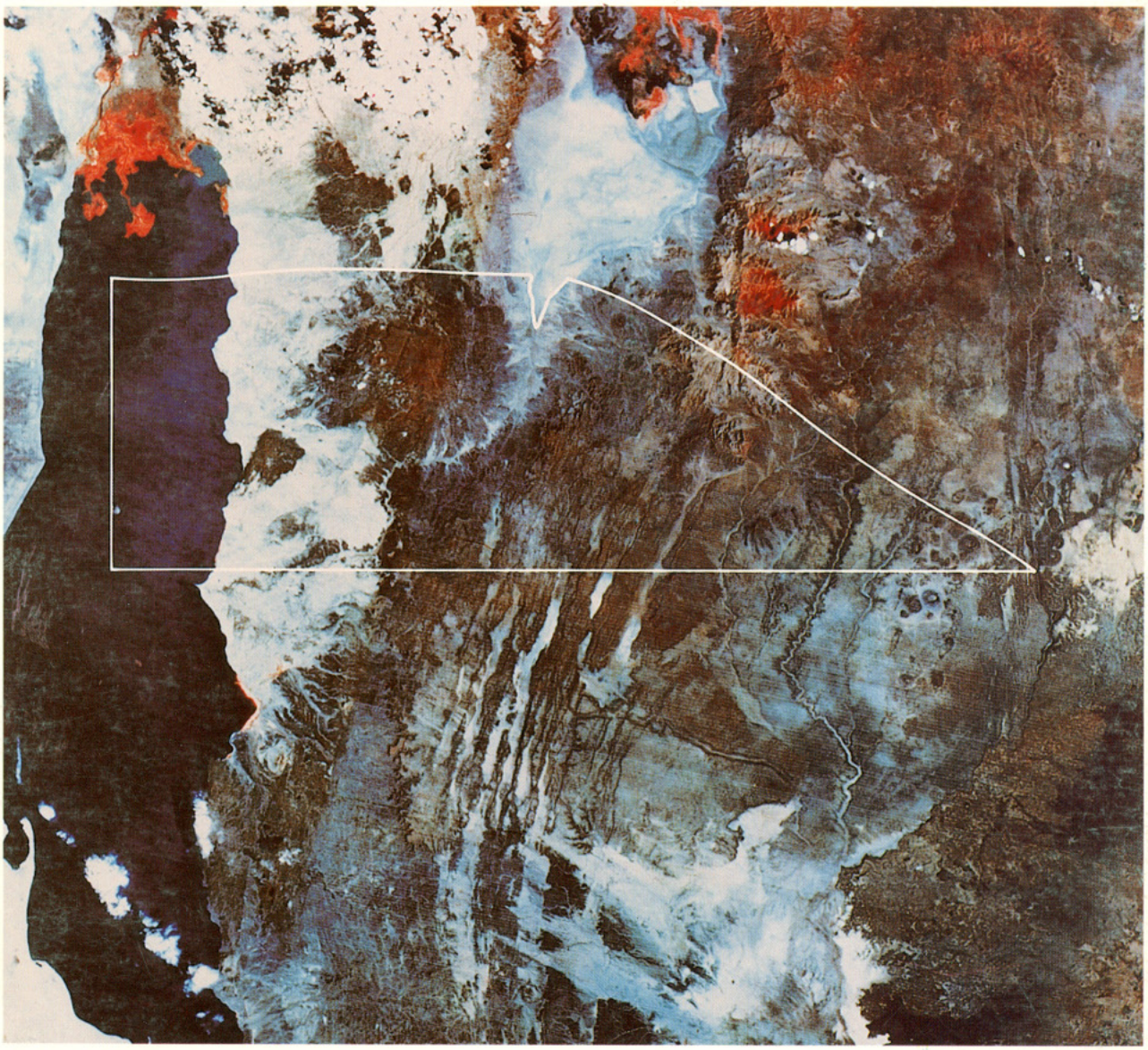
REPUBLIC OF KENYA

MINISTRY OF ENVIRONMENT AND NATURAL RESOURCES

MINES AND GEOLOGICAL DEPARTMENT



# Geology of the Sabarei area

Report 111 (Reconnaissance)

1988

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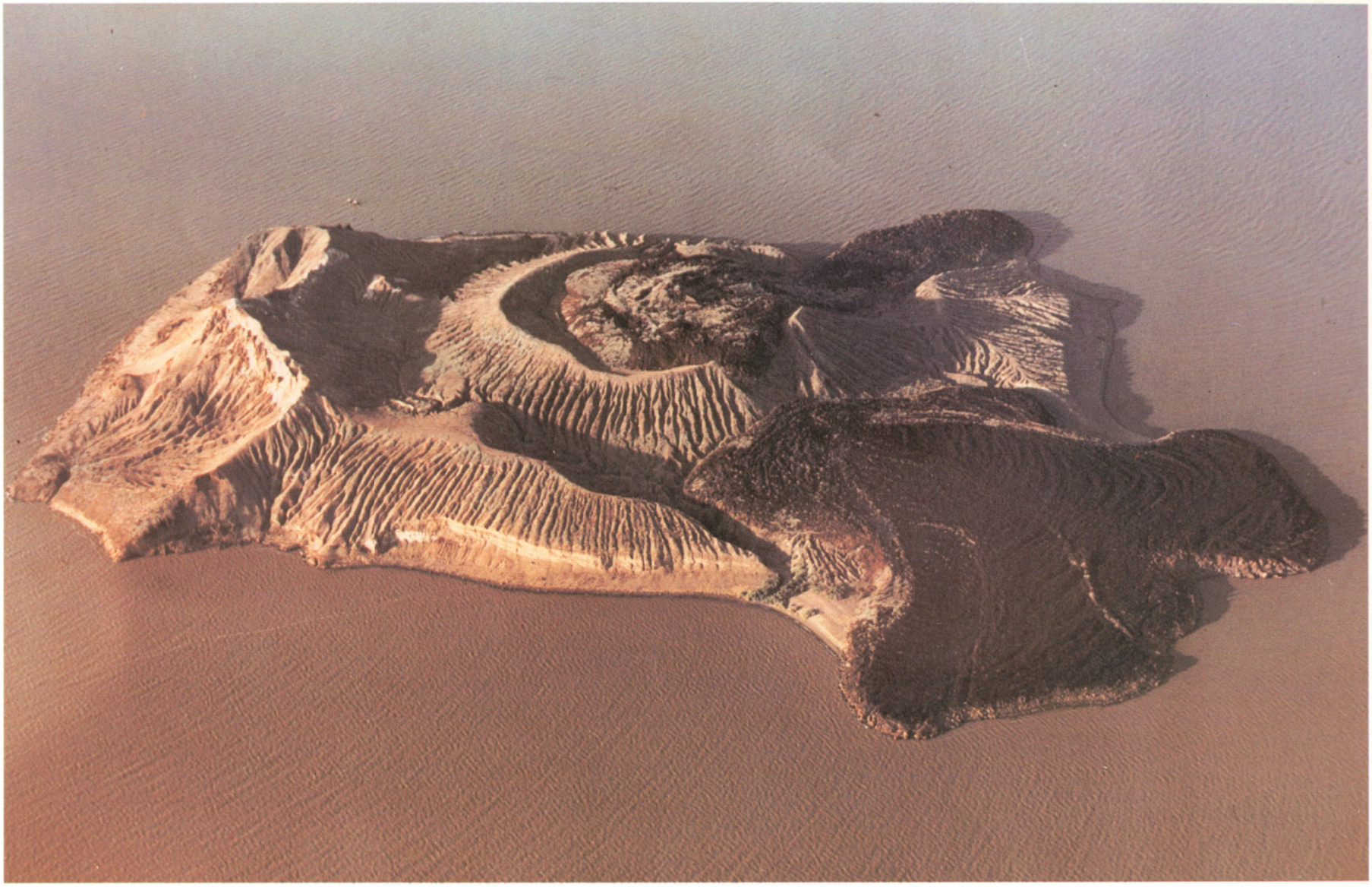


**Plate 1** Aerial view of North Island showing the two youngest lava flows breaching the central crater wall. The deeply weathered diorite plug (Plate 2) is on the left side of the picture. (*Frontispiece*)

MINISTRY OF ENVIRONMENT AND NATURAL RESOURCES
Mines and Geology Department

Report 111 (Reconnaissance)

# Geology of the Sabarei area

Degree Sheets 3 and 4
with coloured 1:250 000 geological map and
results of geochemical exploration

R. M. Key and R. T. Watkins

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MAP

1:250 000 geological map of Degree Sheets 3 and 4 (in back pocket)

# PREFACE

Three years before R. M. Key surveyed most of this remote frontier region by helicopter-supported reconnaissance (in early 1985), R. T. Watkins had mapped the Suregei–Asille area, on the western shoulder of the Lake Stephanie (Chew Bahir) rift, in detail as part of a University of London doctoral research project. It happened that early in 1985 security problems rendered the northern part of Asille inaccessible: therefore it was particularly fortunate that Watkins and Key were able to co-operate in order to compile this reconnaissance report for the entire Sabarei area. Each author is responsible for the descriptions and conclusions of their respective areas. The area includes the northern part of the Koobi Fora Basin containing a sedimentary sequence of world renown on account of the hominid fossil remains. These have been the subject of much detailed research. The integration of such information at disparate levels is no easy task, one involving detailed stratigraphic analyses of limited area, the other ranging rapidly over previously unsurveyed adjacent terrain. The resulting compromise provides the first systematic geological description of a fascinating area, which still remains difficult of access. The discovery of the Jibisa Ring Complex with its related iron formation and high gold values in grab samples, must be noted as a probable future target for mineral exploration.

It is satisfying to record that the joint Kenyan–British mapping and mineral exploration project, together with the results of Dr Watkins' research, have led to the successful completion of this report.

We would also like to express our gratitude to the following people for supporting the field work in difficult and sometimes dangerous circumstances: geological assistants A. Gizuli, G. Injila and driver/mechanic S. Lasoi Bwogo, and especially the helicopter pilot Sebastian Tham.

B. D. Hackman
*Samburu-Marsabit Project Leader*
*22 August 1987*

# Geology of the Sabarei area

**Abstract** The Sabarei area (6290 km²) extends along the Ethiopian border north of latitude 4°N and east of longitude 36°E. To the west is the north-east segment of Lake Turkana including North Island. This account comprises the detailed studies of R.T.W. on the area immediately west of Chew Bahir (Lake Stefanie) as well as the results of a helicopter-supported reconnaissance survey of the whole region by R.M.K. during February 1985. Eight physiographic provinces reflect different ages and eruptive styles of Cenozoic volcanics as well as regional variations in crustal neotectonics. Lake Turkana at 378 m and Hafari at 1692 m represent extremes of elevation above sea level. Four internal drainage basins (Lakes Turkana and Chew Bahir, the Kinu Sogo Rift, and North Chalbi) are controlled by major rift features and contain ephemeral rivers and streams. The area is founded on metamorphic rocks of the Mozambique Orogenic Belt, exposed only in Garba Merille. Three periods of Cenozoic volcanism relate to contemporaneous rifting. Initial late Oligocene through middle Miocene tholeiitic, alkali basaltic and rhyolitic volcanism was progressively confined to the rift systems. End-Miocene through middle-Pliocene volcanism was ubiquitously tholeiitic giving rise to extensive flood lavas. Quaternary volcanism has been confined to the North Island trachytic and Dukana basaltic multicentered volcanoes. Fluvial sedimentary rocks form intercalcations within the earliest volcanic succession. Lacustrine and fluvio-deltaic deposits of 'proto'-Lake Turkana are preserved immediately beneath and within the tholeiitic basalt flood lava succession. During the last 4 Ma, such sediments have accumulated in the Lake Turkana Basin amounting to a thickness of 500 m or more in the well-documented, hominid-fossil bearing succession around Koobi Fora. Continuing volcanic activity during this period is evidenced by the numerous felsic tuffaceous beds present in the sedimentary sequence, although the provenance of the pyroclastic material remains uncertain. Cenozoic intrusive centres include the previously unrecognised alkaline Jibisa Ring Complex of early to middle Miocene age. While rifting indicates tensional crustal conditions, there is no direct evidence of lateral spreading across individual rifts. Dextral strikeslip motion along the Kinu Sogo Rift is indicated by drag folds of Pliocene faults. During the early Miocene the Gregory Rift was continuous northwards into the Lake Stephanie Rift. However in the Pliocene this rifting was confined to three discrete N-trending zones on the northern flank of the Kenya Dome as defined by Saggerson and Baker (1965): Lake Turkana Basin, Kinu Sogo and Ririba Rifts. Rifting has continued into the Quaternary in the Lake Turkana Basin although only the west side (outside the map-area) is fault-controlled. The present local demand for mineral resources is very limited. However the mineral potential of the Jibisa Ring Complex and related extrusives is of national importance (for precious, base and rare-earth metals and industrial minerals). Anomalous gold values are associated with individual lithologies of this province and with an altered diorite plug on North Island. Solfataras on the island indicate an active geothermal system in the underlying volcano.

## 1 Introduction

Sheets 3 and 4 (Sabarei) form the northernmost limit of the Phase 2 area of the Samburu–Marsabit Geological Mapping and Mineral Exploration Project; a technical co-operation between the British (Overseas Development Administration funding of the British Geological Survey) and Kenyan (Mines and Geological Department) Governments. The principal objective of this phase of the project was to produce reconnaissance geological maps (of degree squares), on a scale of 1:250 000, of a previously geologically unsurveyed part of northern Kenya between latitude 2°N and the Ethiopian border, and between longitudes 36° and 38°E. Selected stream sediment samples were collected during the mapping to provide data for regional geochemical atlases. Dr D. I. J. Mallick (BGS Overseas Programme) compiled a Landsat imagery interpretation for the entire Phase 2 area, which was used initially to plan fieldwork and also to finalise delineation of geological boundaries.

Most of the area of Sheets 3 and 4 was covered by helicopter-supported field work in February 1985 from a base camp at Koobi Fora (immediately south of Sheet 3) on the eastern shore of Lake Turkana (Figure 1). Fly camps were established at Ileret and Dukana (immediately south of Sheet 4 on the main track to North Horr) for Landrover-supported traverses to: Jibisa, the Dukana Volcanic Province, Tatesa Hills, Sabarei–Hafari, Buluk and within the Lake Turkana Basin (Figure 2). Data from a doctoral study by Watkins (1983) of the Oligocene-Miocene Asille Group and the Pliocene Gombe Group, forming the western shoulder of the Lake Stephanie Rift, were incorporated in this report.

### 1.1 LOCATION

The area covered by the Sabarei sheet is wedge-shaped situated between latitude 4°N and the Ethiopian border, and longitude 36°E (through Lake Turkana). The eastern extremity is defined by the intersection of the Ethiopian border with latitude 4°N (Figure 1). The length of the southern margin along this line of latitude is 176.5 km, the maximum N–S dimension is 50 km (in the west along the shore of Lake Turkana). The sheet represents a total area of about 6290 km although the western 1225 km² is part of Lake Turkana. North Island lies within the SW corner of the Sabarei Sheet.

