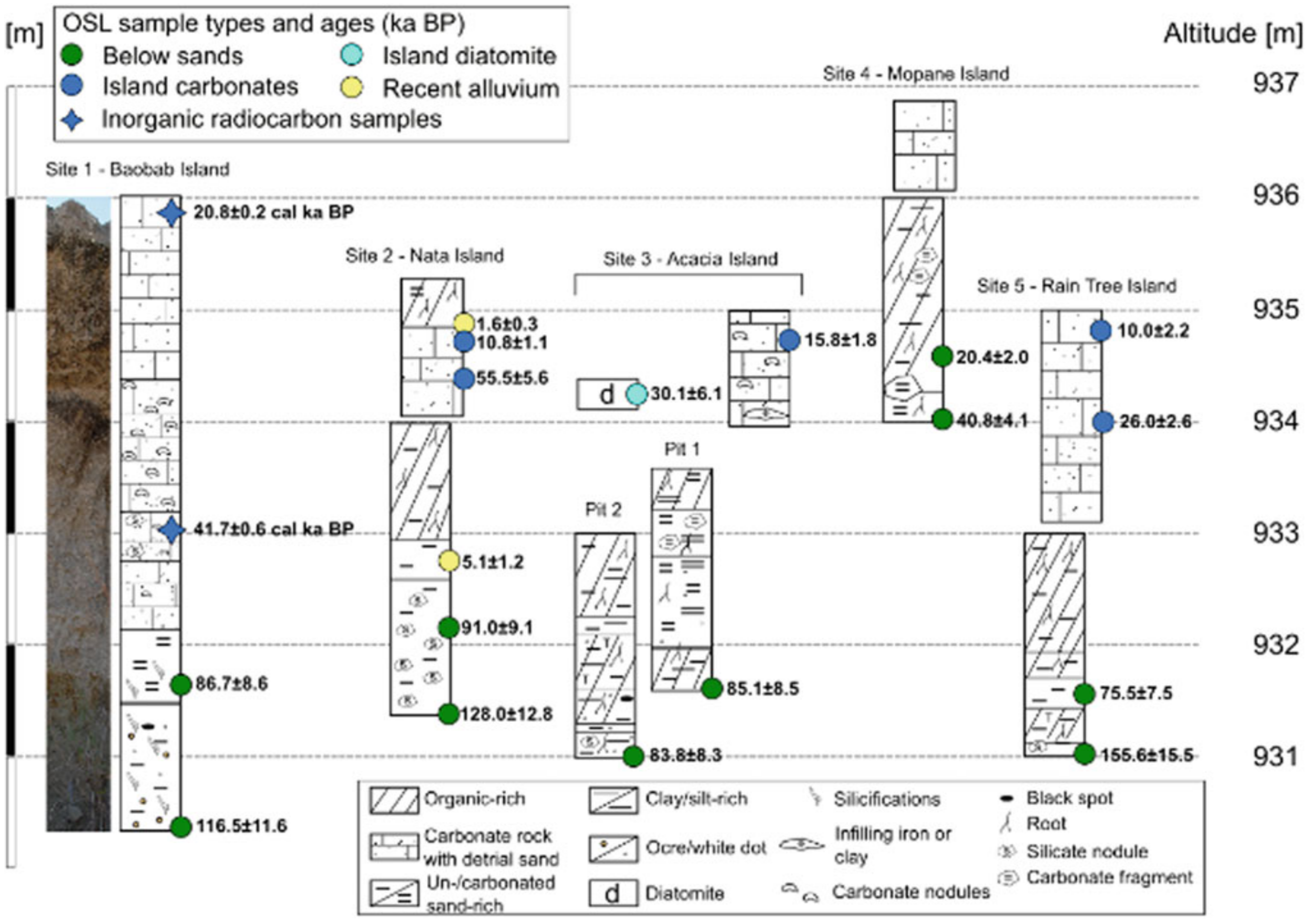

Fig. 6.7 Descriptions of the pits and/or soil profiles in each site represented in Figs. 6.3 and 6.4b, with OSL (circles) and radiocarbon (four-branches stars) ages of the different key sedimentary layers

southern N-S sand ridges. The elevation of the island compared to the surrounding sediments is about 2.5 m. A sandy layer, about 1.60 m thick and enriched in siliceous nodules, forms the bottom part of the pit. It is overlaid by a 40 cm-thick grey sandy layer without any siliceous nodules. An organic-rich sandy layer composes the upper part of the outcrop. The carbonate rock from the bed shows root and vegetation tissue debris and dendritic Fe–Mn features.

Two 2 m-deep pits were dug at site 3 located next to the most southern N-S sand ridge (Fig. 6.4). The first pit is just next to Acacia Island. Its bottom part is mostly an organic-rich and carbonate-free sand 40 cm in thickness. It is overlaid by a carbonate-rich sandy layer 1.10 m thick, without any organic matter. The uppermost part, 30 cm thick, is organic- and carbonate-rich. The second pit is located within the alluvial plain surrounding Acacia Island. The bottom part is a grey sand-rich layer 20 cm thick with abundant siliceous nodules. It is overlain by a thin sandy layer, 10 cm thick, in turn overlain by a 80 cm organic-rich layer and then, a thin sandy layer (15 cm), the top of the outcrop capped with an organic-rich layer, 75 cm thick. The elevation of the carbonate island compared to the surrounding alluvial plain is about 2 m. The carbonate bed studied was 1 m thick. It is a nodular carbonate rock rich in iron/clay infillings. It seems to overlay a diatomite layer (30 cm thick), showing some dark and carbonate veins forming a grid pattern (Fig. 6.6).

Another 2 m deep pit was dug next to Mopane Island located 10 km from the N-S sand ridge in the W (site 4, Fig. 6.3). The elevation of this island compared to the surroundings is about 0.5 m. The pit is characterized by a carbonate-rich layer, 20 cm thick, with small roots and a dominance of a sand size fraction ($\sim$70%). The layer above is 1.80 m thick and is organic-rich, without any carbonate in the fine fraction, but includes carbonate blocks and gravels. The carbonate rock fragments from the bed are very similar to the Nata Island carbonate layer and comprise tissue debris of roots and plants as well as dendritic Fe–Mn features.

A 2 m deep pit was dug next to Rain Tree Island located 10 km from the N-S sand ridge in the W (site 5, Fig. 6.3) in the surrounding alluvial floodplain. The elevation of the carbonate bed compared to the floodplain is about 2.5 to

3 m. The lowest part is a white sandy layer (10 cm thick) having abundant siliceous nodules. It is overlain by an organic-rich sandy layer (30 cm thick), in turn overlain by a grey sandy layer (20 cm thick). Above this light-colored sandy layer, there is a second organic-rich layer (20 cm thick) with an organic content increasing toward the upper part of the pit (1.10 m thick). The carbonate rock from the bed is very similar to those of Nata and Mopane Islands. Tissue debris of roots and plants, as well as dendritic Fe–Mn features, are present.

