Geol. Mag. **141** (*1*), *2004, pp. 41–53.* © *2004 Cambridge University Press*
DOI: 10.1017/S001675680300815X *Printed in the United Kingdom*

Gombe Group basalts and initiation of Pliocene deposition in the Turkana depression, northern Kenya and southern Ethiopia

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(*Received* 6 January 2003; *accepted* 3 July 2003)

Abstract – A little before 4 Ma ago, deposition of Pliocene and Pleistocene strata described as the Omo Group began in the Turkana and Omo basins of northern Kenya and southern Ethiopia. Soon after, basaltic magma erupted as thin lava flows, and intruded as dykes into the oldest Pliocene strata of the basin. These flows and intrusions are similar petrographically and geochemically, and mark a basaltic magmatic event spanning latitudes from 2°45′ N to 6°45′ N at a longitude of about 36° E. By 3.94 Ma, this basaltic magmatic activity had ceased. Previous researchers used these lavas as an important seismic marker in their study of the southern part of the Turkana Basin. Subsequent volcanic eruptions formed North, Central and South islands in Lake Turkana, and the Korath Range in southern Ethiopia. Thus there was a hiatus in basaltic magmatic activity of nearly 4 Ma in the area presently occupied by Lake Turkana and the lower Omo Valley, although volcanism continued on the eastern margin of the basin. Here we review the field occurrences of these basalts, their distinctive petrography, composition, age and significance to Pliocene deposition in the basin.

Keywords: basalts, Africa, Kenya, Ethiopia, geochronology, tectonics.

1. Introduction

The Turkana depression is a vast lowland, incorporating Lake Turkana in northern Kenya and the lower Omo Valley of southern Ethiopia (Fig. 1). The southern (Kenyan) part is known as the Turkana Basin, and the northern (Ethiopian) part is called the Omo Basin. Ebinger *et al*. (2000) distinguished a number of other basins within the Ethiopian part of the region. The Turkana region preserves the longest primate fossil record in Kenya, ranging from Late Oligocene time (Boschetto, Brown & McDougall, 1992) to Late Pleistocene time, and is perhaps best known for its Pliocene and Early Pleistocene mammalian fossils, which include many hominids (e.g. Leakey & Leakey, 1978; Wood, 1991).

The basement of the region is formed by Precambrian metamorphic rocks, commonly exhibiting a Pan-African thermal overprint, that are locally overlain by Cretaceous sedimentary rocks, or by Oligocene and Miocene volcanic rocks with intercalated sedimentary sequences. Each outcrop area of Oligocene and Miocene strata (e.g. Loperot (Joubert, 1966), Lothidok (Boschetto, Brown & McDougall, 1992), Locherangan (Anyonge, 1991), Buluk (Watkins, 1986), Kajong (Savage & Williamson, 1978)) appears to have had an independent history, for there are no obvious relationships between sedimentary units within these older sequences. Thus, correlation between Miocene sedimentary strata in the region is based mainly on dating of volcanic units in the local sections. In contrast, Pliocene and Pleistocene strata of the Shungura, Koobi Fora, Nachukui and Usno formations share a common history, reflecting deposition in a single sedimentary basin or related set of sub-basins. Hence there was a fundamental change in the style of deposition between the Miocene Epoch and the Pliocene Epoch. We believe that this change is tectonic in origin, and that eruption of early Pliocene basalts is a related phenomenon. The present study focuses on the age and composition of these basalts, and their relation to the initiation and evolution of the Omo and Turkana basins. Following Berggren *et al*. (1995), we place the beginning and end of the Pliocene Epoch at 5.3 Ma and 1.77 Ma, respectively.

Pliocene and Pleistocene strata of the Omo Group (de Heinzelin, 1983) have been described as the Mursi Formation, the Nkalabong Formation, the Usno Formation and the Shungura Formation in southern Ethiopia, and as the Nachukui Formation and Koobi Fora Formation in northern Kenya. Useful summary information on lithology, thickness and extent of these formations is given in de Heinzelin (1983), Brown & Feibel (1991), Harris, Brown & Leakey (1988) and McDougall & Feibel (1999).

On the basin margin east of Lake Turkana, Watkins (R. T. Watkins, unpub. Ph.D. thesis, Univ. London,

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1983) recognized two groups of volcanic and sedimentary strata: the Asille Group of Oligocene–Miocene age and the Gombe Group of Pliocene age. These are separated by an angular unconformity in the Suregei region northeast of Lake Turkana. Watkins suggested a minimum duration of 5 Ma for development of the angular unconformity between the Asille Group and the Gombe Group.

The Asille Group (1700 m thick) dips 5–7° to the southwest, but the Gombe Group (200 m thick) is nearly flat-lying and is only slightly faulted in the Suregei region. The Gombe Group consists dominantly of aphyric basalt flows, but also contains thin sedimentary intercalations which Watkins used to divide the group into four formations. From base to top these are the Bas, Chen Alia, Warata and Harr formations. Watkins recognized the distinctive nature of the Gombe Group basalts (see also Watkins, 1986), and named them after the Gombe Plateau that lies east of the northern half of Lake Turkana (Fig. 1). Basalts from other localities in the Turkana Basin are compositionally identical with the Gombe Group basalts described by Watkins. These basalts are of considerable significance to the most recent episode of rifting in the region, as they provide a link between geophysical and geological work. Essential relations between the sedimentary rocks and the Gombe Group basalts discussed here are shown in Figure 2.

Dunkelman, Karson & Rosendahl (1988) depicted the structure beneath Lake Turkana as a series of six half-graben with centrally located Quaternary volcanic centers linked end-to-end by structural accommodation zones, and refer to the assemblage of structural features as the Turkana Rift. Morley *et al*. (1999) delineated the major structures through interpretation of seismic sections produced in conjunction with petroleum exploration during the 1980s, and through reprocessing of the seismic data obtained by Dunkelman, Karson & Rosendahl (1988). Their interpretation agrees reasonably well with that of Dunkelman, Karson & Rosendahl (1988), although the alternating half-graben are not as clearly identifiable. Dunkelman, Karson & Rosendahl (1988) noted a prominent seismic reflector, the ‘Turkana reflector’, beneath the lake, which they believed corresponded to the top of ‘Miocene–Pliocene flood basalts’. Morley *et al*. (1999) also used a prominent reflector in the sections as the top of the Miocene section, and contoured the thickness of sedimentary strata above it. We believe it likely that this reflecting layer constitutes the basalts of the Gombe Group.

2. Distribution and field relations of the Gombe Group basalts

2.a. Surface outcrops

There are two areas of extensive exposure of early Pliocene basalts, one in the northern Omo Valley and the other on low plateaus east of Lake Turkana. These are the Mursi basalt and the type Gombe Group basalts, respectively. In addition to these broad areas of exposure, several smaller exposures of basalts west of Lake Turkana are assigned here to the Gombe Group.