The whole sheet is part of the Marsabit District of

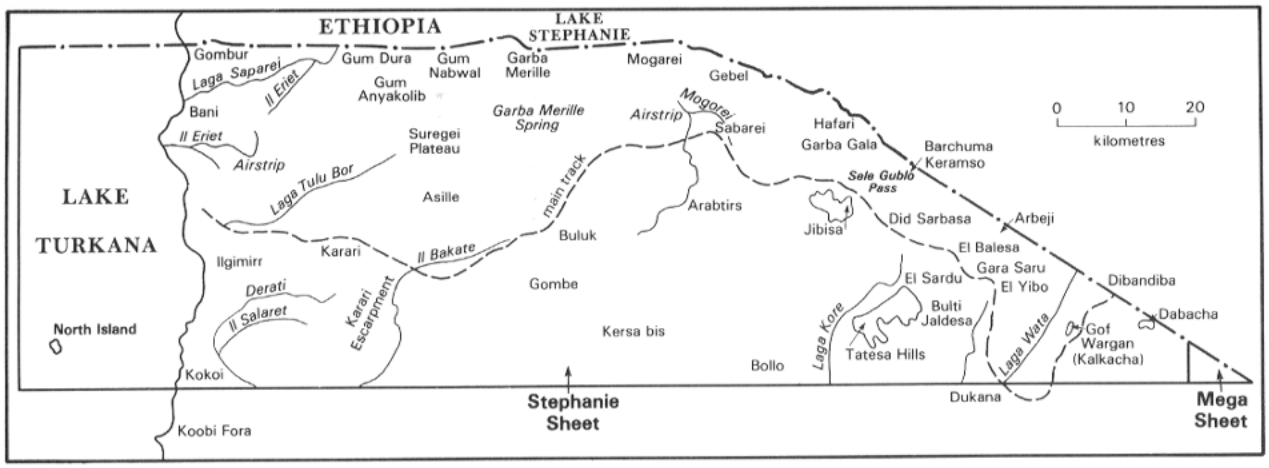



**Figure 1** Location map for the Sabarei sheet with geographical detail

Eastern Province with the district administration headquarters at Marsabit. There is a District Officer at Ileret. Sibiloi National Park, along the eastern shoreline of Lake Turkana extends northwards as far as the Kokoi area where there is a game ranger's office.

The following topographic maps cover the sheet area:

**1:100 000** (incompletely contoured with unnumbered form lines), Series Y633, Edition 1—GSGS, sheets 7 (Jibisa), 6 (Buluk), and 5 (Ileret) form an east (7) to west (5) run. The eastern extremity of the map-area, beyond longitude 37°30°E is not covered by any 1:100 000 sheet.

**1:250 000** (contoured at 1000 ft, intervals) Series Y503, Edition 1—SK, sheet NB-37-13 (Stefanie). This sheet covers the whole area as far east as longitude 37°30°E. A small triangular area east of this meridian is included on sheet NB-37-14 (Mega) which is not available in Kenya (most of the Mega area lies within Ethiopia).

## 1.2 ACCESS

There is a poorly maintained dirt track parallel to, and about 10 km south of, the Ethiopian border which connects Ileret to Dukana (and south to North Horr and Marsabit). This track passes through Buluk, Sabarei, the Sele Gublo Pass (north of Jibisa Mountain) and crosses the Laga Balal at El Yibo, before turning south to Dukana. A lorry track also runs north-north-east from Dukana into Ethiopia. Meandering, poorly signposted, tracks within the Sibiloi National Park connect Kokoi (and north to Ileret on a lorry track) to Koobi Fora and Allia Bay. The track south-east from Buluk to Derati shown on the Stephanie topographic sheet has been eroded at several stream crossings and is presently impassable for motorised vehicles. Dirt airstrips suitable for light aircraft are maintained at Ileret and Sabarei Police Post.

Small fishing boats regularly cross Lake Turkana from the eastern shore around Ileret.

## 1.3 PREVIOUS WORK

The area was first visited by Europeans in 1888 when members of Count Teleki's expedition journeyed to the southern shore of Chew Bahir (Lake Stephanie). Although there is little geological relevance in Hohnel's account of the expedition, he did observe that volcanic rocks formed a covering to much older metamorphics and that these rocks, cropping out to the west of Chew Bahir, have been uplifted relative to those underlying the floor of the lake (Hohnel, 1894). He noticed a variety of felsic units and collected a sample of rock, which Rosiwal (1891) described as a quartz-trachyte and compared to rocks in Shoa Province, southern Ethiopia. The reported strong brick-red colour of the matrix suggests the sample may have been of a distinctive ignimbrite within the upper part of the Langaria Formation.

Subsequently, other exploratory parties (Cavendish, 1898; Harrison, 1901; Hohnel, 1938) passed through the district but did not comment upon the geology which re-

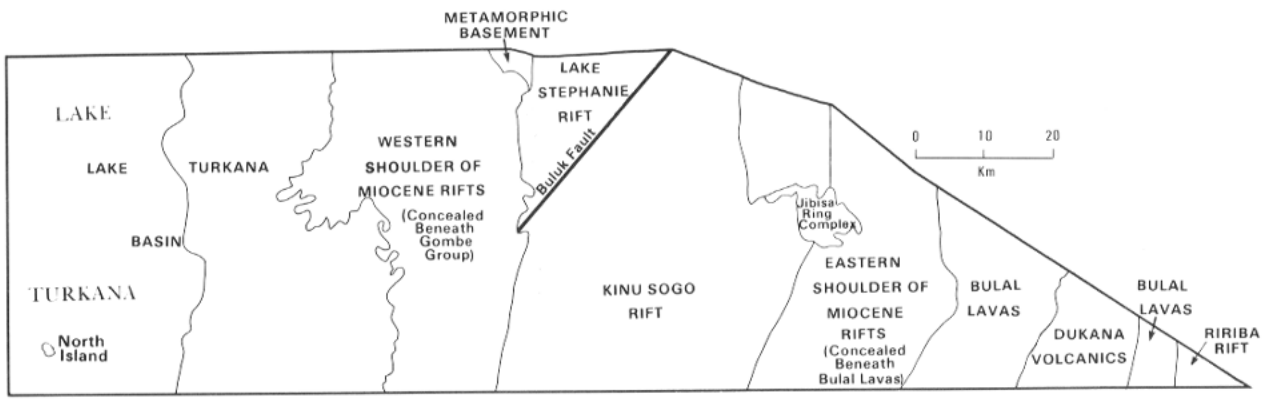



**Figure 2** The major geological provinces of the Sabarei sheet

mained virtually unknown, with the exception of the Lake Turkana Basin, until the studies of the present authors.

The abundance of vertebrate fossils, including hominids, has attracted considerable scientific investigation of the sedimentary and volcanic infill of the Lake Turkana Basin since the original fossil discovery in 1967 (Leakey and others, 1970). A full reference list to work on these Pliocene to Recent rocks is given in Chapter 3.

Regional syntheses of Brotzu and others (1984) on the geochronology and Williams (1978) on the Cenozoic volcanics did not attempt to differentiate the Cenozoic volcanics of this part of northern Kenya.

# 2 Geomorphology

Eight regional physiographic provinces are identified (Figure 3). The diverse landscapes are unstable and reflect different ages and eruptive styles of Cenozoic volcanism as well as regional variations in crustal neotectonics. Lake Turkana forms the northern part of the Quaternary Gregory Rift of Kenya. The present lake level (c.378 m) is relatively low: previously its eastern shoreline extended as much as 30 km eastwards. The central map-area is part of the older (Mio-Pliocene) Lake Stephanie (Chew Bahir*)–Kinu Sogo Rifts. This is a well defined topographic low (c.600 m) along the Ethiopian border where it is flanked by Asille (West) and Hafari (East) massifs of Miocene rock. Southwards the Kinu Sogo Rift is expressed as a series of parallel grabens and blocks defined in plateau lavas. These Mio-Pliocene lavas extend eastwards to form a gently southward dipping plateau away from the Ethiopian Highlands. The impressive Jibisa massif, the Tatesa Hills, and a low range of hills striking south-west from Jibisa form part of an elevated eastern shoulder .to the Miocene Rifts. End-Tertiary through Quaternary volcanism and rifting has modified the eastern landscape to form the eastern Ririba Rift and the Quaternary Dukana Volcanic Province.

Away from the Lake Turkana Basin and the Lake Stephanie–Kinu Sogo Rifts, the map-area generally has elevations of between 600 and 1000 m above sea level. Major exceptions are Jibisa, 1560 m, and Hafari which forms a plateau highland above 1200 m with peaks of about 1692 m in Kenya and exceeding 2150 m across the border in Ethiopia.

Cenozoic rifting and volcanism has been the principal influence on the scenery. There is no evidence of the 'older', pre-Miocene land surfaces recognised elsewhere in Kenya (cf. Saggerson and Baker, 1965; Dixey, 1948) except for the poorly exposed metamorphic basement pavement at Garba Merille. The eight provinces of Figure 3 result from regional volcanic and tectonic events. On a local scale, variations in lithology and individual faults or fractures influence landform.

* Lake Stephanie is renamed Chew Bahir and is referred to as such in this report. However, the former name is retained for the geological structure (the Lake Stephanie Rift).

## 2.1 LAKE TURKANA AND NORTH ISLAND

According to Fuchs (1939) and Dixey (1948), Lake Turkana infills a shallow warped basin, unlike the Gregory Rift to the south which has a well defined fault-controlled central graben as shown by the 1:250 000, Kenya Mines and Geological Survey geological Sheets 27 (Key, 1987a) and 35 (Hackman, in press). Certainly within the map-area the eastern shoreline is not a fault scarp. Rather there is a gentle beach slope to the waters edge and a very gradual increase in water depth westwards. However, crustal warping on the eastern side may be considered a passive response to major normal faulting along the western side of Lake Turkana, the Quaternary Rift being an asymmetric graben in northern Kenya.

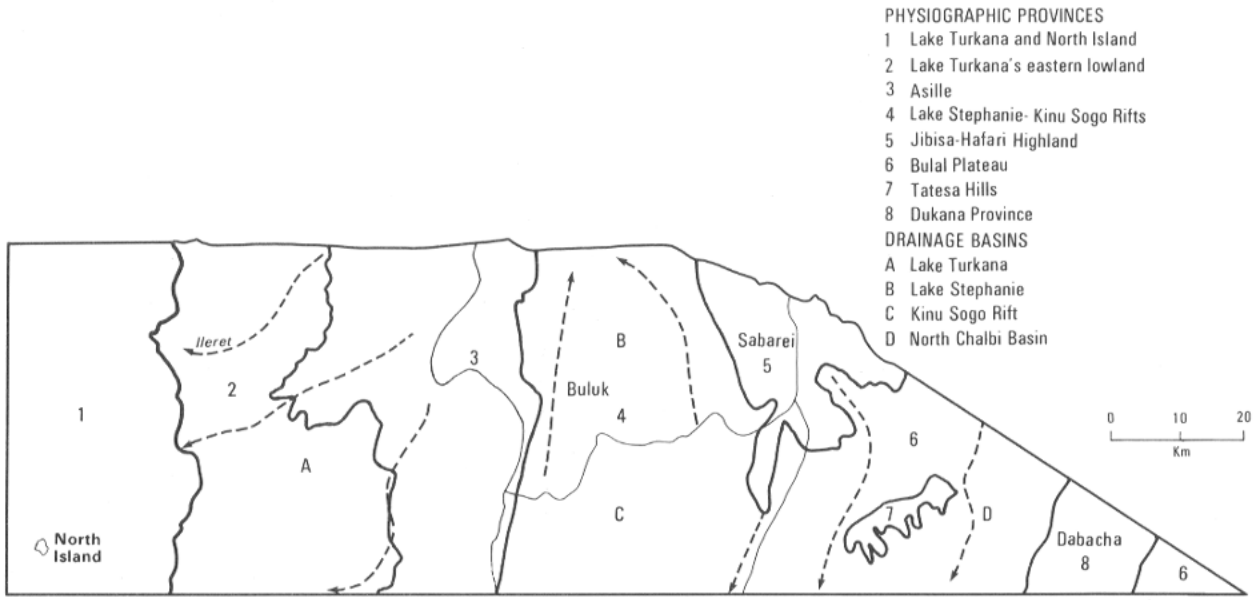



**Figure 3** The eight physiographic provinces and four drainage basins on the Sabarei sheet

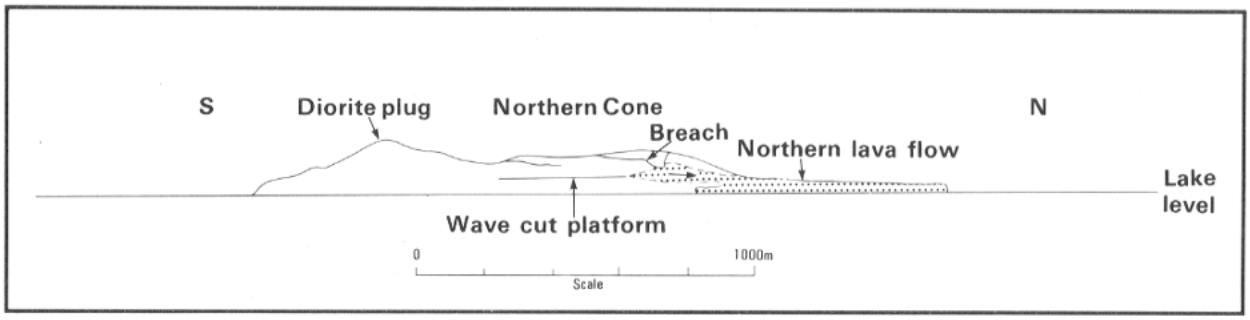



**Figure 4** A N–S profile of North Island

North Island is the summit of the northernmost central trachytic volcano of the Gregory Rift in Kenya. The volcano is active: *solfataras* are present in the south-centre of North Island and two lava flows post-date the most recent high water stand of Lake Turkana dated at less than 10 000 years BP (Brown and Carmichael, 1971). The island has a present surface area of about $2^2$ kilometres with steep cliffs disappearing into deep lake water (Plate 1). Narrow volcanic sand-beaches and banked scree slopes are present locally. The Quaternary volcanism has built a breached crater in the middle of the island which is elevated about 100 m above the lake. The crater has a diameter of several hundred metres with pronounced stream gullying of its outer slopes. A bench terrace is present about 50 m above the present lake level along the east side of the island. Fresh blocky (aa) lava flows breach the west and north sides of the crater. They post-date the wave-cut bench, locally flowing along this feature before descending into the lake. The northern flow (Figure 4) has built a prominent headland whereas the western flow descends a steep crater wall and disappears into the lake at a high angle. The highest part of North Island is the southern hill with an elevation above lake level of about 140 m. This rounded feature is produced by a massive diorite plug infilling an older cone whose summit pyroclastics have eroded away. Geothermal alteration has made the hill's bedrock susceptible to stream erosion producing major gullies on the northern face (Plate 2).