Thin section observations of carbonate rock samples suggest that the carbonate precipitation occurred in palustrine environments. A petrographic study showed that the quartz deposition in the palustrine environment would have been contemporaneous to the carbonate precipitation (Diaz et al. 2019). Consequently, to assess the timing of the carbonate island formation, optically stimulated luminescence dating (OSL) was used on quartz from key sedimentary layers (Fig. 6.5); (i) quartz from sandy layers situated below the carbonate islands (green circles in Fig. 6.5), (ii) quartz from sands trapped within the carbonate rock composing the island carbonate beds (blue circles), and (iii) quartz from sandy layers situated above the island carbonate beds (yellow circles). At Baobab Island, the ages of the carbonate rocks were determined using radiocarbon dating on the inorganic carbon fraction (four-branches blue stars). Finally, quartz extracted from the diatomite described at Acacia Island have been used for OSL dating (light blue circle). It can be hypothesized that OSL and radiocarbon ages from the key sedimentary layers may give an indication of the ages of the respective phases (i) the pre-palustrine carbonate, (ii) the palustrine carbonate, and (iii) the post-palustrine carbonate formations, respectively (Fig. 6.7).

Resulting ages show that the pre-palustrine carbonate formation phase would have occurred before 75 ka BP (before MIS 4). There are two younger samples from site 2 that deposited during MIS 2 (20.4 ka BP) and MIS 3 (40.8 ka BP). Were they already deposited during a different phase? The difference between the carbonate island and the top of the sand pit was only 0.5 m instead of 2 m for the other sites. Could this suggest that erosion was less intense in site 4? And that we are measuring alluvium sands possibly deposited during the palustrine carbonate formation phase? Indeed, the palustrine carbonate formation phase could have spanned from 55.5 to 10.0 ka BP (MIS 3 and MIS 2). Interestingly, quartz trapped in the diatomite are also from this phase (30.1 ka BP). It seems that the chemistry of waterbodies was not homogeneous. Finally, the post-palustrine carbonate phase would have occurred after 10 ka BP (MIS 1; Fig. 6.7, Diaz et al. 2019). Thus, it seems that, while lakes were forming in the Makgadikgadi Basin, palustrine environments were developing in the Chobe Enclave (Fig. 6.8). There are still questions pending regarding the origin of waters, e.g., were these palustrine areas forming before the onset of the activity of the Linyanti fault, allowing waters from the Linyanti to enter directly in the Chobe? Bäumle and Himmelsbach's (2018) work on the Linyanti aquifer proposes that palaeo-lakes along the Linyanti could have formed at the beginning of the extension of the graben system occurring during the Early and Middle Pleistocene, contemporaneously to Palaeo-Lake Makgadikgadi. The continuous development of the graben system during the Late Pleistocene led to the present-day Linyanti swamps. Consequently, the Chobe islands seem to be contemporaneous to the development of the graben system, cutting off more and more the inflow from the Linyanti through time, until a tipping point after which only occasional inflows prevailed, such as today (Fig. 6.8). Moreover, it seems that the water inflow would have been at least sufficient to sustain a palustrine environment and the formation of carbonate until 10 ka BP. After that, the carbonate-rich palustrine system dried out and started to be differentially eroded and solely recent alluvial sands were deposited (MIS 1; Fig. 6.8).

6.3 Geomorphology and Landforms of the Northern Chobe Enclave District

6.3.1 Importance of Hydrology in Shaping Geomorphological Features

The fault systems of the Linyanti and the Chobe control the hydrological settings in the study area. These settings are the main factors controlling the landforms in the enclave. Due to the uplifts linked to the faults, the general topographic slope in the Chobe Basin is <0.02% on average over long distances. The slope is decreasing from SW to NE, as well as from NW to SE. It increases again close to the Chobe strike-slip fault. This slope configuration induces the development of anastomosed river systems that are completely dry in the western part and become ephemeral to continuous going to the east. This hydrological gradient from W to E induces landforms associated with different morphodynamic processes found in three typical areas (Fig. 6.4) as described below: the western palaeo-floodplain, the island area, and the eastern floodplain area.

6.3.1.1 The Western Palaeo-Floodplain

The southwestern part of the northern Chobe Enclave district is characterized by an inherited floodplain composed by old fluvial sediments reworked by aeolian processes (Fig. 6.3). Numerous palaeo-channels are observable, having flow directions to the E in the northeastern part and to the SSW in the southwestern part. A soil and vegetation study of the area (Romanens et al. 2019) showed that there is a sodium-rich layer at the soil surface or under a layer of reworked aeolian

