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STRATIGRAPHY, DEPOSITIONAL ENVIRONMENTS, AND PALAEOGEOGRAPHY OF THE KOOBI FORA FORMATION

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Studies of the geology of the Koobi Fora deposits have been under way for nearly two decades. Over that span of time, a great deal of information has accumulated, and many arguments have been raised and laid to rest. In recent years our understanding of the Koobi Fora sequence has changed dramatically through geochemical correlation of volcanic ashes, improved age control by reliable isotopic age determinations, continued study of the stratigraphical features of the deposits themselves, and, perhaps most importantly, secure correlation of the Koobi Fora sequence with those of the lower Omo Valley and along the west side of Lake Turkana. As a result, we are now able to outline our knowledge of the sequence at Koobi Fora in far greater detail than was possible only a few years ago, and to present a summary of some important revisions to earlier discussions.

The Plio-Pleistocene sedimentary sequence of the Koobi Fora region is discontinuously exposed over an area of some 1200 km^2 along the eastern shores of Lake Turkana (Fig. 1.1). The region was originally referred to as 'East Rudolf,' prior to renaming of the lake in 1975. Subsequently the term Koobi Fora region has been applied, along with the less appropriate 'East Turkana'. Within the Koobi Fora region, three broad geographical entities have been recognized: the Ileret subregion, the Koobi Fora subregion, and the Allia Bay subregion. Drainage systems (sand-bed ephemeral streams, often referred to by the Galla term 'laga') provide a convenient means of further subdividing the region. Most of the larger sand rivers (*Il* in the Dasanetch language) and intermediate sized sand rivers (*Kolom* in Dasanetch) have individual names, as do the waterholes and the more distinctive hills and ridges. Local geographical terms are illustrated in Fig. 1.2. The region has also been divided into a series of collecting areas. These originated as discrete localities (Leakey 1978, Fig. 1.2; Harris 1983, Fig. 1.3) but now cover the entire region (Fig. 1.3).

STRATIGRAPHY

The late Neogene stratigraphical succession in the northern Turkana Basin has been described in terms of two groups of strata (de Heinzelin 1983). The Omo Group consists of the Pliocene and Pleistocene sedimentary units, including the Koobi Fora, Nachukui, Shungura, Usno and Mursi Formations (Fig. 1.4). These strata are locally disconformably overlain by the Late Pleistocene and Holocene deposits of the Turkana Group, which includes the Galana Boi and Kibish Formations.

The stratigraphical framework for the deposits at Koobi Fora was revised by Brown and Feibel (1986) to include the entire Plio-Pleistocene sequence (with the exception of the latest Pleistocene) within the Koobi Fora Formation. This was necessary to alleviate inconsistencies in the previous stratigraphical scheme and to permit a clearer understanding of the deposits. The revised definition of the Koobi Fora Formation is applied to those strata that unconformably overlie Miocene and Pliocene volcanic rocks and that are, in turn, unconformably overlain by late Pleistocene and Holocene diatomaceous lake beds known as the Galana Boi Formation (Owen and Renaut 1986). The Koobi Fora Formation is subdivided into eight members; from bottom to top these are named the Lonyumun, Moiti, Lokochot, Tulu Bor, Burgi, KBS, Okote, and Chari Members. Each member is defined to include a designated basal tuff plus those overlying sediments beneath the base of the next designated tuff. Thus the Tulu Bor Member includes the Tulu Bor Tuff and superjacent deposits that underlie the base of the Burgi Tuff

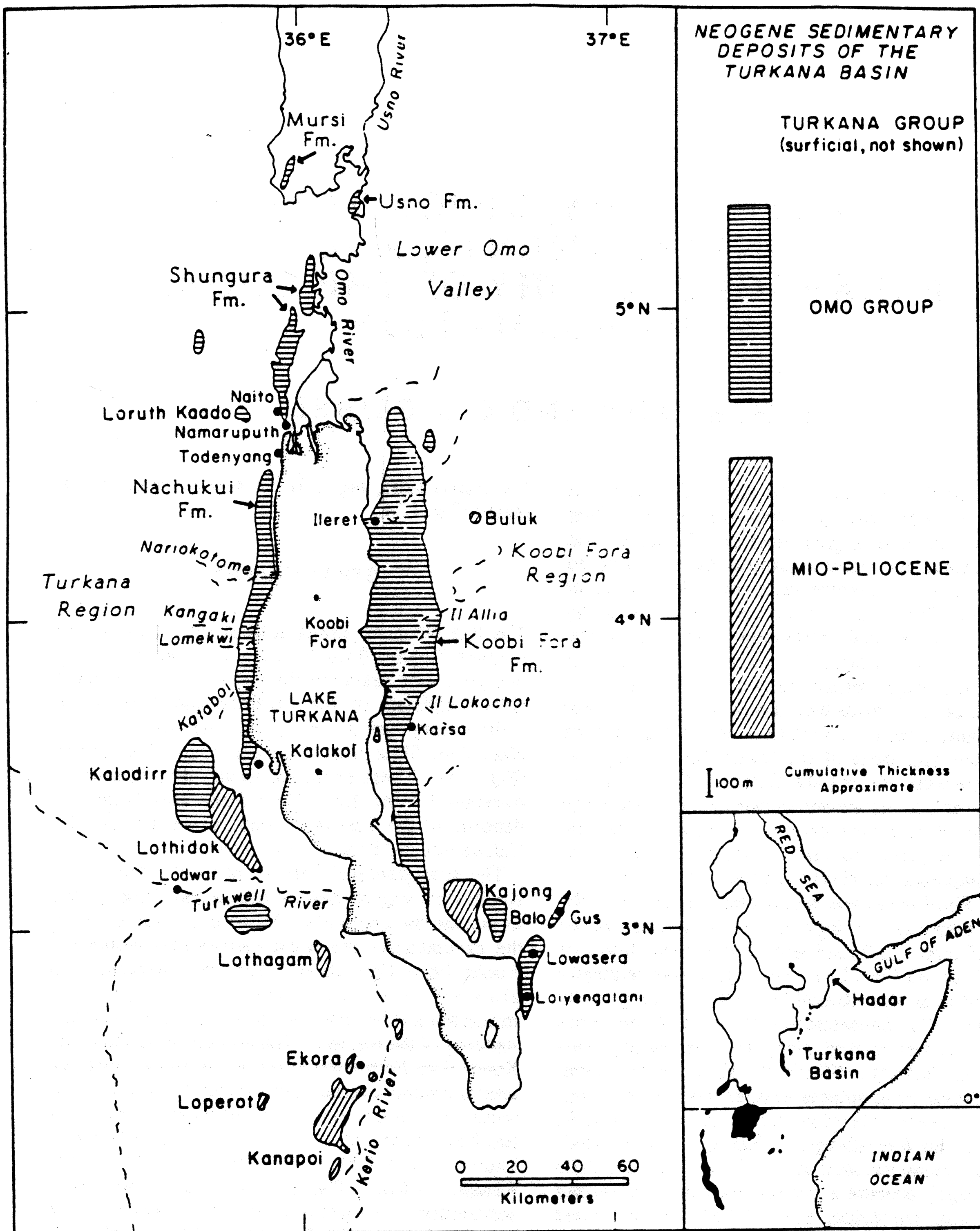

Fig. 1.1. Location map of the Turkana Basin, showing major localities and exposures of Neogene sediments.

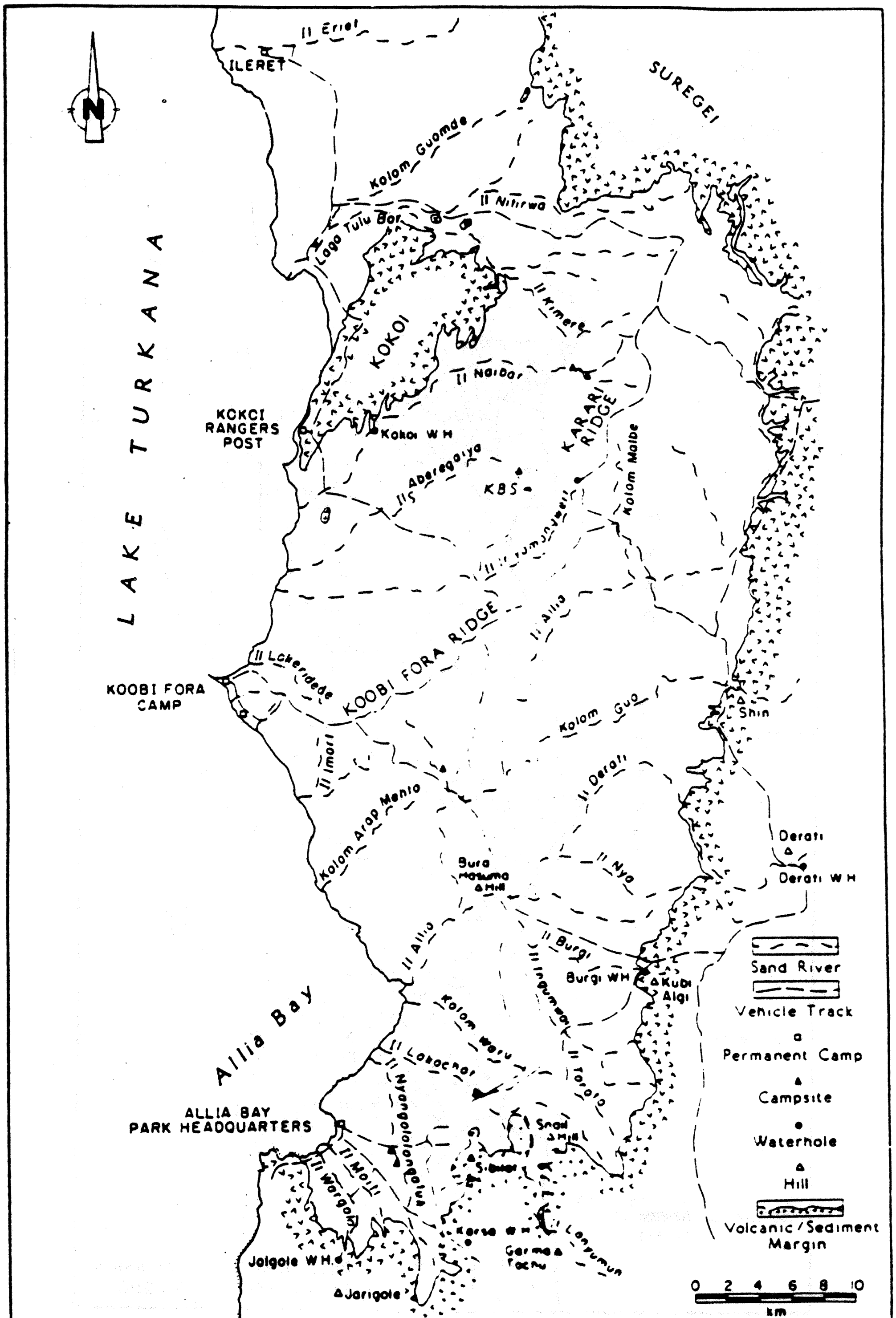

Fig. 1.2. Geographic map of the Koobi Fora region, showing important drainage features and dirt tracks.

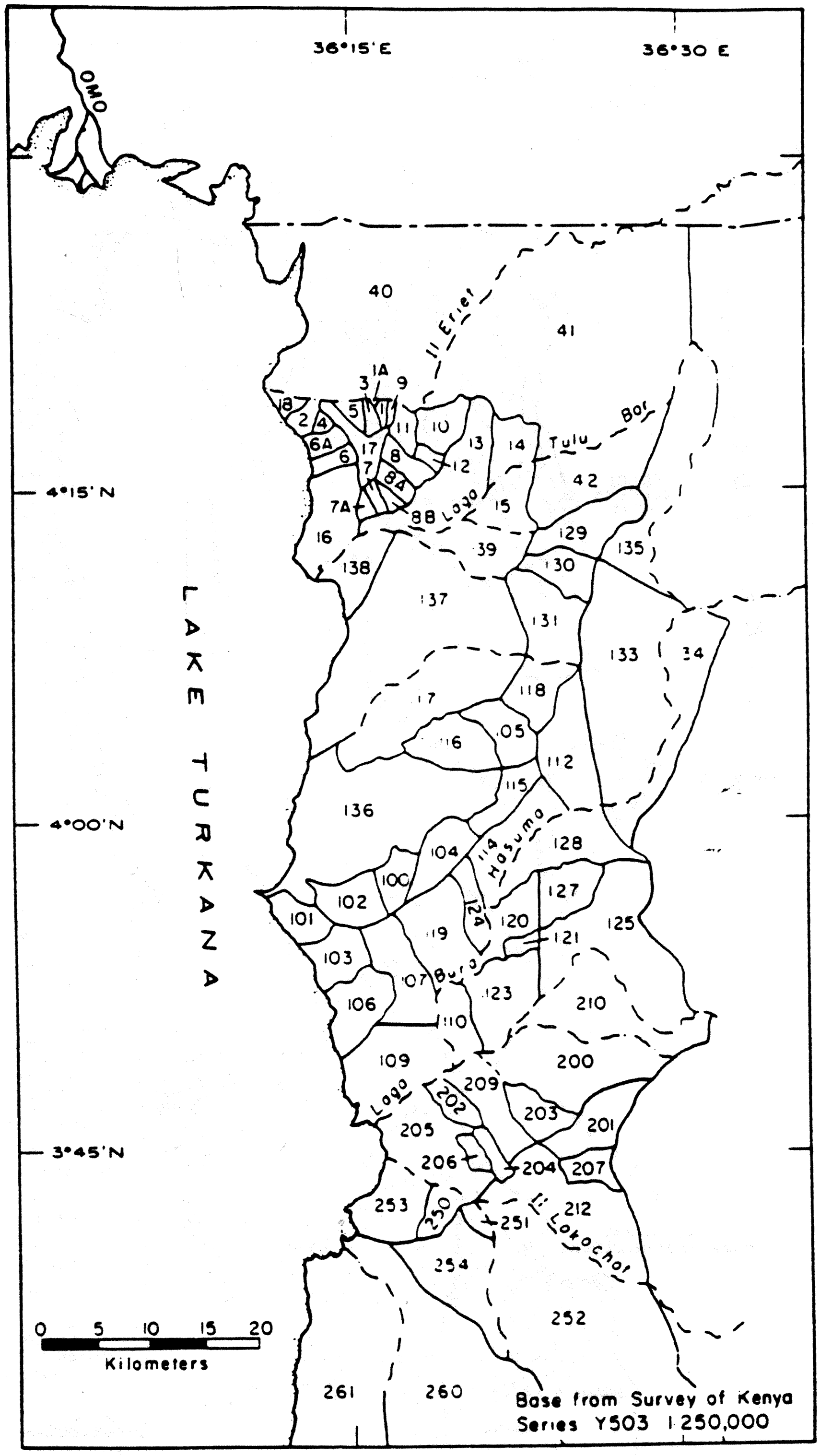

Fig. 1.3. Map of collecting areas of the Koobi Fora region.