## 2.2 LAKE TURKANA'S EASTERN LOWLAND

This province is roughly delineated between the 2000 ft (c.615 m) contour in the east and the present Lake Turkana shoreline. There is a regional westward slope of 1:125 defining the eastern side of the warped basin infilled by Lake Turkana. Incision by the major rivers draining into Lake Turkana has produced a series of low hilly watershed areas above broad shallow valleys partially filled by various unconsolidated deposits. In part the material is related to previous (Pliocene onwards) incursions of Lake Turkana, such as sand bars related to former shorelines. The monotony of this landscape is broken by numerous small cuestas and mesas capped by resistant strata of the Koobi Fora Formation. These features rarely rise more than several metres above the adjacent low ground. However, at Kokoi in the east, and near Ileret, they are more impressive, with steep slopes in excess of several tens of metres. Faulting is a principal control on the morphology of the upstanding areas. The most prominent topographic feature of this province is the basaltic lava-capped horst of Kokoi. The top of this block is about 200 m above the lake level. Preferential weathering along NNE faults and SSW fractures has produced a rugged summit plateau area to the horst.

## 2.3 SUREGEI–ASILLE HIGHLANDS

The broken hilly terrain of the Suregei–Asille highlands at altitudes of between 600 and 900 m separates the major Lake Stephanie and Lake Turkana Rifts. The highest ground is underlain by dissected plateau lavas which are more continuously developed south-eastwards. Low cuestas and mesas capped by resistant volcanic strata of the Asille Group are common in the area north from Buluk to the Ethiopian border. At the border in the east of the area an eroded upstanding pavement of metamorphic 'basement' is unconformably overlain by Oligocene basalts. Circular ridges trace the eroded remnants of minor Miocene vents, e.g. at Gum Anyakolib (GR 28554 98372). Numerous criss-crossing small faults influence the drainage pattern and the formation of cuestas and mesas. Accumulations of coarse clastic debris are found on many hill and scarp slopes, notably along fault scarps, e.g. to the west of Chew Bahir and at the base of the active fault scarp in the south along easting 2935.

## 2.4 LAKE STEPHANIE (CHEW BAHIR)–KINU SOGO RIFTS

This Mio-Pliocene Rift System has two distinct physiographic aspects. The area north of Buluk forms the southern end of the Chew Bahir depression, whereas to the south is an elevated, rifted, lava plain.

The northern depression consists of an interior lacustrine flat flanked by low plains with centripetal dips. The plains have mantles of various superficial debris including major gravel fans related to drainages from surrounding highlands. Laterally stepped NNW–N-trending fault scarps with steep cliff faces up to 50 m high, separate the Asille highlands from the Chew Bahir depression. To the east a number of small, eroded volcanic cones rise above the colluvial plain. Many of the cones appear to lie above the trace of the NE-trending Buluk Fault, which effectively divides the Chew Bahir depression from the southern plateau. The Buluk Gap which is the major river valley into the south end of Chew Bahir is also controlled by the same fault. Eastwards there is a gradual rise in the ground level towards the low hills of the Sabarei area at the base of the Hafari massif. Numerous intersecting faults control the landscape. Originally the eastern edge of the median graben was defined by the ma-

jor fault extending south into Jibisa. However, volcanism related to Jibisa has masked this fault and modified the eastern side of the graben (see geological map).

The southern elevated plain is about 200 m higher than the northern depression; a dissected and eroded scarp related to the Buluk Fault separates the two areas. Residual mesas are present on either side of the Sabarei–Ileret track to the east of Buluk. Numerous (see map cross-section) grabens and horsts in the lavas are controlled by NNE-trending faults. The grabens have pan deposits on their flat bases flanked by scree from the adjacent fault scarps. The horst blocks generally remain horizontal or may be slightly tilted to east or west. The plain has a regional southerly slope. Differences in altitude between adjacent horsts and grabens are generally of the order of several tens of metres. However, two major fault scarps marking the eastern and western margins of the Rift are each about 150 m high.

## 2.5 JIBISA–HAFARI HIGHLANDS

Jibisa rises abruptly about 676 m above its own colluvial spread to a height of about 1560 m above sea level (Plate 3). The mountain consists of a series of arcuate ridges and valleys controlled by the concentric structure of its bedrock (the Jibisa Ring Complex). The four principal lithological components produce distinctive topographic features. The main central ridge is underlain by a coarse aegirine-syenite and sodic granite with good rock outcrop including shear cliff faces to the south and east. A porphyritic microsyenite underlies lower, more rounded ridges, also with good rock outcrop and only a thin scrub cover. The outer slopes and ridges of the main mountain are thickly vegetated and are covered by abundant scree of the underlying mafic rocks. The outermost southern ridge, which is separated from the main mountain by a narrow valley, and a detached chain of hills striking to the SSW, are largely covered by basalt lag but show good summit outcrop. Discordant valleys, mainly on the northern slopes are fault controlled.

The Hafari massif is a most imposing topographic feature overlooking Jibisa. It is underlain by horizontally stratified Miocene lavas which form a series of terraces on scarp slopes, as well as the major summit plateaus. The whole massif is about 25 km long in a ESE direction with the Ethiopian border running along its summit. In Kenya the massif has a maximum altitude of about 1692 m, approximately 800 m above the adjacent lowland. Faults are traced by minor valleys and to the north a major fault scarp separates this massif from the main development of Chew Bahir (Davidson, 1983). The uppermost plateau is protected by a resistant felsic volcanic flow unit.

## 2.6 BULAL PLATEAU

A monotonous plateau surface separates the central Mio-Pliocene Rifts from the Quaternary Dukana volcanic field in the east. The relatively featureless plateau surface at c.1000 m elevation reflects underlying extensive subhorizontal lavas. There is a regional, gentle southerly plateau slope away from the Ethiopian Highlands towards the Chalbi Basin. In the vicinity of Jibisa–Hafari the plateau has a relatively thick colluvial cover washed from the upstanding hills. A similar superficial mantle surrounds the Tatesa Hills to the south-east. Elsewhere the plateau has a thin soil cover with extensive lag gravels. Bollo Hill provides a rounded, positive topographic landmark in the extreme south-centre next to subdued mounds of eroded lavas. Three major fracture-controlled river valleys (Plate 4) dissect the plateau; these are discussed in Section 2.9.

## 2.7 TATESA HILLS

The Tatesa Hills are a NNE-trending chain of strongly weathered, rounded hills rising to a maximum height of about 300 m above the Bulal Plateau. Erosion is controlled by SSE-trending fractures producing a series of subparallel narrow valleys separated by elongate ridges. Finely stratified, horizontal lavas underlie the hills and give rise to a distinctive striped or layer-cake appearance (Plate 5). Unlike the contemporaneous Hafari massif the Tatesa lavas are highly susceptible to erosion.

## 2.8 DUKANA VOLCANIC PROVINCE

East of the Laga Wata the countryside changes; the monotonous Bulal lava plateau is punctuated by cinder cones and *maars* with peripheral low hills underlain by derived lavas and pyroclastics. The largest cone, Dibandiba, and the two largest *maars*, Gof Wargan and Dabacha, have mean elevations and depressions respectively of about 250 m relative to the plateau surface. The *maars* which are conspicuous features from the air, are concealed by their low peripheral rims of finely bedded pyroclastics. The pyroclastics are readily weathered and form extensive colluvial and alluvial spreads. In the extreme east is a subdued 'horst-and-graben' topography related to the N–S Miocene Ririba Rift, the southernmost part of the main Ethiopian Rift (Figure 13).

## 2.9 DRAINAGE

All drainage is ephemeral within four major drainage basins (Figure 3). The watershed delimiting the *Lake Turkana Basin* extends north to south parallel to, and about 40 km east of, the present lake shore. Rivers originate in the eastern (Asille and plateau lava) highlands and flow mostly due westwards down the regional gradient. This drainage is dendritic and controlled by the regional gradient between isolated ridges (Ileret, Koboi), mesas and cuestas. Local tectonic control may be indicated by markedly linear watercourses such as the upper part of the Il Eriet. Along the Buluk Gap river capture has taken place from the drainage into the Lake Stephanie Basin. Evidence for this capture is given by the U-bend in the headwater tributary of the Lake Turkana-directed drainage immediately west of Buluk Police Post.

River profiles change from narrow, steep-sided (ravine) headwater tributaries in the high ground to wide, sand-clogged channels with low, often poorly defined, banks on the western low ground. Locally the narrow headwaters debouch directly into well defined gravel fans generally adjacent to fault scarps.

A majority of the drainage channels terminate inland from the Lake Turkana shoreline. This is due to the present low lake level, a result of incipient subsidence in the Quaternary Rift and the present arid climate of northern Kenya and adjacent Ethiopia. The sporadic water flow in the drainage channels takes the form of flash flooding which rarely utilises the whole channel width on the eastern lowland. Several well defined watercourses

originate in the central high parts of North Island and descend abruptly into Lake Turkana. These cut into the bedded pyroclastics and are deflected by the blocky lava flows.

The *Chew Bahir* (*Lake Stephanie*) *Basin* is characterised by short, linear watercourses emanating off the surrounding higher ground. Unlike Lake Turkana which has permanent, albeit shallow (less than 125 m) water, Chew Bahir is ephemeral forming a mud flat within Kenya. Individual drainage channels rapidly loose their identity on low ground at the margins of the mud flat. Most of the southern and western rivers form major gravel fans at the bases of (fault) scarps. These fans are clearly related to former much wetter climates.

The *Kinu Sogo Rift* is poorly drained. Individual grabens have their own south-directed drainage systems which consist of short, scarp-slope gullies into seasonal pans. These internal drainage networks have no outlet to the south of the map-area. The upstanding horst blocks lack any drainage network.

The eastern map-area is part of the major northern catchment area of the *Chalbi Basin*. The principal drainage channel is the Laga Balal which extends over 200 km from the Ethiopian Highlands into the Chalbi Basin along the western edge of the Huri Shield. The map-area straddles the central part of this major drainage where it forms an impressive gorge. The river is ephemeral with a well defined gravel channel; a relic of former perennial flows such as that recalled in the earlier part of this century by the resident Gabbra community. The Laga Wata and Laga Kurrsa/Jibisa are the major east and west bank tributaries, respectively, of the Laga Balal in the map-area. All these watercourses follow regional NNE to N fractures. The west bank tributary at El Yibo is a relict Miocene drainage cut into the older El Yibo lavas. The stagnant, subsurface water in its coarse alluvium is more saline than well water in the Laga Balal. The interfluvial plateau areas are badly drained: there is a noticeable absence of dendritic-pattern tributaries to the fracture-controlled watercourses.

The Dukana Volcanic Province has a limited internal drainage with the major *maars* collecting seasonal rainfall through direct run-off from their peripheral upstanding rims. The subdued Ririba Rift of the extreme east has south-directed drainage channels which are deflected by the Huri Shield into the Laga Balal. In general, the plateau lava terrain is very poorly drained. Much of the seasonal flow in the large watercourses emanates from Ethiopia. Direct rainfall on the plateau surface percolates through the well jointed surface lavas, feeding shallow aquifers.

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# 3 Stratigraphy

Table 1 outlines the stratigraphy, which is dominated by Cenozoic volcanic rocks associated with the major rifting of East Africa. Present in the Suregei–Asille district and the upper Bakate Valley are diverse strata of distinctive lithologies and wide lateral extent which have facilitated detailed stratigraphic subdivision (Watkins, 1983[3]). Late Oligocene to Miocene rocks unconformably overlying the metamorphosed basement, and characteristically affected by monoclinal folding and abundant faulting, are termed the Asille Group. Overlying the Asille Group, and separated from it by a second angular unconformity, are the much less faulted and almost flat-lying late Miocene to Pliocene basalt lavas of the Gombe Group. Sedimentary strata of the Koobi Fora basin generally overlie the

**Table 1** An outline of the stratigraphy

| Sedimentation | Metamorphism | Volcanism | Intrusion | Geochronology | Age |
|---|---|---|---|---|---|
| Superficial unconsolidatd material | | Trachytic lavas and pyroclastics of North Island trachytes | North Island diorite | | Recent |
| Koobi Fora Formation | | Koobi Fora Formation tuffs Basaltic lavas and pyroclastics of the Dukana Volcanic Province | | 3.6 – 1.28[b] 0.8[c] | Pliocene to Pleistocene |
| – – – – – – – – – – – – – – – – – – unconformity – – – – – – – – – – – – – – – – | | | | | |
| | | Basaltic cinder cones Plateau lavas of Gombe Group and Bulal lavas | | 2.25[c] 5.7 – 3.28[a,d] | Miocene to Pliocene |
| – – – – – – – – – – – – – – – – – – unconformity – – – – – – – – – – – – – – – – | | | | | |
| | | West Jibisa augite-basalt; stratified lavas of Tatesa Hills basalts | | 12.1[c] | Miocene |
| Sedimentary units within Asille Group, notably Il Yia Formation | | Felsic and basic volcanism of Asille Group, Sabarei volcanics and Hafari rhyolites | Jibisa Ring Complex, microsyenite dykes basalt dykes | 28 – 12[a,c,d] | Oligocene to Miocene |
| Red soil above El Yibo lavas – – – – – – – – – – – – – – – – – – unconformity – – – – – – – – – – – – – – – – | | | | | |
| | | El Yibo lavas | | | |
| – – – – – – – – – – – – – – – – – – unconformity – – – – – – – – – – – – – – – – | | | | | |
| Sedimentary rocks subsequently metamorphosed during the Mozambiquian orogeny and exposed west of Lake Stephanie on the Ethiopian border | | | | | Precambrian |

*Geochronology*

a Watkins, R. T., 1983
b Fitch, F. J. and others, 1978
c Teledyne Isotopes, 1985
d British Geological Survey, 1985

Gombe Group disconformably although there is a degree of interdigitation of the flood lavas with the lowermost lacustrine sediments (Findlater, 1976; Watkins, 1983).

In the eastern part of the map area subjected to only reconnaissance study, lavas laterally equivalent to the Gombe basalts are recognised, together with younger, Pleistocene–Holocene volcanics. Additional exposures are clearly contemporaneous with parts of the Asille Group. It is, however, impractical at the present time to make firm correlations between the majority of the exposed rocks and Asille Group strata existing to the west. For this reason, the exposures of Oligocene and Miocene rocks in the map-area are dealt with primarily on the basis of geographical distribution in the following stratigraphic description.

## 3.1 METAMORPHIC ROCKS (mg)

Davidson (1983) mapped a faulted N-trending block of metasedimentary gneisses along the western side of Chew Bahir in Ethiopia, which he traced southwards as far as the Kenyan border. On satellite imagery this metamorphic basement is clearly distinguished by texture and colour from adjacent volcanic rocks and superficial sediments, and can be extended into the Garba Merille area.