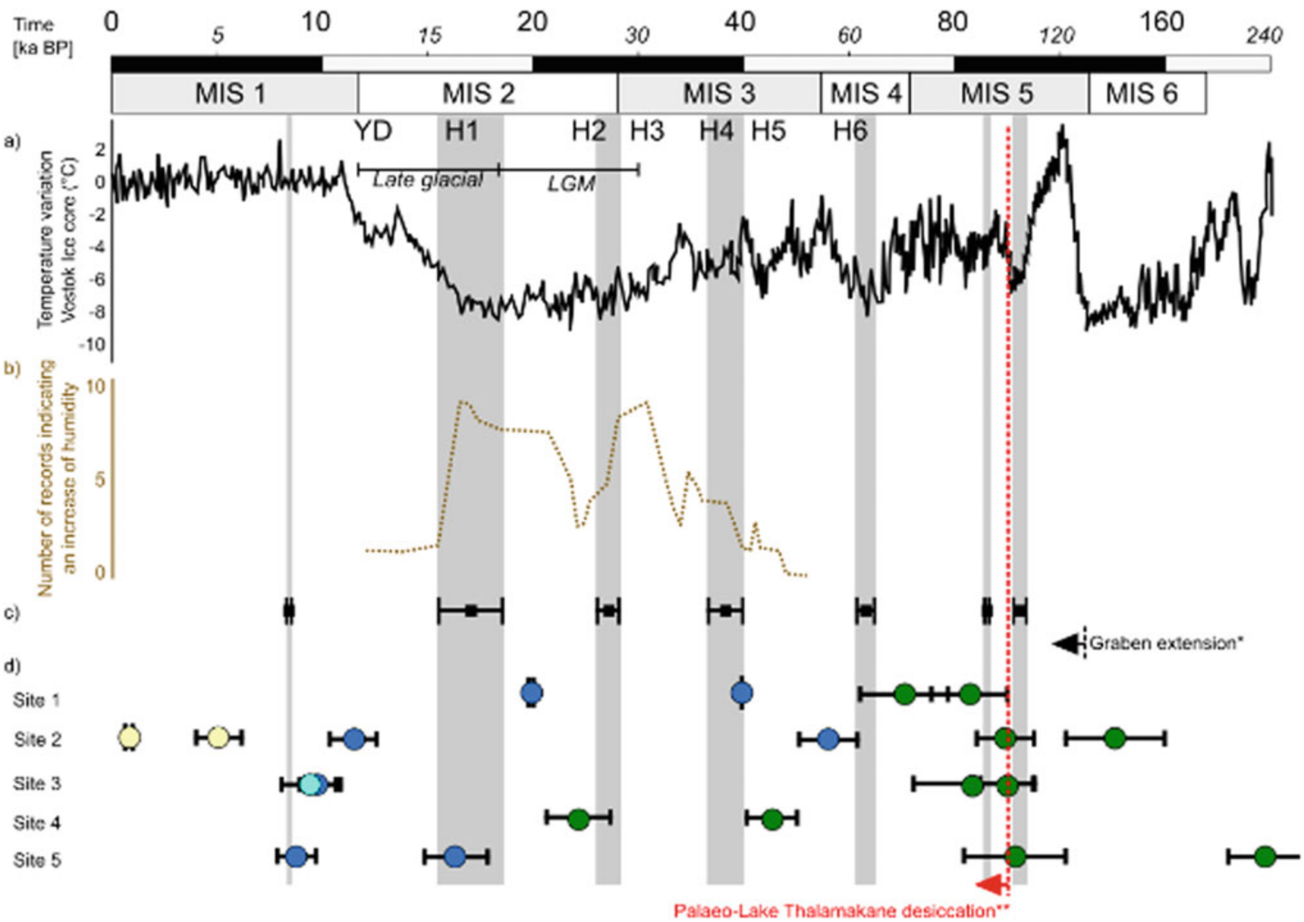

Fig. 6.8 Comparison of the carbonate island ages (circles and four-branches stars) with other types of archives. **a** Temperature variations assessed from ^{2}H in the Vostok Ice Core (Petit et al. 1999); data are available online (https://cdiac.ess-dive.lbl.gov/ftp/trends/temp/vostok/vostok.1999.temp.dat); **b** Numerous records in the Kalahari area are interpreted as an increase in humidity (Chase and Meadows 2007); the grey bars refer to lake highstands of the Mega lake Makgadikgadi (Burrough and Thomas 2008). *Graben extension in the Linyanti fault area (Bäumle and Himmelsbach 2018), **Thamalakane fault activation inducing the formation of the Okavango Delta and the desiccation of the Palaeo-Lake Thamalakane (Moore et al. 2012)

sands and between the palaeo-channels. Such a formation is probably inherited but its extent under the whole palaeo-floodplain remains unclear. Its origin can be attributed to a palaeo-hydrological system similar to what is observed today in the Okavango Delta. A detailed geomorphological map (Fig. 6.4a) shows the Linyanti swamps, and the inherited floodplain delimited by the Linyanti fault (Fig. 6.5b). The palaeo-channel configurations at this boundary suggest an older connection with the Linyanti river.

6.3.1.2 The Island Area

Lying between the palaeo-floodplain in the west and the floodplain in the east, the landscape of the island area (Figs. 6.3 and 6.5d–g) is dominated by the presence of islands the size of several hundred squared meters and a height of a few meters. Ephemeral river channels occasionally surround them and, when the water input is sufficient, small pans form (Fig. 6.4b). Sedimentary rocks of different natures form the islands, i.e., carbonate rocks, diatomite, sepiolite layers, amorphous silica-rich sands, and quartz sands (see above). Some of the islands have been documented (Diaz et al. 2019) but the composition of other islands can only be suggested (dashed lines in Fig. 6.4b) on the basis of field observations; finally, some have an unknown composition (marked with "?" in Fig. 6.4b). The succession of islands and channels forms an undulating landscape, which attenuates and finally disappears completely to the west (black dashed line in Fig. 6.3). In the east, a NNE-SSW sand ridge, around 20 km long and 2 km large, marks the limit with the eastern floodplain (Fig. 6.5c). It is composed by fluvial sands undergoing erosion due to sheet wash and forming triangular morphologies along their western side. On the eastern side, the sand ridge has a straight border possibly associated to an NNW-SSE transferal fault. The sand ridge is split into two halves by a river coming from the eastern floodplain, forming a water gap (Fig. 6.4b).

6.3.1.3 The Eastern Floodplain Area

In the eastern floodplain, smaller sand ridges, having the same characteristics as the one presented above, are present. These ridges are also crosscut by rivers forming sand patches that might have been attached to the ridges before (Fig. 6.4c). The number of ephemeral streams is increasing, and their anastomosed nature creates numerous oxbows. The water can accumulate in multiple pans, which can be connected. Further to the east of the sand ridges, permanent rivers are flowing from Lake Liambezi (from WNW to ESE) and from the Chobe River backflow (from NNE to SSW; Fig. 6.5 a, h). Large swamps are forming around their confluence and sand ridges counter the extent of these swamps in the west (Fig. 6.3).

Finally, two different landscapes are found at the borders of the area. In the north, the Linyanti swamps form around the Linyanti river, flowing along the fault from the Linyanti delta, towards Lake Liambezi. In the south along the Chobe fault, the slope increases abruptly (about 60 m) as illustrated by the elevation profile (Fig. 6.3). At the top of the slope, red sands, a color resulting from aluminum- and iron-rich coatings around quartz grains, are covering the landscape and are likely inherited from past pedogenic weathering. At the foot of the Chobe fault towards the study area, colluvial fans are forming in contact with the alluvial sediments deposited by the Chobe river (Fig. 6.3).

6.3.1.4 The Complexity of the Chobe's Geomorphology

The geomorphology of the drier Linyanti-Chobe basin is thus complex. A relative chronology of events can be proposed, as also supported by the regional Quaternary chronology (see above): first, red sands formed on a palaeo-surface, observed today at the top of the Chobe fault. They can be considered as paleosols, possibly very old and inherited from a period before the Chobe fault activation by the extension of the graben system (Pliocene? before mega-lakes?). The Chobe extension started and led to possible water accumulation along the Chobe fault. While the graben system was still active, the Linyanti fault formed. This event led to palustrine-lacustrine-fluvial environments from which the palaeo-floodplain and the island landforms are inherited. According to OSL dating, carbonate deposition occurred from 55 to 10 ka BP, contemporaneous to Palaeo-Lake Thalamakane retraction in the SSW. Tectonic events might have affected the drainage, inducing drier conditions through time, cutting off the Upper Zambezi and the Linyanti inputs. From 10 to 5 ka BP, the Linyanti inputs should have been stopped, due to continuous lifting, leading to the fossilization of the palaeo floodplain and carbonate-rich palustrine environments. Since this time, relief inversion could have occurred with the preferential erosion of sands around the carbonate-rich formations that are observed today as elevated carbonate islands (Diaz et al. 2019).