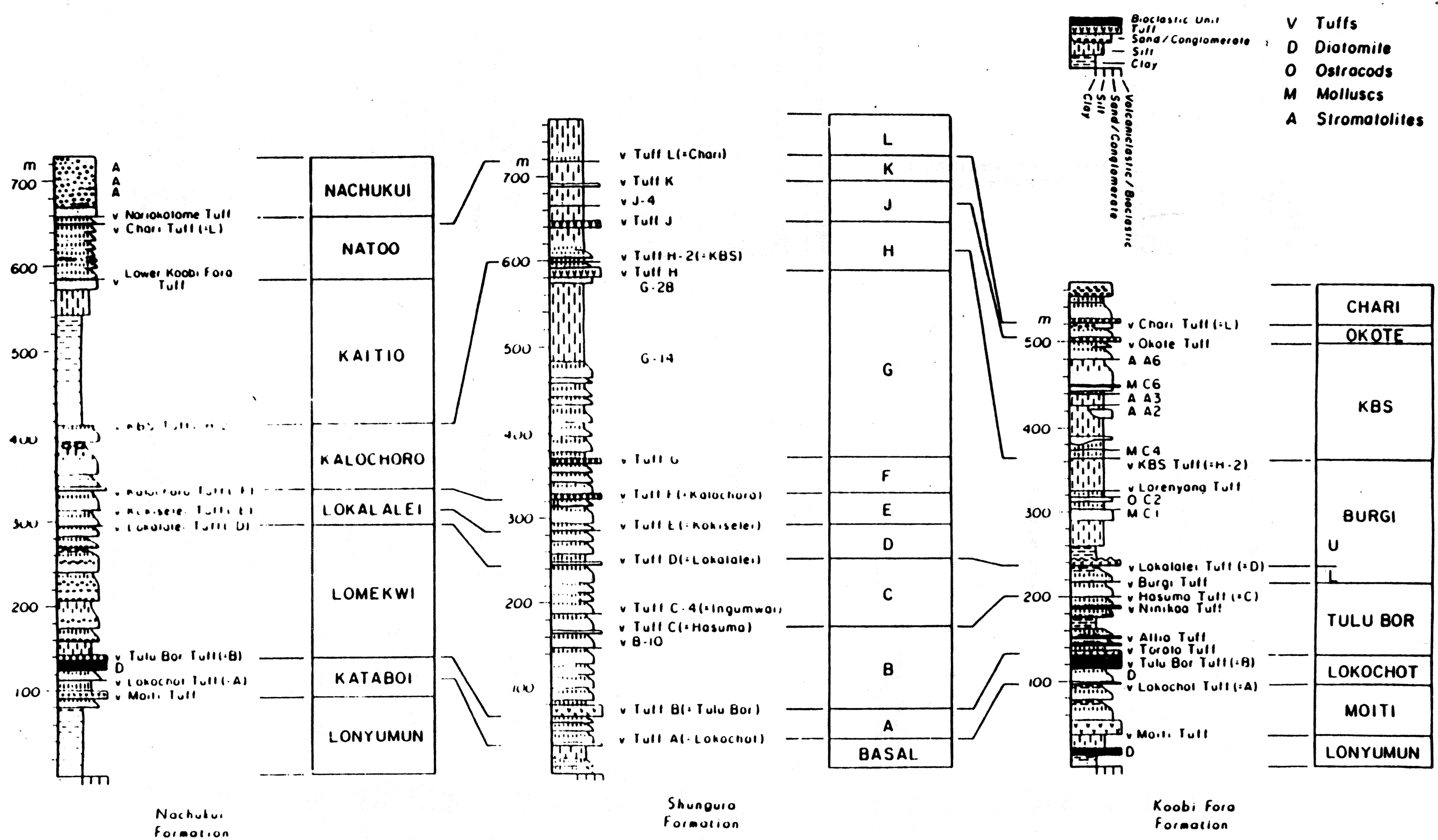

Fig. 1.4. Composite stratigraphic sections of the Omo Group deposits showing correlation of members of the major formations.

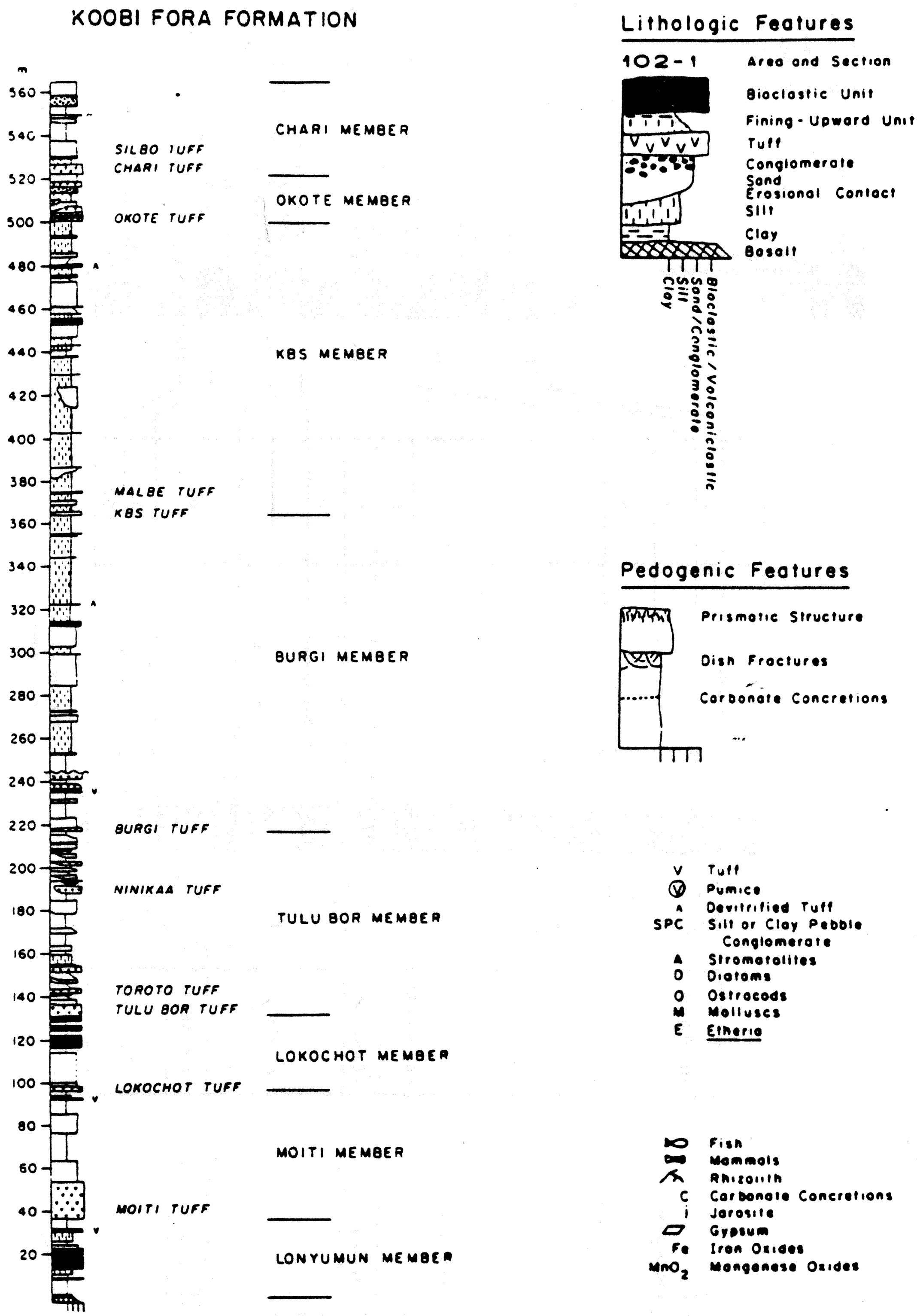

Fig. 1.5. Composite type section of the Koobi Fora Formation (after Brown and Feibel, 1986).

Deposits that underlie the lowest widespread tuff, the Moiti Tuff, are designated the Lonyumun Member. The composite type section of the Koobi Fora Formation is illustrated in Fig. 1.5.

The Koobi Fora Formation is a mostly conformable sequence but is interrupted by the numerous minor diastems typical of fluvio-lacustrine deposits. There are two significant unconformities within the succession. The older of these occurs in the Burgi Member, and is the basis for an informal division into a lower and an upper Burgi Member. The other occurs within the Chari Member, between the Chari and Silbo Tuffs. Brief descriptions of the members are given here, along with temporal estimates from Feibel *et al.* (1989).

In its type locality the Lonyumun Member is about 37 m thick. These sediments are the result of two phases of deposition, an early phase dominated by fine-grained detrital and bioclastic sediments, and a later phase of the upward-fining cycles of detrital clastics with associated pedogenesis indicative of a fluvial environment. The member represents an interval of less than 350000 years.

The type locality of the Moiti Member exposes about 60 m of strata. The member is relatively homogeneous in all its known outcrops, with the cyclical upward-fining detrital sedimentation and associated pedogenesis characteristic of fluvial deposition. It is the only member without evidence of brief lacustrine episodes. It is also notably lacking in erosion-resistant marker beds, and thus has a relatively limited area of exposure. The member represents sedimentation over an interval of about 500 000 years.

The Lokochot Member is some 34 m thick in its type section but is incompletely exposed at any one locality, including the type section. From a regional perspective, however, it is clear that the member is characterized by two phases of deposition. The early phase is typified by relatively coarse-grained detrital clastics and pedogenic alteration indicative of fluvial environments. A second depositional phase consists of finer grained detrital and bioclastic material of a lacustrine episode. The member was deposited over an interval of about 140000 years.

The type section of the Tulu Bor Member exposes about 86 m of sediment. The member is characterized by extremely well-developed upward-fining cycles and pedogenesis, typical of fluvial environments, with minor intervals of fine-grained detrital clastics and bioclastics which indicate lacustrine episodes. The member represents deposition over an interval of some 680000 years.

The Burgi Member is effectively divided into upper and lower portions by a significant unconformity. Local exposures of this member represent intervals of section from either above or below the unconformity; for this reason the type section of this member is defined as a composite of local sections from Areas 207 and 102. The type section of the lower Burgi Member in Area 207 is nearly 27 m thick. Because its upper limit is represented by an unconformity, it is likely that additional younger strata existed (and may still exist) elsewhere in the region. Most of the lower Burgi Member consists of upward-fining detrital clastics with pedogenic alteration reflecting a continuation of the fluvial pattern of the Tulu Bor Member. Near the top of the lower Burgi Member a sequence of interbedded bioclastics and detrital clastics, known as the Lokeridede Complex, indicates a lacustrine episode. The type section of the upper Burgi Member in Area 102 (itself a composite due to local faulting) attains a thickness of over 120 m. The depositional history of the upper Burgi Member is the most complex of any member in the formation. The upper Burgi Member has been further subdivided into five depositional stages by Feibel (1988). Deposits of the upper Burgi Member are dominated by thick upward-coarsening sequences of detrital clastics and interbedded bioclastic units representing a major lacustrine interval with associated deltaic infilling. The entire Burgi Member represents an interval of some 800000 years. It is estimated that the lower Burgi Member strata record some 200000 years, strata of the upper Burgi Member another 100000 years with the remainder (500000 years) represented by the unconformity.

In its type section in Areas 102 and 103, the KBS Member is over 136 m thick. This member marks the beginning of a phase of high lateral variability which characterizes sedimentation in the later members of the Koobi Fora Formation. KBS Member deposits range from interbedded upward-fining detrital clastic and bioclastic units to more homogeneous but coarser detrital clastics with pedogenic modification. These deposits are interpreted to represent a complex of fluvial systems smaller in extent than those recorded from the Tulu Bor Member, along with ephemeral lacustrine environments. The time interval represented by deposits of the KBS Member is about 250000 years.

The type section of the Okote Member is about 22 m thick. This member is characterized by abundant tuffs and tuffaceous siltstones that represent a complex of fluvial deposits. In the Ileret subregion the member includes a prominent reddened analcimic mudstone, a thin dolomite bed, and a bed rich in fish bones overlain by upward-fining detrital clastic cycles; these are interpreted as an intercalated episode of lacustrine deposition in the Ileret subregion. The

Okote Member was deposited over an interval of about 240000 years.

In its type section the Chari Member measures nearly 43 m. It is dominated by thin coarsening-upward detrital clastic sequences with associated fish bone beds, suggesting ephemeral lake environments. It also contains several well-developed large scale upward-fining cycles with associated pedogenic features indicative of major fluvial systems. One of these, near the bottom of the member, is thought to mark a significant unconformity. As a whole, the member spans some 690000 years, but the unconformity most likely accounts for about 500000 years of that interval.

Below we relate the revised stratigraphical terminology to earlier nomenclature, and then discuss regional patterns of stratigraphical variability.

Stratigraphical Background

The earliest stratigraphical terminology used for the Koobi Fora deposits was an informal system devised by Behrensmeyer (1970) and revised by Vondra *et al.* (1971). This scheme was based on exposures between Koobi Fora spit and the KBS site in Area 105. There were four subdivisions: KF I, KF IIA, KF IIB, and KF III (Fig. 1.6). KF I included the predominantly lacustrine sequence now largely incorporated within the upper Burgi Member. The upper limit of KF I was a marker unit known as the Limonite Chunk Conglomerate (LCC; Behrensmeyer 1973). KF IIA spanned the interval from the LCC to the channel complex above the KBS Tuff. This corresponds to the uppermost upper Burgi and lowermost KBS Members in current terminology. The channel complex used to define the upper limit of KF IIA was long thought to represent an 'erosion surface' marking a major break in the sequence, with significant temporal implications. More recent evaluation has not upheld this view, and has shown that it is a complex of channels with little temporal significance, rather than a single channeling event. KF IIB included the dominantly fluvial interval above the channel complex in the Karari Ridge area. This interval is now largely included within the KBS Member. KF III designated the interbedded fluvial and lacustrine sequence exposed near Koobi Fora Spit, mainly in Areas 102 and 103. This is also KBS Member in the revised terminology, being a lateral equivalent of the fluvial deposits to the northeast.