The metamorphic rocks form an incised rocky upland unconformably capped by eroded basalt plateaus of the Nabwal Formation. The dominant lithology is a banded, fine-grained quartzofeldspathic gneiss. Abundant float of dolomitic marble in river channels originated in Ethiopia (Hamar Domain of Davidson, 1983). Traces of biotite, garnet, magnetite, apatite and hornblende occur in the gneisses, which are cut by coarser migmatitic and garnetiferous bands. The marble contains graphite, colourless mica, quartz, apatite, magnetite, biotite, hornblende, tremolite (?) and sulphides. Davidson (1983) records similar lithologies across the border, confirming textural continuity on aerial and satellite photographs. In Kenya the gneisses strike N to NNW with steep dips and N-plunging isoclinal folds.

No other exposures of pre-Cenozoic 'basement' are present. However, the pyroclastic beds surrounding *maars* of the Dukana Volcanic Province contain rounded xenoliths of banded mafic and felsic gneisses. They are medium-grained (0.7 to 2 mm average grain sizes), equigranular rocks. The mafic gneisses are dominated by brown hornblende prisms with subsidiary plagioclase and quartz. The felsic gneisses have quartz and oligoclase bands separated by brown biotite foliae and bands rich in biotite, hornblende, opaques and apatite as well as quartz and feldspar.

The metamorphic rocks are part of the Mozambique Orogenic Belt (Davidson, 1983) and their genesis is outlined in Chapters 6 and 7. These rocks are believed to underlie the whole map-area with the possible exception of Chew Bahir (where new crust may have formed at an early stage in the Cenozoic rifting).

## 3.2 ASILLE GROUP

### Nabwal Formation ($Tvb^N$)

The Nabwal Formation is the oldest division of the Asille Group present in the map area. It comprises a series of basalt lavas which is interrupted in its upper part by relatively thin palaeosols, clastic sediments, and minor pyroclastic units of felsic composition. Having a thickness of 536 m in its type section in the Nabwal Hills, east of the Asille lowland, it constitutes the thickest sequence of volcanic strata of common composition within the Group. Exposure occurs on the dip-slope of the low rounded Nabwal Hills where a lack of cliff exposures makes observation of the lower part of the formation, in particular, difficult. Where the base is observed lying unconformably upon the metamorphic basement, the underlying gneisses are little weathered and only a metre, or so, of mechanically broken material covers their surface. The upper boundary of the formation is sculpted into a mild relief but is readily traceable, normally being overlain by contrasting rock types (pantelleritic ignimbrites and fluvial conglomerates) of the Langaria Formation.

Practically the whole of the sequence is formed of fine-grained aphyric lavas of tholeiitic basalt composition. Some highly porphyritic basalt lavas which predate ignimbrites of the Langaria Formation are included in the uppermost part of the Nabwal Formation despite their clear kinship to subsequent flows of the Irile Member of the Bakate Formation. In the lower part of the formation, individual lava flows appear to be of considerable thickness, interflow weathering appears minimal, and it is evident that accumulation of the lavas was rapid as a result of major flood eruptions. Although flow thicknesses decrease in general up section, there is no significant break in the monotonous lava pile up to a unit of clastic sediments, some 80 m thick, 280 m above the base of the formation. At this level, the underlying lava is strongly weathered to form a clay-rich palaeosol which is succeeded over a sharp boundary by fluvial sediments. The latter consist of cross-bedded, leucocratic, arkosic sandstones and siltstones giving way laterally to indistinctly bedded siltstones and clays, and are interpreted as having formed within a river system as channels and flood plain deposits. Locally, at the top of the sediments there is preserved a 1 to 2-metre thick bed of felsic tuff. The bed contains lenses of well rounded pumice clasts and has clearly suffered reworking in water. The fine-grained vitric tuff is recrystallised towards its top where it is contaminated by terrigenous detritus and appears 'baked' by the overlying basalt lava. The tuff shows alteration to secondary chalcedony and carbonate. Where the tuff is not preserved, there are often bands of chert and massive crystalline limestone which may have formed at its expense. Around the general stratigraphic level of the sediments, there are some distinctive coarsely-porphyritic, feldsparphyric basalt lavas similar to those present in the Irile Member.

A second thin pyroclastic unit, 4 to 5 metres in maximum thickness, is recorded approximately 410 m from the base of the type-section, preserved in part within depressions cut in the underlying lava flow. This deposit is extremely rich in phenocrysts of quartz and oligoclase feldspar, the latter highly fractured and opened along cleavages, a possible indication of violent eruption. Abundant well-rounded basalt and basement quartz pebbles up to 2 to 3 cm in diameter are incorporated into the basal part of the unit where there are also local abundances of fossil wood fragments up to 20 cm in diameter. Ironstone nodules are also a feature of the basal half of the deposit. The generally massive nature of the unit is consistent with it being a product of ash-flow eruption, the pebbles and fossil wood having been swept up from the floors of river valleys in which the pyroclastic unit is now preserved. Some crude bedding seen near the base of the unit and appearing to mantle the underlying topography probably results from processes, such as shearing, which are a feature of ash-flows.

Weathering is increasingly evident between lava flows in the upper part of the formation with clay palaeosols and carbonate-cemented breccias developing. A particularly thick unit of clay-rich siltstone occurs close to the top of the formation upon a deeply weathered vesicular lava. This unit, containing abundant veins and nodules of secondary minerals, is laterally extensive and together with an overlying petrographically distinctive lava, can be readily traced (Watkins, 1983).

### Langaria Formation ($Tmvf^{L}$)

The Langaria Formation comprises a major sequence of pantelleritic ignimbrites (Plate 6) and associated pyroclastic deposits, along with subsidiary amounts of terrigenous sediments and, in one locality, a minor basalt lava flow. The formation everywhere overlies basalt lavas of the Nabwal Formation and is succeeded by basalt lavas of the Irile Member of the Bakate Formation. It is impressively exposed in the type-locality of Langaria cliff on the Il Jimma River (GR 28018 99121) immediately upstream of the confluence with the Il Eriet and adjacent to the Ethiopian border. The type-section records 35.6 m of pantelleritic strata present here in a faulted inlier. A maximum thickness of 48.5 m is recorded further east in Gumun Burrka, a strike orientated chain of low cuestas capped by a highly resistant ignimbrite from the top of the formation. The basal boundary, not exposed at Langaria, is defined from outcrop in the Gumun Burrka Hills (GR 29306 98962). More extensive outcrop occurs further to the east in a north–south zone delimited by faults. The thickness is highly variable with abrupt changes and local elimination of lower units indicating deposition largely within river valleys. Fluvial sediments are interbedded with reworked pyroclastic deposits in lowermost units in the eastern exposures. Steeply inclined, large amplitude cross-bedding and the presence of coarse conglomerates indicate a high energy fluvial depositional environment. The conglomerates contain well-rounded clasts up to 8 cm in diameter and considerably larger dispersed blocks. Abundant rhyolite lava clasts and some of flow-banded obsidian imply erosion of the source volcano of the Langaria pyroclasts or of an older felsic volcanic centre.

The formation is dominated by the products of ash-flow eruption, and includes three major, strongly welded ignimbrites. The lowermost, a blue-grey massive rock, is present only at Langaria, whilst the other two outcrop widely. The lower of these is a dense and glassy, strongly eutaxitic textured unit, approximately 1 m thick at the type-locality, but elsewhere commonly of greater thickness, whence it may be less strongly welded. The uppermost ignimbrite has a highly distinctive pink colour and is the most extensively preserved, probably having erupted after former relief was largely infilled. Localised autobrecciation at the base of this normally massive, resistant rock may indicate its disruption where lain upon wet sediments. Other pyroclastic units, including less strongly welded ash-flows and contemporaneous air-fall deposits, as well as fluvial sediments enriched in pyroclastic debris, are found interbedded with the welded ignimbrites. Fining-upward sequences of sediments at the Langaria cliffs are typical products of fluvial deposition containing lenses of well rounded pumice gravel and thin beds of silt- and claystone at the top.

A characteristic feature of the Langaria Formation is an abundance of secondary iron oxides, forming a preponderance of limonitic nodules and veining, and commonly colouring the rocks red or brown. Ironstone nodules are particularly abundant in the permeable sillars and locally coalesce to form ironstone bands and ultimately ferricrete several metres in thickness. Nodular ferricrete beds are particularly well developed at the top of the formation where the secondary material overlies the massive pink ignimbrite, the two units varying in relative thickness and separated by an irregular boundary. Since secondary nodules are rare within the massive ignimbrite, it is unlikely that the ferricrete has replaced such a rock. More probably, the overlying material was more permeable and may have been a highly pumaceous, little welded deposit formed as a continuation of the eruption responsible for the welded ignimbrite. Elongate, somewhat flattened nodules may be replacing similarly shaped pumic fragments deposited from a 'froth flow'. The metallic bluish grey manganese oxide, cryptomelane, occurs locally within the more permeable pyroclastic rocks.

### Bakate Formation ($Tmvb^{BK}$)

The Bakate Formation comprises three Members with a combined thickness of almost 300 m. The lower, Irile Member and the upper, Il Jimma Member both consist essentially of basaltic lava flows, although these are of different petrography in the two divisions. The intervening Buluk Member is sedimentary, although in places it does contain minor basalt lava flows. Surface hydrothermal deposits, together with secondary calcretes and silcretes form frequent intercalations in the formation, with travertines and sinter beds locally becoming a significant constituent of the Il Jimma Member.

#### Irile Member

The Irile Member, a sequence of lava flows totalling 199 m in type-section, is named from the Il Irile drainage which traverses the main strike outcrop belt on the north-eastern side of the Suregei Plateau. The type-section is recorded over a long surface traverse from (GR 30020 97872) and has a thickness of approximately 200 m. The lavas lie upon Langaria rocks or are locally unconformable upon Nabwal lavas, whence 1 to 2 m of weathering detritus generally marks the boundary. The major part of the member is exposed only in the eastern part of the Surgei–Asille district, although the uppermost lavas crop out extensively across the Asille Plain.

The member comprises multiple lava flows with significant palaeosol intercalations and some minor clastic sedimentary units. A distinctive feature of many of the lavas is their coarsely porphyritic, feldsparphyric texture, in which plagioclase phenocrysts are up to 15 mm in size. Similar lavas occur infrequently in the upper portion of the Nabwal Formation but otherwise appear to be restricted to the Irile Member. However, Irile lavas are not exclusively of this type and other porphyritic and aphyric types are present.

The palaeosols are reddish brown silty-claystones, up to 5 m thick, and seemingly indicate prolonged surface weathering of individual lava flows. Deep weathering was doubtless facilitated by brecciated and vesicular upper parts of the flows, and where amygdaloidal lavas have broken down, large rhombs of calcite and sometimes baryte may be found in the palaeosol. The basalts, which are normally traversed by abundant secondary veining, are rarely free of some degree of low-T alteration. The significant intervals between lava eruptions has been confirmed by K-Ar dating of the bracketing Langaria Formation and Buluk Member pantellerites, which indicates a

time span for the lava succession of around 10 Ma (McDougall and Watkins, in press).

Sequences of aphyric lavas occur intermittently at the base and extensively at the top of the member. The uppermost lavas, recognised by their aphyric texture with dispersed plagioclase phenocrysts, are separated from the underlying flows by a particularly well-developed palaeosol. Clastic sediments are exposed in pockets between the lava flows, some having yielded vertebrate fossils. Thin carbonate and siliceous intercalations are a feature of the upper part of the porphyritic lava sequence. At least some of these deposits may have formed by surface hydrothermal mineral precipitation, although the less substantial outcrops more probably result from groundwater alteration within former soils.

### Buluk Member

The Buluk Member has received considerable attention on account of important well-preserved vertebrate fossils, including hominoids, recovered from fluvial sedimentary rocks in its lower half (Harris and Watkins, 1974; Watkins, 1983; Leakey and Walker, 1985). Pantelleritic tuffs forming the upper part of the Member have been dated as late Early Miocene by the K-Ar techniques (Fitch and others, 1985; McDougall and Watkins, 1985), (see Chapter 5). Zircon crystals present within the tuffs have proved highly satisfactory for fission track dating and the Buluk Member tuffs provide an important potential fission track age standard (Hurford and Watkins, 1987). An account of the geology of the Member has recently been presented by Watkins (in press).

The Buluk Member everywhere overlies lavas of the Irile Member. The type section is recorded from a bluff at the south-eastern foot of the Suregei Plateau (Plate 7) where the lower terrigenous clastic sediments have a thickness of 30 m and the overlying pantelleritic tuffs are 20 m thick. Overall thickness of the Member is highly variable. Outside the type-area of the northern side of the upper Bakate Valley, the lower clastic division is generally absent; in its place is merely a palaeosol which is assigned to the Irile Member. Some vertebrate fossils have been recovered from thin sediments at (GR 29835 97665) and non-fossiliferous conglomerates and sandstones are exposed in the Il Jimma gorge (GR 28124 99007). The thickness of the tuff sequence diminishes in a general way north-westwards, the pyroclastic unit being very thin or absent on the westernmost portion of the Asille Plain.

The lower terrigenous sedimentary rocks are interpreted as having formed within a major river valley. Thick, structureless reddish brown claystones representing flood plain alluvium constitutes two-thirds of the total thickness of terrigenous sediments in the type-section. The remaining sediments include cross-stratified channel sandstones and conglomerates. Maximum clast size is approximately 3 cm in the type-area but to the north (GR 29400 98328) unusually coarse deposits (up to 25 cm max. clast dimension) occur. Large trough sections, approximately 15 to 20 m in width and 4 m deep are exposed and there is a general fining upwards within units. Finer-grained sandstones and siltstones exhibit small scale cross-bedding and may show root casts, bioturbation and the development of small algal structures, such as have been interpreted elsewhere as forming in over-bank ponds (B. E. Bowen, *pers. comm.*; Watkins, in press). The sediments associated with the channels contain abundant silicified wood as well as vertebrate fossils. A faunal list has been presented by Leakey and Walker, 1985. The fossils represent a diverse fauna and are frequently well preserved. They are found locally in dense concentrations, and at one site form a fossil 'bone bed' (R. E. Leakey, *pers. comm.*; Watkins, in press).

The base of the overlying tuff sequence is sharp and there is no interfingering of pyroclastic and detrital sediment. The tuffs nevertheless appear to have experienced fluvial reworking either in discrete rivers and streams or by sheet wash. Evidence for the reworking includes ubiquitous cross-lamination with occasionally steeply inclined cross-sets; an abundance of small cut-and-fill structures; and the presence of well-rounded pumice fragments that have been sorted into discrete bands and lenses. It is apparent that the local drainage basin rapidly became choked with pantelleritic ash, preventing the contamination of the tuff beds by 'normal' detrital materials. Only at the very top of the Member do terrigenous sediments become incorporated in the tuffs. On the Asille Plain, the tuffs are seen to infill small valleys whence they may sometimes show steep dip on the underlying basalts. However, features of primary air-fall bedding generally are absent and comprehensive reworking is believed to have taken place. The far greater and uninterrupted spread of the tuff sequence, when compared with the terrigenous sediments of the lower part of the Member, might suggest that the reworking took place by sheet-wash over a broad area of subdued relief and was not restricted to a discrete river valley.

