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THE PLIO-PLEISTOCENE: SEDIMENTS, ENVIRONMENTS, AND GEOCHRONOLOGY ALONG THE KARARI ESCARPMENT, EAST TURKANA, KENYA

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The sedimentary sequence of East Turkana, Kenya, provides an excellent opportunity to document change contemporary with the Plio-Pleistocene boundary in East Africa. Within exposures along the Karari Ridge, sediments of the Fluvio-lacustrine Koobi Fora Formation record a dramatic change in depositional regime from low to high energy attributable to climate change and/or tectonic activity. The correlation of this abrupt transition with the beginning of the Pleistocene epoch is not without question, but controversy resulting from radiometric dating of volcanic ash units, paleomagnetic zonation, and paleontologic data exists and indicates that resolution of the Plio-Pleistocene boundary at East Turkana will be possible only through additional study.

† † †

INTRODUCTION

Since 1968, the Upper Cenozoic sediments surrounding Lake Turkana (formerly Lake Rudolf) in the northwestern portion of Kenya, have been the focus of a comprehensive interdisciplinary research effort related to the search for evidence of early man in East Africa (Coppens, et al., 1976). In particular the Plio-Pleistocene fluvial and lacustrine sediments of the East Turkana Basin (Fig. 1) have yielded abundant hominid remains and stone artifacts (Leakey, et al., 1970; Leakey, 1971, 1972, 1973, 1976; Isaac, et al., 1971; Isaac, 1976; Isaac, et al., 1976; Harris and Isaac, 1976). Corresponding geologic (Leakey, et al., 1970); Vondra, et al., 1971; Bowen and Vondra, 1973; Vondra and Bowen, 1976, 1977; Findlater, 1976a and b; and Vondra and Burggraf, 1978), paleomagnetic (Brock and Issac, 1974, 1976; Hillhouse, et al., 1977), radiometric (Fitch, et al., 1974; Curtis, et al., 1975; Fitch, et al., 1976; Hurford, et al., 1976), and paleontologic (Maglio, 1971, 1972; Cooke and Maglio, 1972; Cooke, 1976; Harris and White, 1977) studies have been and are being completed by investigators in the United States, Great Britain, Kenya, and Australia. Thus, the occurrence of sediments spanning the late Pliocene and early Pleistocene epochs, the

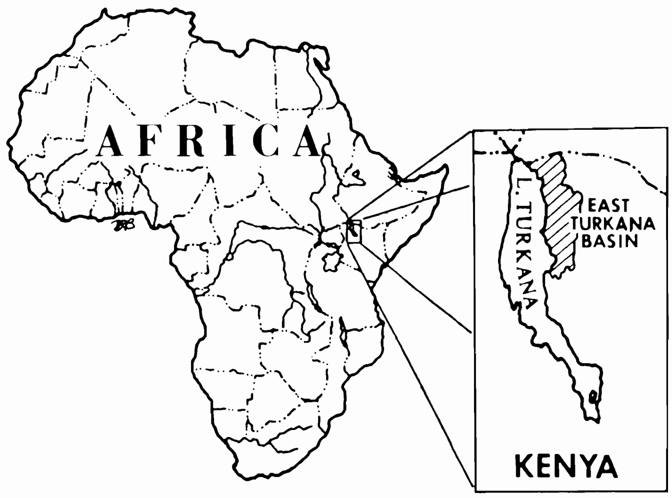


Figure 1. Location of Lake Turkana in northwest Kenya and position of the East Turkana Basin. By decree of the Kenyan government in 1975, the name of the lake was officially changed from Lake Rudolf to Lake Turkana, the northernmost extension of the lake into Ethiopia retaining the name, Lake Rudolf.

abundant and well-preserved fossil remains, the existence of interbedded radioisotopically datable volcanic material, and the integration of paleomagnetic reversal stratigraphy make the deposits of the East Turkana Basin notable to the study of the Plio-Pleistocene boundary in East Africa.

TECTONIC SETTING

Lake Turkana occupies a tectonic depression associated with and marginal to a segment of the Gregory or Eastern Rift System where it passes through northern Kenya (Fig. 2). The

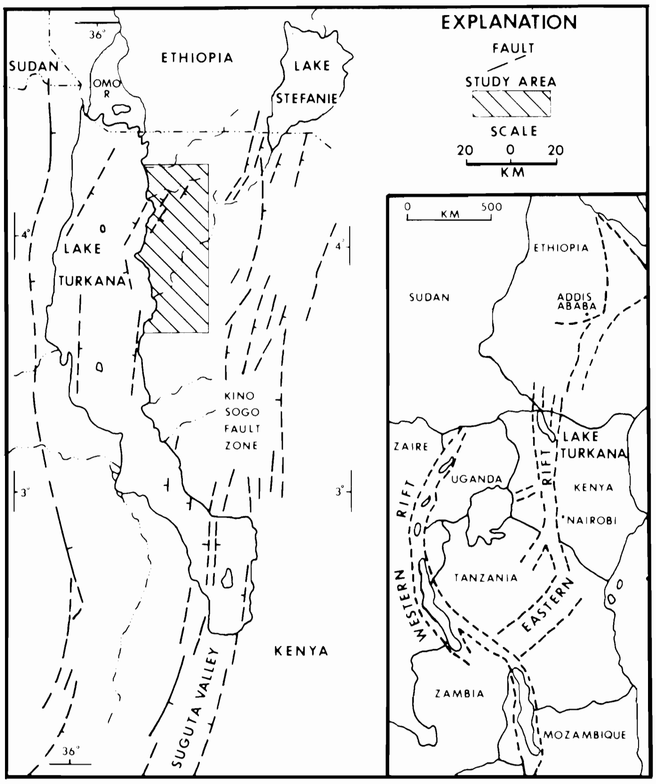


Figure 2. Tectonic map of the East African Rift System (insert) illustrating the marginal rift valley setting of Lake Turkana (after White, 1976).

lake is bounded on the west by a normal fault and on the east by gently dipping sediments lying on middle Cenozoic volcanic strata of a westwardly tilted monoclinal flexure. Initial rift-related deformation began with uplift of the Arabo-Ethiopian dome during the late Eocene and continued intermittently with early Miocene domal warping of the crust in Kenya and adjacent parts of Ethiopia and Tanzania (Baker, et al., 1972). Major rifting and associated faulting during the late Miocene and Pliocene were centered along the crestal regions of the domes resulting in the formation of a somewhat discontinuous curvilinear fracture belt. The Turkana depression of northern Kenya may be the result of initial rifting, later abandoned as crustal deformation proceeded to the southeast, forming the Kino Sogo fault zone (Cerling and Powers, 1977). Within the depression, extensive Miocene (17.2 ± 1.8 Myr) volcanics are presently manifest as basalt flows, ignimbrites and tuffs, with related interbedded sediments (Fitch and Miller, 1976).

Pliocene and younger deformations are recorded in the sediments within the East Turkana Basin. Periodic late Cenozoic to Recent tectonic activity is evidenced by numerous northeasterly-trending normal faults, by bed attitudes which indicate a post-Pleistocene local uplifting of the Kokoi Horst, and by abundant volcaniclastic sediments in the Plio-Pleistocene and Recent basin deposits.

Previous Work

Exposures of the Pliocene through Recent sediments in the East Turkana Basin were first reconnoitered in 1969 by Behrensmeyer and other members of the East Rudolf Research Expedition organized by R. E. F. Leakey and the National Museums of Kenya. An informal stratigraphic nomenclature was developed (Leakey, et al., 1970; Vondra, et al., 1971) and later formal stratigraphic units were established through measurement of stratigraphic sections and mapping of marker horizons (Bowen and Vondra, 1973). Continued regional mapping at a scale of 1:24,000, stratigraphic sectioning, and sample collection and analysis led to the development of a comprehensive stratigraphic framework within which archeological and anthropological discoveries could be located (Bowen, 1974; Acuff, 1976; Vondra and Bowen, 1976, 1977).