As the early terminology began to be applied to exposures farther away from the original localities, a greater portion of the stratigraphical sequence was included in the defined units. Thus when many deposits of the Allia Bay subregion were referred to KF I, the unit came to include sediments actually belonging to the Lonyumun, Moiti, Lokochot, Tulu Bor, and lower Burgi Members, as well as the original deposits now assigned to the upper Burgi Member. An understanding of this early system is necessary, for in the years before aerial photographs (first flown for the Koobi Fora region in 1970), fossil specimens were sometimes documented only relative to stratigraphical context in this older system.

Following extensive reconnaissance work by Bowen (1974), a formal stratigraphical nomenclature was proposed by Bowen and Vondra. They named three formations, a Kubi Algi Formation exposed largely in the Allia Bay subregion, a Koobi Fora Formation which encompassed most of what had been referred to KF I, II, and III, and a Guomde Formation to include overlying deposits of the Ileret subregion (Bowen and Vondra 1973). The key element in this stratigraphical subdivision was the use of boundary stratotypes, tuffs, which delimited the units. As these markers were not continuously exposed over the Koobi Fora region, an implicit element in the definition was the ability to correctly correlate disparate exposures of tuff. This proved to be the undoing of this system, and led to prolonged debate amongst the researchers involved in the study.

As defined, the Kubi Algi Formation included all sediments from the base of the Plio-Pleistocene section to the base of the Suregei Tuff Complex. In practice this was taken to include most of the Lonyumun, Moiti, Lokochot, Tulu Bor, and lower Burgi Member strata as they are now understood (see maps in Bowen 1974). What was called the Suregei Tuff Complex, however, turned out to include a variety of stratigraphical entities. The type locality is the diatomite sequence in the Lonyumun Member (now referred to as the Suregei Complex). Also included as inferred correlates were the Burgi Tuff and diatomites of the Lokeridede Complex in the Burgi Member (Cerling and Brown 1982).

The original definition of the Koobi Fora Formation included all strata from the base of the Suregei Tuff Complex to the top of the Chari Tuff. For the most part this corresponds to deposits now considered part of the upper Burgi, KBS, and Okote Members. A serious miscorrelation in the early work equated the Tulu Bor Tuff with a tuff exposed near Koobi Fora spit, the Lorenyang Tuff, which lies in the upper Burgi Member. As a result, strata associated with the Tulu Bor Tuff (Lokochot and Tulu Bor Members) were correlated with strata of the upper Burgi Member. As representative fossil collections accumulated from the relevant sections, the miscorrelations became strikingly obvious, at least to the palaeontologists (Cooke and Maglio 1972; Harris

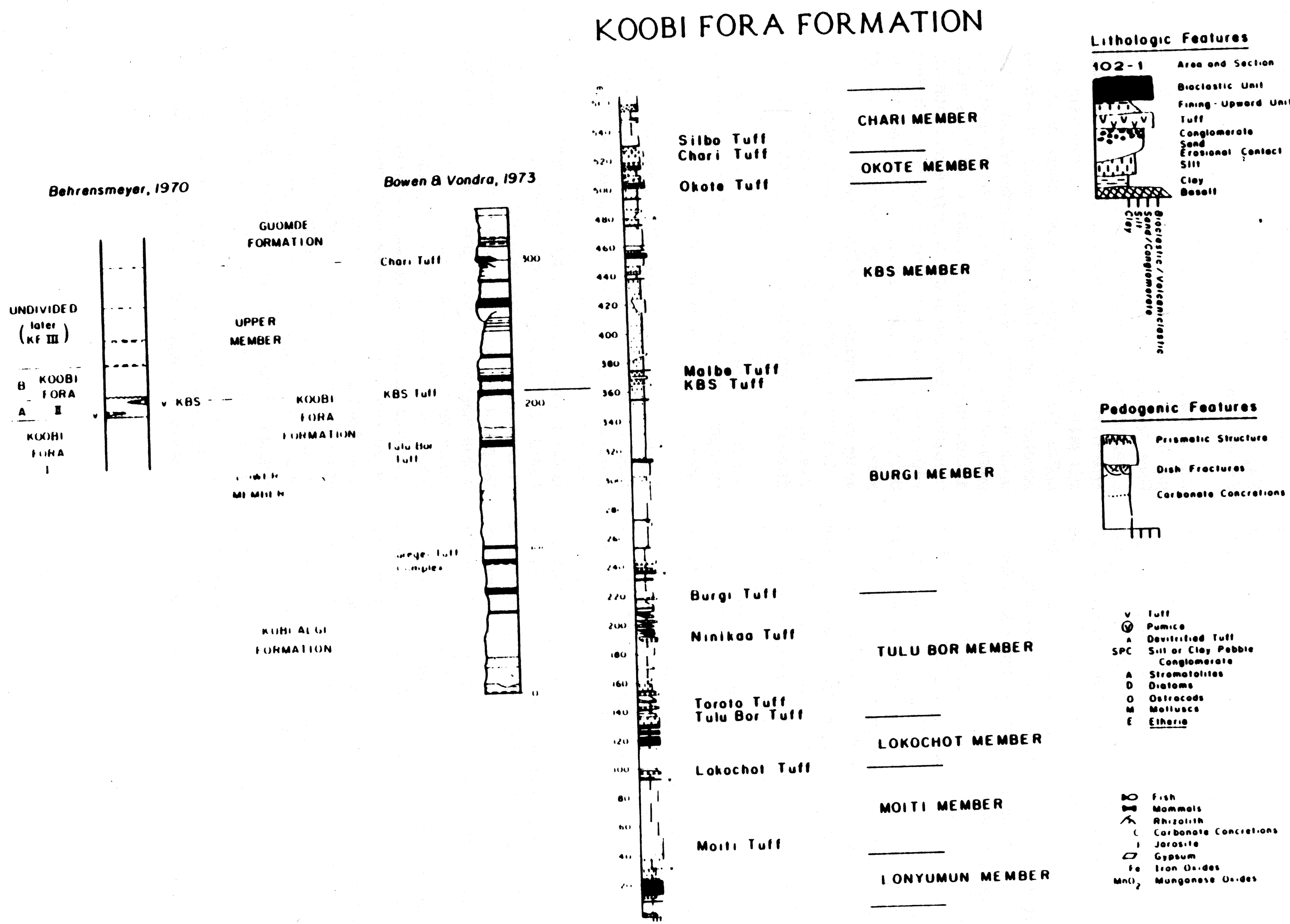

Fig 1.6 Comparison of early stratigraphic terminology to type section of the Koobi Fora Formation.

and White 1979; Williamson 1982). The Koobi Fora Formation was subdivided into members. Initially, a break was placed at the top of the KBS Tuff, delimiting a Lower Member and an Upper Member. A third, Ileret Member, was designated in the Ileret subregion, as lying between the top of the KBS Tuff and the top of the Chari Tuff, thus being a lateral equivalent of the Upper Member. Later on, these units were modified slightly. The boundary between Upper and Lower Members was moved to the base of the conglomeratic sand channel above the KBS Tuff, and the Ileret Member was subsumed into the Upper Member. Thus, the Lower Member came to equate with the original KF I and KF IIA, while the Upper Member encompassed KF IIB and KF III. In current terminology, the Lower Member is included within the upper Burgi and lowermost KBS Members, while the Upper Member falls within the upper KBS, Okote and basal Chari Members.

The Guomde Formation, as originally defined, included the sequence from the top of the Chari Tuff to the base of the Galana Boi Beds. It was believed to rest on an unconformity. More recent work suggests that the unconformity inferred at the base of the Guomde Formation is a fluvial erosion surface not unlike many others throughout the sequence. There is a significant disconformity slightly higher in the section, but there is little stratigraphical evidence of its presence. The strata referred originally to the Guomde Formation are now largely included within the Chari Member of the Koobi Fora Formation, although some strata of the former Guomde Formation are considered to be much younger, part of the late Pleistocene Galana Boi Formation.

With the development of tephrostratigraphy as a tool for correlation, geological evidence was added to the growing faunal documentation of significant miscorrelations in the established stratigraphical sequence (Brown and Cerling 1982; Cerling and Brown 1982). In order to revise the stratigraphy, however, it was first necessary to establish the correct sequence of tuffaceous units and to document in detail the intervening lithological sequence. This goal was realized in the revision of the stratigraphical nomenclature for the Koobi Fora region presented by Brown and Feibel (1986) and outlined above. An immediate advantage of the revised stratigraphy was a clear accommodation of the faunal evidence without conflicts (Fig. 1.7), and straightforward concordance with the stratigraphical sequence of the lower Omo Valley (de Heinzelin 1983), and subsequently with the Nachukui Formation (Harris *et al.* 1988*a*, 1988*b*). Based on the revised stratigraphical sequence and close correlation with related deposits around the Turkana Basin (the Plio-Pleistocene sequence of the basin is collectively referred to as the Omo Group), Feibel and co-workers (1989) were able to develop an integrated chronology for these deposits with a high degree of resolution. This integrated chronology is presented in Fig. 1.8 for reference.

Regional stratigraphy of the Koobi Fora Formation

One of the initial problems encountered in establishing the correct stratigraphical sequence of the Koobi Fora deposits was the discontinuity of exposures. Local sections are separated by areas of thorn scrub or sand rivers, or are interrupted by faulting. Whereas this presented a serious obstacle in attempting to define the stratigraphy, it does assist in its description. The isolated areas of exposure tend to occur in patches with a fairly coherent stratigraphical unity, and thus they become useful subdivisions for a discussion of the vertical and lateral variability of the formation. Because of the stratigraphical continuity within these areas, the fossil faunas collected from them tend to represent distinct assemblages but the areas do not in any way conform to distinct geographical entities of the Plio-Pleistocene.

The three subregions (Ileret, Koobi Fora, Allia Bay) used by earlier workers no longer suffice for detailed interpretation of the stratigraphy. We here consider the major features of Koobi Fora stratigraphy in terms of eight subregions which represent coherent stratigraphical entities, but which we define only loosely; their locations are indicated on Fig. 1.9.

Southern Allia Bay Plains: This subregion extends from the western margin of Sibilot to the eastern margin of Jarigole, and includes Areas 253, 254, 260, and 261. It is characterized by very subdued topography in which exposures are largely confined to the banks of major streams (Wargolo, Il Moiti, Nyongololongatuk), and small tributary gullies. On the plains the Koobi Fora Formation is covered with a thin mantle (1-2 m) of gravel and finer grained shelly deposits of Late Pleistocene or Holocene age. The type sections of the Lonyumun, Moiti, and Lokochot Members are located here, as are exposures of the Tulu Bor Tuff and the lower part of the Tulu Bor Member. Over most of the subregion the beds strike to the north-east, and dip gently to the north-west, but on the flanks of Jarigole, where the Koobi Fora Formation is in fault contact with Miocene volcanic rocks the strata strike north-east, and dip to the east. A composite section for this subregion is given in Fig. 1.10.

The basal, fine-grained sequence of the Lonyumun Member was deposited on a surface of at least 36 m

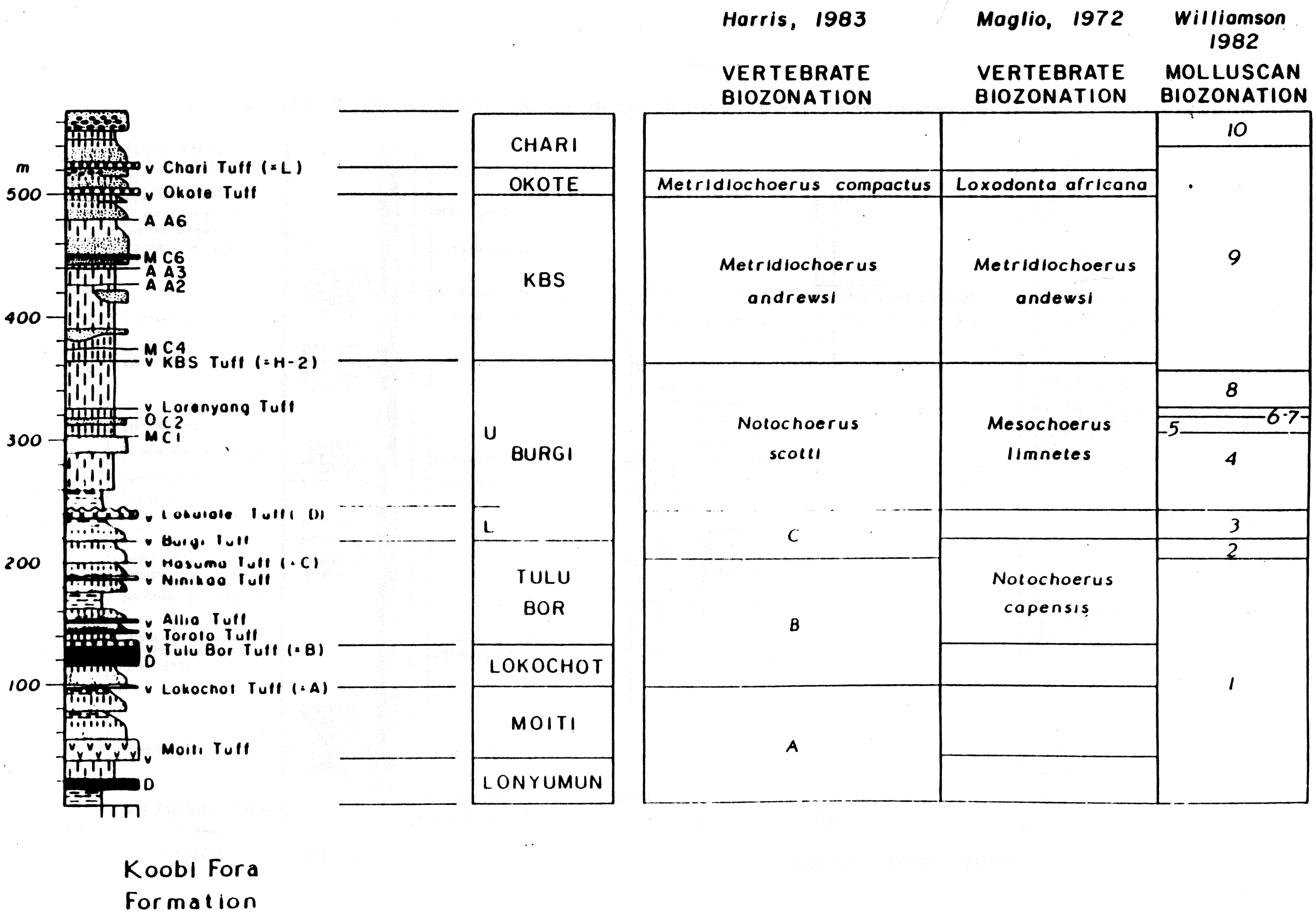

Fig 1 7 Relationship between faunal zonation schemes and stratigraphic succession of the Koobi Fora Formation.