The tuff sequence recorded in the type section can be considered the result of a single volcanic episode. The uppermost bed in most localities is a distinctively salmon-pink coloured, fine sandstone which may have been partially 'baked' by the overlying lava. However, this resistant marker bed, is locally overlain by a further thin sequence of reworked tuffs suggesting that subaerial exposure, rather than 'baking', was responsible for its increased induration. The overlying tuffs probably represent additional reworking and redeposition of the original pyroclastic deposits, although additional closely contemporaneous pyroclastic volcanism cannot be ruled out.

The exposure of the Buluk Member in the lower Il Jimma Valley (GR 28125 99010) comprises a maximum of 6 m of siltstones and sandstones with dispersed pebbles, fining upwards into siltstones and claystones which are capped by a maximum of 3 m of silt-contaminated tuffs.

### Il Jimma Member

The upper member of the Bakate Formation has extensive and continuous outcrop over a large part of the western Asille Plain and in a strike belt extending south-south-eastwards along the base of the eastern wall of the Suregei Plateau and into the upper Bakate Valley. Where leucocratic tuffs of the Buluk Member are absent, the base of the Il Jimma Member is normally indicated by a significant palaeosol developed upon the upper Irile lavas. In places, the lowest Il Jimma lava flows are overlain by further pantelleritic tuffs of the Buluk Member. The boundary between the two members, however, is generally flat, there having been little downcutting into the relatively soft tuffs prior to the basaltic eruptions.

The type locality is a gorge on the Il Jimma River, immediately south-east of Langaria cliff, where the type-section on the wall of the gorge (GR 28087 99019) records the maximum measured thickness of 47 m. Here the member comprises strongly columnar-jointed basalt lavas that are exclusively fine-grained and aphyric. The flows are thick and massive with well formed 'colonades', and

show little vesiculation. Red-brown silt and clay palaeosols lie upon the upper flows. Elsewhere, the member is generally much thinner and includes glassy, scoriaceous lavas, the basalts remaining exclusively aphyric. Flow structures, including some small lava tubes, may be found in the upper part of the youngest flow, where 'spindle' and other lava bombs occur in the clinkery lava. Preservation of these structures indicates a lack of erosion prior to the laying down of the overlying Gum Dura rocks, supporting the conclusion of K-Ar dating studies that the pantelleritic and basaltic volcanism of the Buluk and Il Jimma Members and the Gum Dura Formation were closely spaced in time (McDougall and Watkins, 1985; in press). The scoriaceous lava and projectiles are locally cemented by calcite, zeolite, chalcedony and other low-T materials. Some oxidation and breakdown of the glassy lava has produced a bluish purple weathering deposit which has proved a useful stratigraphic marker of the Bakate/Gum Dura Formation boundary.

At one locality on the Ethiopian border, close to Langaria cliff, highly coloured siliceous sinters mark the former site of surface hot-springs. The deposits include delicate fossil plant remains. Rather similar rocks occur at numerous other localities and the formation of the Il Jimma lavas appears to have been accompanied by widespread, localised surface hydrothermal activity. In each case the underlying Il Jimma lavas are themselves deeply altered with veining and infilling of cavities by low-T precipitates. In the eastern exposures of the Il Jimma Member, dolomite and siliceous beds, each up to 6 m thick, are frequently intercalated with the basalt lavas. The dolomites are normally massive, structureless rocks of yellow to orange colour, whilst the siliceous units frequently exhibit complex banding. The basal contacts of these units is normally sharp and an origin of the carbonate/silica beds through *in situ* weathering processes appears unlikely. A hydrothermal origin is again suspected.

**Gum Dura Formation** ($Tmvf^{GD}$)

The Gum Dura Formation is the thickest series of pantelleritic rocks within the Asille Group of the map area, comprising 68 m of pyroclastic deposits. It is named from the Gum Dura Hills, a low undulating plateau capped by the highly resistant, welded ignimbrites of the formation's Middle Member. Other exposures occur on the central part of the Asille Plain, around the eastern wall of the Suregei Plateau, and in the eastern part of the upper Bakate Valley, the resistant ignimbrites giving rise to numerous small mesas and cuestas.

The pyroclastic pantellerite sequence is divided into three parts. In the absence of suitable local geographical nomenclature, these divisions are formally defined, but as yet informally named the Lower, Middle and Upper Members. The Upper Member includes some basalt lava flows at its base. Additionally, there are substantial units of carbonate/silica rocks which are termed the Manadure Member. The type section of the formation is a composite of those of the individual Members. The formation everywhere overlies the Il Jimma Member of the Bakate Formation. The Lower and Middle Members are absent from the exposures in the extreme north-west of the Asille plain.

Lower Member

The Lower Member is 24.8 m thick in its type section in the Il Dura Valley (GR 29109 98841) and is constructed of numerous, generally thin, pyroclastic beds. Thickness varies considerably, however, between 0 and a maximum measurement of 48 m, with greater accumulation within pre-existing valleys. A number of characteristic rock types are present. At the base are weakly indurated, blue-grey, laminated and cross-laminated vitric tuffs. Grain-size varies from very fine ash to lapilli. Rounded pumice gravels may be coarser-grained (5 to 10 mm). The basal bed locally has a purple coloration owing to incorporation of material weathered from the underlying basalt, but subsequent beds are notably free of extraneous sediment. The presence of cross-stratification with steep foresets and abundant cut-and-fill channelling; concentration of very well rounded pumice fragments into bands and lenses; and the fact that the beds infill rather than mantle irregularities in the underlying topography; all suggest comprehensive reworking of original air-fall material (cf. Buluk tuffs). Reworking would have taken place shortly after the pantelleritic pyroclastic eruptions and, as in the case of the Buluk Member, must have entirely blanketed the drainages to account for the absence of contamination by terrigenous sediment.

The generally fine-grained tuffs are overlain by, and thence interbedded with, coarser deposits of different character. These are normally rich in angular and subangular lithic and juvenile glass clasts with rarer pieces of pumice, lying in a matrix composed of crystals and finely fragmented glass, the latter decaying to a yellowish clay. The units fine upwards from a maximum clast size of 10 to 15 cm at the base. The lithic content includes pieces of the underlying Il Jimma basalt lavas, but more frequent are pieces of a flow-banded rhyolite not recorded from the underlying succession. The pumice component, in contrast to that from the fine tuffs, is angular. The juvenile glass component includes rare perfectly isotopic obsidian, but is generally flow-banded, crystallite-bearing glass showing partial spherulitic devitrification. A perlitic fracture within the glass accounts for its break up into angular fragments, yet some large blocks persist indicating a proximal volcanic source.

Each graded unit represents a single pulse of deposition with the pronounced cross-bedding ruling out a mass-flow or mud-flow origin. Silicified logs present in places in the bases of the units suggest an ash-flow origin and a hot emplacement may be further indicated by common association of minor fumarolic/hydrothermal deposits in the form of lenses of orange-coloured sinter, less than 1 m thick. Alternatively, the breccias may represent fluvial flood deposits on the flank of an active volcano. In either case, the abundant clasts of glass and flow-banded rhyolite almost certainly reflect the erosion of contemporaneous lavas. The repetition of units and their interbedding with fine tuffs might indicate the rapid destruction of a series of lava domes, or merely cyclical changes in the fluvial regime on the flanks of the volcano.

A third common component of the Lower Member is an ignimbrite of light-grey colour. It is poorly welded, being composed of quite equidimensional pieces of pumice usually less than 10 mm in diameter. These vary in colour from dark grey to white, giving the rock a mottled appearance. This ignimbrite which contains xenoliths up to 50 cm in size, is well exposed in the Gum Dura hills, e.g. (GR 28347 98576). Other ignimbrites are exposed locally within the Lower Member such as at (GR 29879 97124), south-east of the Suregei Plateau, where a weakly welded ash-flow is up to 6 m thick.

Fine-grained tuffs occurring towards the top of the member are less distinctly laminated than the lower tuffs



**Plate 2** Gullying on the northern side of the main peak of North Island. Stream erosion has been facilitated by solfatara alteration of the dioritic bedrock.



**Plate 3** Jibisa Mountain viewed from Hafari.

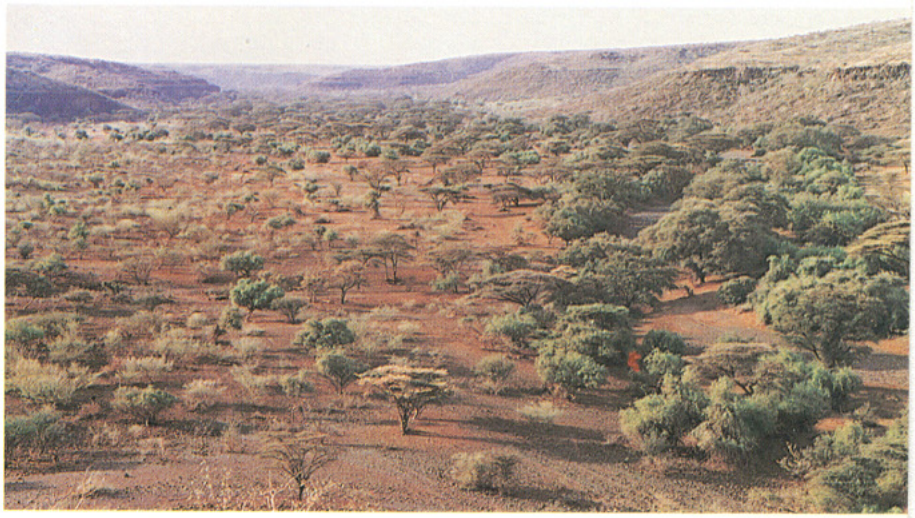


**Plate 4** The Laga Balal gorge at El Yibo. Note the well defined planar contact between the basal El Yibo lavas and Bulal lavas about half way up the gorge walls.



**Plate 5** Fine stratification of the Tatesa Hills.



**Plate 6** The Langaria Formation at the type locality of Langaria Cliff on the Il Jimma River (GR 28005 99124).

Four ignimbrites are seen forming the river bed, the small cliff (behind man), the dark unit, and a pink sheet near top of cliff which has here been leached to a cream colour. The dark ignimbrite and this uppermost ignimbrite are preserved extensively in the Suregei-Asille district and form important stratigraphic markers. Deposits between ignimbrites are co-eval air-fall tuffs and water reworked material. Also evident in the picture is a well developed nodular ferricrete, brown in colour, which overlies and partly replaces the upper pink ignimbrite.



**Plate 7** Middle part of the Bakate Formation at the type locality on the south-eastern flanks of the Suregei Plateau (GR 29945 96952).

In the foreground are exposed fluvial channel deposits from which hominoid fossils have been recovered. Overlying these and completing the Buluk Member are largely structureless, reddish-brown, flood plain claystones and bluish-grey, fluvial reworked pantelleritic tuffs. The hillock is capped by resistant basaltic lavas of the Il Jimma Member.



**Plate 8** Upper member of the Gum Dura Formation, north of Suregei Plateau (GR 29075 98452).

The lowermost of the ignimbrite sheets is seen to thicken towards the axis of a shallow valley (right) cut into the basalt lava flow.



**Plate 9** Hafari massif with thick porphyritic flows of the Hafari rhyolites overlying the well bedded Sabarei volcanics.

and generally show a degree of secondary alteration and contamination by non-juvenile sediment. This suggests a lull in pyroclastic activity prior to the major eruptions responsible for the ignimbrites of the Middle Member. In a few localities, very coarse obsidian breccias with blocks up to 50 cm or more across are seen at the top of the Member. One such occurrence (GR 29194 98552) has obsidian boulders up to 2 m in diameter and is overlain by a brecciated flow-banded, sanidine-rich rhyolite. These deposits again indicate the proximity of a volcanic vent and the generation of rhyolite lava domes.

### Middle Member

The high resistance to erosion and thus wide preservation of strongly welded ignimbrites of the Middle Member, coupled with their distinctive appearances, makes them an invaluable stratigraphic marker within the Asille Group. They are of additional importance having provided K-Ar and $^{40}Ar/^{39}Ar$ isochron ages (Fitch and others, 1975, 1985; McDougall and Watkins, 1985). At the type section of the Middle Member on the Gum Dura Plateau (GR 28336 98519), the two ignimbrites have a combined thickness of 8.9 m. Elsewhere this thickness may be considerably reduced and one, or other, of the flows may be missing. Probably, the ignimbrite sheets have complex geometry with the younger flow lying in part within valleys cut into the older sheet. At the type-locality the boundary between the sheets is marked by up to 1.5 cm of crystal-rich pyroclastic material or fine chert. Up to 10 cm of poorly consolidated ash or soil is locally present between the ignimbrites in the upper Bakate Valley.

The lower ignimbrite at the type locality is mid-grey to very dark grey in colour, being darker where more strongly welded. It has a eutaxitic texture with 'fiamme' of fresh 'black' glass up to 30 cm in length. Where the base of the flow can be observed, it appears generally flat-lying over the underlying tuffs which had infilled previous topography, accounting for the wide lateral spread of the quite thin ignimbrite sheet. Fossil wood and extraneous lithic fragments are only locally abundant within the ignimbrite. The degree of welding varies, often being reduced in the lower part or where the sheet is thickened.

The second ignimbrite is readily distinguished by its greenish blue colour. It is similarly a crystal-rich, strongly welded rock. In this case, the large glass 'fiamme' have frequently decayed to secondary clay and may have been eroded to leave elongate voids subparallel to flow. The fine matrix of the rock has devitrified and there is ample evidence of secondary silicification with quartz lining the cavities and forming abundant thin veins. In some localities, pyroclastic textures have been lost and the rock appears as a fine-grained, massive rock. This is the case in the western part of the Gum Dura Hills where there appear to be two flow-units, the discontinuity having been exploited by a basaltic sill.

### Manadure Member

Like the Bakate Formation, the Gum Dura Formation has associated with it considerable amounts of carbonate/silica rock. These form extensive beds immediately below and above the Middle Member, being particularly thickly developed on the central part of the Asille Plain. The numerous carbonate/silica units at this stratigraphical level are collectively termed the Manadure Member after the Il Manadure which has its confluence with the Il Jimma alongside some prominent cliff exposures of the limestones and cherts.

The lower unit recorded in the type-section on this cliff (GR 28524 99155) underlies an ignimbrite of the Middle Member. It is a dense, white, extremely fine-grained dolomite rock which contains dispersed small (< 1 mm) crystals of sanidine and quartz. Mostly it is massive and structureless, although it is broken by subconchoidal, 'slickensided' fractures. Towards its base it may be somewhat cavernous. Underlying siltstones are enriched in carbonate, but there is a distinct boundary between these and the Manadure unit. Nodular textures or the incorporation of lithic debris within the base of the limestone, such as commonly exhibited by calcretes, are not found.