Detailed stratigraphic studies of the most productive fossil-bearing areas along the Karari Escarpment were accomplished through the measurement of additional stratigraphic sections and by mapping of local marker horizons on photographic bases at a scale of 1:6,000 (Bainbridge, 1976; Burggraf, 1976; Frank, 1976; White, 1976). Each horizon was physically traced and documented as to stratigraphic position through the measured sections and also sampled for later laboratory analyses to aid in environmental reconstruction.

Continued sedimentologic and stratigraphic (Findlater, 1976a; Vondra and Burggraf, 1978, in press), mineralogic (Mathisen, 1977), and geochemical (Cerling, 1977; Cerling et al., 1977) studies have further documented changes in climate and depositional environments through the Plio-Pleistocene succession in the East Turkana Basin. A general trend toward more arid conditions is suggested by paleoalkalinity measurements of the lacustrine sediments (Cerling, et al., 1977) and by variations in heavy mineral textures (Mathisen, 1977). A corresponding change in the fluvial regime from a dominantly low sinuosity braided system to a high sinuosity meandering system has been reported by Vondra and Burggraf (in press).

An absolute time framework based on radioisotopic analyses of selected minerals of volcanogenic origin was developed by Fitch, et al. (1974). Major tuff levels were dated using conventional K-Ar (Fitch, et al., 1974; Fitch and Miller, 1976; Curtis, et al., 1975), $^{40}Ar/^{39}Ar$ age spectrum technique (Fitch and Miller, 1976; Fitch, et al., 1976), and fission track analyses (Hurford, et al., 1976).

Finally, paleomagnetic (Brock and Isaac, 1974; Hillhouse, et al., 1977) and paleontologic (Harris, 1976; Harris and White, 1977) studies have led to some controversy relating biostratigraphic, time-stratigraphic, and lithostratigraphic frameworks between separate areas within the basin and between the East Turkana Basin and the Omo Basin farther north (see in particular, Harris and White, 1977; Hillhouse, et al., 1977). This paper will focus on the current understanding of the Plio-Pleistocene boundary in the East Turkana sequence with emphasis on the stratigraphic evidence accrued to date.

REGIONAL STRATIGRAPHY

The Upper Cenozoic strata of the East Turkana Basin constitute a 325 meter fluvio-lacustrine complex which unconformably overlies a series of Miocene and Pliocene volcanics and records a general westward regression of the Paleo-Lake Turkana shoreline. Bowen and Vondra (1973) and Bowen (1974) divided the sequence into three lithostratigraphic units of formational rank which are: (1) the Pliocene Kubi Algi Formation, (2) the Plio-Pleistocene Koobi Fora Formation, and (3) the Pleistocene Guomde Formation (Fig. 3). Additionally, discontinuous deposits of diatomaceous siltstones of lacustrine origin occurring sporadically adjacent to the present lake margin were informally designated the Galana Boi Beds. The distribution of each of these units was mapped by Bowen (1974; Fig. 4).

Kubi Algi Formation: The Kubi Algi Formation consists predominantly of coarse-grained sediments–conglomerate and conglomeratic sublitharenite–with interbedded, fine-grained sublitharenites, drab siltstones and claystones, and tuffs (Bowen, 1974; Acuff, 1976). It varies greatly in thickness but includes 98 meters of sediment at the type locality. The Kubi Algi Formation occurs only in the southern third of the basin and is conformably overlain by the Koobi Fora Formation.

Koobi Fora Formation: The Koobi Fora Formation occurs throughout all but the southernmost part of the basin and includes a sequence of as much as 210 meters of granule to boulder conglomerate, fine- to coarse-grained feldspathic litharenite and lithic arkose, varigated siltstone and claystone, bioclastic carbonate, and tuff (Bowen, 1974; Acuff, 1976; Bainbridge, 1976; Burggraf, 1976; Frank, 1976; and White, 1976). Because of their extensive lateral occurrence, two major tuff horizons (the Suregei and Koobi Fora Tuffs) have been defined as the lower and upper boundary marker beds at the type section of the Koobi Fora Formation. Two other laterally extensive although discontinuous tuff horizons (the Tulu Bor and the KBS Tuffs) were utilized as marker beds for purposes of regional correlation.

Along the Karari Escarpment, the two members of the Koobi Fora Formation are well exposed. The Lower Member consists primarily of lacustrine and transitional-lacustrine, fine-grained deposits, while the Upper Member is composed primarily of fluvially-derived conglomerate, sandstone, and mudrocks (See "Stratigraphy of the Koobi Fora Formation along the Karari Escarpment").

Guomde Formation: The Guomde Formation consists of a series of tuffs, laminated siltstones, and intercalated, thin, bioclastic carbonates disconformably overlying the Koobi Fora Formation along the Ileret Ridge. Faulting and Recent erosion have greatly modified the total thickness of the formation, but in general, the Guomde thickens slightly toward the south from 32 to greater than 37 meters.

MAJOR LITHOFACIES

Bowen (1974) and Vondra and Bowen (1976) identified four major lithofacies within the deposits of the East Turkana Basin. These include: (1) the laminated siltstone facies; (2) the arenaceous bioclastic carbonate facies; (3) the lenticular, fine-grained sandstone and lenticular-bedded, siltstone facies; and (4) the lenticular conglomerate, sandstone, and mudstone facies. Based on lithologies, bed geometries and lateral extents, primary sedimentary structures, and fossil content, depositional environments were interpreted for each lithofacies and include: (1) prodelta and shallow shelf lacustrine; (2) littoral lacustrine–beach, barrier beach, and associated barrier lagoons; (3) delta plain–distributary channel, levee, and interdistributary flood basin; and (4) fluvial channel and flood basin.

Complex intertonguing and lateral gradation of one facies with another is related to westward regression of the lake due to variations in climate and/or tectonic movement. Drainage throughout the Late Cenozoic is believed to have been primarily southwestward; therefore, the major lithofacies occur as approximately north-south trending belts which migrate through time relative to the fluctuating shoreline (Vondra and Bowen, 1976, 1977).

STRATIGRAPHY OF THE KOOBI FORA FORMATION ALONG THE KARARI ESCARPMENT

The Koobi Fora Formation was divided into two members in the Ileret and Koobi Fora areas based on lithlogic heterogeneity and distribution (Bowen and Vondra, 1973). Terminology for each area was different owing to the lack of secure correlation between the two areas. Vondra and Bowen (1977) unified the member terminology and redefined the contact between the two as the base of the conglomeratic channel sandstone complex which occurs stratigraphically immediately above the KBS Tuff in the Koobi Fora area. Along the Karari Escarpment, the Lower and Upper Members are well defined and have been studied in detail by Bainbridge (1976), Burggraf (1976), Frank (1976), and White (1976) (Fig. 5).

Lower Member: The Lower Member along the Karari Escarpment includes portions of all of the four major lithofacies of Bowen and Vondra (1973). The laminated siltstone facies occurs primarily at the base of the formation along the Karari Escarpment and includes parallel, laminated- to