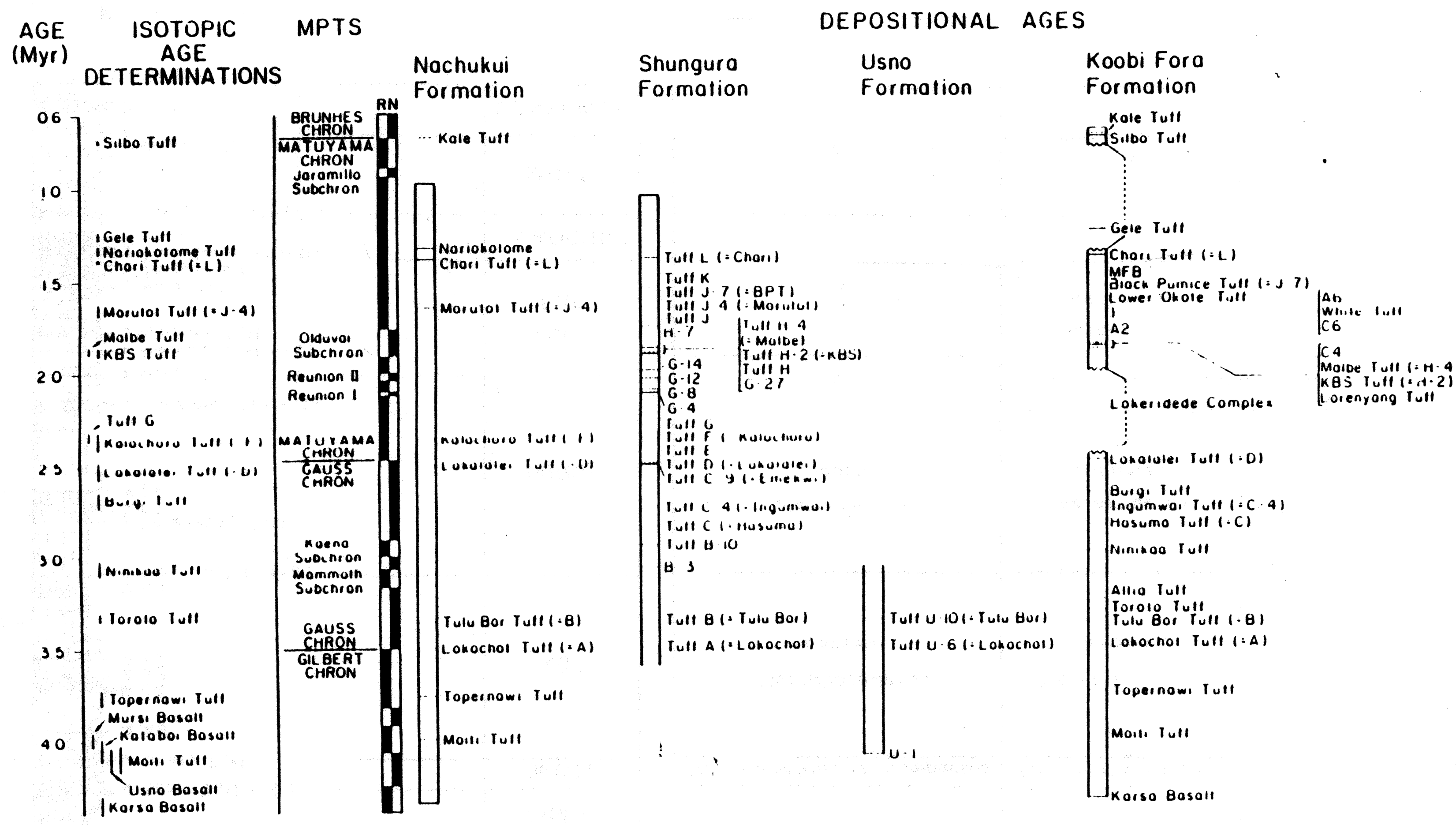

Fig. 1.8. Integrated chronometric framework for the Omo Group deposits of the Turkana Basin (from Feibel *et al.* 1989)

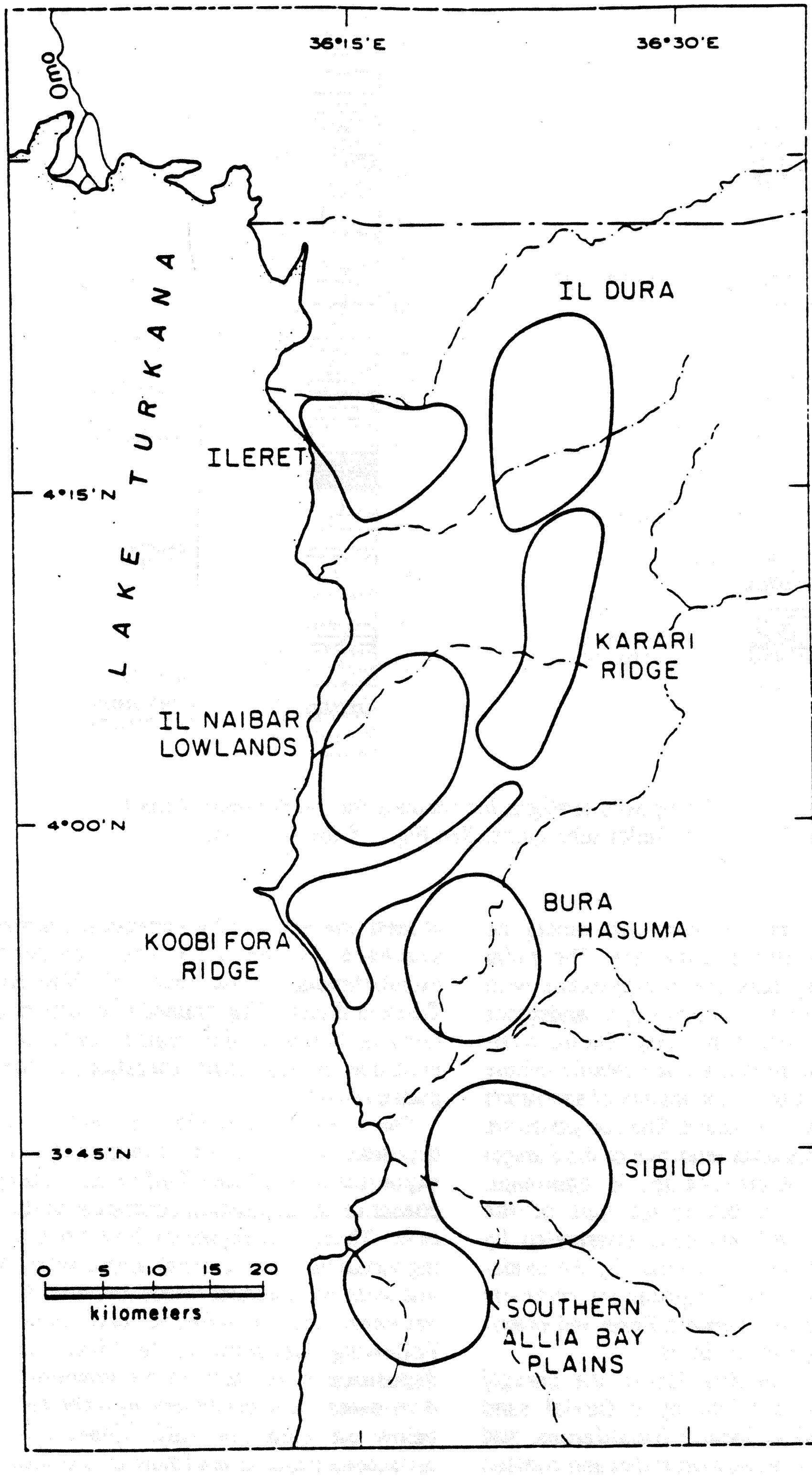

Fig. 1.9. Map of the major regions of exposure of the Koobi Fora Formation discussed in the text.

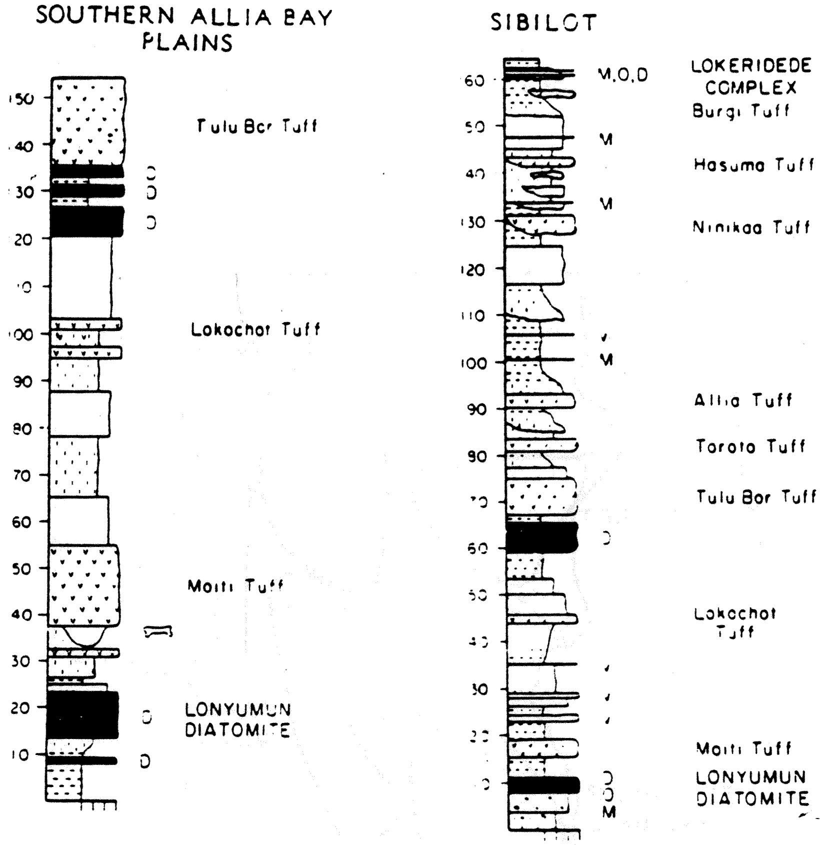

Fig. 1.10. Composite stratigraphic sections for the Southern Allia Bay Plains and Sibilot subregions. See Fig. 1.5 for symbols.

relief, which implies that the basin had undergone erosion prior to emplacement of the lake. The initial deposits are primarily pale green claystones with scattered ostracods, but also contain thin sandstones rich in ostracods and small fish bones. On the western flank of Sibilot there is a basal volcanic pebble conglomerate, overlain by a few meters of sandstone before claystone deposition ensued. The claystones are overlain by the diatomaceous sequence of the Suregei Complex, in which *Melosira* sp. is dominant. Gypsum is common in the lower half of the Lonyumun Member, and has been interpreted by Cerling (1977) as having been formed by the oxidation of sulphides. These basal deposits are extremely widespread throughout the Turkana Basin and extend beyond its present-day eastern limits.

On the Southern Allia Bay Plains, the Suregei Complex is directly overlain by a fluvial sand sequence. This sand is locally fossiliferous, and although the material is somewhat rolled and consists primarily of teeth and bone fragments, it has produced at least one noteworthy vertebrate assemblage. This sequence is the first good evidence of the establishment of the ancestral Omo River in the Turkana Basin. Fine grained overbank deposits overlying or lateral to the channel sands show the first evidence of pedogenic alteration, in the form of a palaeovertisol.

The Moiti Tuff is the first widespread fluvially deposited tephra occurring in the sequence. In most exposures of the Moiti Tuff in this subregion, three phases of sedimentation contribute to its 17 m thickness. These may represent three flood events, washing airfall ash from a distal source area. It is unlikely that these were annual floods, because rhizoliths occur between the prominent tuffaceous intervals. Following deposition of the Moiti Tuff, a style of deposition characterized by upward-fining cycles dominates, and continues upward to about 15 m below the Tulu Bor Tuff. Other less widespread tuffaceous deposits are found in this interval, most of which are reworked and confined to channels, but at least one of which contains accretionary lapilli and most likely represents deposition of airfall ash on a land surface. Vertebrate fossils recovered from this part of the section come mostly from channel sands.

Laminated diatomites of the upper third of the Lokochot Member record the presence of a short-lived lake of limited aerial extent. These are very well exposed below the Tulu Bor Tuff in Areas 250, 260, and 261, and along the western slopes of Sibilot. Diatoms are generally poorly preserved in these sediments, but the flora is known from precisely correlative strata near Kataboi west of Lake Turkana. There, the assemblages are again dominated by *Melosira* sp. Vertebrate fossils from this interval consist mainly of fish, but also include scattered, well-preserved mammalian remains near its base.

The highest stratigraphical level normally reached in this subregion is the Tulu Bor Tuff, although locally a few metres of strata are preserved above it. In this subregion this tuff is usually very coarse grained, very thick (5–17 m), and characterized by large-scale trough cross bedding at the base of an upward-fining cycle. Finer intervals within the tuff indicate that it was deposited in three major events, probably very closely spaced in time.

Sibilot Subregion: This subregion includes the area around Sibilot (Areas 250, 251, and 252), and exposures between Il Lokochot and Il Ingumwai to the north (Areas 202, 203, 204, 207, and 212). A composite stratigraphical section for this subregion is shown on Fig. 1.10. Over much of this subregion the Koobi Fora Formation is capped by coarse volcanic cobble conglomerates, many of which probably date from the Late Pleistocene, but one of which can be shown to be of Pliocene or early Pleistocene age locally. Minor outcrops of the shelly Galana Boi Formation are also present. This subregion includes the type sections of the Tulu Bor Member and of the lower Burgi Member, but also includes strata of the Lonyumun, Moiti, and Lokochot Members. In general, exposures are somewhat better than in the Southern Allia Bay subregion, but the broad active channels of Il Lokochot and Il Ingumwai, and the gravel covered surfaces, obscure relations between local sections. The Koobi Fora Formation is faulted against Miocene volcanic rocks in the eastern part of this subregion (especially good exposures of this relationship are seen in northern Area 207), and covered by alluvium and sub-Recent lake deposits along the lake shore. The strata here are broken by several large faults that strike roughly north–south. Dips may be quite steep adjacent to the faults, but otherwise trend gently to the north-west. Von Höhnel must have camped in this subregion when he hunted elephant at Allia Bay in 1888; his map shows the lake as being far more extensive at that time (von Höhnel 1890).