6.4 Landscape Components of the Chobe Enclave

6.4.1 Water in the Linyanti-Chobe Region

Both surface and groundwaters of the study region can be compared to the Okavango and the Zambezi (Dyer 2017). While the Okavango Delta hydrology has been particularly well studied over the last few decades (e.g., McCarthy et al. 1991; McCarthy and Ellery 1998; McCarthy et al. 1998; 2000; McCarthy 2006; Atekwana et al. 2016; Akondi et al. 2019), the Kwando and Zambezi rivers, which may also be of importance to the Chobe region, have been less studied so far (Tooth et al. this book). By analogy to the previous studies in the region (McCarthy, 2006; Atekwana et al. 2016; Akondi et al. 2019), the surface waters of the region can be classified as calcium-magnesium-(sodium)-bicarbonate type waters for the major rivers (Okavango, Kwando/Linyanti, and Zambezi; Fig. 6.9). The Okavango waters, however, have a larger proportion of potassium, whereas the Zambezi waters have a higher electrical conductivity. While the dissolved elemental proportions in the groundwaters do not change markedly relative to the surface waters, they have a higher overall conductivity (ion concentrations and total dissolved ions) compared to the large river entries. Consequently, a more marked change is noted for the Makgadikgadi pan surface waters and extremely evaporated terminal swamp waters of the Linyanti/Chobe as these become sodium-(potassium)-chloride type waters with locally also important sulfate concentrations, notably in the Linyanti/Chobe area. These highly evaporated terminal waters also reach calcite and/or anhydrite saturation (Dyer 2017; Atekwana et al. 2016). Zambezi river waters have a higher Ca/Mg compared to the Okavango River waters, which may indicate geologically slightly different catchments. As the pH of the Zambezi River is about 7.5, while that of the Okavango is about 6.9, this could indicate the presence of carbonate-bearing rocks in the Zambezi basin. The Okavango waters, in contrast, are more concentrated in sodium and potassium compared to the Zambezi River, suggesting that sodium and potassium as well as Ca are likely derived from silicate (feldspar) weathering. These trends in the major ion compositions are clearly represented in a modified Gibbs diagram (Fig. 6.9).

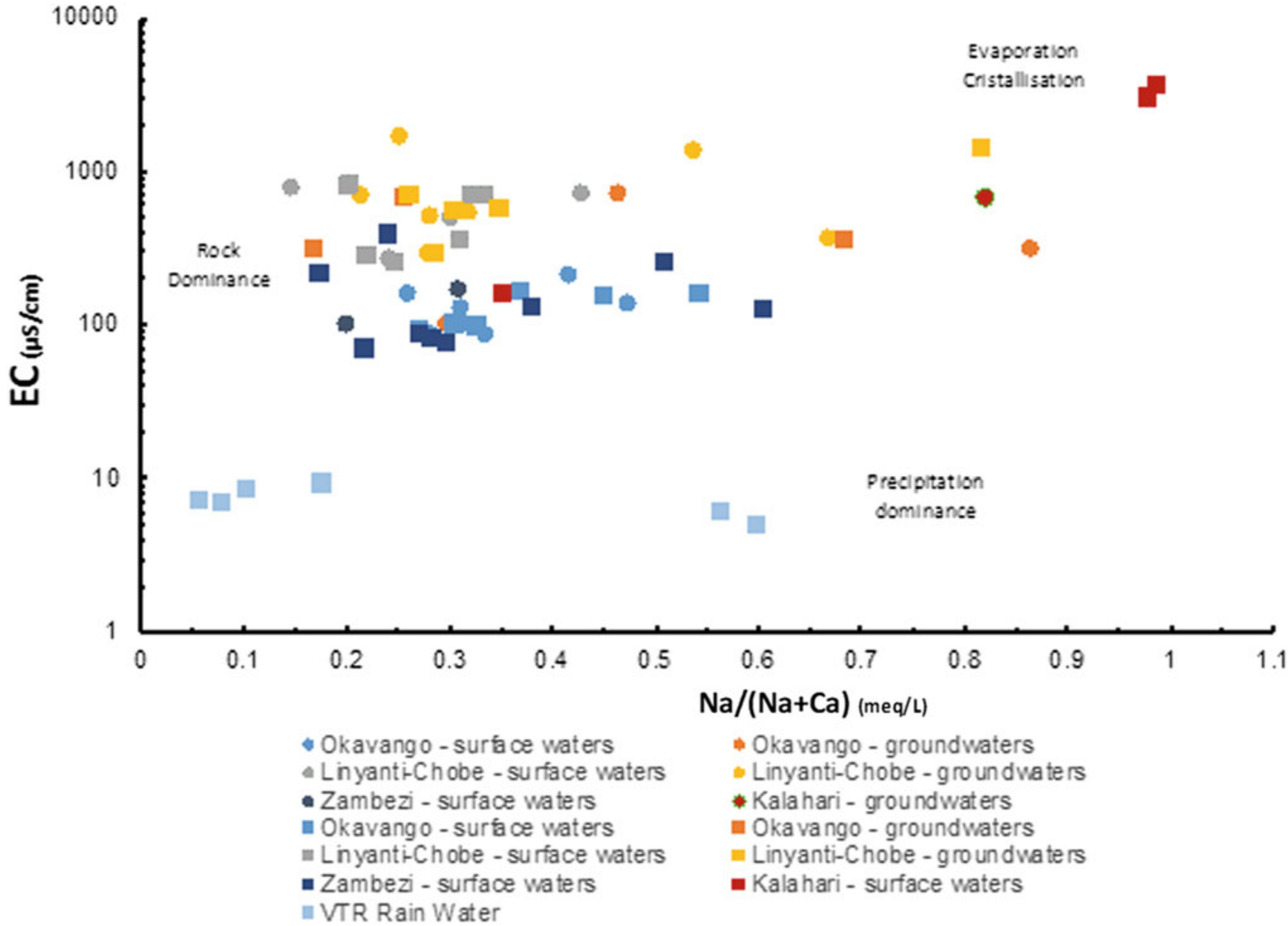

Fig. 6.9 Modified Gibbs diagram illustrating the changes in Na and Ca concentrations as a function of source terrain, evaporation-crystallization processes in the studied region

6.4.2 Vegetation and Soils

Eight plant communities can be distinguished in the Chobe Enclave (Vittoz et al. 2020): three grassland units, three woodland units, and two forest units (Fig. 6.10; Table 6.2). Their spatial distribution depends on the fine microtopography of the northern Chobe region. Based on the vegetation map (Fig. 6.11) and the geomorphology, the study area can be divided into four areas: (1) the floodplains along the Linyanti river, characterized by a gradient of flood duration and frequency; (2) a mosaic of *Combretum* woodlands and dambo grasslands in the island area, with contrasted topography positions and substrates; (3) a mosaic of mopane woodlands and sandveld in the western palaeo-floodplain, similarly related to topography; (4) and the *Baikiaea* forests on red sands, located on the eastern side of the Chobe Fault, extending to the Chobe Forest Reserve.

In a parallel study, five soil types were recognized in the study region (Romanens et al. 2019). The following soil types were found (Figs. 6.12 and 6.13), named according to the WRB (IUSS Working Group WRB, 2014):

- Arenosols are largely dominated by sand along the whole profile, generally poor in nutrients and organic matter (Skarpe et al. 2014).
- Kastanozems are characterized by a dark, organic topsoil, saturated with bases, and by a calcium-rich horizon.
- Chernozems and Phaeozems have a high organic content in most of their profile, with a black topsoil horizon; compared to Phaeozems, the Chernozems are characterized by a calcium-rich horizon.
- Solonchaks and Solonetz are characterized by a salt- or sodium-rich horizon in their profile, the second being richer in clays.