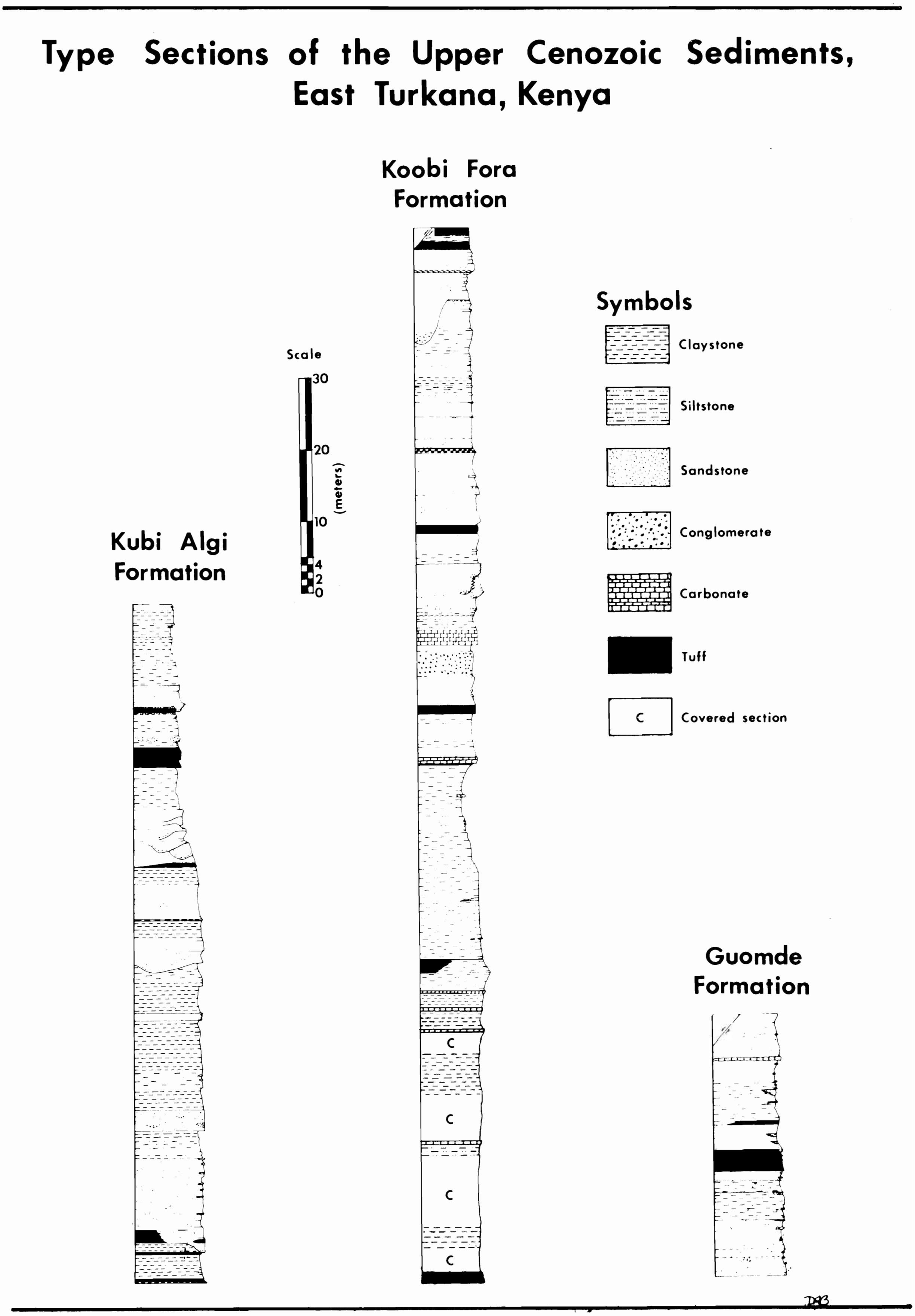


Figure 3. Type sections of the East Turkana sequence (after Bowen and Vondra, 1973; Bowen, 1974).

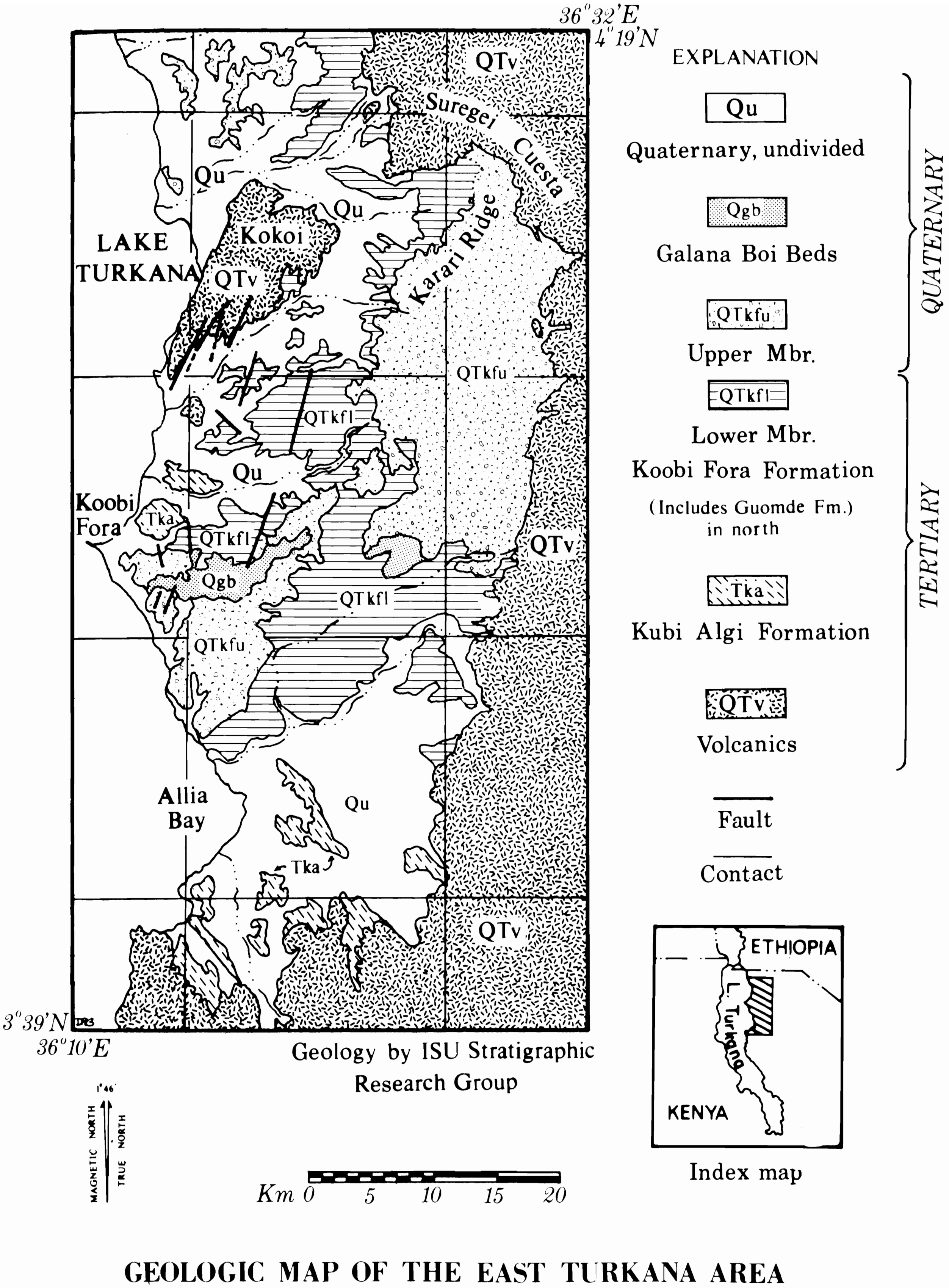


Figure 4. Geologic map of the East Turkana Basin.