South of Il Lokochot, exposures consist dominantly of fine grained sediments of the Lonyumun Member and of fluvial sediments of the Moiti and Lokochot Members. It is here that relief on the surface below the Koobi Fora Formation can be shown to have been at least 36 m, because a continuous mollusc-packed sandstone is exposed over that vertical extent in contact with Miocene volcanic rocks. The Moiti Tuff in this area consists of a basal pure vitric tuff 20–30 cm in thickness, overlain by fluvially reworked ash deposits that contain large, but rare, pumice clasts in the upper part. As noted by Brown *et al.* (1985), the pumice clasts differ compositionally from the Moiti Tuff, and may represent products of an earlier eruption. There is a good cliff exposure along Il Lokochot (Area 252) in which the Moiti, Topernawi, and Lokochot Tuffs are exposed in sequence. The Moiti Member in this section is only about 16 m thick, roughly one quarter of the thickness in its type section. The Topernawi Tuff is present only in this section at Koobi Fora, so that internal correlations within the Moiti Member from this subregion to the southern Allia Bay Plains are not straightforward.

Exposures in Area 212, are limited to short sections exposed in stream banks. By contrast Area 207 contains superb exposures of the lower part of the Tulu Bor Member along Il Toroto, and of the lower Burgi Member along Il Ingumwai. The Tulu Bor Member in this area includes the Tulu Bor Tuff, the Toroto Tuff (deposited in two layers about 4 m apart separated by a fluvial sequence), the Allia Tuff, and the Waru Tuff, the latter deposited in a lacustrine sequence. It appears that the lower part of the section (the Lonyumun, Moiti, and Lokochot Members) is absent, or very thin in the southeastern part of the area, for the Tulu Bor Tuff lies only 10 m above a basalt flow.

A cliff about 60 m high along Il Ingumwai provides the type section of the lower Burgi Member, and also exposes the upper part of the Tulu Bor Member. Strata in this interval are exclusively fluvial, and several well-developed palaeosols are clearly displayed at the tops of upward-fining sequences.

The lower part of the type section of the Tulu Bor Member is located in Area 204, and the upper part lies in Area 202, where the section continues upward into the lower part of the Burgi Member. Four tuffs present in the lower part of the Tulu Bor Member of Area 207 are also exposed here, the lowest three of which were referred to the Allia Tuff by Findlater

(1976), but later distinguished and renamed by Brown and Cerling (1982). The Ninikaa Tuff is present filling a channel higher in the section, and the Hasuma Tuff also occurs as a channel fill near the boundary between Areas 204 and 202. The Nile oyster, *Etheria*, is first encountered in the section at the base of channel sands above the Tototo Tuff in Area 204, and is common in similar situations higher in the member. The section is dominantly made up of fluvial channel sandstones with associated finer sediments, but there is evidence for minor lacustrine deposition between the Allia Tuff and the Ninikaa Tuff. This lacustrine episode is marked by poorly preserved molluscs and an airfall deposit of the Waru Tuff.

In Area 202, the Burgi Tuff is present, and a meter above it is a distinctive molluscan assemblage that lies within the Lokeridede Complex (Williamson 1981, 1982; Brown and Cerling 1982). We believe that, although this assemblage is closely associated with the unconformity between the lower and upper Burgi Members, it lies within the lower Burgi Member. Deposits of clearly lacustrine origin are not well known from this time interval but do extend as far south as Loiyengalani.

Bura Hasuma: This subregion encompasses the exposures spread broadly to the north of Bura Hasuma Hill, including most of Areas 107, 110, 119, 120, 121, 123, and 124. Fig. 1.11 depicts a composite stratigraphical section from this subregion. The upper Burgi and KBS Members are exposed here, and the Okote Tuff Complex is is present in the north (Area 119). Strata here are largely flat-lying, although minor faulting is locally abundant.

The upper Burgi Member strata belong to Stages 4 and 5 (Feibel 1988), and contain many beds with abundant invertebrate fossils, predominantly ostracods and molluscs. Vertebrate fossils are generally rare except in the uppermost part of the upper Burgi Member. The *Elephas recki atavus* skeleton preserved *in situ* in a National Museums exhibit lies in the basal part of the KBS Member. The KBS Member here includes the sequence of marker beds recognized from the Koobi Fora Ridge (Feibel 1983), including cryptalgal biolithites and bioclastic sandstones, but in a much condensed section. These marker beds extend well to the east but they continue to become thinner and disappear before the basin margin is reached. Tuffs are poorly preserved in this subregion, but the KBS and Malbe Tuffs are firmly linked to the bioclastic marker horizons that are more useful in this subregion. Vertebrate fossils occur mainly in the upper part of the KBS Member in this subregion, between marker beds A4 and A5, and include some relatively complete skeletons. The *Euthecodon* skeleton in Area 119, exposed by the National Museums of Kenya for *in situ* exhibition, lies between marker beds A4 and A5; a fossil tortoise (*Geochelone* sp.), excavated nearby for the same purpose, lies just above marker bed A5. The Okote Member cropping out in Area 119 includes the Black Pumice Tuff and an overlying tuff of the Okote Complex (Brown and Feibel 1985).

That portion of the eastern basin margin between Shin and Kubi Algi has been little studied. Bowen (1974) first noted that the disconformity between the lower Burgi Member and the upper Burgi Member is well represented in this vicinity. Strata from the interval between the Moiti and Burgi Tuffs lie below the disconformity, and overlying strata represent the upper Burgi, KBS, and Okote Members. The upper Burgi Member here is comprised of laminated siltstones, with sparse, thin mollusc-rich layers, overlain by planar bedded sandstones and siltstones that continue through the transition to the KBS Member. Conglomerates and coarse sandstones, with one prominent stromatolite layer, occur immediately beneath a tuff of the Okote Complex and are assigned to the KBS Member. Basalt cobble conglomerates, which occur in Area 200, imply proximal relief on the basin margin to the east of this area during KBS Member times.

Koobi Fora Ridge: This subregion consists mainly of the north and north-west facing scarps stretching from Area 103 to Area 115 (see Fig. 1.3). A composite stratigraphical section for this subregion is given in Fig. 1.11. Strata along the escarpment belong to the lower Burgi, upper Burgi, and KBS Members, and include the type sections of the latter two units. The upper Burgi and KBS Members are thicker and more complete in Area 102 than in any other part of the Koobi Fora region. A series of low hills in the northern part of Area 102 have yielded suid fossils that demonstrate the presence of strata belonging to the Tulu Bor Member, and the upper part of this sequence contains molluscan fossils that belong to the 'Suregei Isolate' of Williamson (personal communication). Faulting is locally intense in this subregion, and the attitude of strata accordingly variable. Farther from the lake, however, the frequency of faulting decreases and strata are mostly flat-lying.

Upper Burgi Member strata (Stages 1–4 of Feibel 1988) here rest disconformably on the older strata, and represent deeper lake portions of the upper Burgi lake and a series of infillings by deltaic lobes. It is one of these lobes in which the Lorenyang Tuff, formerly confused with the Tulu Bor Tuff, was deposited.

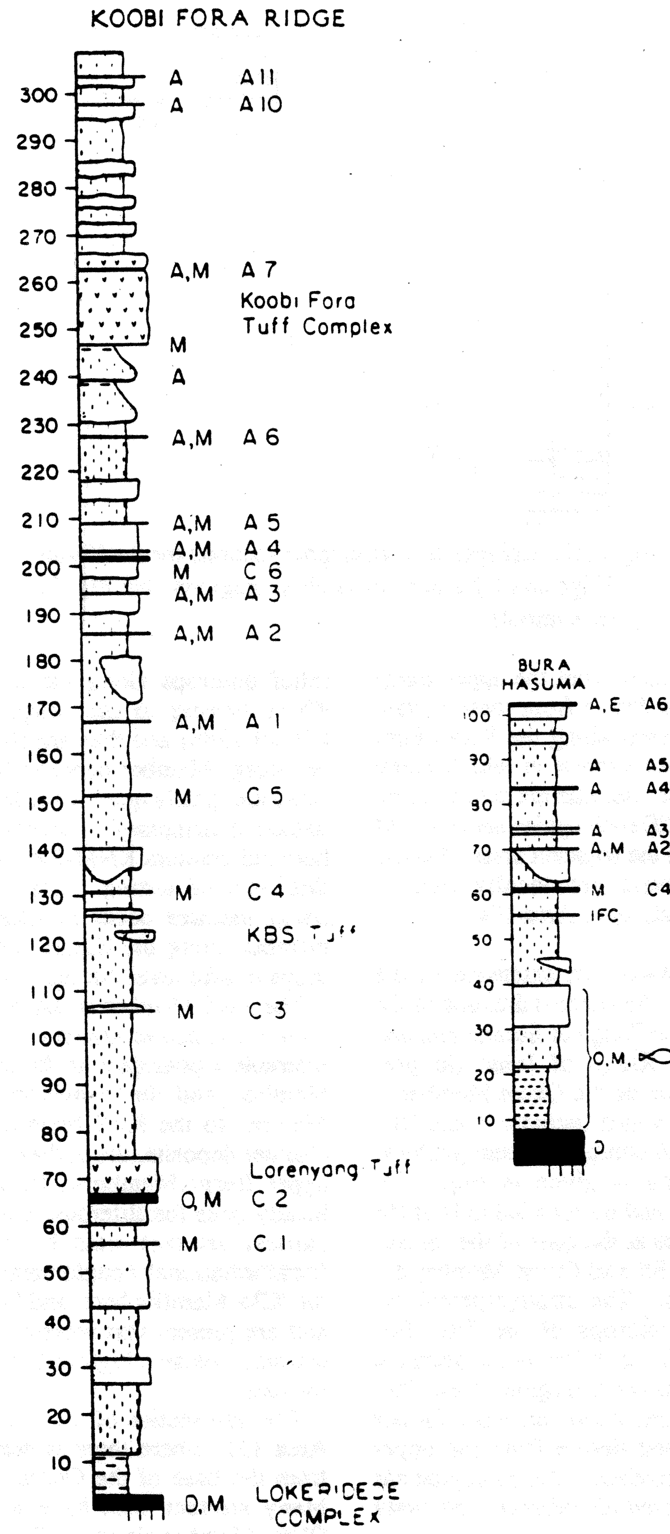

Fig. 1.11. Composite stratigraphic sections for the Bura Hasuma and Koobi Fora Ridge subregions. See Fig. 1.5 for symbols.

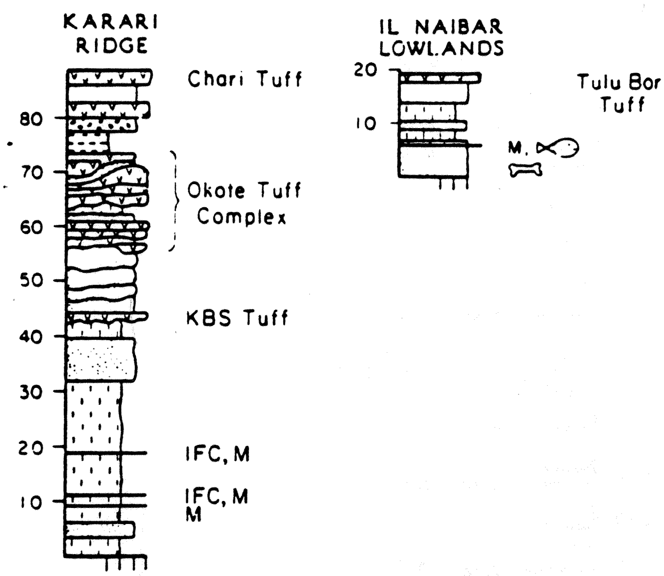

Fig. 1.12. Composite stratigraphic sections for the Karari Ridge and Il Naibar Lowlands subregions. See Fig. 1.5 for symbols.

Almost all mammalian fossils from the upper Burgi Member in this subregion derive from these deltaic lobes. KBS Member deposits along the Koobi Fora Ridge represent a complex system in which many shallow (1–2 m) lacustrine intervals are separated by fluvial sequences. About 20 such cycles are recorded in Area 102, with fewer in the adjacent areas. Fluvial channels in these intervals are locally rich in mammalian fossils, particularly in Area 104.

Karari Ridge: The north-eastern extension of the Koobi Fora Ridge, exposed by streams draining to the west, is known as the Karari Ridge or Escarpment and includes the Aberegaya Ridge of Isaac (in preparation). The type section of the Okote Member is located in this subregion, which includes Areas 105, 118, 129, 130, and 131. A composite stratigraphical section for this subregion is given in Fig. 1.12. Exposures are here dominated by rounded hills of the upper Burgi Member strata at the base of the escarpment, with badlands of KBS and Okote Member deposits forming the scarp. The stratigraphical sequence includes small outcrops of the Tulu Bor Member and older strata in north facing exposures at the northern termination of the subregion (Area 129). Most vertebrate fossils are, however, from farther south in the subregion and derive from the upper Burgi, KBS, and Okote Members. The subregion has seen relatively little structural disturbance, and strata are nearly flat-lying.

The upper Burgi Member deposits occur as low-relief outcrops along the base of the escarpment. These belong to the Stage 5 depositional phase (Feibel 1988) and thus are slightly younger than upper Burgi Member deposits farther south. Here the strata are locally quite fossiliferous, particularly where deltaic distributary channels are represented. The hominid cranium KNM-ER 1470 was recovered from this type of setting. Distinctive intraformational conglomerates occur in lake margin facies of the member along the Karari Ridge, and attest to brief drops in lake level through this interval.

The KBS Member is characterized by fluvial sands with a distinctive gravel component. There is considerable topography on the basal surface of the KBS Member, and the transition from the upper Burgi Member to the KBS Member is marked by fluvial channel deposits, the earliest of which lie within the upper Burgi Member. These fluvial channels are locally quite fossiliferous, and are associated with the earliest artefact sites known from Koobi Fora. Intraformational conglomerates are common within the KBS Member here, and basalt pebble conglomerates are present that represent deposition by streams draining volcanic highlands along the basin margin to the east.