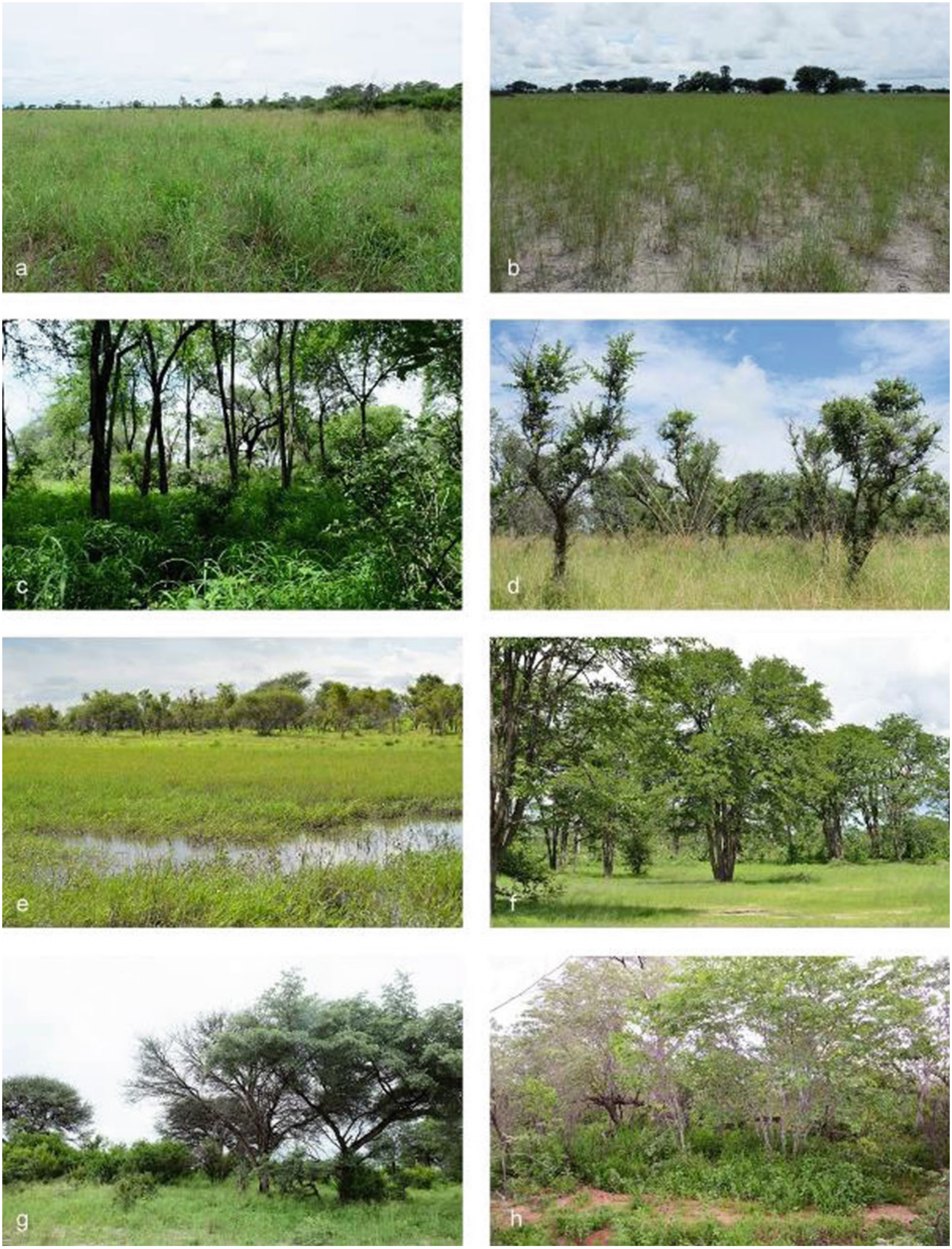

Fig. 6.10 Illustrations of the vegetation units. **a** Wet floodplain; **b** Dry floodplain grassland; **c** Riverine forest; **d** *Combretum* woodland; **e** Dambo grassland; **f** Mopane woodland; **g** Sandveld; **h** *Baikiaea* forest (Photos: F. Pellacani, a, b, c, d, g; P. Vittoz, e, f, h)

Table 6.2 Scientific names used in Vittoz et al. (in press) for the different plant communities

Vegetation unit	Scientific name (Vittoz et al. 2020)
Dry floodplain grassland	*Aristida junciformis–Aristida meridionalis* Grassland
Riverine forest	*Croton megalobotrys–Setaria verticillata* Forest
Combretum woodland	*Eragrostis superba–Combretum hereroense* Woodland
Dambo grassland	*Geigeria schinzii–Setaria sphacelata* Grassland
Mopane woodland	*Colophospermum mopane–Jasminum stenolobum* Woodland
Sandveld	*Ipomoea chloroneura–Oxygonum alatum* Woodland
Baikiaea forest	*Baikiaea plurijuga–Baphia massaiensis* Forest

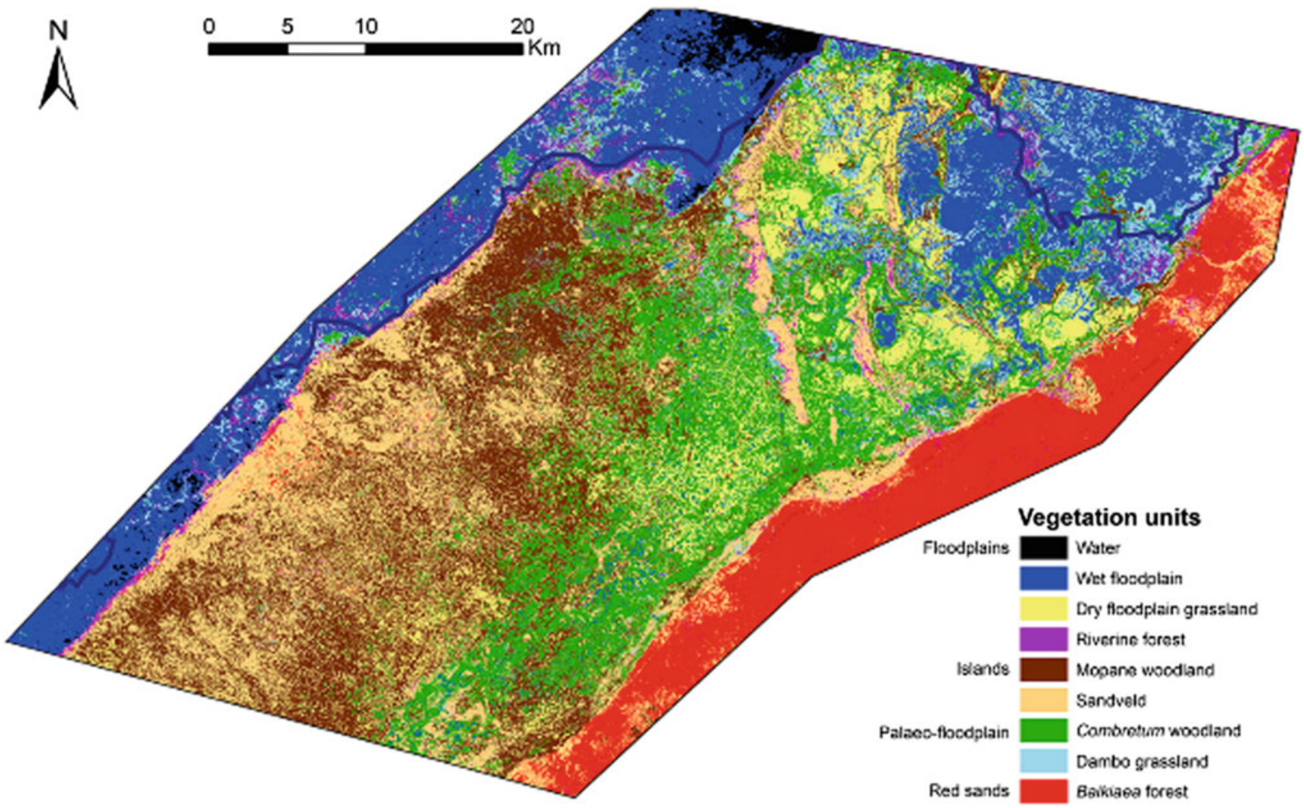

Fig. 6.11 Vegetation map of the Chobe Enclave, with vegetation unit groups according to landscape areas. The Linyanti River is delineated by the dark blue line

- Calcisols contain calcium-rich horizons but compared to Kastanozems, are not particularly rich in organic carbon.