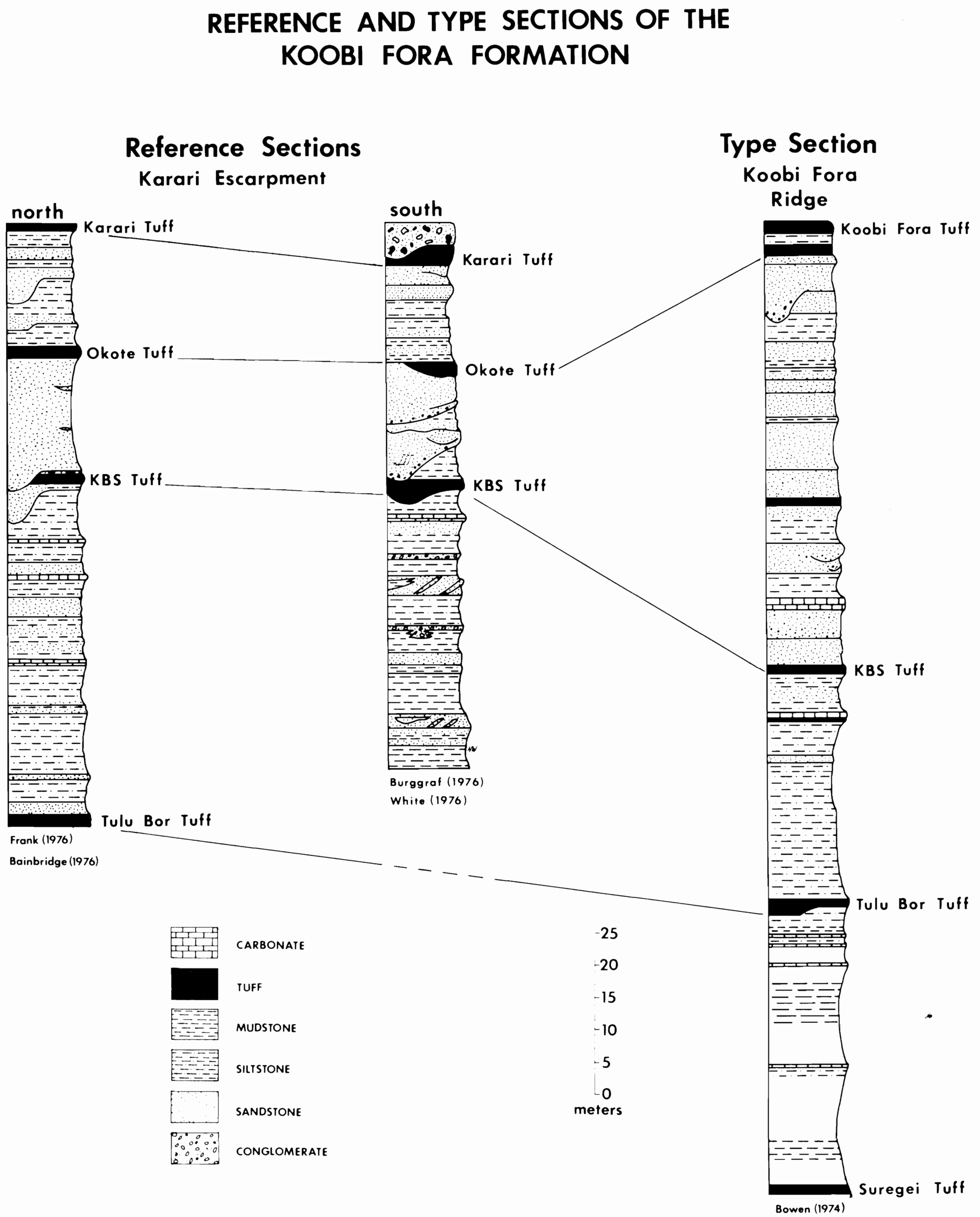
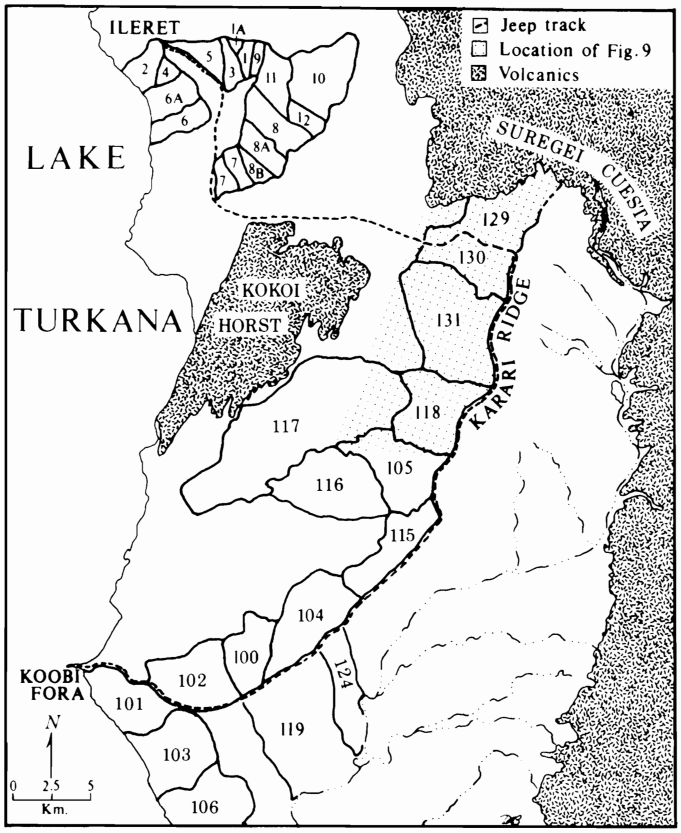


Figure 5. Correlation of reference sections of the Koobi Fora Formation along the Karari Escarpment with the type section loca directly east of Koobi Fora Spit.

thin-bedded mudrocks with alternations of ripple-laminated strata and lenticular, bedding-plane concentrations of limonitic siltstone nodules varying in thickness from 1 to 5 centimeters. This facies grades upward and interfingers with the arenaceous, bioclastic, carbonate facies which is characterized by coarsening-upward, beach sequences of parallel- to ripple-laminated, silty litharenites which grade vertically into planar and trough, cross-bedded, feldspathic litharenites. These are often capped by arenaceous, packed gastropod carbonates. Intertonguing with these deposits is the lenticular, fine-grained sandstone and lenticular-bedded, siltstone facies which makes up the greatest portion of the Lower Member along the Karari Escarpment. It is composed of 20 to 25 meters of deltaic, parallel-laminated to thinly-bedded mudrocks and thin ripple-laminated, fine-grained, sandstone lenses with occasional distributary channels (4 to 6 meters thick) of large-scale planar and small-scale trough, cross-bedded litharenite. Conformably overlying this sequence is the fluvial, lenticular conglomerate, sandstone, and mudstone facies marked at its base by the moderate-brown claystone enclosing the KBS Tuff throughout areas 131, 118, and 105 (Fig. 6). These sediments and the mudrocks and fine-grained, ripple-laminated and small-scale, cross-bedded sandstones lying above the tuff but below the arkosic sandstones of the basal portion of the Upper Member of the Koobi Fora Formation reflect fluvial deposition on a very low-energy floodplain.

The moderate-brown claystone is characterized by irregular fracture, lack of primary sedimentary structures, and abundant white calcaresous concretions 1 to 3 centimeters in length increasing in diameter vertically. Red and pale-green oxidation-reduction of iron and black manganese dioxide stain commonly coat fracture surfaces. Textural uniformity and the nature of the concretionary carbonate content present a notable contrast with the underlying sediments. This mudstone and an identical unit above the tuff consistently occur with the KBS throughout areas 118 and 105 and much of area 131. Near the Suregei Cuesta to the north, the KBS Tuff directly overlies lenticular, distributary channel sandstones and overbank levee and floodbasin mudrocks, indicating that deposition took place on an erosional surface over which a maximum downcutting of about 4 meters has been observed in the field.

The KBS Tuff outcrops widely through areas 130 and 131 but only sporadically between areas 118 and 105, thus correlation must be based on relationships other than the simple tracing of a horizon through the later areas. The tuff itself represents deposition in a variety of subenvironments but with few exceptions displays characteristic zonation expressed as color bands. This sequence includes a lower white to light-gray, ripple-laminated to small-scale trough, cross-bedded, fine- to medium-grained tuff which grades upward to a structureless or indistinctly-parallel laminated, grayish pink, fine-grained, silty tuff which often contains root casts and crotovina. In turn, the tuff is overlain by a sequence of



Fossil Collecting Localities

Figure 6. Outline map of fossil collecting localities for the northern East Turkana Basin showing area of interest adjoining the Karari Ridge.

mudstone, siltstone, and fine-grained sandstone which, along much of the exposure of the KBS Tuff through areas 131 and 130, have been removed by later fluvial activity and replaced by the conglomeratic sandstones of the basal part of the Upper Member. The moderate-brown concretionary mudstones enclosing the KBS Tuff occur nowhere else in the Lower Member sequence.