The northern extent of the Gombe Group basalts is the northern limit of the Mursi basalt as mapped by Davidson (1983). He describes these rocks as ‘flood basalts that extend as an immense sheet northward beyond the boundaries of the project area at 6°45′ N, a distance in excess of 150 km, and having a total thickness probably less than 100 m. Throughout its extent the Mursi basalt is composed of a relatively few, thin, columnar flows of dark gray basalt, locally carrying scattered plagioclase phenocrysts and chlorite filled amygdules’. WoldeGabriel & Aronson (1987) report a petrographically similar basalt above the Moiti Tuff in the Omo River Canyon which has an age of 3.94 ± 0.20 Ma. This basalt, however, is markedly different in composition from basalts of the Gombe Group (see analysis of sample ET62 in G. WoldeGabriel, unpub. Ph.D. thesis, Case Western Reserve Univ., 1987).

Sedimentary strata of the upper part of the Mursi Formation are strongly oxidized immediately below the Mursi basalt (Butzer, 1976; Haesaerts, Stoops & van Vliet-Lanoë, 1983). Basaltic dykes cut early flows of the Mursi basalt in the southern part of its extent, but no dykes are known to intrude the Mursi Formation. Near the type locality of the Mursi Formation (de Heinzelin, 1983), the basalts are overlain by the Nkalabong Formation which we now know includes the Tulu Bor Tuff (3.4 Ma) and the Lokochot Tuff (3.6 Ma).

Further east in the lower Omo Valley a basalt 5–10 m thick lies at the base of the Usno Formation (de Heinzelin, 1983). Over 10 m of sedimentary strata underlie this basalt, and a basalt dyke projects into the Omo River near the same locality.

The most extensive outcrops of Gombe Group basalts occur east of Lake Turkana (see Fig. 1), where they cap the Suregei and Gombe plateaus. Two basalts have been formally named within this group (Key & Watkins, 1988): the Harr Basalt and the Chen Alia Basalt. Hooker & Watkins (1980) describe the Harr Basalt as a unit capping the Miocene Asille Group. Watkins (unpub. Ph.D. thesis, Univ. London, 1983) distinguished the Harr Basalt from the underlying Chen Alia Basalt on the basis of widespread, thin bioclastic limestone and sandstone beds that lie between them. Outliers of the Harr Basalt occur as small plateaus in southern Ethiopia (Davidson & Rex, 1980; Davidson, 1983). Both of these basalts were assigned to the Gombe Group by Watkins in his doctoral thesis. Gombe Group lavas extend from the Ethiopian border to Gus, and crop out far to the east of Lake Turkana (see Fig. 1). They are generally less than 50 m thick. Key & Watkins (1988) believe that the Bulal lavas, which were defined

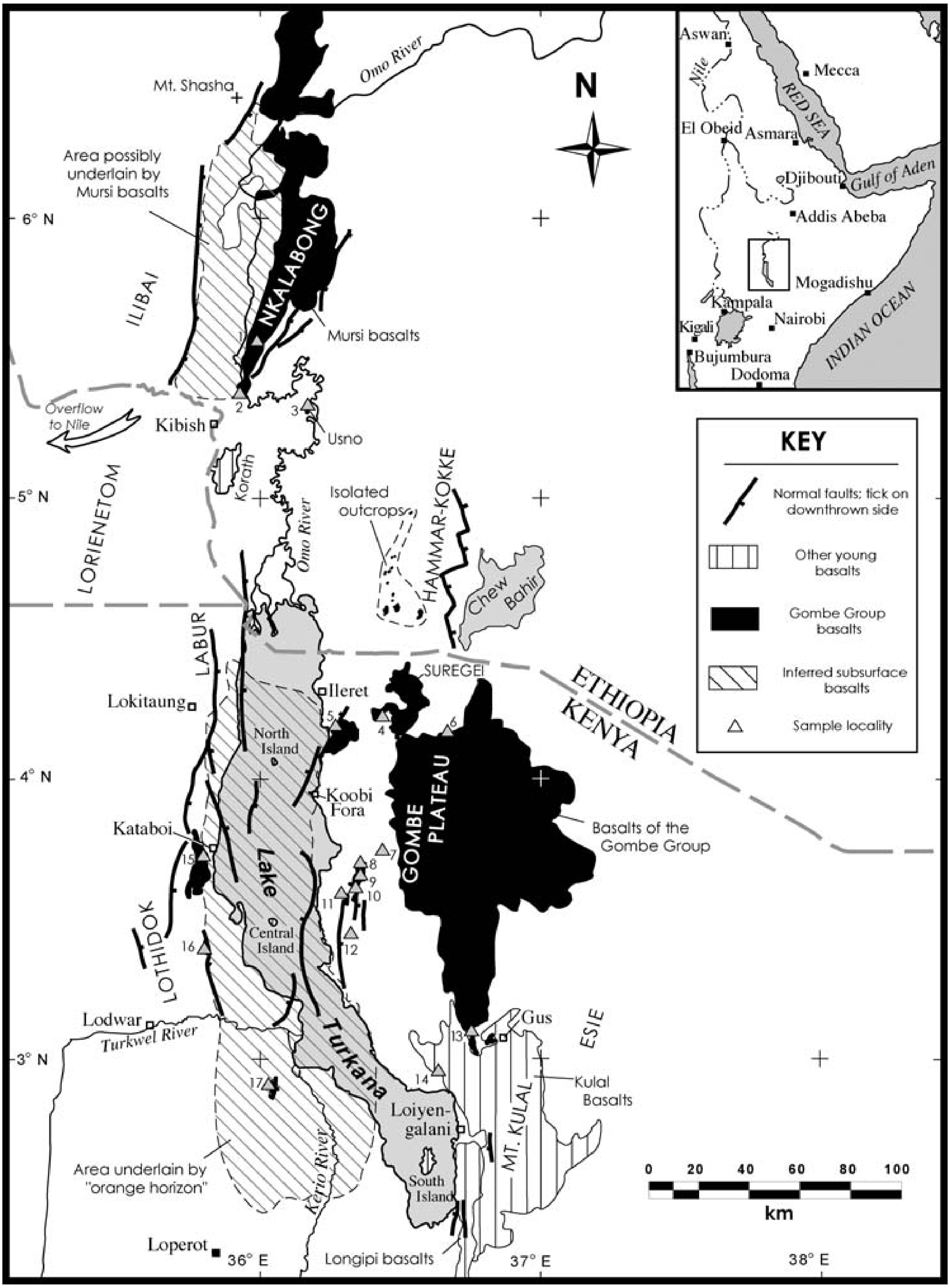

Figure 1. Distribution of basalts discussed in this paper. Triangles show locations of analysed basalts. Numbered localities are: 1 – Mursi (NK-51); 2 – Mursi (KIB01–138); 3 – Usno (WS-1); 4 – Harr (K83-1530); 5 – Kokoi (IL02-81); 6 – Bakate (ANU81-170); 7 – Il Ingumwai (ANU81-187); 8 – Lokochot (K00-6133, K00-6135); 9 – Garma Tochu (K83-1551); 10 – Karsa (ANU 92-409; ANU78-1023, 1025); 11 – Il Moiti (K83-1694); 12 – Basalt dyke SE of Jarigole (K84-1815); 13 – Furaful (K90-4580); 14 – Lenderit (K91-3538); 15 – Kataboi (K81-530, K89-3448); 16 – Lodwar-Kalakol Road (K86-2764); 17 – Lothagam (K86-2764, K86-2897, K86-2899, K86-3000). Sources: Baker (1963); Charlesley (1987); Davidson (1983); Dodson (1963); Key (1987); Key & Watkins (1988); Ochieng' *et al*. (1988); Wilkinson (1988); Walsh & Dodson (1969); Morley *et al*. (1999).