6.4.2.1 Floodplains

The wettest parts of the floodplains were not considered in detail. However, a high diversity of plant communities is expected, close to the one observed in the Okavango Delta (Sianga and Fynn 2017). These plant communities are related to the water depth in permanently flooded areas (Ellery and Ellery this book) and the importance and recurrence of flood events by the Linyanti river in the seasonal swamps (Murray-Hudson et al. 2011; Pricope 2013).

The dry floodplain grasslands are present on higher topographic positions, which are flooded at a low frequency, possibly not each year or not at all, but are still under the seasonal influence of a high water table. This plant community has a low cover of grasses ($\sim$40%), established on former alluvial deposits, with a very high sand proportion ($\sim$98%; soils classified as Arenosols) related to a very low Cation Exchange Capacity (CEC). The dominant species are the grasses *Aristida meridionalis*, *A. junciformis,* and *Hyperthelia dissoluta*, and the forbs *Dicerocaryum eriocarpum* and *Meeremia tridentata* are some of the typical species encountered. These dry floodplain grasslands are

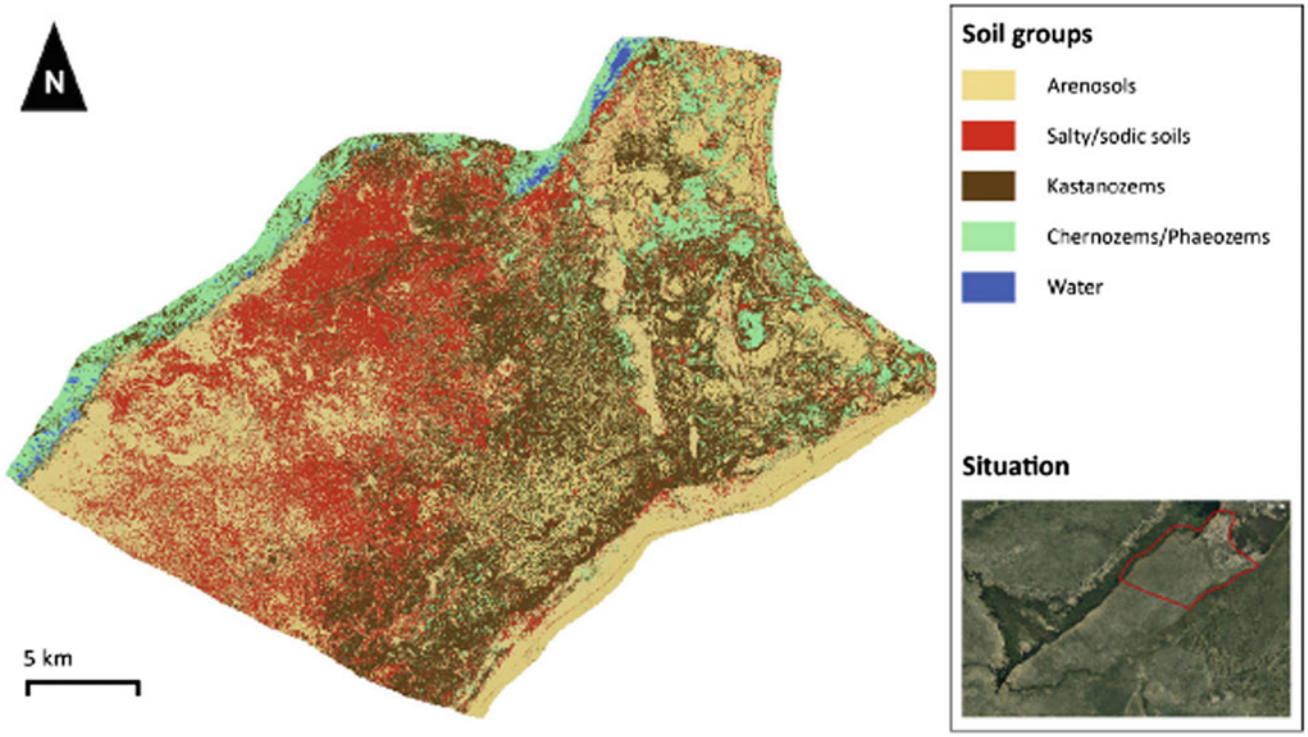

Fig. 6.12 Soil map of the northeastern part of the Chobe Enclave. This map was produced by conducting parallel soil (Romanens et al. 2019) and vegetation studies (Vittoz et al. 2020) and the soil map was partly based on the vegetation map

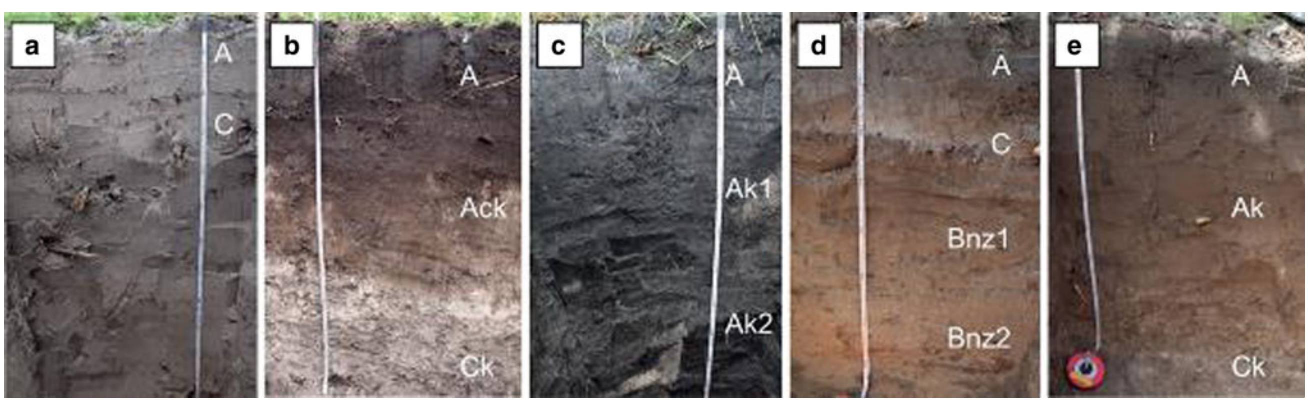

Fig. 6.13 Examples of soils found in the Chobe Enclave. **a** Arenosol: soil of floodplains and wind-reworked areas. **b** Kastanozem: soil found on sandy areas and diatomitic/carbonate islands. **c** Chernozem: soil of the humid depressions (dambos) with some hydromorphic properties. **d** Solonetz: soil often developed under Mopane woodlands. **e** Calcisol: soil easily formed on carbonate islands. Letters on pictures refer to the nomenclature of horizons (for more details, see IUSS Working Group WRB 2014 and Romanens et al. 2019)

characterized by numerous termite mounds (Mujinya et al. 2010, 2014), often occupied by trees (McCarthy 1998; McCarthy et al. 1998), some of them found almost exclusively in such situations (e.g., *Phoenix reclinata*, *Imperata cylindrica;* Roodt 1998).