The type section of the Okote Member is located in Area 131, where there is nearly complete exposure from the base of the Okote Tuff to the Chari Tuff. Many artefact sites have been located within the Okote Member along the Karari Ridge, and the member is quite fossiliferous. Lithologically the member is the most heterogeneous of any in the Koobi Fora Formation. Small channels filled with basalt pebble conglomerates are widespread, and tuffs and tuffaceous sandstones that grade into siltstones are also abundant. Tabular layers of calcium carbonate occur at the tops of many of the finer grained beds, and are interpreted as soil carbonates (caliche horizons). The Chari Tuff is the highest unit exposed in this subregion, and contains large pumice clasts (up to 60 cm in diameter) at nearly all outcrops.

Il Naibar Lowlands: The broad reach of low relief (and little exposure) between the Kokoi and the Karari Ridge is here termed the Il Naibar Lowlands, after the major sand river on its north-west side. The subregion includes most of Areas 116, 117, northern Area 136, and the south-eastern edge of Area 137. A composite section from this subregion (Fig. 1.12) ranges from the Lonyumun Member through the upper Burgi Member. The strata exposed become generally younger from west to east, as the elevation increases.

The Lonyumun Member is represented by the Suregei Complex diatomites. The Moiti Member is poorly represented by strata that include a tuff similar in composition to pumices from the Moiti Tuff farther south. The Lokochot Member is exposed in the centre of the subregion and represented by two distinct facies. To the west (Area 117), the upper Lokochot Member deposits are laminated silts and fine sands in the form of coarsening-upward cycles; these are interpreted as representing the deltaic infilling of a short-lived lake. Diatomites, so prominent below the Tulu Bor Tuff in the Allia Bay subregion, are absent. Farther east, there is no evidence of lacustrine deposits; instead a series of fluvial upward-fining cycles occur beneath the Tulu Bor Tuff. These represent the north-eastern limit of the Lokochot Member Lake. The Tulu Bor Tuff is fairly widespread in the subregion, but superjacent deposits of the Tulu Bor Member are only locally preserved. They typically occur as a stacked series of upward-fining cycles, locally very fossiliferous. The only named tuff, other than the Tulu Bor itself, occurring here in this member is the Ninikaa Tuff. The lower Burgi Member is represented by the distinctive sequence of the Lokeridede Complex in northern Area 116. The Burgi Member unconformity has eroded down to the level of the Burgi, Ninikaa, or Tulu Bor Tuffs in different parts of the subregion. To the east, upper Burgi Member sediments deposited above the unconformity include a thick prodeltaic and marginal lacustrine sequence similar to that of the Karari Ridge.

Strata belonging to the Lonyumun, Moiti, and Lokochot Members are poorly exposed on the lower parts of the Kokoi highland north of Il Naibar. These dip to the south-east, having been elevated along with the Kokoi fault block. The basal contact of the Koobi Fora Formation here consists of dark brown claystones resting on basalt, and these are overlain by the altered diatomites of the lower Lonyumun Member. The Lokochot Tuff in these exposures contains very well preserved accretionary lapilli.

Ileret: The Ileret subregion forms a prominent topographical high south of the Il Eriet, surrounded by a dissected rim of exposures. The subregion is bounded to the south and east by the channel complex of the Kolom Guomde. It includes the cluster of collecting Areas 1–12. A composite section of strata exposed here is given in Fig. 1.13. These sediments represent the KBS, Okote, and Chari Members, including the type section of the latter. Exposures in this subregion are heavily faulted, and attitudes fluctuate accordingly.

Strata of the KBS Member, typically exposed far down the dissecting drainages, consist of a series of upward-fining cycles, with several intercalated mollusc-bearing sandstones in the lower part of the member. These exposures are distinctively banded, with light yellow sands alternating with brown mudstones. Channel-form tuffs within the brown mudstones mark the lowest part of the Okote Member. Slightly higher in the section, the Okote Member strata take on a distinctive appearance with tuff-filled channels and tuffaceous silts lateral to them. Near the top of the Okote Member, a distinctive yellow fluvial sand is overlain by reddened mudstones. This channel sand is the source of many of the fossils from the Ileret subregion. The reddened mudstones are overlain by a series of upward-fining cycles capped by the Chari Tuff. The Chari Member exposures in the subregion include several thick upward-fining cycles, and a series of thin coarsening-upward sequences with abundant fish remains in the capping sands. Several molluscan and ostracod sandstones associated with the latter appear to represent shallow lake fillings.

The plains south of Ileret Ridge expose little section, except near the Kokoi highland, where strata of the upper part of the Chari Member are tilted almost vertically along the fault contact with the Kokoi basalts. These strata include the Silbo Tuff and contain a well-preserved endemic molluscan fauna (Williamson 1981).

Il Dura: The western flanks of the Suregei Plateau expose a sequence of sediments over what is here termed the Il Dura subregion, named for the promin-

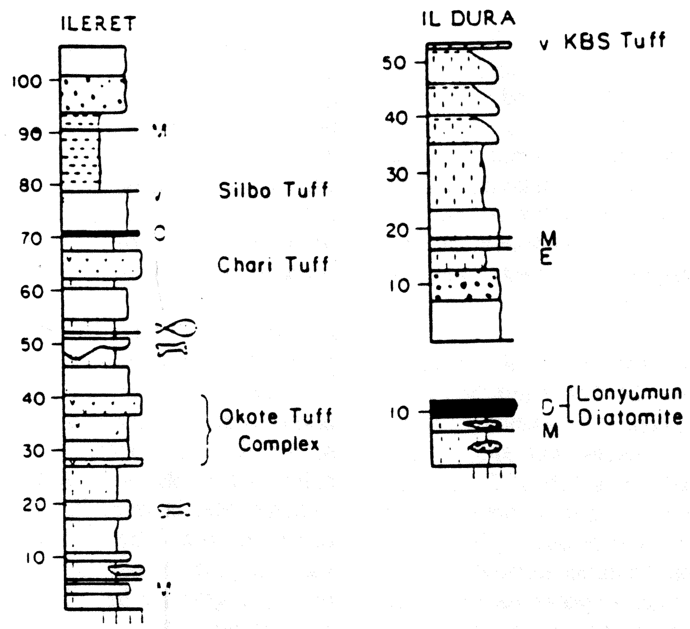

Fig. 1.13. Composite stratigraphic sections of the Ileret and Il Dura subregions. See Fig. 1.5 for symbols.

ent sand river which drains it. The subregion includes Areas 14, 15, 41, and 42. A composite section of this subregion is given in Fig. 1.13. The stratigraphical sequence in this subregion includes widespread exposures of the Lonyumun Member and patches of upper Burgi and KBS Member strata. These sediments were deposited unconformably over a partially eroded basalt terrain. Lower portions of the sedimentary sequence show effects of mantling this pre-existing topography. The northern limits of these exposures have not yet been examined.

The Lonyumun Member is represented primarily by fine-grained deposits with local lenticular interbeds of basalt conglomerate near hills on the Pliocene basal surface. Locally the basalt surface is mantled by thick accumulations of gastropods, similar to those seen east of Sibilot. The most visible interval of the Lonyumun Member in this subregion is the diatomite of the Suregei Complex, which includes a relatively well-preserved and diverse diatom flora. Upper Burgi Member deposits lie disconformably on the Lonyumun Member, and include laminated siltstones and molluscan sandstones. These are overlain by the KBS Tuff and an interval of thin upward-fining cycles. A thick conglomerate lies 23 m below the KBS Tuff in a section north of Il Biliet, and probably represents deposition on an ancient alluvial fan of the Il Eriet.

DEPOSITIONAL ENVIRONMENTS

The depositional environments preserved in the Koobi Fora Formation furnish many data for interpreting the palaeoenvironments present in the region during the Pliocene and Pleistocene. Although a balanced palaeoenvironmental reconstruction needs to integrate contributions from geological and palaeontological studies, the intent here is to present the geological evidence for such a goal. The ensuing discussion of depositional environments is based on a sedimentary facies analysis (Feibel and Brown 1986; Feibel 1988), but is couched in terms of environmental facies. This allows us to develop a more appropriate context for discussion of the fossil faunas elsewhere in this volume. Five major environments are represented by the deposits of the Koobi Fora Formation: fluvial channel, fluvial floodplain, delta, lake margin, and lake basin.

Fluvial channel facies: The fluvial channel deposits of the Koobi Fora Formation originated either as products of the large perennial channels of the axial drainage network (the ancestral Omo River), or as those of the smaller ephemeral systems draining the local basin margins. Both are dominated by coarse clastic detritus, sands and gravels, but may be distinguished by the relative proportions of clast sizes and by different clast lithologies and palaeochannel sizes.

The axial drainage system dominated the basin geography during the interval between 4 to 2.5 Ma. After a subsequent lacustrine episode, it reappeared episodically in modified form from 1.9 Ma to about 0.6 Ma. The most common deposits of this system are upward-fining cycles, in which erosional basal scour is overlain by gravel lag and trough cross-bedded coarse to medium sands, and then capped by horizontally laminated and ripple-laminated fine sands and silts. Gravel lags associated with these channels are invariably composed of granule to small pebble sized clasts of quartz and perthitic feldspar, and of silt or clay pebbles derived from erosion of mud drapes and the adjacent floodplain deposits. This sequence grades upwards into a relatively thick fine-grained sequence representing floodplain deposits. The best example of channel geometry is illustrated in exposures of the Tulu Bor Tuff in Area 250. There the magnitude of the ancient channel can be determined as being about 9.5 m in depth and some 220 m across (comparable in size to the channel of the modern Omo River). These deposits represent the lateral migration of a large meandering river across a floodplain. A second, less common variant of the axial drainage system is dominated by planar crossbedded sands with occasional interbeds of horizontally laminated sand. The erosional basal scour is only shallowly developed, and superposed fine-grained floodplain deposits are very thin. This variant represents the channels of a braided stream. The two styles of fluvial channel deposits were first recognized by Vondra and Burggraf (1978) from their studies along the Karari Ridge, but are well expressed throughout the formation.

Lateral drainages issuing from the eastern basin margin are first recognized in the upper Burgi Member, and persist locally throughout the remainder of the formation. The primary depositional features represented in these marginal drainages are similar to those of the axial system but smaller in scale. However, the braided channel form is more common in these deposits. Their most striking characteristic is the relatively high proportion of basalt clasts, which range in size from large cobbles to fine sand. At the basin margin most of the clasts are orthomictic basalt, but the proportion of quartzo-feldspathic clasts introduced by the axial system increases rapidly towards the basin centre. Where the lateral drainages join the axial system, gravel bars are formed from the volcanic components, which are significantly larger than anything carried by the axial drainage. The lateral drainages did not extend far into the basin during deposition of the Koobi Fora Formation, in striking contrast to the present when such channels extend nearly to the modern lake. The marginal drainage channels are relatively small (metres in depth, tens of metres in width), and have very little associated floodplain material.

Fluvial floodplain facies: The floodplains that developed lateral to the fluvial systems vary in accordance with the character of the rivers and, to a certain extent, with time. They are characterized by relatively structureless fine-grained deposits, often with associated white carbonate accumulations. They have been affected by varying degrees of pedogenesis (soil formation) of several kinds.

The most common form of pedogenesis recorded in the floodplain facies is a combination of structural features with a zone of large dish-shaped fractures with slickensided surfaces at the base that is overlain by a zone of sand-filled crack networks, occasionally with a gilgai microtopography at the ancient soil surface. This combination of features represents a palaeovertisol, or ancient black cotton soil (Feibel 1988) and is the most common feature of the floodplain deposits throughout the formation.

Hydromorphic palaeosols, those in which the downward movement of water in the soil profile has dominated development, are relatively rare at Koobi Fora. They are characterized by a strongly developed prismatic structure, and a concentration of clays in the subsoil (Bt) horizon. The best developed of these palaeosols occurs in the middle of the Lokochot Member; several others are present in the upper Burgi Member.

Accumulations of soil carbonate characterize many exposures of the floodplain facies at Koobi Fora. These range from discrete nodules several centimetres in diameter to caliche horizons more than a meter thick. The most extensive soil carbonate accumulations are associated with the marginal channels of the KBS and Okote Members along the Karari Ridge. Although elsewhere interpreted in terms of and climatic implications, the well-developed carbonate-rich soils of the Koobi Fora Formation show strong association with basin margin fluvial systems that drained calcium-rich source areas.

Delta facies: Deltaic accumulations at Koobi Fora are very restricted in distribution and are characterized by subaqueous deposits in which a high clastic input, commonly manifested by coarsening-upward cycles, dominates sedimentation. The more distal, or prodeltaic, portions of the sequence are typically orange-brown mudstones, often laminated, with thin interbeds of fine sand. Grain size and bedding coarsen upwards, and channel-form sand bodies are locally incised into the mudstones. A common feature of the

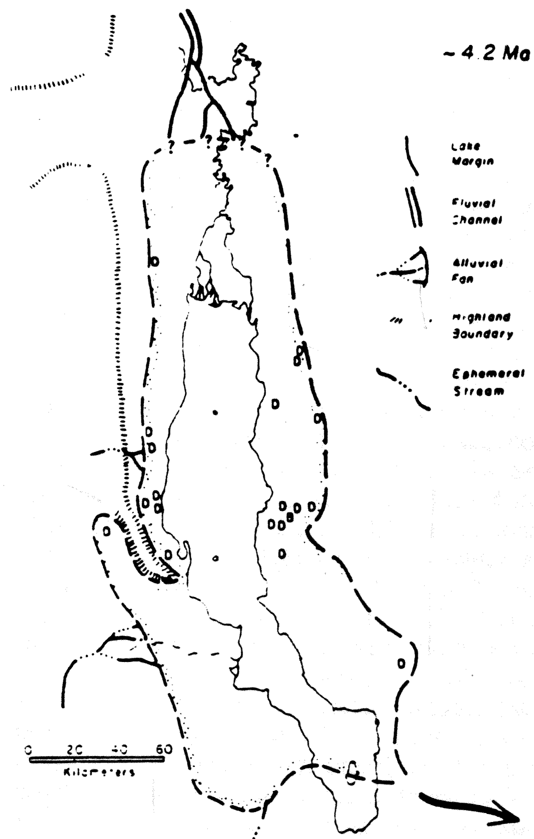

Fig. 1.14. Palaeogeographic reconstruction of the Turkana Basin at 4.2 Ma. Outline of modern Lake Turkana included for reference. D indicates exposures of diatomite, B indicates exposures of beach sequences.