Elsewhere, the lower unit may be partly siliceous showing horizontal banding/bedding, giving way laterally to grey siltstones with thin, sometimes powdery, carbonate beds. In many places, the limestones are thinly interbedded with pantelleritic pyroclastic beds of the Lower Member.

The upper units, post-dating the Middle Member ignimbrites, are usually more siliceous and at the type locality dolomite is restricted to minor bands and lenses. The rock varies from translucent pale blue to orange-brown chalcedony, may be massive in part, but is normally strongly banded subparallel to the unit boundaries. Discrete bedding planes divide units which are up to 3.5 m thick. These upper Manadure beds have their greatest development on the central Asille Plain where they are interbedded with pantelleritic ignimbrites of the Upper Member.

Pedogenic processes, producing calcrete and silcrete, are a clear possibility in the genesis of carbonate/silica beds in tropical environments. However, numerous features of the rocks of the Manadure Member, such as the absence of decaying lithic fragments in the rocks, the commonly sharp unit boundaries and interbedding of strata showing no calcification or silicification, and the uniformity of texture and composition, mitigate against such an origin for several of the major units. Those rocks within the vicinity of the type locality, in particular, are very similar to rocks preserved as cappings to the Anyakolib vents which are most readily explained as the products of surface hydrothermal precipitation. It is considered that significant parts of the Manadure Member share a similar origin. Remaining carbonate and silica deposits probably formed as products of groundwater mineralisation in areas peripheral to active hot-springs. Clearly there is gradation between hydrothermal conditions and those of cooler groundwater alteration, making the recognition of the different precipitates difficult. (A detailed discussion on the character and genesis of the various carbonate and silica deposits is given in Watkins, 1983.)

### Upper Member (Plate 8)

The Upper Member is defined by a composite type section recorded on the northern flank of the Suregei Plateau (GR 29082 98437; 28997 98449) having a thickness of 34.5 m. It comprises a further sequence of pantelleritic pyroclastic rocks but includes some basalt lavas at its base. The lower boundary is easily recognised, overlying Middle Member ignimbrites or units of the Manadure Member. The upper boundary, however, is frequently gradational as the pyroclasts have been reworked by streams during Il Yia times. To simplify the stratigraphy, reworked tuffaceous units wholly enveloped by Il Yia sediments have been assigned to the sedimentary formation; unreworked ignimbrites similarly placed, however,

belong to the Upper Member of the Gum Dura Formation which consequently interdigitates to a degree with the overlying formation.

The basaltic unit present at the base of the Upper Member is invariably a single lava flow. In the vicinity of the type-section, the lava flow is 13.3 m thick, with a very fine-grained to glassy texture and a distinctive reddish weathering. It has a platy, horizontal fracture towards its base and is amygdaloidal in its upper part. This aphyric lava is exposed in numerous outcrops north of the Suregei Plateau, but is unlikely to be the same flow as thin, finely-porphyritic lavas intermittently exposed within the upper Bakate Valley. Other basalt lavas cropping out in the area of Gum Dura are porphyritic, with some predating the Middle Member.

At the type-section, the lowermost pantelleritic ignimbrite of the Upper Member is seen to thicken as it infills a valley cut into the basalt lava flow, thin silt and claystone being preserved beneath it. The light-grey ignimbrite is crystal-rich and displays varying degrees of welding. However, even when containing flattened pumice 'fiamme' up to 30 cm in length, it remains a relatively soft rock. It is the most voluminous ignimbrite of the Suregei–Asille district, having overwhelmed existing valleys before spreading out laterally over the adjacent terrain. In the western part of the Asille Plain, where the older members of the Gum Dura Formation are absent or reduced to some thin tuffs and tuffaceous sediments, the ignimbrite lies disconformably upon lavas of the Il Jimma Member, Bakate Formation. There is evidence of the destructive flow of the ignimbrite with fossil logs and much sediment having been incorporated into the pyroclastic unit during its flow along river valleys. Impressive silicified logs, a metre or more in diameter, and extremely dense concentrations of well round cobbles occur in the basal parts of the flow, whilst finer lithic and fossil wood debris persists throughout the vertical and lateral extent of the flow.

A second ignimbrite of darker grey coloration is, in contrast, almost devoid of extraneous material, indicating the carpeting of its range by the previous flow and the swift succession of this second unit. Often there is an almost imperceptible lithological break with the underlying flow, although in the type-section 0.6 m of tuffaceous silt containing reworked pumice fragments is seen. The upper ignimbrite, which is thinner and less extensive than the first, contains few phenocrysts but has a large content of small, angular pieces of obsidian. Pumice commonly retains a dark grey colour, unlike that of cream to white colour in the lower ignimbrite.

East of the Suregei Plateau and in the Bakate Valley, the lower portion of the Upper Member is represented by two equally distinctive ignimbrites. The lower is a light green, crystal-rich rock, while the upper has fewer phenocrysts and is grey to cream in colour. A similar sequence is seen at the eastern end of Langaria cliff (GR 28040 99104) where the upper flow is manifestly the lower, light grey ignimbrite of the type-section. It is possible, therefore, that the upper of the two ignimbrites in the eastern exposures is a lateral extension of this major flow. However, numerous thin ash-flows are present in the Upper Member on the Asille Plain, and such a correlation at this time is highly tentative.

The lower ignimbrite in the Bakate Valley contains abundant lithic and fossil wood fragments in its basal part. Locally some large segments of fossil log, 4 m or more in length, are preserved within the flow. Aqueous solutions permeating the ignimbrite have given rise to silicified zones which are more resistant to erosion and stand up from the outcrop as isolated columns and wall-like structures. The silicification may have occurred soon after deposition of the pyroclastic unit, driven by heat retained from the eruption. Thin siliceous beds locally capping the ignimbrite may have a similar hydrothermal origin. The upstanding columns, which are polygonal in cross-section and up to 50 cm across, appear particularly coarse-grained and rich in angular lithic fragments. They may, consequently, represent sites of former gas-drilled pipes which provided preferential channels for water percolation. Similar silicified columns are developed to a lesser extent within the upper ignimbrite in the Bakate Valley.

The uppermost part of the Upper Member comprises a varying thickness of relatively thin ash-flows and fluvial reworked pyroclastic deposits. The latter are mostly fine-grained tuffs but are accompanied in places by coarser, cross-bedded units similar to those of the Lower Member, although having a lesser content of juvenile glass. Cobbles of Middle Member ignimbrites are found in the bases of thin reworked units and the whole sequence is marked by significant decomposition of the fragmental glass to secondary clays and the contamination of pantelleritic deposits by terrigenous sediment. Nodules of iron-stained carbonate and some thin secondary ironstone bands are present.

### Il Yia Formation ($Tms^{Y}$)

The Il Yia Formation is named from a tributary of the Il Dura which has excavated a broad valley in the loosely consolidated sediments on the northern side of the Suregei Plateau. The type-section of the major part of the formation is recorded in this area (from sections at GR 28770 98571; 28765 98507 and 28823 98433). Other outcrops occur west of the Plateau, where the type-section of the uppermost part of the formation is recorded on the north-western slope of Nakwele Hill (GR 28812 97252), and in the upper Bakate Valley. The composite type-section records a thickness of 125 m. Minor exposures of the formation exist close to Derati and Kubi Algi Mountains south of the map area (Watkins, 1987). Whilst formed exclusively of sedimentary rocks, the formation interdigitates in its lower part with pantelleritic ignimbrites of the Gum Dura Formation and basalt lavas of the Biliet Member of the Bakate Formation. Numerous units of carbonate/silica rock, originating from surface hydrothermal or pedogenic processes, are assigned in each instance to the most appropriate of the stratigraphic divisions.

The Il Yia Formation is characterised by coarse-grained fluvial deposits (see Plate 11). It commonly exhibits a gradational lower boundary, underlying pantelleritic pyroclastic rocks of the Gum Dura Formation having been reworked and incorporated into the basal beds. Individual units dominated by pantelleritic pyroclastic material recur at higher levels in the formation where they are thin and of only local extent. They generally reflect erosion and redeposition of older Gum Dura rocks, but some component from contemporary eruptions cannot be ruled out since isolated outcrops of ash-flow pantellerite (assigned to the Upper Member of the Gum Dura Formation) exist within the conglomerates.

The formation is built largely of substantial units of coarse cobble conglomerate. The clasts are well rounded, normally bladed in form and show a strong imbricate structure. Individual units appear poorly sorted with clasts ranging from pebbles to cobbles 20 to 25 cm in max-

imum diameter, with less frequent, much larger, blocks. They include a variety of rock types of which pantelleritic ignimbrites (Langaria and Gum Dura Formations) and basement granitic rocks are easily recognisable. Quartz derived from the metamorphic basement is prevalent. Basaltic clasts are generally of small size and are deeply altered. The abundant matrix is normally crystal-rich and of coarse sand to granule gravel grade. In the lowest parts of the formation it contains a major component of the underlying Gum Dura pyroclastic deposits which imparts a greenish colour. This contamination and the green tinge decrease upwards.

The sedimentary pile is built of large lenticular bodies of sediment, each showing a general fining upwards from conglomerate, through arkosic sandstones, to thin siltstones and claystones at the top, representing deposition from a waning flood. Frequently, clast supported conglomerates pass upwards into sandstones containing dispersed rounded cobbles and pebbles. The coarseness of the conglomerates masks all but rudimentary bedding. However, the finer-grained deposits display planar and cross stratification and abundant cut-and-fill channelling. Reddened oxidised siltstones occur extensively beneath interbedded Biliet lavas, seemingly indicating periods of reduced deposition and possible soil formation prior to the basaltic outpourings. The Il Yia sediments throughout are only poorly consolidated, and are easily removed by stream erosion. There is some cementation of bands and lenses by secondary ironstone, and there is wider development of small secondary nodules of limonite and calcite. North-east of Nakwele (GR 29159 97335) fine-grained sediments of the upper part of the formation exhibit mudcracks and possibly root casts and burrowing, as well as abundant secondary carbonate nodules. Fossil wood, including portions of tree trunks several metres in length and more than a metre in diameter, occur within the conglomerates. The fracture and separation of segments of log and the dispersal of large matching fossilised wood fragments, is evidence that silicification took place prior to incorporation in the sediments. The silicified wood originated in ignimbrites of the Gum Dura Formation, particularly the less strongly welded units of the Upper Member from which the highly resistant fossilised wood is readily eroded. The depositional environment of the conglomerates is not conducive to preservation of vertebrate fossils. Accordingly, thus far, only few, generally rounded, pieces of bone have been recovered from the formation.

A fluvial environment of greater energy than that producing the earlier Buluk Member sediments, is required for the transport and deposition of the Il Yia rocks. Given an apparent absence of significant conglomerates of older age that may have been reworked, it would appear from the abundance of basement clasts that the rivers had an important distal source. Composition of the conglomerates does not appear to alter after the eruption of successive Biliet lavas, showing that the basaltic flows were of comparatively local extent and did not bury the source regions of the clasts. The necessarily rapid erosion of source rocks implies a marked change in drainage profile at the beginning of Il Yia times. Production of the conglomerate units was very probably triggered by major faulting giving rise to fault scarps, the erosion of which is known to lead to rapid accumulation of coarse deposits over wide areas. Without detailed study of the geometry of the bed forms, it is not possible to say whether the sediments reflect the growth of large fans or the existence of a major braided river.

Carbonate/silica rocks assigned to the Il Yia Formation are mostly associated with fine-grained siltstones and claystones and form thin but often laterally persistent beds closely beneath the intercalated Biliet lavas. They include friable, powdery deposits but are usually denser rocks in which the carbonate is accompanied by chalcedony. Texture is highly variable from irregular banding to brecciated and nodular textures. Unlike several carbonate/silicate units placed within the Biliet Member of the Nakwele Formation, those of the Il Yia Formation normally incorporate amounts of silt and coarser lithic detritus, and are interpreted as duricrusts. In the upper part of the Formation, limestone beds are deformed into small conical structures and larger, elongate, sinuous 'anticlinal' features with steeply dipping limbs, which may be analogous to the 'caliche buckles' described by Goudie (1972).

### Nakwele Formation ($Tmvb^{NA}$)

The Nakwele Formation comprises the youngest strata of the Asille Group that are exposed within the map-area. The sequence of basalt lavas is 300 m thick in the area but may be significantly extended when correlation with units exposed in the southern margin of the Koobi Fora Basin is completed. It takes its name from Nakwele Hill (GR 28914 97189) which lies at the southern end of an extensive, 125 $km^2$ outcrop of the formation in the Gum Warata upland. It is subdivided into two Members: the Biliet Member which comprises those lavas in the lower part of the formation that are interbedded with Il Yia sediments; and the Nakoyen Member which makes up the succeeding major part of the formation. Throughout, the formation is constructed of lava flows of alkali olivine basalt composition which show varying textures from very fine-grained aphyric rocks to coarsely porphyritic types. A majority of flows are porphyritic with well-formed augite being the predominant phenocrysts. These may be accompanied by plagioclase feldspar and/or olivine. Some flows are extremely rich in olivine phenocrysts. At a locality in the Il Biliet Valley, there is a dense concentration of well-rounded ultramafic nodules exposed within a deeply altered lava flow. These are pyroxenite and serpentinite after dunite.

#### Biliet Member

The member is well exposed in the upper Bakate Valley and has scattered outcrops upon the central part of the Asille Plain. The type-section of the member is recorded on a promontary (GR 28519 97964) and thence on a prominent outcrop of limestone/chalcedony rock (GR 28777 97428) overlooking the head streams of the Il Biliet; it records a thickness of 92 m. The basalt flows typically are thicker and less deeply weathered than subsequent lavas in the main part of the Nakwele Formation. The lowermost flow, present only in the upper Bakate Valley, is a distinctive, extremely fine-grained, aphyric rock having a 'sooty' appearance and exhibiting a strong horizontal fracturing. Subsequent Biliet lavas include some finely porphyritic types which have an equally distinctive 'speckled' appearance. This feature, which is particularly marked on weathered surfaces, results from the coexistence of dark augite phenocrysts with plagioclase phenocrysts which have altered to white secondary minerals. Other Biliet flows may be either aphyric or porphyritic, with some especially coarsely porphyritic ones containing euhedral augite phenocrysts 1.5 to 2.5 cm in size together with abundant unaltered olivine.

The concurrence of basaltic lava eruptions and accum-

ulation of Il Yia fluvial sediments has given rise to abundant hydrothermal products in the form of amygdales and vug fillings and more extensive travertine and sinter deposits. Those carbonate/silica rocks with brecciated or nodular textures are interpreted as duricrusts. However, where decaying lithic debris is entirely absent from the limestones and cherts direct precipitation from surface hydrothermal waters is suspected. Such units locally may be of considerable thickness, as in the case of exposures at (GR 28567 97783) where 19 m is recorded. The wide range of structural and textural variations displayed by the carbonate/silica units has been described by Watkins (1983).