Considering the Lower Member sediments as a whole, a general pattern of delta progradation is evident. A reconstruction of paleogeographic conditions (Fig. 7) illustrates the southwestward progradation of the delta complex represented by the strata of the Lower Member of the Koobi Fora Formation. Consecutively younger shoreline positions (Fig. 7a) are shown by the dashed lines numbered 1 through 5, inclusively. Isochronous surfaces for each of these instances (Fig. 7b) reflect the successively younger age of the sediments toward the southwest, but they cannot be recognized in the field since sedimentation occurred continuously so that the majority of lithostratigraphic units are time-transgressive. Isochronous units are, however, represented by the tuff beds periodically distributed throughout the formation. The KBS Tuff, for example, was deposited over a surface such as that

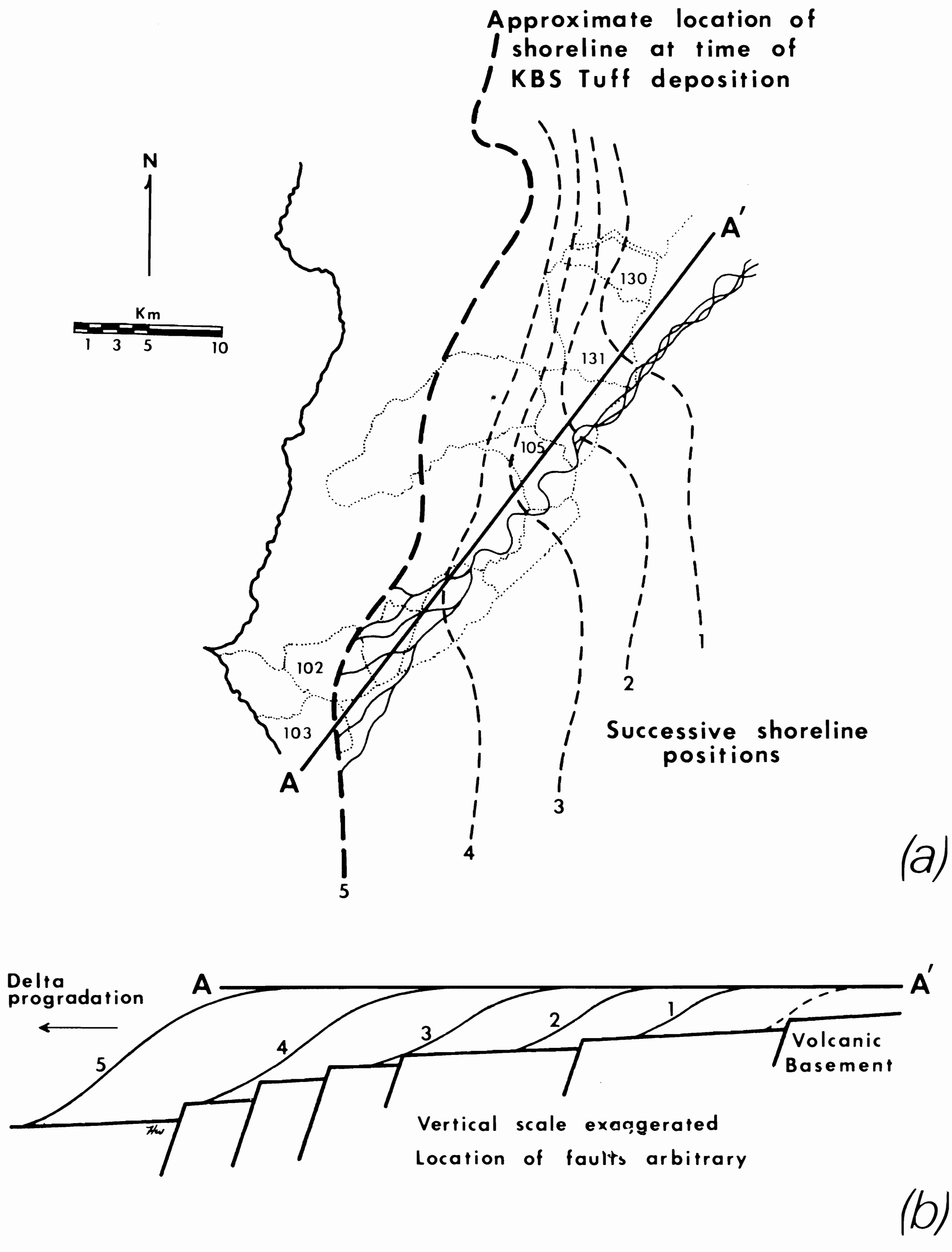


Figure 7. (a) Schematic representation of shoreline positions during delta progradation, just after time of Tulu Bor Tuff de tion, to time of KBS Tuff deposition.

(b) Inferred cross section (A-A′) illustrating progressively younger time lines (1 through 5) through the delta com Faulting and deplacement of the volcanic basement, though highly schematized, is interpreted from the regional tec setting.

illustrated in Fig. 7. The structureless, moderate-brown claystone underlying the tuff is indicative of a subaerially exposed horizon with a weakly developed caliche profile similar to those recently reported by Hubert (1977) in overbank red mudstones of the New Haven Arkose. It is interpreted, then, that this horizon was developed across a broad area overlying the prograding delta sediments throughout areas 105, 118, and 131. Tuffaceous debris brought into the basin by ash-laden flood waters spilled out over the channel banks across a featureless plain in area 131 and portions of area 130 and filled swales and abandoned channel crescents proximal to the channel and in the area of the distributary channel system on the delta plain in areas 118 and 105 and southward.

Color zonation within the tuff is believed to be due partly to depositional segregation of the volcanic and clastic detritus in fining-upward sequences and partly to weathering of the upper portion of the tuff prior to burial by additional sediment. Root casts, crotovina, and carbonate concretions attest to the pedogenic reworking of the upper portion of many of the tuff beds. Subsequent fluvial activity preserved the ash in some areas, depositing overbank sands, silts and clays above it, while in other locales, the ash was removed by migrating channel-systems and ephemeral streams. In many instances throughout the Karari Ridge exposures, the KBS Tuff was cut out by coarse-grained, conglomeratic sandstones of the Upper Member (Fig. 8).

Upper Member: Throughout the Karari Escarpment, the Upper Member of the Koobi Fora Formation is distinct from the Lower Member lithologically, and because of the erosional nature of the Upper Member basal contact [as much as 11 meters of downcutting has been recorded (Burggraf, 1976)], the boundary is readily apparent in the field. The basal portion of the Upper Member of the Koobi Fora Formation consists of conglomeratic arkoses, lithic arkoses, and feldspathic litharenites. These occur as lenticular, fining-upward sequences and include abundant clay galls, carbonate concretions and crotovina, and occasional interbedded arenaceous and tuffaceous siltstones (Burggraf, 1976). This sequence reaches a maximum thickness of about 20 meters in central area 131 (see Fig. 5) and through areas 118 and 105 averages approximately 15 meters thick. It is conformably overlain by fine-grained arkoses and feldspathic litharenites, highly tuffaceous siltstones and mudrocks, and tuffs which occur, in general, as three, fining-upward cycles where a maximum thickness of about 19 meters is present (central area 131).

The Upper Member also includes well-rounded, basalt cobble conglomerate lenses to 4 meters thick and several tens of meters wide which are found primarily in the lower half of the Member laterally grading into conglomeratic feldspathic litharenites. Along the basin margin in area 129, similar deposits include pebbly mudstones and thin- to thick-bedded tabular basalt cobble conglomerates (Frank, 1976) which unconformably overlie Miocene volcanics and grade basin-ward into conglomeratic feldspathic litharenites.

Vertebrate fossil remains occur infrequently throughout the Upper Member although several hominid cranial and post-cranial specimens (Leakey, 1976; Walker, 1976) and abundant stone artifacts (Isaac, et al., 1976; Isaac, 1976) have been found (Fig. 9).