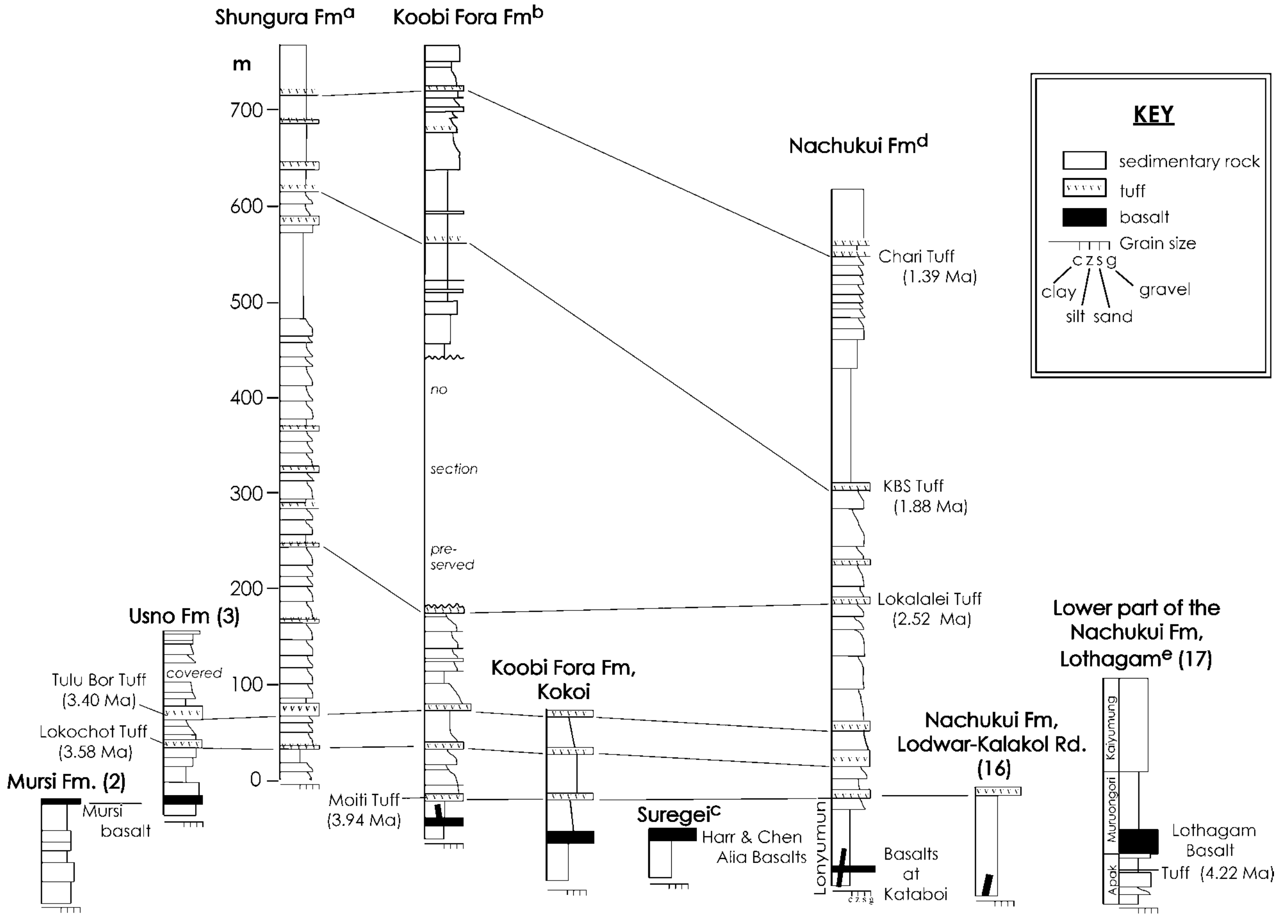

Figure 2. Stratigraphic columns at localities where Gombe Group basalts have been identified, showing that they are confined to the lowest part of the Pliocene sections (see also Watkins, 1986). The Lonyumun Member of the Koobi Fora and Nachukui formations lies below the Moiti Tuff. Parenthetic numbers following column titles refer to the location numbers on Figure 1. Notes: a – Shungura Formation column after de Heinzelin (1983); b – Koobi Fora Formation column after Brown & Feibel (1991); c – Nachukui Formation column after Harris, Brown & Leakey (1988); d – Lothagam section after McDougall & Feibel (1999).

by Davidson (1983) and extend to at least 37°35′ E, also belong to the Gombe Group.

Besides the extensive basalts on the Suregei and Gombe plateaus, many smaller exposures of Gombe Group basalts exist east of Lake Turkana. The northernmost of these is on the Kokoi Highland, an upfaulted block of the Koobi Fora Formation on which strata of the Lonyumun Member are intruded by basalts. A fault, which extends into the lake (Morley, Harper & Wigger, 1999), defines the northwestern extent of the structure. Narrow valleys expose dykes with 1–2 cm black glassy margins, and lava flows with hexagonal columnar jointing. The Tulu Bor, Lokochot and Moiti Tuffs crop out on the southern flanks of the block. These basalts are also exposed in low hills south of the Kokoi Highland.

At the southern edge of outcrops of the Koobi Fora Formation near Allia Bay, basalt dykes intrude a mollusc-packed sandstone in the Lonyumun Member of the Koobi Fora Formation (Fig. 2). The intrusive relation is evident from the chilled margins of the basalt, from the thermal alteration of the mollusc shells, and from the crushing of mollusc shells orthogonal to the contact. The dykes do not extend to stratigraphic levels as high as the Moiti Tuff, which Leakey *et al.* (2001) dated at 3.94 ± 0.03 Ma (see Fig. 2). Therefore, the vertical extent of the mollusc bed at this locality provides no basis for estimating the minimum depth of the Lonyumun Lake (*contra* Brown & Feibel, 1991).

At Karsa, a columnar basalt overlies a thin sequence of sedimentary strata that yielded molluscs of Zone 1 of Williamson (1981). The basalt was believed to underlie the Lonyumun Member of the Koobi Fora Formation, but instead it may be coeval. Not far to the northeast, a similar lava overlies a sedimentary section about 15 m thick, and at nearby Garma Tochu, basalt dykes and flows of the Gombe Group are exposed. Much further south, at Furaful, is a 20-metre section with basal basaltic tuffs overlain by diatomite and capped by a basalt flow (Fig. 1). Gombe Group basalts extend at least as far as Lenderit (location 14, Fig. 1), where a flow is interbedded in a pyroclastic complex, and is overlain disconformably by the Tulu Bor Tuff and associated sediments. Other petrographically similar plateau basalts also occur in this area, but analysed samples have lower TiO_2 contents, and those which have been dated are younger than the Gombe Group basalts.