Riverine forests form a species-rich community, encountered as small fragments on prominences in or along the floodplain of the Linyanti river. It has the highest mean tree cover (~50%) and the tallest trees (~8.5 m) of all vegetation units, their density and size related to their possibility to reach a permanent water table with their roots. The alkaline soils (Kastanozems) are dominated by sand but with 15–33% of silt, probably due to the high density of termite mounds. *Berchemia discolor, Vachellia tortilis*, *Croton megalobotrys*, *Combretum mossambicense,* and *Dichrostachys cinerea* are some of the typical trees and shrubs, often climbed by *Capparis tomentosa*. *Kigelia africana, Terminalia prunioides,* and *Ficus thonningii* are less frequent woody species but encountered only in riverine forests. Some typical species in the understory are the grasses *Setaria verticillata*, *Panicum maximum,* and *Urochloa trichopus* and the forbs *Ipomoea dichroa, Acalypha fimbriata, Justicia heterocarpa, Achyranthes aspera, Dicliptera paniculata, Commelina petersii, Cyathula orthacantha,* and *Leonotis nepetifolia*.

6.4.2.2 Islands

Two plant communities are distributed in this area according to the topography. The *Combretum* woodlands is found on small islands underlain by carbonate beds*.* The soils (Kastanozems) are characterized by the highest clay content of the study area (14–20%) that favors a high CEC dominated by calcium, promoting an alkaline pH (7.8–8.8). The typical trees are *Combretum hereroense*, *Sclerocarya birrea* subsp. *caffra*, *Senegalia nigrescens,* and *Vachellia hebeclada*. The grasses *Eragrostis superba*, *Heteropogon contortus*, *Digitaria eriantha,* and *Cenchrus ciliaris*, and the forbs *Rhynchosia minima, Hoslundia opposita* and *Hibiscus caesius,* and *Solanum panduriforme* are frequently encountered.

The dambo grasslands are found in pans. This community occupies shallow, depressions that are waterlogged during the wet season (Acres et al. 1985). Without a river draining them, fine mineral materials tend to accumulate, and the waterlogging restrains organic matter mineralization. Hence, the soils (Chernozems or Phaeozems) show the lowest sand content (19–29%) but the highest silt (59–68%) and organic carbon (3–5%) contents of all the plant communities in the Chobe Enclave. The high CEC is dominated by calcium, resulting in a neutral to alkaline soil pH. Grasses largely dominate the community, with *Setaria sphacelata, Digitaria milanjiana, Hyparrhenia rufa, Bothriochloa bladhii, Cymbopogon caesius, Panicum repens,* and *Trachypogon spicatus* as frequent or dominant species.

6.4.2.3 Western Palaeo-Floodplain

The western palaeo-floodplain is similarly characterized by a mosaic of two vegetation units according to topography. The area is a fossil delta with islands separated by channels inherited from ancient hydrological dynamics, before being separated from the Linyanti river by tectonic activity (Kinabo et al. 2008; Fynn et al. 2014).

The islands are occupied by the mopane woodlands. Some areas are characterized by tall *Colophospermum mopane* trees, while in other places the species is limited to stunted specimens, with a low tree cover. It seems that these sizes are related to different soil conditions: stunted mopanes grow on silty soils characterized by a high CEC, saturated by sodium, and with high alkaline pH (9.5–10), classified as Solonchaks. In contrast, tall trees occupy soils with a sandy upper horizon, with moderate alkaline pH (7.0–8.2), but with the presence of a sodic horizon below 30 cm (Solonetz or other soil types, like Calcisols, with a *salic* horizons). These high sodium concentrations suggest that these areas have an origin comparable to the present islands in the Okavango Delta (Ramberg and Wolski 2008; McCarthy et al. 2012). Besides mopane, the typical species are the grasses *Schmidtia pappophoroides*, *Aristida adscensionis,* and *Digitaria eriantha*; the sedge *Kyllinga buchananii*; and the forbs *Ipomoea coptica* and *Chamaecrista absus*.

The palaeo-channels are filled with leached, river-washed sands (Arenosols) and occupied by sandvelds. Compared to dry floodplain grasslands, this plant community has a higher topographic position, with a complete absence of visible water table during the wet season. Tree and shrub layers are characterized by the presence or dominance of *Senegalia cinerea*, *Terminalia sericea*, *Grewia retinervis,* and *Dicrostachys cinerea*, and the herb layer by the grasses *Panicum maximum*, *Urochloa trichopus,* and *Digitaria eriantha*, and the forbs *Acanthosicyos naudinianus, Spermacoce senensis, Indigofera flavicans*, *Gisekia africana*, *Cleome hirta,* and *Pavonia senegalensis*.

6.4.2.4 Baikiaea Forest

Southeast of the Chobe Fault, the *Baikiaea* forests have a dense woody stratum. This vegetation unit, which covers a large area in Chobe National Park (*Baikiaea plurijuga–Combretum apiculatum* woodland) according to Skarpe et al. (2014), grows on thick, reddish Kalahari sands. The soils are classified as Arenosols and their red color is due to an iron and aluminum-rich oxyhydroxide coating, which is responsible for acidic conditions (pH 5.7–6.0). This area is completely disconnected from the rivers and the plants have to rely only on rainwater. Trees and shrubs show high diversity, the most frequent species being *Combretum elaeagnoides, Baikiaea plurijuga, Baphia massaiensis, Croton gratissimus,* and *Bauhinia petersiana*. The grasses *Dactyloctenium giganteum, Panicum maximum, Urochloa trichopus,* and *Digitaria eriantha,* and the forbs *Jacquemontia tamnifolia, Ipomoea pes-tigridis*, *Vigna unguiculata*, *Chamaecrista absus,* and *Harpagophytum zeyheri* are constant or locally important species.