Based on detailed stratigraphic and sedimentologic study (Burggraf, 1976; Frank, 1976), the sediments of the Upper Member of the Koobi Fora Formation have been divided into four, inter-fingering subfacies (Vondra and Burggraf, 1978, in press). These are the: (1) interbedded, basalt-clast conglomerate and pebbly mudstone subfacies; (2) lenticular, basalt-clast, conglomerate subfacies; (3) polymictic conglomerate and sandstone subfacies; and (4) interbedded sandstone and tuffaceous, siltstone subfacies, respectively representing: (a) alluvial fan, channel and debris flow; (b) high-energy channel (bar-core and gravel sheet); (c) lower energy channel (bar-side, transverse bar, and point-bar); and (d) flood basin, depositional environments.

The sediments of the Upper Member record a changing, fluvial regime documented by overall lithologic heterogeneity and by sequences of primary sedimentary structures indicative of hydrodynamic, flow conditions during deposition. The lower half of the Upper Member is interpreted to represent deposition from a sandy, low-sinuosity, braided stream system. Dominant primary structures are plane bedding and large-scale planar cross-stratification. Other structures include isolated large-scale trough cross-stratification and thin sequences of ripple-drift cross-lamination. These structures are believed to represent longitudinal and transverse bars of a braided sand-bed, ephemeral stream (Vondra and Burggraf, in press) entering the basin from the north and northeast and trending primarily in a southwesterly direction toward the paleo-Lake Turkana.

Roughly, the upper half of the Upper Member is dominated by tabular beds of fine-grained, plane-bedded to ripple-bedded fine sand and silt. Coarse- to fine-grained sand units consist primarily of large-scale, trough cross-stratified, coarse-grained sand fining upward to plane-bedded medium to fine sand and/or small-scale, trough cross-stratified, fine sand. The entire sequence is often capped by a thin bed of argillaceous siltstone and is interpreted to represent deposition as a point bar in an intermediate to high-sinuosity, meandering river (Vondra and Burggraf, in press).

GEOCHRONOLOGY AND THE PLIO–PLEISTOCENE BOUNDARY

Several radioisotopic-dating techniques (Fitch and Miller, 1976) have been applied to a variety of primary mineral species extracted from pumice cobbles and boulders associated with the respective tuff horizons. The results of $^{40}Ar/^{39}Ar$-dating techniques (Fitch and Miller, 1976; Fitch, et al., 1976)

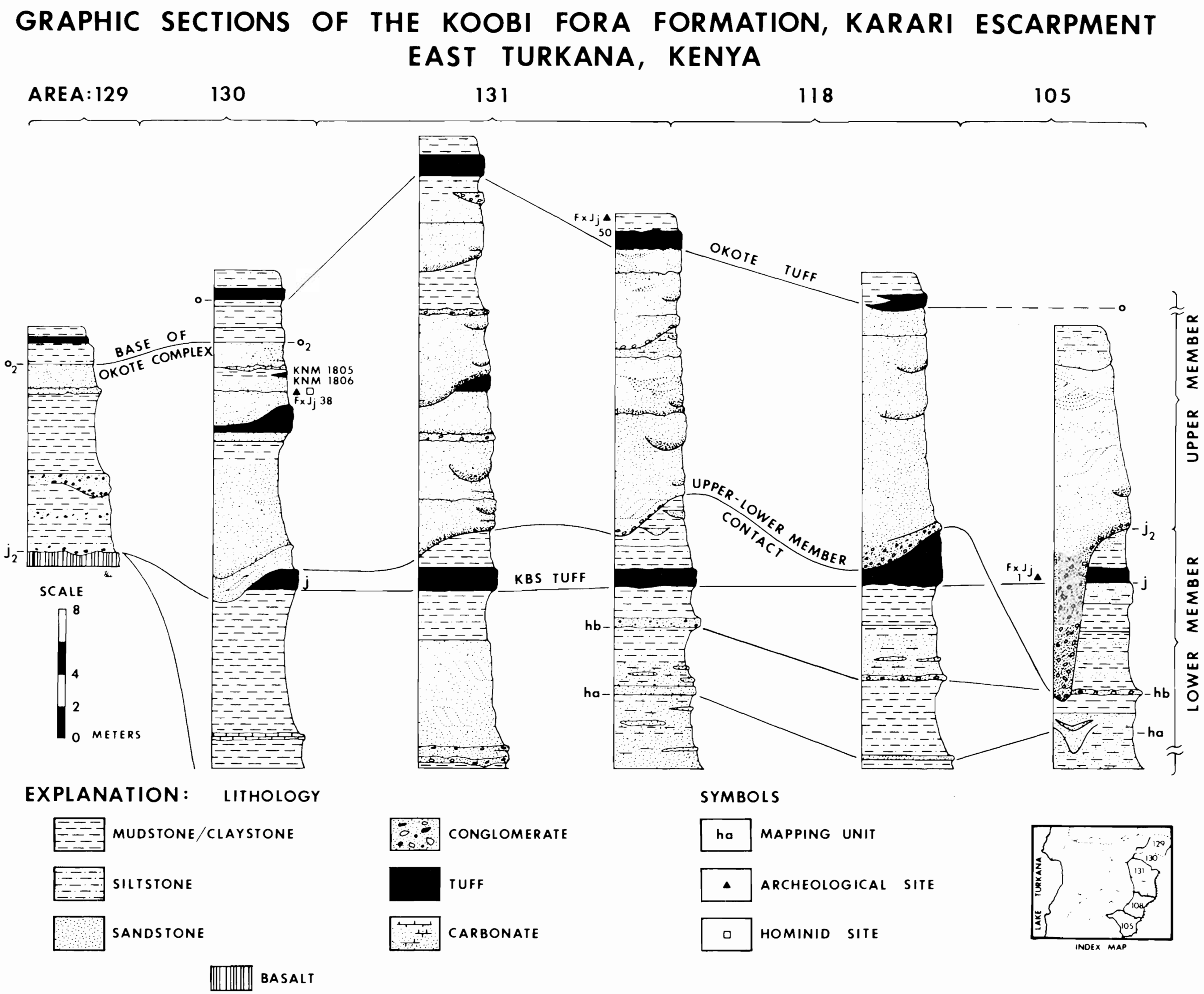


Figure 8. Graphic sections of the Koobi Fora Formation exposed along the Karari Escarpment. Only that portion of the sequence bordering the Lower-Upper Member contact is shown depicting the dramatic change in lithlogic character across the unconformity. Mapping units designating the Lower-Upper Member boundary