Basalts with similar petrography and composition occur at three localities west of Lake Turkana: Kataboi, Lodwar–Kalakol Road and Lothagam. The best exposures are in a gorge west of the town of Kataboi (Fig. 1), where Lonyumun Member strata about 10 m thick are confined between columnar basalts, each a few tens of metres thick with columns about 1 m across. Dykes associated with the younger basalt sheet are 0.5–2 m thick, have a general N–S strike and are nearly vertical. The base of the upper basalt is spheroidally weathered as is also the case at Lothagam. Southwest of Kataboi, basalt dykes intrude Lonyumun Member strata, but do not extend upward to the level of the Moiti Tuff (Harris, Brown & Leakey, 1988). Similar intrusive features are seen at an exposure along the road between Lodwar and Kalakol, on the eastern edge of the Lothidok Range. Again the Moiti Tuff is present near the top of the section (see Fig. 2), but it is not affected by the basalt.

McDougall & Feibel (1999) named the Lothagam Basalt that forms the ridges of the western half of Lothagam Hill. It is about 50 m thick, has well-developed columns, local chilled margins and spheroidal weathering. It lies between the Apak Member and the (lacustrine) Muruongori Member of the Nachukui Formation (McDougall & Feibel, 1999). Patterson, Behrensmeyer & Sill (1970) originally described this basalt as a sill, but McDougall & Feibel (1999) suggest that it was 'emplaced as a flow moving into soft sediments . . . '. The Muruongori Member is lithologically similar to the Lonyumun Member of the Nachukui Formation, and is of approximately the same age.

The basalt at Lothagam marks the known southern extent of Gombe Group basalts west of Lake Turkana. However, a petrographically similar basalt, the Kalokwanya Basalt, lies above the Pliocene strata at Kanapoi about 70 km south of Lothagam. The Kalokwanya Basalt and other basalts near the Kalabata River are included with the Kaling–Kanapoi basalts, which extend south to 2° N latitude (Ochieng' *et al*. 1988). Like the Lothagam basalt, the Kalokwanya basalt is of reversed paleomagnetic polarity (D. Powers, unpub. Ph.D. thesis, Princeton Univ., 1976). The single sample available, however, is strikingly different in composition from the Gombe Group basalts, so this basalt is not discussed further.

2.b. Probable subsurface occurrences

Two areas in the Turkana Depression are believed to be underlain by similar basalts, one in the northern part of the lower Omo River valley, and the other beneath Lake Turkana. On the basis of a small inlier near Mt Shasha (Fig. 1), Davidson (1983) suggested that 'the Mursi basalt underlies much if not all, of the cover of the middle Omo plain, and its buried extension may continue southward beyond the Nyalibong [Nkalabong] range'. Further south, Morley *et al*. (1999) used a strong reflection on seismic data for correlation of seismic horizons in the Kerio Basin, and called the reflector the Orange horizon (or 'O horizon'). This layer was penetrated in the Eliye Springs well, from which a sample of an igneous rock close to the location of the reflection gave an age of 5.1 ± 0.2 Ma. They note also that 'the O horizon correlates to a lava flow dated at 4.3 ± 1.9 Ma'. Morley, Harper & Wigger (1999) contoured the depth to the O horizon, and show its spatial distribution; they also show the depth to the orange horizon decreasing to zero at Lothagam Hill, where the Lothagam Basalt crops out (see also fig. 22 in Morley *et al*. 1999). On the basis of this information, these areas are included as possible extensions of the surface outcrops of the Gombe Group basalts in Figure 1. Two observations in the seismic sections are important to this account: (1) there is only one strong reflector in sections beneath Lake Turkana, and (2) the reflector is thin, probably not exceeding 300 m (see next section).

3. Geochronology of Gombe Group basalts

Samples dated for this study used techniques that are already well described elsewhere (e.g. McDougall & Schmincke, 1977; McDougall, 1985; McDougall & Feibel, 1999). Precision of the K/Ar ages is generally about 1% standard deviation. New K/Ar data and previously published K/Ar data are compiled in Table 1, excluding analyses with $> 5\%$ error. Previous K/Ar ages are reported by Fitch & Miller (1976); Brown & Nash (1976); Brown *et al*. (1985); Feibel, Brown & McDougall (1989); McDougall & Watkins (1985, 1988); Key & Watkins (1988); Davidson & Rex (1980); Watkins (unpub. Ph.D. Thesis, Univ. London, 1983); Wilkinson (1988) and Zanettin *et al*. (1983). All ages used here were calculated using published or measured analytical data on potassium and argon, and the decay constants of Steiger & Jäger (1977). Where noted in Table 1 they disagree significantly with the numerical ages reported in the original publications.

Although reasonably precise ages have been determined for many Gombe Group basalts, in some samples the ages may not reflect their time of eruption or intrusion because of extraneous argon leading to ages which are anomalously old (e.g. the Lothagam Basalt). In other samples, the original glassy groundmass has been partly altered to chlorite, and the measured ages may be too young. The groundmass of these rocks probably contains much of the potassium, and this is where much of the radiogenic argon is likely to be generated. However, basaltic glasses, especially if altered, are unlikely to retain radiogenic argon quantitatively, leading to younger apparent ages. Thus, the effect of alteration on K/Ar ages depends critically

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Table 1. K/Ar dates and analytical information for Gombe Group lavas from the Turkana and Omo Basins, northern Kenya and southern Ethiopia