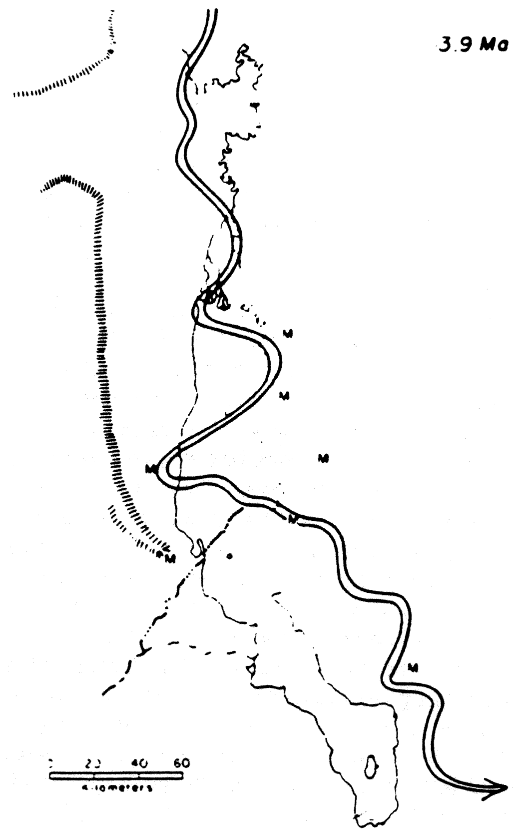

Fig. 1.15. Palaeogeographic reconstruction of the Turkana Basin at 3.9 Ma. M indicates exposures of the Moiti Tuff. Outline of modern Lake Turkana included for reference.

deltaic facies in the upper Burgi Member is a sequence of inclined sands and bioclastic carbonates which represent the foreset beds of Gilbert-type deltas. The channel-form sand bodies of deltaic distributaries are important sources of vertebrate fossils in some areas.

The subaerial portions of the 'delta plain' are, in practice, impossible to distinguish from the laterally equivalent fluvial deposits, and no attempt to do so will be made here. The term 'deltaic' has been so over-used in previous discussion of the Koobi Fora deposits that we here use this term only with extreme reluctance. Part of the reason for earlier usage may have been that modern geography assumed a disproportionate role in interpretation of past environments — exposures near the modern lake were considered to have been proximal to a lake in the past and thus to represent deltas. In this way interpretation of some of the more complex intervals has been obscured, and the stratigraphical record was over-interpreted. As used here, 'deltaic' refers only to the subaqueous sequence described above.

Lake margin facies: Lake margin environments varied considerably over the time-span represented by the Koobi Fora deposits. The Koobi Fora lake margin facies are dominantly low-energy accumulations that contrast sharply with the high-energy deposits in the Nachukui Formation west of the modern lake (Harris *et al*. 1988*a*). The most prominent marginal deposits are molluscan packstones, cryptalgal biolithites, and calcareous sandstones. These are interbedded with laminated and/or massive finer grained clastics. The sandy beach facies can be important sources of vertebrate fossils, such as the well-preserved, articulated *Euthecodon brumpti* skeleton of Area 119. Lake margin facies are not known from the Lokochot Member lake or from the short lacustrine interval within the Tulu Bor Member.

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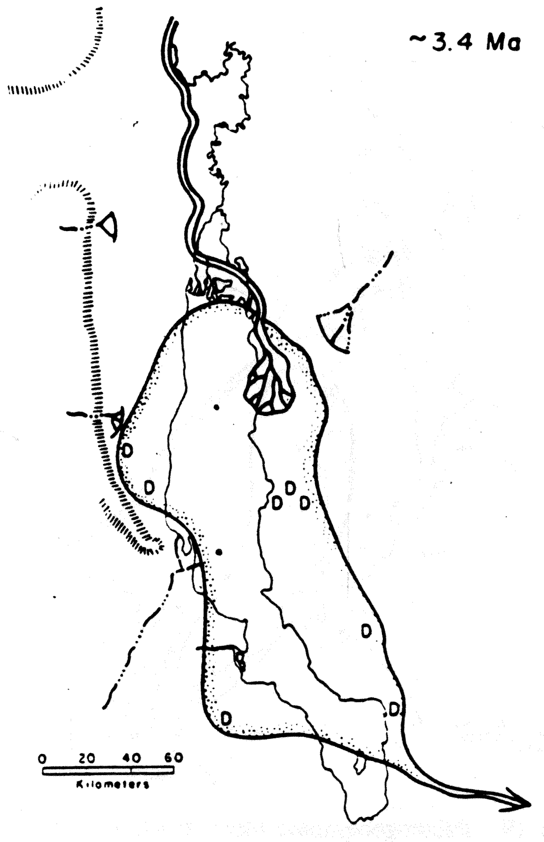

Fig. 1.16. Palaeogeographic reconstruction of the Turkana Basin at 3.4 Ma. D indicates exposures of diatomite. Outline of modern Lake Turkana included for reference.

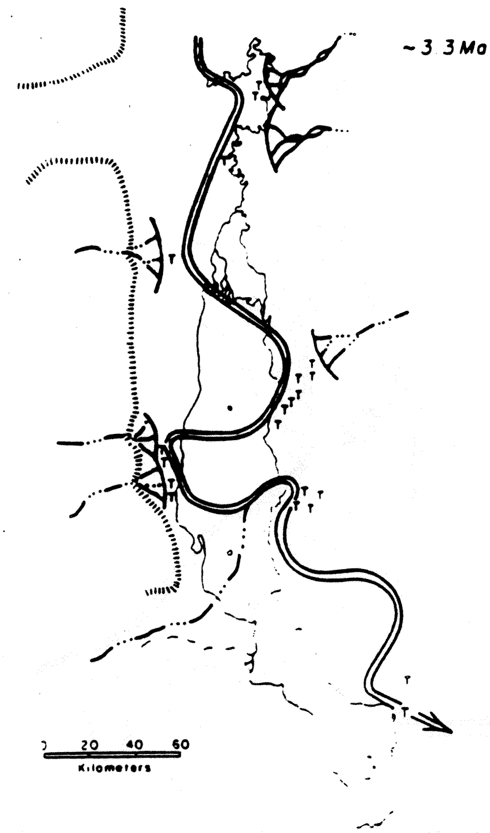

Fig. 1.17. Palaeogeographic reconstruction of the Turkana Basin at 3.3 Ma. T indicates exposures of the Tulu Bor Tuff. Outline of modern Lake Turkana included for reference.

Lake basin facies: Basinal lacustrine deposits are relatively rare at Koobi Fora. They are characterized by pale green claystones, often with scattered ostracods. Vertebrate fossils other than fish are almost totally lacking from these deposits, which represent portions of the lake in which clastic input, either from deltaic sources or biogenic production, is low. Such conditions are most typical of areas of relatively deep water (sub-littoral) away from the lake shore. The scarcity of deep lake deposits suggests that lakes were generally shallow in the Koobi Fora region.

PALAEOGEOGRAPHY

With the integration of the Plio-Pleistocene deposits of the Koobi Fora, Shungura, Usno, and Nachukui Formations (the major Omo Group deposits) into a basin-wide sequence representing a single depositional system, reconstruction of the regional palaeogeography became a possibility for the first time. The resultant topographical reconstructions, first presented by Brown and Feibel (1988), differ significantly from earlier attempts. Our interpretations are presented here as a series of topographical maps constructed for the region at 4.2 Ma, 3.9 Ma, 3.4 Ma, 3.3 Ma, 2.3–2.4 Ma, 2.2 Ma, 2.0 Ma, 1.89 Ma, 1.88 Ma, 1.76 Ma, 1.65 Ma, 1.49 Ma, and 1.39 Ma (Figs 1.14-1.26).

The temporal intervals chosen for illustration depict a sequence of topographic changes within the basin. The maps may be somewhat misleading because they emphasize the lacustrine phases of basinal history. This is mainly due to the well-preserved record we have for these intervals, and to the dramatic changes they represent. The fluvial phases were far more dominant in the long term, representing the main depositional agent for over 85 percent of the time represented by deposits in the basin.

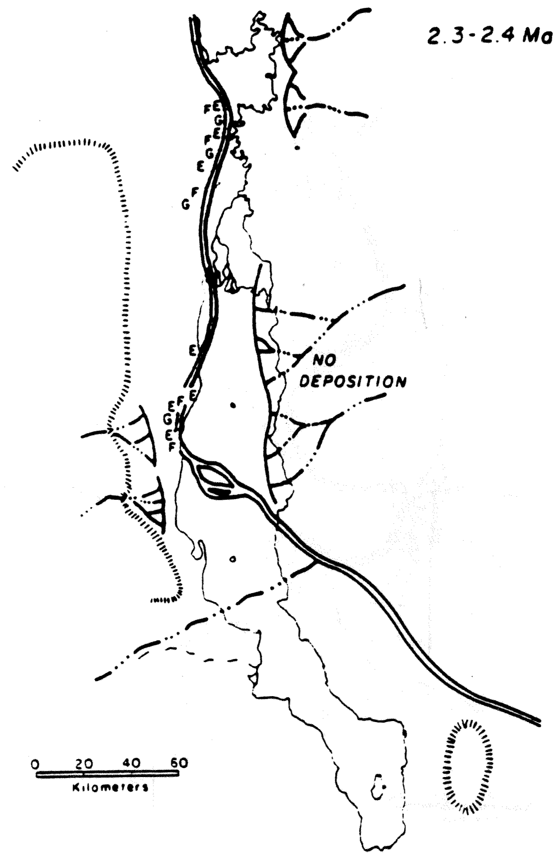

Fig. 1.18. Palaeogeographic reconstruction of the Turkana Basin at 2.3-2.4 Ma. E, F and G indicate exposures of Tuffs E, F, and G respectively. Outline of modern Lake Turkana included for reference.

Fig. 1.19. Palaeogeographic reconstruction of the Turkana Basin at 2.2 Ma. Dots represent lacustrine sediments. Outline of modern Lake Turkana included for reference.

The history of hydrographical linkages for the Turkana Basin is difficult to reconstruct. The Omo River has been flowing down from the Ethiopian Highlands for over 5 million years (Wolde-Gabriel and Aronson 1987) and for most of that interval it has flowed into the axial Turkana Basin. For the first 2.5 million years represented by Omo Group deposits, the river probably flowed through the basin and exited to the south-east. This exiting river has been termed the Turkana River (Feibel 1988). The only times at which the presence of the Turkana River can be demonstrated to occur are at 3.3 Ma (Fig. 1.17), when the Tulu Bor Tuff was deposited in the Loiyengalani region, and at 1.9 Ma (Fig. 1.20) when an immigrant stingray, *Dasyatis africana* (Feibel 1988), entered the basin from the Indian Ocean. At some time after 1.9 Ma, the basin's southeast outlet was closed. The inflow of the Omo River into the basin also seems to have decreased episodically after about 1.7 Ma, and particularly between about 0.6 to 0.1 Ma. The most likely cause of such decrease is temporary diversion of the Omo into headwaters of the Nile, perhaps via the north-western edge of the Turkana Basin. There appears to have been sporadic sedimentation in the basin during the Middle and Late Pleistocene with floods opening channels into the basin and then choking them with sediment. This has resulted in a very complex, and as yet not fully resolved, stratigraphical record for the later portion of the basinal history.

DISCUSSION

At the present time, Pliocene and Pleistocene strata are found both east and west of Lake Turkana at elevations of up to 160–165 m above the present lake level. The bold escarpment formed by Miocene and older rocks of the Labur and Murua Rith ranges west

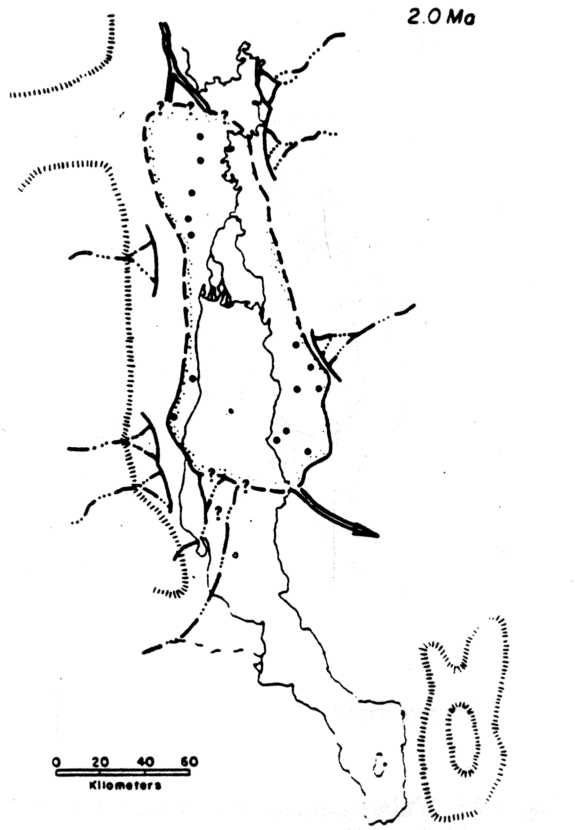

Fig. 1.20. Palaeogeographic reconstruction of the Turkana Basin at 2.0 Ma. Dots represent lacustrine sediments. Outline of modern Lake Turkana included for reference.