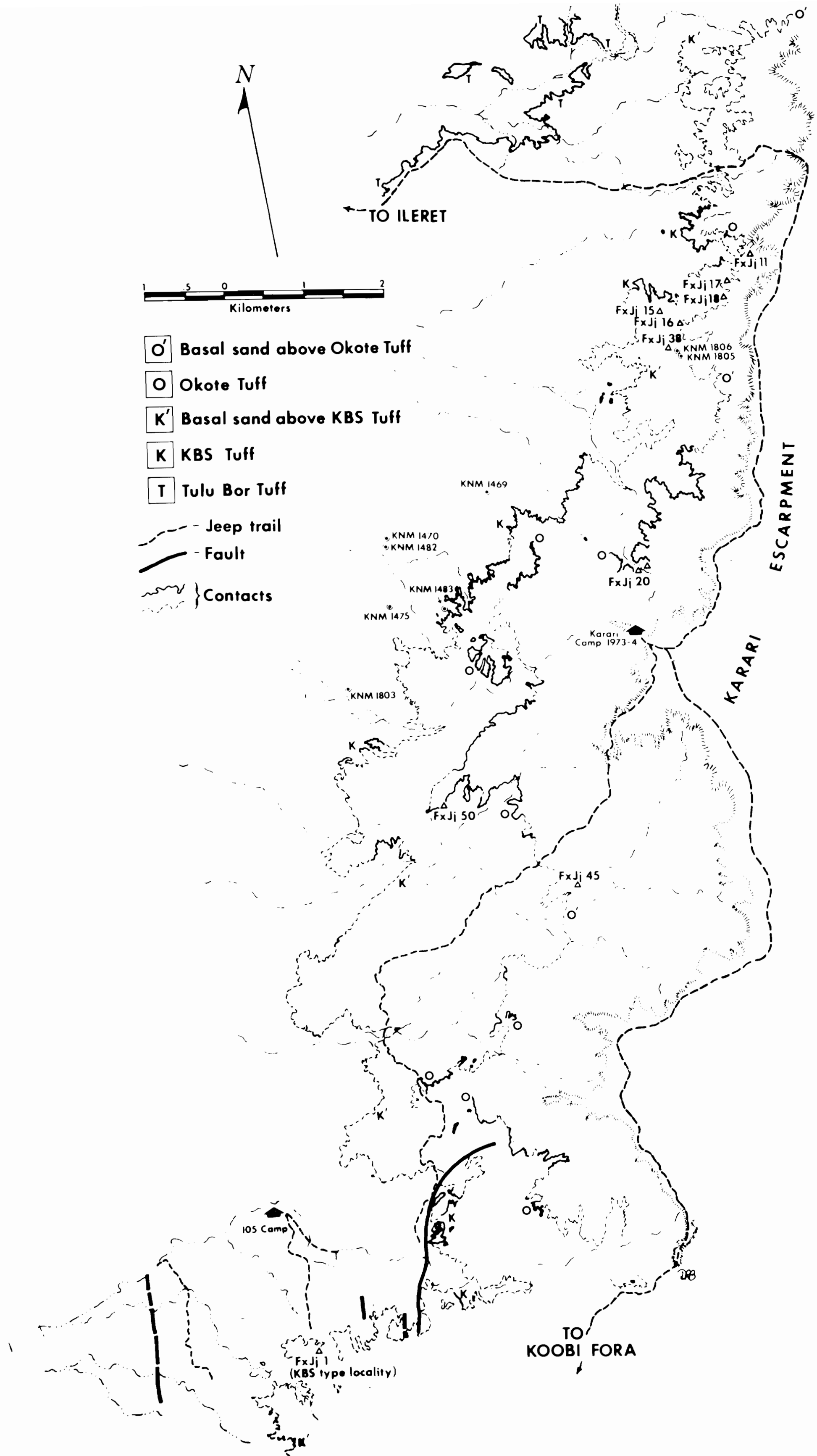


Figure 9. Outline geologic map of the Karari Escarpment area.

suggest the following best apparent ages for each tuff: (1) Chari Tuff–1.28 ± 0.23 Myr, (2) Karari Tuff–1.23 ± 0.1 Myr, (3) Koobi Fora Tuff–1.57 ± 0.0 Myr, (4) Okote Tuff–1.56 ± 0.02 Myr, (5) KBS Tuff–2.42 ± 0.01 Myr, and (6) Tulu Bor Tuff–3.18 ± 0.09 Myr. However, conventional K-Ar dating methods have been applied to samples collected from the KBS Tuff in areas 105, 131, and 10 with the resulting determination of two dates–1.82 ± 0.04 Myr for area 131 and 1.60 ± 0.05 Myr for areas 105 and 10 (Curtis, et al., 1975). The discrepancy in results obtained by applying different dating techniques to the KBS samples has been discussed by the respective investigators (Curtis, et al., 1975; Fitch, et al., 1976) and will not be further argued here. However, the suggestion by Curtis, et al. (1975) that more than one tuff is present at the stratigraphic level of the KBS Tuff from area 131 southward to area 105 is untenable. The KBS Tuff and subjacent and superjacent strata were physically traced and mapped continuously along the Karari Escarpment at a scale of 1:6,000 as mentioned above, from area 129 through areas 130, 131 and 118 to area 105. The KBS Tuff in area 131 is the KBS Tuff of area 105. Relative ages determined by careful application of the laws of superposition and original continuity will always provide a more firm basis for dating than radiometric, paleontologic or paleomagnetic evidence of age (Bishop, 1972).

Confusion has also arisen because of the disparity in ages of similar fauna between the Omo and East Turkana (Rudolf) areas (Cooke, 1976; Harris and White, 1977). In particular, collections of the suid *Mesochoerus limnetes* from sediment beneath the KBS Tuff indicate an evolutionary stage similar to that of Member G of the Shungura Formation of the Omo Sequence and to that of Bed I in Olduvai Gorge (suggesting a corresponding date of approximately 1.8 Myr for the KBS Tuff). Cooke (1976) reports the possibility that ecological differences between the two areas may have accounted for the dichotomy in populations but that such a barrier is difficult to conceive of. Additionally, Harris (personal communication, 1977) reports the progressive evolutionary advancement in the suid fauna recovered from the sediments beneath the KBS Tuff from area 131 southward to 105 and southwestward to areas 102 and 103. It has been suggested, then, that the overlying tuff (KBS) cannot be the same in each area but that instead the tuff (KBS) in areas 131, 105, and 102 and 103 represent successively younger eruptions.

This premise is not sound under geological considerations and, in fact, demonstrates the danger of attempting to "date" an overlying unit by its relationship to underlying, non-isochronous strata.

Instead, it is proposed that the KBS Tuff, representing a short interval of volcanic activity, was deposited on a low-relief floodplain formed over a long period of time by a southwestwardly, prograding, delta complex. Thus, the sequence of delta, delta plain, and distributary channel sediments occ ring in the Lower Member of the Koobi Fora Formation, a beneath the KBS Tuff, decreases in age to the southwest.

Paleomagnetic measurements of the polarity of sediments of the Koobi Fora Formation were applied a check for the radioisotopic dates previously obtained. Br and Isaac (1974) reported on the magnetostratigraphy of exposures of the Formation at Ileret along the Karari Esca ment in area 130, along the Koobi Fora Ridge in area 1 and near Lake Tarkana in area 103. They calibrated th results with the original K-Ar and $^{40}Ar/^{39}Ar$ dates of 2.6 0.26 Myr obtained by Fitch and Miller (1976) for the K Tuff. This date was recomputed by Fitch, et al. (1976) 2.42 ± 0.01 Myr and corroborated by Hurford, et al. (19 with a fission track date of 2.44 ± 0.08 Myr and later c firmed by Gleadow (personal communication, Fitch, 19 in an independent laboratory in Australia. The KBS Tuff the sediments of the uppermost portion of the Lower Mem of the Koobi Fora Formation [as redefined by Vondra Bowen (1977)] occurring above the KBS up to the unc formable contact with the Upper Member are characteri by normal polarity. The sediments in area 105 between Tulu Bor Tuff dated at 3.18 ± 0.09 Myr (Fitch and Mil 1976) and the KBS Tuff are predominantly normal with exception of two short intervals of reversed polarity. T normal magnetozone was correlated with the Gauss nor epoch. The two reversed intervals were considered to rep sent the Mammoth and Kaena events.

The paleomagnetic results of the sediments of the Up Member are more difficult to interpret. They are fluvial origin with several superimposed cut-and-fill structures. basal portion of the Upper Member above the unconforma basal contact in areas 105 and 103 displays normal polar and was assigned to the Olduvai event of Matuyama age. H ever, deposits in area 130 above the unconformity recor reversed polarity. These were interpreted as belonging to reversed interval ranging from 1.61 - 0.95 Myr occurr above the Oluvai event. The absence of a normal magnetoz representing the Olduvai event was explained as resulting fr a longer duration of erosion recorded by the unconform between the Lower and Upper Members. The uncertainty the length of the hiatus represented by the unconformity weak point, however, Maglio's (1972) faunal zonation and estimates based on the original Omo radiometric age de minations of Brown and Lajoie (1971) and the subsequ age determination on the Okote Tuff (Fitch and Miller, 19 seemed to confirm this interpretation. Conversely, the magne stratigraphy seemed to provide a check of the KBS Tuff determination previously obtained by Fitch and Miller (197 However, Hillhouse et al. (1977) suggested an alternate in pretation of the paleomagnetic properties displayed by Koobi Fora Formation on the basis of new samples and two younger dates of 1.60 ± 0.05 Myr and 1.82 ± 0.05 Myr the KBS Tuff by Curtis et al. (1975). They proposed that