	Location no. on Figure 1	K (%)	40Ar* (mol/g)	40Ar* (%)	Age (Ma)	Error (Ma)	Source reference
Mursi basalt							
KA-2094	2	0.83	5.967E-12	24.5	4.16	0.2	1
ANU82- 24	2	0.926, 0.916	6.39E-12	54	3.99	0.04	2
Usno basalt							
LKA- 2	3	0.707	3.918E-12	64.6	3.19	0.15	4
ANU83- 23-1	3	0.691, 0.698	4.89E-12	23.6	4.08	0.06	2
ANU83- 23-2	3	0.691, 0.698	4.92E-12	24.8	4.06	0.06	2
Kokoi basalts							
ANU81-114	5	0.649, 0.646	6.03E-12	38.3	5.36	0.06	5
ANU81-119	5	0.721, 0.721	7.31E-12	25.3	5.83	0.08	5
ANU81-113	5	0.684, 0.685	7.30E-12	38.2	6.14	0.07	5
ANU99-240	5	1.041, 1.060	7.17E-12	27.3	3.93	0.06	This paper
Chen Alia basalt							
ANU81-172	6	0.995, 0.993	6.441E-12	27.8	3.73	0.05	3
ANU81-170	6	0.994, 0.992	6.503E-12	33.9	3.77	0.05	3
Harr Basalt, Feijej							
89FJ-3	–	0.86	0.5445	42.0	4.42	0.07	9
Gombe Group Lavas, Suregei Plateau							
FM7936	–	1.27	7.05E-12	22.2	3.18	0.10	8
FM7936	–	1.27	7.36E-12	22.4	3.34	0.10	8
SL07	–	0.86	7.32E-12	23.6	4.88	0.09	8
SL07	–	0.86	7.14E-12	31.6	4.75	0.09	8
HR07	–	0.88	7.45E-12	36.2	4.88	0.09	8
HR07	–	0.88	7.45E-12	36.2	4.88	0.09	8
PA22	–	0.61	5.67E012	18.5	5.31	0.14	8
PA22	–	0.61	5.71E012	17.2	5.35	0.14	8
Gombe Group Lavas, Il Ingumwai							
ANU81-187	7	0.816, 0.814	5.956E-12	46.4	4.21	0.05	3
ANU81-189	7	0.764, 0.767	5.984E-12	34.5	4.50	0.05	3
Lokochot basalt							
ANU00-300	8	0.758, 0.754	5.33E-12	31.8	4.06	0.05	This paper
Garma Tochu basalt							
ANU92-411	9	0.855, 0.850	6.11E-12	22.9	4.12	0.05	This paper
ANU92-410	9	0.701, 0.702	5.86E-12	35.8	4.81	0.05	This paper
Karsa basalt							
ANU92-409	10	0.811, 0 815	6.63E-12	17	4.69	0.07	This paper
ANU92-409 (2)	10	0.811, 0.815	6.70E-12	24.4	4.74	0.05	This paper
ANU78-1025	10	0.804, 0.803	5.54E-12	29.2	3.97	0.05	5
ANU78-1023	10	0.829, 0.826	6.25E-12	43.8	4.35	0.05	5
Basalt dyke SE of Allia Bay							
ANU92-412	12	0.839, 0.834	4.69E-12	26.5	3.23	0.04	This paper
Furaful basalt							
ANU92-424	13	0.816, 0.816	5.53E-12	52.4	3.91	0.04	This paper
Lenderit basalt							
ANU92-437	14	0.704, 0.704	5.16E-12	18.4	4.22	0.06	This paper
Kataboi basalt							
ANU92-414	15	0.904, 0.910	6.21E-12	46.3	3.94	0.04	This paper
ANU83-19-1	15	0.679, 0.684	4.83E-12	26.8	4.08	0.05	2
ANU83-19-2	15	0.679, 0.684	4.78E-12	23.2	4.04	0.06	2
Lodwar-Kalokol Road basalt							
K263	16	0.86	1.043E-11	11	6.98	0.3	7
Lothagam Basalt							
Plagioclase separates							
ANU87-3-1(2)	17	0.328, 0.331	2.694E-12	20.7	4.70	0.09	6
ANU93-1056	17	0.327, 0.330	2.78E-12	13.6	4.87	0.09	6
Whole rock basalts							
ANU87-3-1	17	1.014, 1.011	8.858E-12	35.9	5.04	0.06	6
ANU87-18-1	17	1.138, 1.141	1.059E-11	35.7	5.35	0.07	6
ANU93-1056	17	0.857, 0.862	6.81E-12	56.3	4.56	0.05	6
ANU93-1058	17	0.751, 0.751	7.92E-12	39.1	6.07	0.06	6

$\lambda_\varepsilon + \lambda_{\varepsilon'} = 0.581 \times 10^{-10}$ a^{-1}; $\lambda_\beta = 4.962 \times 10^{-10}$ a^{-1}; ^{40}K/K $= 1.167 \times 10^{-4}$ mol mol^{-1}.
Sources: 1. Brown, 1969; 2. Brown *et al.* 1985; 3. McDougall & Watkins, 1988; 4. Brown, de Heinzelin & Howell, 1970; 5. McDougall, 1985; 6. McDougall & Feibel, 1999; 7. Zanettin *et al.* 1983; 8. Watkins (unpub. Ph.D. thesis, Univ. London, 1983); 9. Asfaw *et al.* 1991.

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on the nature and history of the rocks. There is sufficient variation in the age of single basalts such as those at Karsa and Lothagam that all of the measured ages cannot be correct. The total age range shown in Table 1 is 3.18–6.98 Ma, but there is considerable geological evidence to suggest that this range is far too large.

We accept only a minority of measured K/Ar ages (13 out of 42 listed in Table 1) as possibly valid. Our position is somewhat less radical than that taken by McDougall & Feibel (1999), who excluded all of the K/Ar ages on Lothagam Basalt because the ages were inconsistent with other stratigraphic and geochronological control on that unit.

Three lines of evidence suggest that the Gombe Group basalts were erupted over a reasonably short period of time. First is that in 12 seismic sections presented in Morley (1999, appendix A), the Orange ('O') horizon appears to be a single layer or a closely associated pair of layers. This is in accord with observations in outcrops, where the Gombe Group basalts are seen to consist of only one flow or only a few flows with associated dykes. If the Gombe Group basalts were erupted over the long time period suggested by the large dispersion in the K/Ar ages, then one would expect to see multiple strong reflectors within the basinal sediments, However, these do not occur. Moreover, as the youngest ages (~ 3 Ma) are about half the value of the oldest ages (~ 6 Ma), one would expect that strong seismic reflectors would occur through half the section, but they do not.

The second line of evidence is that over the entire area where the Orange horizon is recognized, it probably does not exceed 200 to 300 m in thickness, corresponding to two-way travel times of < 0.1 s, using a velocity of 3300 m/s, typical of shallow fractured basalts. This observation agrees with the sections accessible around the basin margin. Even including local thickening caused by topography at the base of the Gombe Group, the entire group, including the underlying sediments, does not exceed 220 m (R. T. Watkins, unpub. Ph.D. thesis, Univ. London, 1983).

The third line of evidence is that all Gombe Group basalts in the Turkana Basin observed thus far predate the Moiti Tuff, which establishes a minimum age of 3.94 Ma for the entire group. It is more difficult to place a maximum age on the basalts using stratigraphic arguments, but at two localities, tuffs believed to underlie the Lonyumun Member afford some constraints. At Kanapoi, fluvial conditions persisted until at least 4.12 Ma ago, when a thin lacustrine sequence was deposited, and then capped by a channel filled with a tuff dated at 4.07 ± 0.02 Ma (Leakey *et al.* 1998). Thus, at this locality lacustrine deposition began around 4.1 Ma ago. At Lothagam, McDougall & Feibel (1999) measured an age of 4.22 ± 0.03 Ma on anorthoclase from pumice clasts in a bed beneath the lacustrine sequence of the Muruongori Member. Thus, here too, the lacustrine phase that marks the Lonyumun Member of the Nachukui and Koobi Fora formations post-dates 4.2 Ma, consistent with the information from Kanapoi. As the Gombe Group basalts intrude the lowest part of the Lonyumun Member, they must be somewhat younger than 4.2 Ma in age.

Another consideration is that Gombe Group basalts are of reversed palaeomagnetic polarity wherever the polarity has been measured. Thus it is likely that they were erupted during the later part of the Gilbert Chron (3.58–4.18 Ma; Cande & Kent, 1995), and this together with the stratigraphic constraint of the overlying Moiti Tuff suggests that most lie in the narrow temporal range 3.9–4.18 Ma.