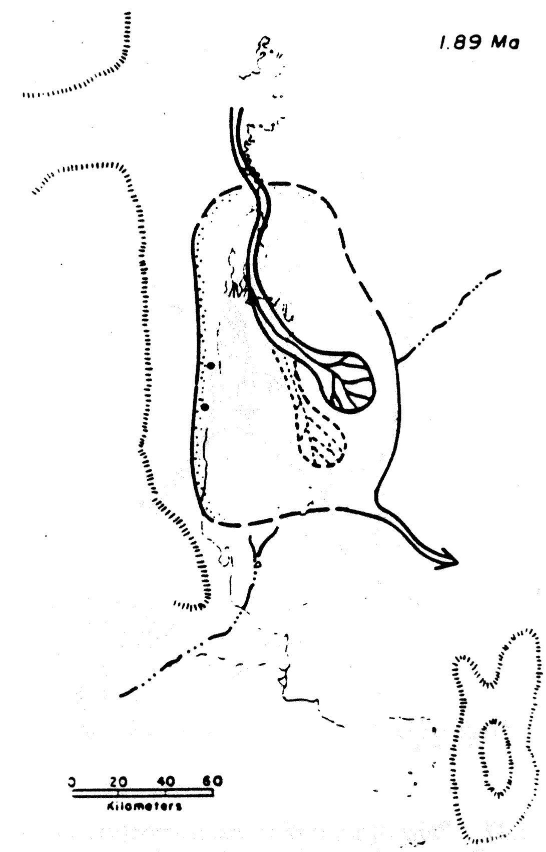

Fig. 1.21. Palaeogeographic reconstruction of the Turkana Basin at 1.89 Ma. Dots represent lacustrine sediments. Outline of modern Lake Turkana included for reference.

of the lake rises from a series of faults along the range front that were active sometime after deposition of the Chari Tuff (1.4 Ma). East of the lake this was not the case, and strata appear to have lapped onto the older lavas of the eastern part of the Karari Ridge. The highest elevations of Pliocene and Pleistocene strata are nearly the same on both sides of the lake. This is perhaps most easily explained by depression of the centre of the basin relative to the edges following deposition of Middle Pleistocene strata. The present east-west gradients should therefore be steeper than those of the past. The magnitude of this difference is perhaps in the order of 50 m from the centre of the basin to the margins, but may have been substantially more. This difference alone would have modified the distribution of plant communities and would have affected the ease with which the ancestral Omo River could traverse its floodplain. During the time of deposition of the Lokochot and Tulu Bor Members, it appears that sedimentation rates in the lower Omo Valley and the Koobi Fora region were nearly equal. This in turn suggests that north-south gradients were very low, and that sedimentation responded to an external control. With very low north-south and east-west topographical gradients, floodplains along the Omo River should have been broader and more extensive than they are today. With less slope to enhance their flow, marginal streams are likely to have been confined to regions nearer their upland sources. The present gradient enables streams to incise into older and compositionally different strata, thereby creating naturally disturbed areas that support different vegetational communities from those that exist on adjacent plains (Carr 1976). A lesser gradient in the past would have produced less incision, affording the possibility of a more uniform vegetational cover.

Most of the upstanding topographic areas within

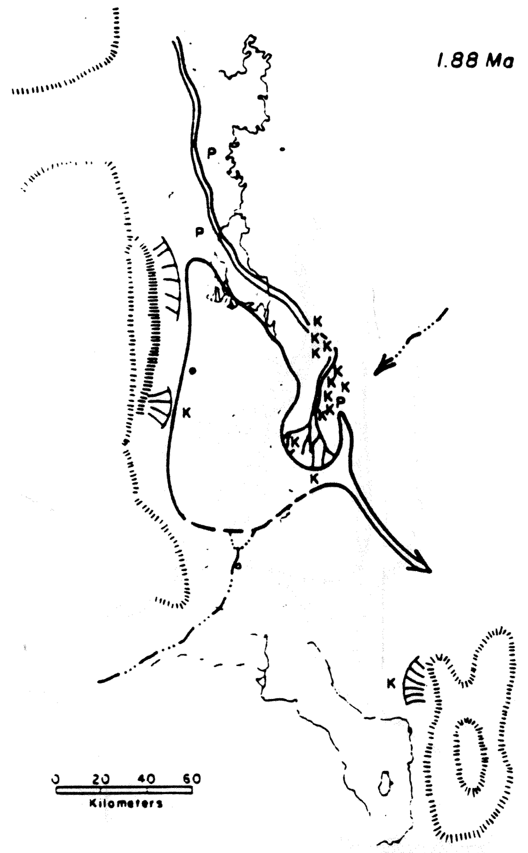

Fig. 1.22. Palaeogeographic reconstruction of the Turkana Basin at 1.88 Ma. Dots represent lacustrine sediments. K and P indicate exposures of the KBS Tuff and occurrences of KBS pumices. Outline of modern Lake Turkana included for reference.

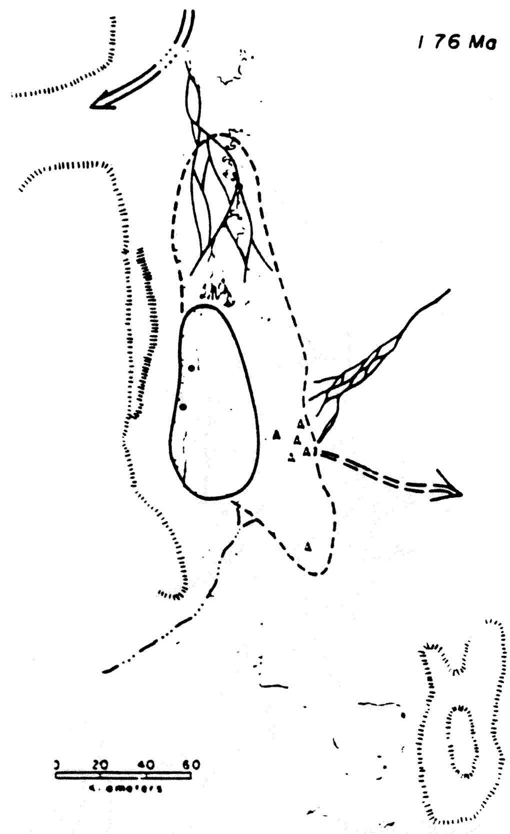

Fig. 1.23. Palaeogeographic reconstruction of the Turkana Basin at 1.76 Ma. Dots represent lacustrine sediments. A indicates exposures of algal biolithites. Outline of modern Lake Turkana included for reference.

the Turkana Basin were formed rather late in time. There is clear evidence that elevation of the Kokoi highland postdates deposition of the Silbo Tuff (0.74 Ma), and depiction of this element as a highland in some earlier reconstructions (e.g. Findlater 1976) is erroneous. Likewise, Jarigole was elevated after deposition of the Tulu Bor Tuff, probably after deposition of the Lokalalei Tuff, and most likely as late as the Middle or Late Pleistocene. The present eastern basin margin in Area 207 also postdates the Lokalalei Tuff, but probably predates the Okote Tuff. Other relatively young geomorphic features in the region are Mt Kulal (built around 2–2.5 Ma), Asie (built between 2.7–0.5 Ma), Mt Marsabit (built around 1 Ma), the Huri Hills (*c*. 3–0.5 Ma), the Korath range in the lower Omo Valley, and the islands of Lake Turkana (Recent). For much of the Plio-Pleistocene the basin comprised a broad, nearly flat area, extending at least from the vicinity of Allia Bay to the northern end of the lower Omo Valley, with no internal obstructions except those formed by constructional processes along the basin margin. With these differences in mind, one can envisage at some times a broad basin, with a large perennial river more or less centrally located that was slightly elevated above the surrounding plain, and that wandered to and fro across the plain as time passed. Broad floodbasins would have been situated lateral to this river that gave way distally to alluvial fans of low slope in which the substrate was generally coarser toward the mountains. At other times the central part of the basin would have been occupied by a lake, the eastern and western shorelines of which would impinge more or less directly on the distal parts of the alluvial fans, largely eliminating the vegetational communities associated with the flood

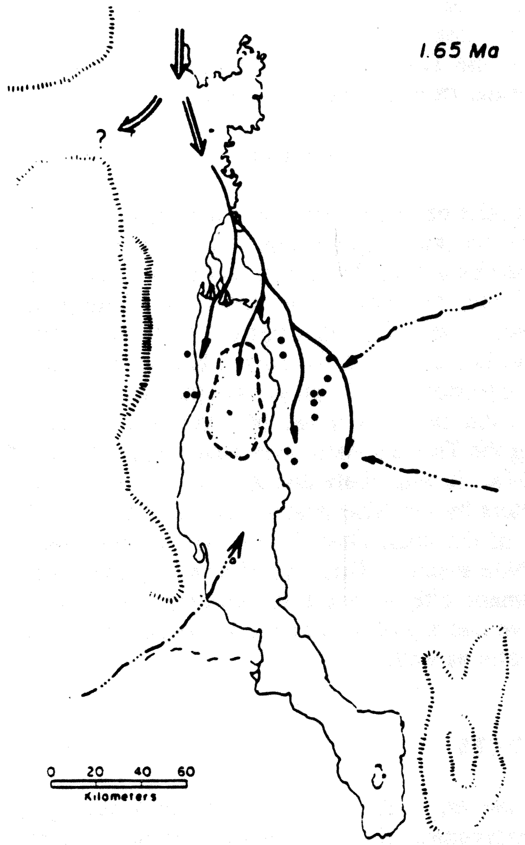

Fig. 1.24. Palaeogeographic reconstruction of the Turkana Basin at 1.65 Ma. Dots represent exposures of the Lower Okote Tuff. Outline of modern Lake Turkana included for reference.

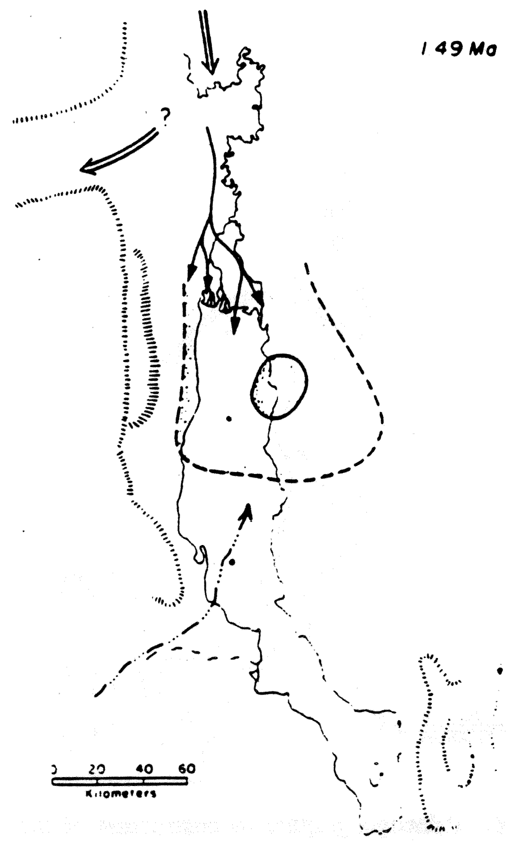

Fig. 1.25. Palaeogeographic reconstruction of the Turkana Basin at 1.49 Ma. Outline of modern Lake Turkana included for reference.

basins. At these latter times, however, the ancestral Omo River, would have formed a broad delta at the northern end of the lake, with its own characteristic faunal and floral communities.

Much has been written about the environmental context of fossils based on reconstructions of depositional environments. From the discussion of environmental facies presented above, it can be seen that such a discussion can only be valid if the depositional context is well understood and correctly interpreted. Otherwise errors of omission and commission render the environmental categories too broad to be of much use, or so detailed that they are of only local significance. Note for example that few of the assemblages attributed to deltaic deposits by Behrensmeyer (1975) are considered to derive from deltaic depositional environments as the term is employed here. Rather a variety of environments was lumped together in Behrensmeyer's analysis, obscuring the complexity of the depositional system.

Deposition during high lake stands of the Holocene tends to obscure the topographical relief that had developed on older strata from *c.* 0.6 to 0.1 Ma ago. In the lower Omo Valley, Holocene features include extremely extensive, elongate beach ridges that largely control the distribution of floristic communities (Carr 1976). This is also true to a lesser extent in the Koobi Fora region where, at the margins of broad alluvial surfaces, beach ridges create internally drained pans that support impoverished floristic communities uncommon elsewhere.

For the past 4 million years the regional climate has been arid or semi-arid and dominated by a strongly seasonal moisture regime. This is supported by the almost ubiquitous occurrence of palaeovertisols in the floodplain deposits at Koobi Fora and in the

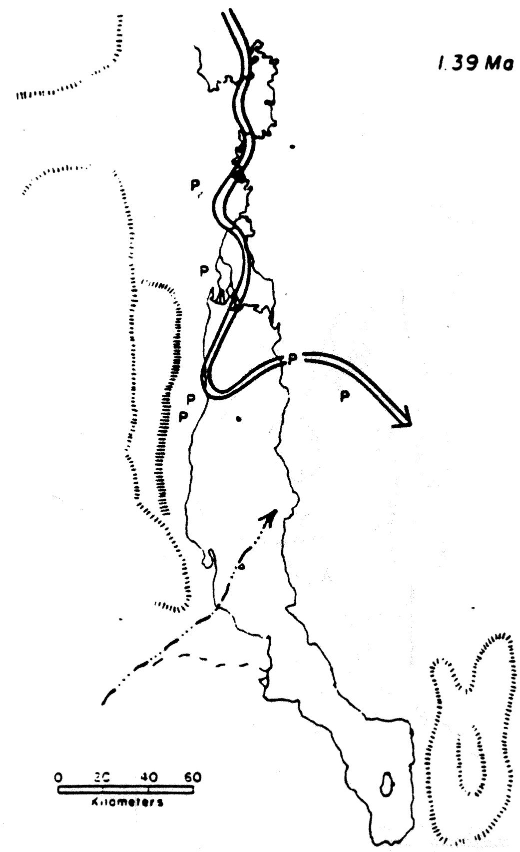

Fig. 1.26. Palaeogeographic reconstruction of the Turkana Basin at 1.39 Ma. P indicates exposures of the Chari Tuff with pumices. Outline of modern Lake Turkana included for reference.

Nachukui and Shungura Formations.The preservation of volcanic landforms in the region also supports the view of long-term regional aridity (Charsley 1987). This seasonality was probably reflected in both the local rainfall, and in the strong seasonal variations in flow of the ancestral Omo River. Contrary to earlier views (Vondra and Bowen 1978), lake margin settings were unusual in the history of the basin (reflecting less than 15 per cent of the time span recorded). They provided a source of water over intervals of a few hundred thousand years and then vanished. The major source for water in the basin has always been rivers rising in more distant regions, most importantly the ancestral Omo River. The spatial and temporal distribution of rivers must have affected the composition and distribution of past communities in the basin. However, stream piracy of Omo River headwaters by tributaries of the Nile should be reflected in the Turkana Basin biota and might also result in a similar signal. Numerous studies have suggested a significant event at about 1.8 Ma (e.g. Cerling *et al.* 1988), but have attributed it to a regional climatic event. Certainly a part of the impact in the Turkana Basin was likely a periodical loss of water from the Omo drainage.

SUMMARY

The deposits of the Koobi Fora Formation indicate that over the past 4 million years the region has been dominated by a large fluvial system flowing down the axis of the Turkana Basin from the Ethiopian Highlands. Periodical development of lakes within the basin has produced short-lived bodies of water, with a disproportionately large stratigraphical record. A reduction in the volume of Omo River water reaching the Turkana Basin may have begun as early as 1.9 Ma. It was likely due to piracy of the Omo headwaters by the Blue Nile system, and periodical capture of the main Omo flow by the Pibor-Sobat-White Nile system. This loss of water by the basin had dramatic effects on the depositional system, and may have had a profound effect on the communities of the basin as well.

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