#### Nakoyen Member

The main part of the Nakwele Formation is constructed of very numerous thin lava flows. In this respect, the sequence of alkali olivine basalts can be contrasted with similarly thick sequences of tholeiitic lavas comprising the Nabwal Formation and the Gombe Group. It is evident, however, from the K-Ar dating study of McDougall and Watkins (in press) that the generation of alkali olivine basalt magma, overall, was no slower than that of the tholeiite, but merely the mode of eruption was different. Individual alkali basalt Nakwele flows being thinner (< 5 m), of smaller volume, and more local in extent. Weathering of the rubbly upper parts of the flows has produced interbedded oxidised palaeosols which are characteristic of the Nakwele sequence. In some cliff sections, up to 30 per cent of the sequence is made up of the weathered material and the remaining lavas are deeply altered. In the case of some thick, easily recognisable lavas, that build the upper part of Nakwele Hill and to the south provide a useful stratigraphical marker, ubiquitous alteration of olivine phenocrysts to red-brown iddingsite and the infilling of strings of micro-vesicles by calcite can be attributed to hydrothermal processes active shortly after eruption rather than to subsequent surface weathering. It is interesting, that with the arrival of clay-rich oxidised palaeosols in the sequence, there is no longer frequent development of carbonate/silica units, the stratigraphically highest being a marked calcrete deposit exposed on the eastern flank of Nakwele Hill.

The eruptive style of the Nakwele sequence may be illustrated by the rocks forming the hill of Nakoyen, which lies close to the south-western edge of the Suregei Plateau (GR 28842 97539). This steep-sided ( ~ 30°) hill is a partial remnant of a basaltic cinder cone and is the source of a lava flow which can be traced a little distance to the south. The 80 m high hill is composed of bedded pyroclasts and interbedded thin lavas, which for the most part are steeply-inclined. Bedding direction is highly variable, reflecting the complex growth of the cone, and it is clear that only a small portion of the original structure is preserved. Lapilli form beds, 0.5 m in average thickness, in which the vesiculated glass fragments are generally well-preserved but in some beds lie in a fine palagonitised matrix or have been cemented by calcite. Units exhibit reversed grading and there is a general reduction in pyroclast size upwards through the cone. Towards the top, where clast size is less than 5 mm, cross-bedding is pronounced. A fine-grained and glassy, aphyric lava flow appears to issue from below the scoria pile and so may predate the main phase of Strombolian activity. The lavas of similar character occurring within the pyroclastic sequence are thin and highly vesicular. They have 'spindle' and 'bread-crust' bombs incorporated in their upper parts, indicating that they are indeed surface lavas rather than intrusive sills, and very likely formed by the agglutination and flow of fountaining liquid spatter.

#### Gurro Member (**Sibilot Formation**)

In his original stratigraphic account, Watkins (1983) informally assigned two pantelleritic ignimbrites and associated minor siltstones and carbonate/silica beds cropping out in the upper Bakate Valley to the Gurro Member. These rocks are exposed to the east of a circular extension of the alluvial plain known as the Gurro. They lie between basalt lavas of the Nakwele Formation around the stratigraphic level of the Biliet Member and were informally assigned to this formation. Thin rhyolitic pyroclastic deposits, often contaminated by terrigenous detrital sediment, occurring within the Nakwele lavas further to the north were also placed within the Member, although direct lateral correlation was uncertain. Isotopic dating by McDougall and Watkins (in press) has subsequently shown the Gurro rocks to belong to a major episode of pantelleritic volcanism, the products of which are preserved widely around the margin of the Koobi Fora Basin, of around 13 Ma in age. Although precise relationships between strata exposed in various Miocene inliers is still unclear, the pantelleritic pyroclastic rocks are most logically grouped as the Sibilot Formation (McDougall and Watkins, op. cit.), to which the Gurro Member is here formally assigned.

The lower of the two ignimbrites present in the Bakate Valley is exposed only locally and is probably confined to a narrow valley cut in underlying lavas. It is a light grey, crystal-rich rock which has in places suffered secondary silicification with pumice fragments having been replaced by concentric growths of chalcedony. Such alteration increases towards the top of the ignimbrite and there is evidence to suggest that surface hydrothermal activity occurred in the area of present outcrop around the time of formation of the ignimbrite. The overlying ignimbrite, which has a distinctive light green colour, is more extensively preserved and may be traced southwards in a large inlier of Miocene rocks south of the Bakate Gap. There is nevertheless evidence to suggest that this ignimbritic ash-flow, too, was initially confined to discrete river valleys. In the type-section of the member on the southern side of the Bakate flood plain (GR 29249 95861) this ignimbrite is 36 m thick and is underlain by carbonate-enriched siltstones. It is extremely pumaceous in parts and shows varying degrees of welding, but with pumices commonly remaining as subrounded, little flattened fragments. Decomposition of these fragments to easily removed clays has resulted in a vacuous 'open-textured' rock. Secondary silicification may give rise to a brittle, highly resistant rock. The ignimbrite is generally massive, although indeterminate bedding is sometimes apparent close to the base where there are also local concentrations of finely comminuted fossil wood. A dramatic increase in the degree of welding within the top metre of the sheet has produced a striking brick-red, dense rock containing delicate, highly elongated, dark glass 'fiamme'.

### 3.3 **EL YIBO LAVAS** ($Tv^{EY}$)

These Cenozoic volcanics are named after the El Yibo area where the Sabarei to Dukana track crosses the Laga Balal. Here exposures are confined to the lower parts of deep gorge walls of the Laga Balal and the subdued hills to

the west. However, this unit is also thought to be widely distributed beneath the Bulal lavas in the eastern half of the map-area: the area of exposure is a regional topographic low (see map cross section).

Morphologically the El Yibo lavas are distinct from the younger plateau lavas in forming low hills with a fossil drainage network on the west bank of the Laga Balal. A reference locality is the eastern wall of the Laga Balal gorge immediately downstream of the El Yibo wells.

The age of the lavas is imprecisely known. They are unconformably overlain, with a red soil interface, by the Tatesa Hills lavas (dated at c.12.1 Ma for an upper flow) and the Mio-Pliocene Bulal lavas. The El Yibo lavas are subhorizontal and the red soil horizon is ubiquitously developed. Flows are cut by a microsyenite dyke related to the Jibisa Ring Complex; syenite phases dated at c.20 to c.14 Ma. Similar dykes to the south-west have also been dated at c.16 Ma (Wilkinson, 1988).

The total thickness of the formation is not known, as a basal contact (presumably with metamorphic basement) is never seen. Up to four flow units, individually from 20 to 40 m thick, are exposed at El Yibo.

The lavas are basaltic with highly vesicular, fissile, aphyric and porphyritic textures. All are strongly weathered with common secondary carbonate along fractures and in vesicles. Columnar joints are confined to thick flows (Plate 10).

Petrographically the lavas are hawaiites: plagioclase compositions range from An 45 to An 50. Plagioclase is the ubiquitous phenocryst phase, locally having ankaramitic textures with equigranular clinopyroxene with or without olivine. This last mineral is always strongly replaced by iddingsite. The opaque phase is ilmenite with secondary limonite.

The El Yibo lavas may be part of an early (Oligocene or early Miocene) phase of basic plateau lavas in northern Kenya east of the Kinu Sogo Rift. Alternatively, they may be a basal part of the Tatesa Hills basalts in which red soil zones are common. To the west this early phase of basaltic volcanism is represented by the Nabwal Formation of the Asille Group.

### 3.4 **SABAREI VOLCANICS** ($Tmv^S$)

This unit forms the low gentle countryside cut by the broad valleys of the major headwater tributaries of the Mogorei River passing through Arbatirs and Sabarei. It also forms the lower slopes of the Hafari massif to the east. To the south it disappears beneath the horizontal Mio-Pliocene plateau lavas. However, it is continuously exposed for more than 50 km northwards in Ethiopia, where Davidson (1983) recognises undifferentiated Miocene basalts, rhyolites, trachytes, phonolites and ignimbrites. The Sabarei volcanics, in fact, constitute the lower part of the east shoulder of the Lake Stephanie Rift.

Within the catchment area of the Mogorei River the landscape is pleasantly varied with low hills and ridges defined by narrow flow units and small cuestas, mesas and pyramid-shaped pinnacles capped by outliers of the Hafari rhyolites. Outcrop is good and the exposures besides the motorable track through Sabarei towards Jibisa illustrate the assorted lithologies of this unit.

A lower lava on the south-east side of the Hafari massif has been dated (K-Ar whole rock, Appendix 1) at about 16.2 ± 1.0 Ma. The regional geology also suggests an early to middle Miocene age for the unit (cf. Davidson, 1983) which is cut by microsyenite dykes related to the middle Miocene stages of the Jibisa Ring Complex.

The Sabarei volcanics are overlain, with apparent conformity, by the Hafari rhyolites. Good exposures of the contact are exposed on scarp slopes of the various mesas, cuestas, and pinnacles of the Mogorei River valley as well as on the major scarp slopes of the Hafari massif. To the west, the Sabarei sequence forms low ground with Nakwele Formation rocks present on higher slopes although no contacts between the two units have been located. North-west of Sabarei a series of Pliocene cinder cones puncture the Sabarei volcanics. In the south the sequence is overlain by various colluvial and alluvial spreads, and abuts the SSW-trending range of hills of the west Jibisa augite-basalt. To the east of the Hafari massif, extensively altered (to hematite-rich rocks) members of the Sabarei volcanics are faulted against Bulal lavas.

The base of the volcanic sequence is not exposed. Both Davidson (1983) to the north, and Watkins (1983) to the west, recognise basal basalts within the shoulders of the Lake Stephanie Rift: the Teltele basalts and Nabwal Formation, respectively. In the Sabarei area, lower basalt lavas are recognised and incorporated into the Sabarei volcanics but not differentiated as a specific formation because of the reconnaissance nature of the fieldwork. The upper surface of the basalt forms a well defined sloping pavement at the base of the Hafari massif. However, its base is unexposed.

Numerous Miocene and Pliocene faults cut and tilt all lithologies. As a basal contact is not exposed the total thickness of the Sabarei volcanics is unknown. About 200 m of Sabarei strata overlie the basal basalt on the west side of the Hafari massif. A water borehole (C 5001) at Sabarei penetrated 100 m of assorted felsic volcanics according to the drillers' log (Appendix 2).

Various basaltic lavas, rhyolitic ignimbrites, phonolites, trachytes, trachyandesites and rhomb porphyries (including a microlarvikite) are exposed with varying degrees of secondary iron alteration and brecciation. The intense faulting precludes establishing a rigid stratigraphy. Basal units are basaltic and upper units rhyolitic with an impersistent uppermost basalt on the east side of Hafari.

The basal basalts are variably fissile, flaggy to massive, dark grey to purple black, blocky and scoriaceous, weathered lavas. They tend to give rise to low plateaus. A single thin section of a plagioclase-porphyritic lava showed twinned and zoned andesine phenocrysts in a matrix of clinopyroxene, untwinned plagioclase, hematite with secondary limonite and clay (devitrified glass?).

Within the middle part of the sequence, exposed on the south-western slopes of Hafari, are several thin (10 to 20 m) but laterally continuous flows. These are aphanitic, massive rocks with extensive hematite replacement imparting a crimson colour to weathered surfaces. In thin section, euhedral nepheline grains were noted in a hematite-rich groundmass; these lavas may be altered phonolites.

Aphanitic, locally highly vesicular, basalts cap fault-controlled hills immediately south of Sabarei. These lavas have also been variably replaced by hematite to produce grey, yellow, brown and crimson weathered surfaces. Pipe amygdales are infilled by yellow and green clays, calcite and agate (which form extensive float at the bases of the hills). The basalts are variably porphyritic with olivine, augite and labradorite phenocrysts in a ground mass of microlitic clinopyroxene, labradorite and opaques.

Two hills, 4.5 km south of Sabarei, on the north side of the motorable track are underlain by mottled (olive-green

with pale brown spots) lavas containing euhedral feldspar laths, small lithic fragments, and welded textures in summit exposures. These N-trending hills are noticeably darker on aerial photographs than adjacent countryside and may represent erosional remnants of a narrow N–S flow unit. One thin section of this rhomb-porphyry revealed anorthoclase phenocrysts and highly altered lithic clasts in a microcrystalline matrix partly composed of lithic and glass fragments.

Other felsic volcanics, variably replaced by hematite, include pink, homogeneous rhyolites with rare feldspar phenocrysts and blue-grey ignimbrites. The latter rocks are well exposed in the area between Arabtirs and Sabarei. They have a strong welded fabric, locally imparting a horizontal fissility, with flattened lithic and pumice fragments. Rounded orthoclase crystals, with or without quartz phenocrysts, deflect the welded fabric.

Systematic sampling of the lower slopes of Garba Gala on the east side of Hafari revealed three rhomb porphyries, one phonolite, two trachytes, one trachyandesite and six hawaiitic or ankaramitic basalts over 200 m of volcanic thickness.

An uppermost, thin (< 10 m) pink-grey ignimbrite overlying a 20 cm thick olive green tuff was identified. Several microsyenite dykes cut this sequence. A chemical analysis of the trachyandesite from Garba Gala is included in Table 2. The high $Fe_2O_3$ content reflects secondary hematite. A similar petrogenetic model for the volcanic sequence within the eastern shoulder of the Lake Stephanie Rift is proposed as for the Rift's western shoulder (previous section). This is described in Chapter 4 which deals with the Jibisa Ring Complex, a major source of the eastern volcanics.

### 3.5 **HAFARI RHYOLITES** ($Tmvf^{HF}$)

This formation is named after the Hafari massif which straddles the Ethiopian border to the east of Chew Bahir. Felsic volcanics of the Hafari rhyolites cap the whole of this massif above an elevation of 1046 m and small, lower outliers to the west in the Mogorei Valley. Similar felsic extrusives continue to underlie high ground northwards into Ethiopia: Davidson (1983) incorporated them into an undifferentiated Miocene volcanic sequence.

The subhorizontal massive felsic flows dipping very gently ESE which underlie the tablelands are clearly distinguished both in the field and on aerial photographs. Dyke swarms and major fractures produce positive and negative morphological and photogeological lineaments, respectively.

Field traverses were completed down the southern slopes of Garba Gala and Barchum Keramso, and spot observations made along the summit plateau of Hafari. Good outcrop in each of these localities could be used as reference sections.

The Hafari rhyolites overlie the early to middle Miocene Sabarei volcanics. They are situated close to the Jibisa Ring Complex, a major centre of felsic volcanism on the eastern shoulder of the Chew Bahir rift. It is probable that the major flows, which are unusually extensive for those of rhyolitic composition, were erupted from the Jibisa Volcano. Felsic intrusives of the ring complex have minimum K-Ar whole-rock ages of 20.7 ± 1.4 and 14.4 ± 0.7 Ma. Late microsyenite dykes from the complex cut the Hafari rhyolite flows and, hence, the latter are almost certainly of middle Miocene age.