K/Ar ages on Gombe Group basalts with less than 5 % analytical error are compiled in Table 1. Given that ages on these basalts are highly suspect, as one may conclude from the work at Lothagam, there is little point in discussing the individual results. Many ages fall within the interval 3.95 to 4.22 Ma, but this may be fortuitous. Rather, ages in that interval are simply consistent with the stratigraphic control that we have. Our experience with dating these lavas suggests that conventional K/Ar dates do not yield good estimates of the time of eruption of these particular lavas. Careful collection of additional samples in stratigraphic context may resolve the age range of these lavas, and $^{40}Ar/^{39}Ar$ step heating experiments on selected samples also would assist in determining the reliability of measured ages.

4. Petrography

The Gombe Group basalts are remarkably uniform petrographically. They are medium grained (0.1–0.2 mm), and composed primarily of plagioclase (~ 50 %), clinopyroxene (~ 20 %), glass (~ 20 %) and Fe–Ti oxides. They are either aphyric or contain sparse small phenocrysts of plagioclase, clinopyroxene and rarely olivine. Plagioclase forms a medium-grained matrix of interlocking laths up to 1 mm long. The grains are sub- to euhedral, and some crystals show normal zoning and multiple albite twinning, with Carlsbad-albite twinning in larger grains. Clinopyroxenes form euhedral crystals up to 0.5 mm across in some cases. Where present, interstitial glass is gray, dark brown, or light purplish brown, and may contain oxide dendrites. In most samples, the plagioclase, clinopyroxene and iron oxide appear unaltered, but olivine, where present, commonly shows incipient alteration. Glass may be quite fresh and isotropic, or altered to a green mineral of low birefringence, probably chlorite. McDougall & Feibel (1999) describe anhedral clinopyroxene crystals that represent a glomerocryst or a xenocrystic aggregate. Other samples have glomerocrystic aggregates of plagioclase and pyroxene. Some glassy margins of dykes contain glomerocrystic aggregates, or phenocrysts of sharply euhedral plagioclase, euhedral pyroxene and Fe–Ti oxides.

Table 2. Major and trace element analyses of representative Gombe Group basalts from the Turkana and Omo Basins, northern Kenya and southern Ethiopia

Sample	NK-51	KIB01-138	WS-1	K83-1530	K00-6133	K83-1694	K84-1815	K81-530	K86-2764	K90-4661	309F[b]
Site name	Nkalabong	Mursi	Usno	Suregei	Il Lokochot	Il Moiti	SE of Jarigole	Kataboi	Lodwar-Kalakol Rd.	Lothagam	Mursi
No. on Fig. 1	1	2	3	4	8	11	12	15	16	17	–
SiO_2	48.72	49.06	49.49	50.5	50.64	50.61	49.26	49.68	50.13	49.62	49.80
TiO_2	3.38	3.44	3.46	3.64	3.56	3.63	3.67	3.69	3.71	3.53	3.56
Al_2O_3	13.14	13.35	13.37	13.28	13.14	13.19	13.02	13.47	13.14	13.28	13.63
$Fe_2O_3^a$	14.64	14.77	14.84	14.65	14.39	14.62	15.26	13.96	15.05	14.39	15.11
MnO	0.21	0.21	0.22	0.22	0.21	0.21	0.22	0.21	0.20	0.22	0.21
MgO	4.7	4.86	4.85	4.36	4.17	4.05	4.54	3.67	4.35	4.38	4.57
CaO	8.61	8.45	8.64	8.14	8.07	8.1	8.7	8.6	8.11	8.18	8.82
Na_2O	3.06	3.21	3.17	3.46	3.52	3.35	3.15	3.33	3.28	3.3	3.00
K_2O	1.01	0.98	0.85	0.8	0.88	0.98	0.85	0.84	1.02	0.85	0.86
P_2O_5	0.48	0.47	0.49	0.58	0.61	0.60	0.51	0.60	0.56	0.57	0.52
LOI	1.41	0.83	0.42	0.29	0.61	0.96	0.65	1.66	0.57	0.9	0.23
Total	99.36	99.64	99.82	99.92	99.8	100.3	99.82	99.7	100.12	99.23	100.31
Normative Minerals (wt %, assuming Fe^{3+}/Total iron = 0.1)											
Quartz	0.36	0.00	0.68	2.15	2.36	2.85	1.07	2.83	1.88	1.92	1.49
Plagioclase	46.60	47.47	47.48	48.28	48.26	47.36	46.54	49.08	46.71	48.26	46.99
Orthoclase	6.21	5.97	5.14	4.79	5.32	5.91	5.14	5.14	6.15	5.20	5.14
Diopside	17.67	16.89	16.93	15.47	15.93	15.57	17.96	16.85	15.75	15.44	16.26
Hypersthene	19.21	19.14	19.72	18.43	17.64	17.70	18.70	15.35	18.79	18.75	19.82
Olivine	0.00	0.52		0.00							
Ilmenite	6.61	6.70	6.70	7.01	6.91	7.03	7.12	7.24	7.18	6.91	6.82
Magnetite	2.20	2.20	2.19	2.15	2.13	2.16	2.26	2.09	2.22	2.15	2.28
Apatite	1.16	1.11	1.16	1.37	1.44	1.41	1.20	1.44	1.32	1.37	1.20
Trace elements (ppm)											
V	377	376	383	362	348	363	412	378	372	347	380
Cr	23	<20	23	<20	<20	<20	<20	<20	<20	<20	35
Co	39	41	40	34	33	36	40	35	36	35	56
Ni	56	86	51	77	78	83	90	94	79	81	20
Cu	67	87	85	29	27	267	47	52	34	38	61
Zn	129	140	136	149	139	153	156	164	133	135	151
Ga	23	22	23	24	24	24	23	24	23	24	23
Ge	2	2	1	2	2	2	2	2	2	1	1
Rb	17	18	18	22	21	20	19	18	21	20	17.3
Sr	369	364	381	371	358	368	358	377	346	367	387
Y	40	37	39	45	47	47	42	46	43	45	46.8
Zr	225	227	232	262	268	271	242	262	245	255	256
Nb	25	27	26	29	29	31	27	29	28	29	30
Ba	322	303	322	312	310	324	276	303	294	295	263
La	29.0	28.7	28.0	31.9	33.3	33.3	29.7	32.5	30.6	31.9	27.7
Ce	63.8	61.9	61.7	69.6	73.0	72.4	65.1	70.8	66.9	68.7	63.4
Pr	8.3	8.0	8.1	9.1	9.6	9.5	8.5	9.2	8.7	9.0	
Nd	36.7	35.6	35.5	40.2	42.3	41.3	37.4	41.1	38.4	39.7	36.2
Sm	8.4	8.1	8.2	9.5	9.8	9.6	8.9	9.5	9.1	9.3	8.75
Eu	2.9	2.8	2.8	3.1	3.3	3.2	3.0	3.1	3.0	3.1	2.8
Gd	8.4	8.2	8.3	9.4	9.9	9.7	8.9	9.6	9.2	9.2	
Tb	1.4	1.3	1.3	1.5	1.6	1.6	1.5	1.6	1.5	1.5	1.41
Dy	7.9	7.7	7.9	9.1	9.4	9.4	8.6	9.1	8.8	8.8	
Ho	1.5	1.5	1.5	1.7	1.8	1.8	1.7	1.8	1.6	1.7	
Er	4.2	4.0	4.1	4.8	4.9	5.0	4.5	4.8	4.6	4.7	
Tm	0.6	0.5	0.6	0.6	0.7	0.7	0.6	0.7	0.6	0.6	
Yb	3.4	3.3	3.4	3.9	4.1	4.0	3.7	3.9	3.7	3.8	3.89
Lu	0.5	0.4	0.5	0.5	0.6	0.5	0.5	0.5	0.5	0.5	0.61
Hf	6.2	6.0	6.0	6.7	6.8	7.0	6.5	7.0	6.5	6.5	6.36
Ta	2.0	1.9	2.0	2.2	2.3	2.2	2.0	2.2	2.1	2.1	1.93
Pb	7	6	5	6	6	7	6	8	<5	<5	12
Th	2.9	2.6	2.8	3.0	3.2	3.2	2.8	3.1	2.9	2.9	2.8
U	0.76	0.68	0.74	0.76	0.82	0.82	0.73	0.80	0.75	0.77	0.9

a – Total iron as Fe_2O_3; b – analysis from Stewart & Rogers (1996).