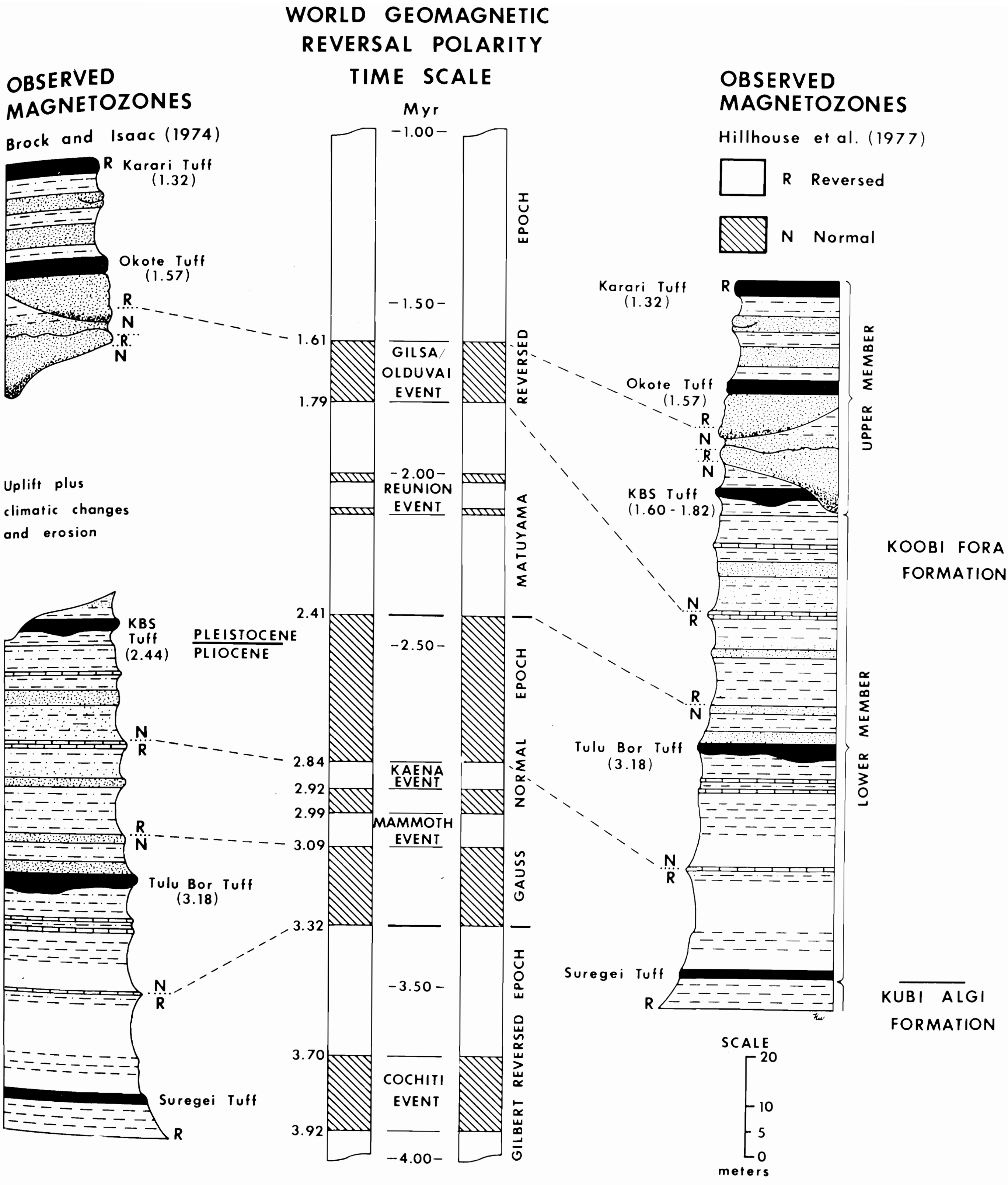


Figure 10. Suggested correlation of observed magnetozones in the Koobi Fora Formation with the geomagnetic polarity time scale, modified after Brock and Isaac (1974) and Hillhouse et al. (1977). The stratigraphic section presented represents a composite section of the Koobi Fora; positioning of the observed magnetozones is tentative. The age of the Plio-Pleistocene boundary used is that of Boellstorff (1977) and the age of the KBS Tuff are those of Hurford et al. (1976, 2.44 Myr) and Curtis et al. (1975, 1.60-1.82 Myr).

reversed magnetozone beneath the KBS Tuff correlates with the lowermost reversed interval of Matuyama age which includes the Olduvai event and ranges from 2.41 to 1.79 Myr. The Tulu Bor Tuff is included in the uppermost normal interval of the Gauss age contradicting the 3.18 ± 0.09 Myr age determined by Fitch and Miller (1976). The upper portion of the Lower Member including the KBS Tuff and the lower portion of the Upper Member were equated with the Olduvai event. This suggests that the unconformity separating the two members and locally recording the erosion of up to 11 meters of the Lower Member represents an insignificant gap in time. This alternate interpretation was an attempt to satisfy the Curtis et al. (1975) date for the KBS Tuff and suid-based correlations [suggested by Cooke (1976), and established by Harris and White (1977)] of strata immediately beneath the KBS Tuff with Member G of the Shungura Formation in the Omo, and Bed I at Olduvai Gorge (See Fig. 10).

Obviously, there are major problems with each interpretation. Both are calibrated against radiometric ages of the KBS Tuff, so resolution of the controversy concerning the age of the KBS will tend to clarify the interpretations. The Brock and Isaac (1974) interpretation suggests that the Plio-Pleistocene boundary more or less corresponds with the unconformity between the Lower and Upper Member regardless of whether the age estimate of Berggren (1971) of about 1.85 Myr or Zagwijn (1974) of about 2.5 Myr or the recent Boellstorff (1977) date of 2.5 ± 0.1 Myr is taken as the correct date for the Plio-Pleistocene Boundary as defined and accepted in 1948 by the General Assembly of the 18th Geological Congress (Intern. Geol. Congress, 1948). The unconformity resulted from regional uplift and a significant climatic change–a change from more humid conditions than exist in the area today (Bonnefille, 1976; Cerling, et al., 1977) to arid conditions. This is documented by the regression of paleo-Lake Turkana (Bowen, 1974; Vondra and Bowen, 1976, 1977) and a change to a fluvial regime (Vondra and Burggraf, 1978), by a change in composition and texture of the sandstone (Bowen, 1974), by a change in heavy minerals (Mathisen, 1977), by a change in oxygen-isotope values of carbonates and the presence of certain zeolites (Cerling, et al., 1977), and by the fossil fauna (Harris, 1976; Cooke, 1976).

If the base of the Pleistocene is climatically defined in East Africa, the unconformable contact between the Lower and Upper Members of the Koobi Fora Formation is a logical boundary between the Pliocene and Pleistocene. And if the recomputed Fitch, et al. (1976) date for the KBS is correct, this boundary is compatible with the fission track date obtained by Boellstorff (1977) for the Plio-Pleistocene boundary (base of the Calabrian) at the Calabrian stratotype area in southern Italy.

SUMMARY

The identification and understanding of relationships between time-transgressive facies and associated isochronous units can be achieved only through careful and detailed geologic study. The prograding delta model presented above demonstrates how adjacent, apparently conformable, lithostratigraphic units actually converge in a time sense. Therefore, the dating of subjacent or superjacent strata relative to an isochronous horizon is largely meaningless except in a purely relative sense. It is certainly ill-advised to attempt to date the isochronous horizon by the strata which underly it.

The unconformity separating the Lower and Upper Members of the Koobi Fora Formation documents a dramatic change in climate and tectonic activity in East Africa. It is here suggested that this break corresponds with the Pliocene/Pleistocene boundary regardless of which previous estimate is used as the age of the base of the Calabrian at its stratotype area. Assuming that the unconformity is, in fact, the Pliocene/Pleistocene boundary at East Turkana, the recent fission track date of 2.5 Myr (Boellstroff, 1977) for the base of the Calabrian provides a best fit with the chronology established by Fitch and Miller (1976).

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