6.4.3 The Role of Termites

The most notable species of termites encountered in the northern Chobe Enclave district are the fungus-growing termites (FGT) (subfamily Macrotermitinae, Isoptera). They are important for two reasons: (i) their large iconus epigeal mounds can be seen from afar and (ii) their ability to modify the distribution of sediments and soils in order to build their nest (Jouquet and Lepage 2002; Mujinya et al. 2011). Fungus-growing termites are a subfamily composed of 12 genera and 345 species. One of these genera, *Odontotermes* (142 species) present in this region, tends to build cryptic subterranean fungus-chambers 20–50 cm below ground. An important species found in the area is the *Trinervitermes trinervoides*, also commonly known as snouted harverster termite for the protruding portion of their face. These are non fungus-growing termites and do not build large fungus-chamber but only small scattered domed mound of only a few centimeters high. The most common species observed in the study area is *Macrotermes michaelseni*, a FGT species part of he *Macrotermes* genus (65 species). It has long been considered as an ecosystem engineer (Jones 1990) for the modifications it brings to the soil, its ability to concentrate nutrients, and its capacity to create patches of fertile land in tropical and subtropical savanna ecosystems (Jouquet and Tavernier 2005; Corenblit et al. 2016).

6.4.3.1 The Potential Impact of Termites in Savanna Ecosystems of the Chobe Enclave

Fungus-growing termites have shared an exosymbiosis since 30 Ma (Roberts et al. 2016) with a fungi belonging to the *Termitomyces* genus. In order to maintain the symbiosis, *Macrotermitinae* must maintain specific hydric and thermic conditions in their nests. Therefore, fungus-growing termites build large biogenic structures in which they are able to increase the alkalinity of soils by an order of magnitude of 3–4. They also increase their carbonate content, the C/N ratio, and concentrate nutrients such as potassium and phosphorus. FGT modify the chemical compositions and mineralogical propreties of clays through the process of selection and transportation of sand grains in their buccal cavity, where they are mixed with saliva (Jouquet et al. 2011). They also act as accelerating agents of clay alteration and chemical weathering in tropical ecosystems. The activities of FGT tend to slightly raise the land surface locally, providing some recolonization advantages. They form a pattern of fertile lands by conentrating nutrients, and enhance the growth of vegetation by creating islands of fertility (see above; Dangerfield et al. 1998). The selection of very fine sands, in order to meet the construction requirement for their mounds, create patches of clayey sands that have the property to retain water for long periods of time, producing scattered pockets of water in semiarid regions such as the Chobe-Linyanti region (Pennisi et al. 2015).

Different stages of FGT mounds can be observed in the northern Chobe Enclave district (i) active termite mounds (Fig. 6.14a), generally between 50 cm and a few meters high; they are homogenous and hard consolidated forms; (ii) mounds that have been recently abandoned by the termite colony, due to a predator or a natural hazard; their shape appears as similar to a flattened promontoir with broken pieces laying on the adjacent ground; they can be recolonized by the same species of termites; (iii) relic termite mounds (Fig. 6.14b); these mounds have seen many stages of occupation over a relatively long period of time, from a few dozen years up to hundreds of years and their size can reach 18 m in diameter and be 7 m high. These relic mounds can probably be considered as geomorphological objects by their size, their impact on the environment, and their role in the ecosystem. They are distinctive features in the landscape of which many animals take advantage, such as snakes, monkeys, leopoards, lions, elephants, as well as plants (including trees).

Like in many parts of Africa, people from the the study region are using mounds for different purposes, e.g., soil amendment, geophagy (Mills 2008), or building material. Their volume is considerable: a single mound in the Chobe had an above ground volume of 1,350 m^3. They also can remain for long periods of time: FGT relic mound ages have been estimated by optically stimulated luminescence (Kristensen et al. 2015) to be 4,000 yrs old in Ghana and to be 2,200 yrs old using ^{14}C in Congo (Erens et al. 2015).

6.4.3.2 Hydric Gradients Influencing FGT Activities

In regard to termite activities, two gradients are observed in the Chobe Enclave: (i) an EW gradient assigned to the topographic flooding pattern and (ii) a NS gradient attributed to the distance to the Linyanti river in the north. Distribution of *Macrotermes* mounds in the landscape represent these gradients fairly well. In the eastern part of the enclave, large active mounds, similar to the one encountered in the center of the Okavango Delta today, dominate the topography. Moving further west, these big mounds become less common and after a short distance toward the center of the enclave, they totally disappear. In reference to the NS gradient, within the riverine woodland of the Linyanti river, all mound stages are observed (active, abandoned, or relic). Moving south, the active termite mounds are rarer, most of them being abandoned; finally, in the middle part of the enclave, there are no active *Macrotermes* mounds.

6.4.3.3 Termites Influence the Evolution of the Landscape

Regarding the formation of islands, Dangerfield et al. (1998) were the first to describe the role of FGT in the context of the Okavango Delta. A similar process may have existed for most of the northern Chobe Enclave's past. FGT are probably still involved in the geomorphological shaping of the region through this specific island formation process found in its eastern part. The sediment composition of the islands being an important factor, a recent study quantified the way in which FGT modify local soil grain-size distributions to make the building material for their nest (Van Thuyne et al. 2021). When built on a coarse sand soil, FGT tend to

Fig. 6.14 **a** Young active termite mound of the *Macrotermes michaelseni* species, situated on a diatomaceous soil in an open grassland on the eastern side of the Chobe Enclave. **b** A non-re-colonized relic termite mound of the *Macrotermes michaelseni* species. Its size is impressive

moderately increase the upper part with finer material and tend to increase substantially the lower part in finer material, whereas, when built on a silt soil, the opposite is observed. Once abandoned and destroyed by erosion, the altered soil material consequently modifies the grain-size distribution of the local sediments.

The above-mentioned modifications brought by FGT have considerable importance and impact on this semiarid region. For example, the soil pH is often low and the soil remains deficient in many nutritive elements: the mounds are able to provide better soil conditions with a more alkaline pH and a higher nutritive content. The mounds have also the capacity to retain water for long period of time, (i) favoring a de-aridification process (Pennisi et al. 2015), (ii) offering a fire protection in these savanna-prone fire zones, and (iii) contributing to higher stability and resilience of the ecosystem (Pringle et al. 2010).

6.5 Challenges and Conclusions

The northern Chobe Enclave district constitutes a fascinating location in order to understand the complex relationships between the development of landforms in a high African plateau under the influence of trans-tensional pull-apart basins and the various external agents at work during the Quaternary. At the crossroads of various rivers and influenced by neotectonics, large parts of the landscape are not yet explained: issues such as (i) how did the syntectonic depocenters evolve through the Quaternary? Or (ii) what are the ages of the various faults and water gaps affecting the region? The presence of large palustrine carbonate deposits is not totally explained: how were such deposits possible in the past? And what would have been the environmental conditions that enabled such calcium carbonate accumulations in an essentially siliceous basin? In addition, if the source of carbon can easily be identified (CO_2 and organic matter), where is the source of calcium?

This chapter presented a limited state of the art of the science conducted in the area. It is clear that the present-day composition of the various water bodies cannot totally explain the diversity of the geochemical nature of the sediments. Nevertheless, it seems likely that the large presence of sodium in soils and sediments could be related to some water bodies, associated with efficient evaporation processes. The vegetation is perfectly adapted to the landscape constraints. Another important point is the role of termites: they redistribute the particle sizes of sediments at the surface of the geomorphological features and partly shape the landscape. Today, the main exogenous parameters affecting the landscape seem to be wind, water balance (rain and the prominent role of rivers and floodplains), as well as the biological impact of the living realm. But the conditions prevailing today were probably different during the past, making the reconstruction of the landscape history challenging.

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