5. Chemical composition

Representative basalt samples were selected for major and trace element analysis, excluding those with obvious alteration. Major elements were analysed by standard techniques of X-ray fluorescence spectroscopy. Trace elements and rare earth elements were measured by inductively coupled plasma mass spectrometry (ICP-MS) at Activation Laboratories Ltd, Ancaster, Ontario. Precision is generally on the order of 1% of the amount present. Twenty-seven samples of Gombe Group basalts were analysed, and representative analyses are given in Table 2. There is little variation in the elemental composition of the

Gombe Group basalts, even though samples come from as much as 300 km apart.

Compositionally these rocks are quartz normative, high-titanium basalts with rather low MgO that lie near the boundary between alkaline basalts and tholeiitic basalts (Le Bas *et al*. 1986), and may be classified as transitional basalts.

Of about the same age are the Tirr Tirr Basalts, first reported by Baker (1963). These are situated at about 1° N, west of Baragoi, Kenya. They are 3.6–3.9 Ma in age and lie with angular unconformity on older volcanic rocks in the area (Sawada *et al*. 1998). Kabeto *et al*. (2001) give analyses of two of these, one of which (KY9701305 in their table 2) is compositionally similar to the lavas under discussion from Turkana, but in general these lavas are distinctly lower in SiO_2, and higher in iron, MgO and CaO.

Late Miocene and Pliocene basaltic rocks from the Middle Ethiopian Rift Valley reported by Hart *et al*. (1989) are lower in TiO_2 content than the Gombe Group basalts. Incompatible element ratios show other differences, with samples from the Middle Ethiopian Rift Valley generally having lower values of Ba/La, Ba/Th, Ba/Ta, K/Ba, Th/Ta, La/Th, La/Ta, Sr/Nd and Ti/Zr. Similarly, basaltic glasses from the Shaban Deep (Red Sea), are distinct from the Gombe Group basalts in having lower TiO_2 (Haase, Mühe & Stoffers, 2000).

Although the compositional range in these lavas is small, there is a highly significant positive correlation between P_2O_5 and Zr, which suggests that the basalts are the product of simple partial melting followed by crystal fractionation (Hooper, 1997). MgO is positively correlated with CaO, suggesting that the abundance of these elements may be controlled by the abundance of clinopyroxene. Neither Al_2O_3 nor iron content is significantly correlated with CaO content.

6. Discussion

From the description above, it appears that basalts erupted about 4 Ma ago over an area extending from the lower Omo Valley (6°45′ N) to the Kerio Valley (2°45′ N) and from 35°30′ to 36°45′ E. The duration of this episode appears to have been relatively short. Nowhere have these basalts been found to intrude or overlie the Moiti Tuff (3.94 ± 0.03 Ma: Leakey *et al*. 2001), and at Lothagam the basalts are clearly less than 4.22 Ma in age (McDougall & Feibel, 1999). Although we cannot presently exclude samples of similar composition that have yielded older K/Ar ages, wherever there is stratigraphic constraint, these lavas are younger than the base of the Lonyumun Member and older than the Moiti Tuff, so the probable maximum duration is about 300 000 years. Because of the close correspondence in composition, it is possible that the duration was considerably less.

Prior to intrusion and extrusion of the basalts, a thin sequence of lacustrine deposits had been deposited over nearly the entire area eventually covered by Gombe Group basalts. Near the western edge of the basin, at least 120 m of sedimentary strata, mainly lacustrine siltstones and claystones, were deposited before intrusion of the basalts. Such sediments can be deposited very rapidly. For example, near Koobi Fora, 100 m of lacustrine strata below the KBS Tuff (1.88 ± 0.02 Ma: McDougall, 1985) were laid down in about 100 000 years, yielding an accumulation rate of ~ 1 m/1000 years. Thus the interval between the inception of Pliocene deposition and intrusion of dykes and eruption of the lavas was probably brief.

Sedimentary strata into which the basalts were intruded, or over which they were extruded, include the Lonyumun Member of the Koobi Fora Formation, the Lonyumun Member and Apak Member of the Nachukui Formation, the Kanapoi Formation and the Mursi Formation. The basalts themselves have been locally named in many places: the Mursi basalt in southern Ethiopia, the Harr Basalt, Chen Alia Basalt and Gombe Group basalts to the east of Lake Turkana and the Lothagam Basalt to the west. Given that these basalts are macroscopically distinctive, it is useful to refer to all of them by a single name, for which Gombe Group basalts seems appropriate. Chronologically they are younger than the oldest parts of the Koobi Fora, Nachukui and Mursi formations, but as they are lithologically distinct from these formations, Gombe Group basalts should be treated as a unit that interdigitates with and/or cuts across sedimentary strata of about the same age.

The ultimate cause of the eruptions of basaltic material is most likely stretching of the lithosphere in the area, which has been modelled by Hendrie *et al*. (1994) and also by Morley *et al*. (1992). Their models concentrate on transects across the southern part of the Turkana depression, near Lothagam. Hendrie *et al*. (1994) estimate that at the end of the Miocene Epoch, extension was 25–30 km across the profile, and by the end of the Pliocene Epoch, total extension is estimated at 35–40 km. In other words, extension of about 10 km occurred from about 5 Ma to the present day in the southern part of the basin. Morley, Harper & Wigger (1999) show that subsidence associated with this extension is concentrated in a relatively narrow zone approximating 25–30 km in width. In the central part of the basin, this zone is mainly situated between the faults that form the eastern boundary of the Lothidok Range and faults that lie near the eastern side of the lake. In the northern part of the basin, the extensional zone is bounded mainly by faults near the eastern and western edges of Lake Turkana itself. If we assume that all of the extension is accommodated in this narrow zone, then stretching factors (β) would be in the order of 1.5–1.6. This is almost certainly an overestimate, because faults bounding the Nachukui Formation west of the lake, and faults within that formation, account for some of the extension. At the latitude of Koobi Fora, faults through the Koobi Formation are minor. In both the Koobi Fora Formation and in the Nachukui

Formation, some major faults involve units as young as the Nariokotome Tuff (1.33 Ma), and the Silbo Tuff (0.74 Ma). At Kalochoro, at least 350 m of motion is necessary in the past 1.3 Ma to bring the Kokiselei Tuff in contact with the Nariokotome Tuff. North of the Kokoi, at least 450 m of motion is necessary in the past 0.74 Ma to bring the Silbo Tuff in contact with the lower part of the Lonyumun Member of the Koobi Fora Formation. If we include the boundary fault west of the lake in the zone of extension, its width would be 35–40 km, yielding stretching factors of 1.3–1.4.

Morley, Harper & Wigger (1999) show a maximum 2-way travel time of $\sim$3.5 seconds above the 'orange' reflector on profiles between Koobi Fora and the western edge of the lake, but the average 2-way travel time is about 2.5 seconds. Using their seismic velocity depth conversion, this corresponds to a maximum thickness of 5 km of sedimentary section, and an average section thickness of about 3.5 km.

The amount of tectonic subsidence is given by the sediment thickness divided by the amplification factor, F, where

$$F = (\rho_m - \rho_f)/(\rho_m - \rho_s)$$

and ρ_m is the density of the mantle, ρ_f the density of the fluid filling the basin, and ρ_s the density of sediment fill in the basin (cf. Archimedes *in* Commandino, 1565). If we assume the initial basin was subaerial, and take $\rho_m = 3.3$, $\rho_f = 0$ and $\rho_s = 2.3$, then F would be a maximum of 3.3, so that the initial tectonic subsidence would be between 1000 and 1500 m. According to White & McKenzie (1989), this implies that the stretching factor, β, should be between 1.2 and 1.8, within the range estimated above (1.2–1.5).

Lake Turkana today is at an elevation of $\sim$400 m, and there is reasonable geological evidence that waters from the area flowed to the Indian Ocean for much of the past 4 Ma (Brown & Feibel, 1991). This implies that the region was above sea-level. In order to maintain this elevation, the central part of the basin must have been filled very rapidly as it subsided. Before $\sim$4.5 Ma, the Omo River probably flowed into the Nile system through the gap between Lorienetom and Ilibai (Fig. 1). Until the beginning of the Pliocene Epoch, there is no evidence that the Omo River flowed into the basin, but the only sensible source of vast amounts of fine-grained sediment needed to fill the subsiding depression is the Omo River. Thus it is likely that subsidence caused by stretching of the lithosphere in the region was responsible for capture of the Omo River by the Turkana depression. Once captured, sedimentation could have easily kept pace with subsidence, maintaining the basin at approximately its present elevation. The area underlain by sediment 3000 m thick has dimensions of 30×100 km, so the volume that must be filled is 9×10^{12} m^3. If this is deposited over the 3 Ma period from 4 to 1 Ma, then 3×10^6 m^3 of sediment must have been deposited each year, which, at a density of 2.5×10^6 kg/m^3 corresponds to a sediment mass of 7.5×10^{12} g. Cerling (1986) estimated the annual mass sediment input at 10–20 $\times 10^{12}$ g from the Omo River, so an area twice this large could easily have been filled from the Omo alone. The depression need only have been $\sim$100 m below the overflow level at the time it began to be filled by the Omo River. Eruption and intrusion of the Gombe Group basalts acted to preserve the initial lacustrine sediments beneath them, and near the edges of the ancient lake, vertebrate fossils are likewise preserved. After the depression was filled by detritus from the Omo River, the fluvial system extended over it, eventually finding its way out of the basin through the low point somewhere in the vicinity of Mt Kulal about 3.9 Ma ago. Volcanic activity in the region of Mt Kulal (itself between 2.5 and 2 Ma old: unpub. data) may have severely constricted this route of escape, so that at various times (e.g. 3.5 Ma, 2 Ma) the river waters were impounded to form lakes for brief periods of time. Still later, construction of Longipi volcano $\sim$0.7 Ma ago (unpub. data) may have provided the final plug in the basin. Climatically induced highstands of the lake after closure caused waters of the Omo to flow into the Nile system several times. However, following each highstand of the lake, the Omo River flowed southward to the region now occupied by Lake Turkana.

The volume of the Gombe Group basalts is quite small. Even if we assume that the entire floor of the basin is underlain by similar basalts, which is hinted at by Davidson (1983), the area covered is about 500 km by 100 km, and as the thickness is normally <100 m, the volume must be <5000 km^3. Thus the volcanism that followed inception of Pliocene sedimentation was not only of limited duration, but also of limited quantity.

7. Conclusions

Watkins (1986) first noted the association of basaltic volcanism and sedimentation during the Pliocene Epoch in the Koobi Fora area east of Lake Turkana. Here we have shown that basalts of very similar composition erupted very widely in the Turkana Basin at about the same time that Pliocene deposition began in the region. On the basis of current dates, it is possible that this volcanism extended over a period of $\sim$2.5 Ma as suggested by Watkins (1986), yet most of the volcanism appears to have occurred over a much shorter period of no more than 0.3 Ma, coincident with the beginning of deposition of the Omo Group. This is part of what gives these basalts their special significance. Many of these basalts have local names (e.g. Mursi, Usno, Lothagam, etc.), but all can be assigned to a single group: the Gombe Group, implicitly defined by Watkins (unpub. Ph.D. thesis, Univ. London, 1983), which is the most widely used name for these related

https://doi.org/10.1017/S001675680300815X Published online by Cambridge University Press

lavas. East of Lake Turkana, the Gombe Group consists of at least three flows separated by thin lacustrine sedimentary sequences (Watkins, 1986). This report greatly expands the area over which these basalts are known to occur, and also shows that in many instances dykes were the source of the flows in the central and western parts of the basin, including the Kokoi highland in the Koobi Fora region. The Gombe Group interdigitates with the basal part of the Omo Group of Pliocene and Pleistocene sedimentary strata, and it appears that the inception of deposition and eruption of basalts are two related responses to regional tectonic evolution. Individual age determinations on Gombe Group lavas should be treated with caution, for there are many examples of dates which are younger, and other dates that are older than is possible given the stratigraphic control and the precise ages on anorthoclase feldspar from tuffs within the stratigraphic succession. Finally, recognition of the compositional similarity of this group of basalts, their occurrence in the Mursi and Usno formations in the Omo Valley, and their occurrence west of Lake Turkana provide strong support for the identification of the Turkana reflector (the Orange horizon of Morley *et al*. 1999) as the Gombe Group basalts, which was originally proposed by Dunkelman, Karson & Rosendahl (1988), although they did not use this term.

Acknowledgements. Field work was supported by the National Science Foundation through grants to F. Clark Howell for the International Omo Research Expedition, by grants from the National Science Foundation, the National Geographic Society, and the L. S. B. Leakey Foundation to F. H. Brown for work on the Koobi Fora and Nachukui formations, by grants to John G. Fleagle and F. H. Brown for work on the Kibish Formation, and by grants to Meave G. Leakey for work at Koobi Fora, Lothagam and Kanapoi. Analytical work for K/Ar and $^{40}Ar/^{39}Ar$ dating was supported by the Research School of Earth Sciences, The Australian National University. We thank Peter Hooper, Ronald Watkins, and Giday WoldeGabriel for constructive comments on an earlier version of this manuscript.

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