The contact between the Hafari and Sabarei units is taken as the base of the lowermost massive and thick rhomb-porphyry flow (see Plate 9). On the east side of Hafari this also coincides with the top of the uppermost basalt unit in the Sabarei volcanics. Essentially the Hafari rhyolites are wholly felsic, whereas the Sabarei volcanics are a mixed volcanic suite. The upper contact of the Hafari rhyolites is never seen. Much of the Hafari massif is unfaulted, probably a reflection of rift-related faults decreasing in number eastwards and, in part, because the capping rhyolites post-date much of the faulting. The major bounding fault along the eastern side of Lake Stephanie in Ethiopia can be traced on satellite imagery into Kenya as far as Jibisa. In Ethiopia throws of over 1000 m are recorded on this fault (Davidson, 1983) whereas in Kenya it does not displace the Hafari rhyolites. The surface trace is a poorly defined valley on the summit plateau but a major valley on the scarp slope.

About 600 m of felsic volcanics are preserved on Hafari and constitute the present maximum thickness of the Hafari rhyolites. The stratigraphy is dominated by two, thick (over 100 m), laterally extensive massive porphyritic ignimbrites which form cliff faces overlooking Sabarei. The upper flow has a uniform pink hue due to extensive hematite alteration. However, the underlying flow is more erratically replaced by hematite and is generally a creamy grey colour. Both flows are massive, fine-grained and homogeneous except for local developments of brecciation and/or welding (Plate 12). The resistant cap rock to the outliers in the Mogorei Valley is a fine-grained, grey, variably welded porphyritic ignimbrite. It is laterally extensive relative to the narrow felsic flows of the Sabarei volcanics which is the reason for including it in the Hafari rhyolites.

All ten samples collected from the Hafari rhyolites on Hafari proved to be orthoclase porphyritic flows. The rock matrix is fine-grained, usually microcrystalline with minor, altered mafic phases, clay, limonite, and hematite. Quartz and microlitic feldspar form discrete grains in this matrix. Orthoclase phenocrysts, up to 4 mm in size, are fragmented euhedral crystals showing simple twinning or else are untwinned (rarely microperthitic).

Aegirine occurs in the orthoclase porphyry capping the plateau at GR 3329 9696 in the Mogorei Valley. Sanidine is the phenocryst phase in the porphyry underlying the pyramid shaped hill at GR 3344 9712.

A single whole rock analysis of the upper pink flow is given in Table 3. The analysis is peralkaline, and pantelleritic (over 10 per cent normative quartz). In part this may be due to secondary sodium (as albite) in the microcrystalline matrix (compare the analysis with the pantellerite from Mayor Island in Ewart and others, 1968) although pantellerites are the characteristic rhyolite material of the Rift Valley. The petrogenesis of the Hafari rhyolites is related to the Jibisa Ring Complex and discussed in the next chapter. According to Bowden (1985) the volcanic sequences related to African alkaline (A type) granite ring complexes are ignimbritic rhyolites. The considerable lateral extent of the two major flows is unusual for felsic rhyolitic (or pantelleritic) extrusives, although Oftedal (1960) describes similar flows from the older (Permian) Oslo volcanic field. The rhomb-porphyries may represent subaerial ash fall.

### 3.6 **TATESA HILLS BASALTS** ($Tmvb^{T}$)

This formation is named after the isolated NE-trending range of low hills to the west of the Laga Bulal which are

clearly visible from the main Dukana to Sabarei track. Outcrop of the basalts is confined to the map-area and their full geographical distribution is therefore shown on the geological map. Individually, the Tatesa Hills are rounded, and strongly incised (fracture controlled), rising to maximum elevations of c.1100 m above sea level, which is about 300 m above the adjacent plateau. Their most characteristic feature is the uniform, horizontal, volcanic stratification which gives each hill a pronounced striping. This is visible, both on the ground and on aerial photographs despite the thick scrub which covers the hills. Outwash from the hills mantles adjacent plateaus producing pale areas on aerial photographs that contrast with the dark plateau basalts.

Traverses up either of the two major southern hills could be used to define future type/reference sections. During the fieldwork a single traverse was made up and along a smaller northern, more accessible, hill underlain by only six lava flows. About 30 flow units underlie the main hills to indicate a mean flow thickness of about 10 m.

A lava flow from the summit plateau of the traversed northern hill (i.e. about one-fifth of the way up the total exposed succession) gave a K-Ar whole rock age of c.12.1 Ma (Appendix 1). The total age range of the complete succession is not known; numerous red-weathered tops to flows suggest long periods without volcanic activity. The volcanism may consequently have spanned a long period of the middle Miocene.

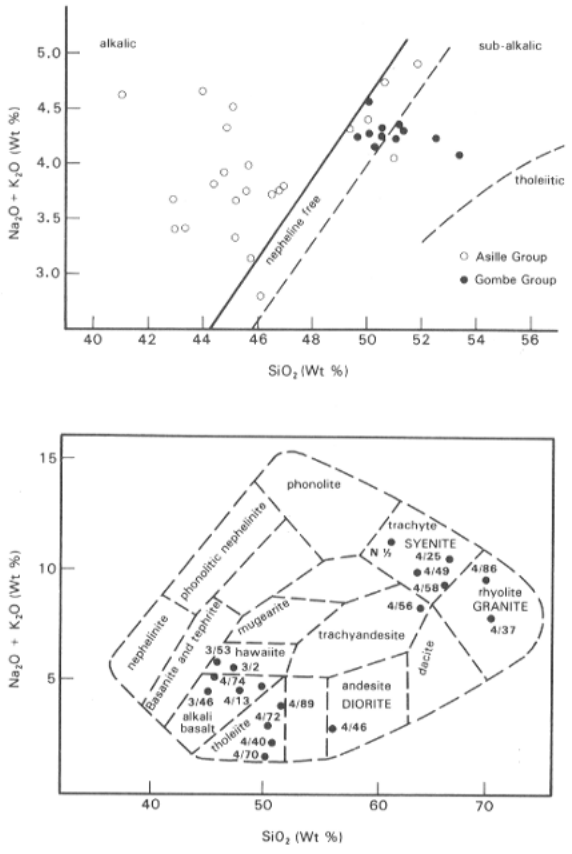



**Figure 5** ($Na_2O + K_2O$) vs $SiO_2$ plots of whole rock analyses, after Cox and others (1979). Thick line separates nepheline-bearing and nepheline-free rocks (Hughes, 1982, fig. 10.3)

The Tatesa Hills basalts unconformably overlie a weathered surface of the El Yibo lavas. The hills form a series of inliers encircled by Bulal lavas, although actual contacts are masked beneath colluvium.

All the flows appear to be basalts (i.e. no unusual weathering horizons were noted during a helicopter circuit of the largest hill). On the traversed hill, coarse-grained basalts are interlayered with scoriaceous flows beneath fissile, aphanitic upper lavas. Most flows are dull grey, variably weathering to shades of crimson (notably on upper surfaces). Individual flows are invariably c.10 m thick. Olivine phenocrysts are confined to discrete flows; grey clay infilling to vesicles was observed in one scoriaceous lava. Float of an unusual plagioclase-phyric lava is part of the scree mantle to the largest southern hill. A dolerite (feeder) dyke is exposed on the summit of the traversed northern hill. It trends, with a parallel fissility, towards 023°, i.e. subparallel to the NE alignment of the Tatesa Hills.

The lavas are petrographically varied, with olivine basalts, ankaramitic basalts, plagioclase porphyries and hawaiites. Plagioclase compositions vary between An46 and An56 except in the feeder dyke (An64). Subhedral olivine phenocrysts are strongly altered to iddingsite. Augite phenocrysts are rounded and zoned but are not titaniferous. Plagioclase, ore and clinopyroxene (microlitic) constitute the groundmass along with secondary limonite and green clay replacing devitrified glass.

An ankaramitic basalt and an aphyric lava from the Tatesa Hills basalts were analysed (Table 3) and both plot in the tholeiitic basalt field of Figure 5. The ankaramite composition is relatively unfractionated as it plots close to source on an A'F'M diagram (Figure 6).

The Tatesa Hills basalts are regarded as part of the volcanic succession which formed the eastern shoulder of the Lake Stephanie–Kinu Sogo Rifts during the Miocene. The basalts were erupted more or less contemporaneously with felsic flows from Jibisa. The pronounced NE alignment of present outcrop is regarded as a primary feature controlled by the (NE) trend of feeder dyke sources.

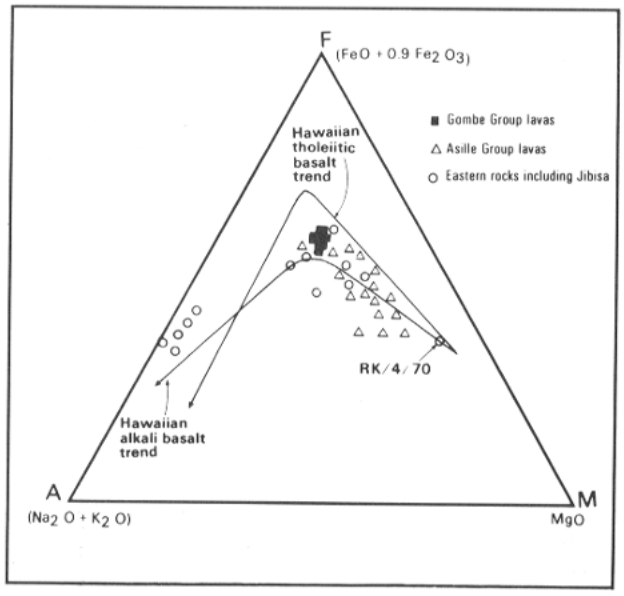



**Figure 6** Whole rock analyses plotted on an A'F'M diagram

### 3.7 **WEST JIBISA ANKARAMITE** (augite basalt, $Tvb^a$)

The unnamed low range of hills trending SSW away from the west side of Jibisa is composed of a black, coarsely porphyritic basalt. The phenocryst phase is augite so the unit is informally referred to as the West Jibisa ankaramite (augite basalt), and is spatially confined to the map-area. Each hill is low, never more than 200 m in elevation above adjacent low ground, and covered by boulder scree of the black basalt. The hills consequently have characteristically dark phototones. The basalt is not stratified but massively homogeneous throughout its outcrop area. Therefore, the summit outcrop of the most accessible northern hill (GR 3432 9697) is a suitable reference locality. Lower outcrops adjacent to the motorable track are strongly altered.

The exact age of this unit was not determined. The highly eroded nature of the hills (cf. Tatesa Hills) and their structural setting on the eastern shoulder of the Kinu Sogo Rift suggest a Miocene age. They are juxtaposed to Jibisa, and strike into this Miocene centre, which is apparently the source of the basalts. The augite basalt appears to overlie the intrusive basalt phase of Jibisa. Several fault sets, including the NNE Pliocene system, cut each hill controlling the weathering. The contact of the augite basalt with the younger plateau lavas to the south is partly faulted.

The present exact thickness of the augite basalt is unknown but may be of the order of 100 m. The dominant lithology is the black ankaramite although major secondary alteration adjacent to Jibisa modifies the field appearance. In thin section the phenocrysts are seen to be strongly zoned titaniferous augite; as corroded grains up to 1 mm long showing minor serpentine or chlorite alteration. Minor and variable proportions of fragmented olivine phenocrysts are similarly altered. The groundmass is dominated by tiny titaniferous augite grains with interstitial plagioclase, olivine and ore. Accessory phlogopite and nepheline may be present with secondary serpentine, chlorite, sericite, calcite, and quartz increasing towards Jibisa.

The augite basalt is unique in the map-area in having titaniferous augite as a principal phase. This is reflected by the high $TiO_2$ content of three analysed samples with respect to other basalts (Table 3). Two of the analyses plot as alkali basalts and the other as a hawaiite on Figure 5. However, they all lie nicely on an alkali basalt fractionation trend (Figure 6) with the rest of the Jibisa related volcanics.

The balance of available evidence suggests that augite basalt is part of the volcanic activity related to Jibisa.

### 3.8 **BULAL LAVAS** ($Tv^B$)

Flood lavas forming the plateau east of the Kinu Sogo Rift are well exposed in section in the Laga Bulal (alternatively spelt Balal) gorge and on the interior slopes of Gof Wargan in the extreme east. The Bulal gorge may be taken as the type-locality of the lavas in Kenya. Davidson (1983) has referred to their northern continuation into Ethiopia as the Bulal basalt. The Bulal lavas are easily distinguished from adjacent volcanic units and their geographical extent and age relationships are easily defined (Figure 7). The plateau forming lavas extend south to the Chalbi Basin (Charsley, 1987). In Ethiopia they have an extensive thin soil cover whereas in Kenya the plateau surface is extensively mantled by a thin lag gravel. Hills standing above the plateau surface are formed of different lithostratigraphic units, e.g. Tatesa Hills, Bollo.

The Bulul lavas are clearly equivalent to the lavas of the Gombe Group (Watkins, 1983; Watkins and Kirkwood, in press) to the west. Contemporaneity of the lavas capping the plateaus on either side of the Kinu Sogo Rift is shown by the K-Ar geochronology. Brotzu and others (1984) reported K-Ar whole rock ages of between 6.1 ± 0.6 and 4.5 ± 0.3 Ma for the southernmost Bulal lavas, essentially the same timespan recorded from the Gombe lavas at the margin of the Koobi Fora Basin (Fitch and others, 1985; McDougall, 1985; McDougall and Watkins, in press). The Bulal lavas, therefore, can be confidently assigned to the Gombe Group.

The flood lavas unconformably overlie the El Yibo lavas, there being a distinctive red soil developed upon the weathered surface of the older sequence. The contact is well seen in the walls of the Laga Bulal gorge at El Yibo. The almost flat-lying lavas abut eroded Miocene rocks of the Tatesa Hills basalt, Sabarei volcanics and the Jibisa Ring Complex. In the east they provide the foundation for the Dukana Volcanic Province being perforated by the Quaternary *maar* craters. The lava pile thickens eastwards and westwards from the topographic low at Laga Bulal: less than 100 m are recorded at the laga whilst thicknesses in excess of 200 m occur beneath the Dukana volcanics and on the western shoulder of the Kinu Sogo Rift.

The walls of Gof Wargan expose six lava floods, individually up to 60 m but mostly less than 20 m thick. Flow units of the aphyric tholeiitic lava generally have vesicular, locally scoriaceous or blocky, tops and brecciated, blocky bases. Flow unit cores may be massive or fissile, and flow structure is often evident in field exposures. Fissile units have closely spaced joints which disrupt the lava into small angular blocks. Substantial columnar joints are restricted to thick flows. Interflow soils are absent. Secondary carbonate lines joints and fractures and infills vesicles.

The lavas are mostly dark grey to black and of fresh appearance in the field (cf. Davidson, 1983) with very thin, film-like weathering on exposed surfaces. Five out of six lavas examined in thin section contained abundant plagioclase feldspar as microphenocrysts and within the matrix. Individual feldspar laths show variable degrees of twinning and occasionally are strongly zoned. Clinopyroxene, Ti-rich magnetite and secondary clay (replacing devitrified glass) accompany feldspar in the very fine-grained equigranular groundmasses. Olivine microphenocrysts are rare.

### 3.9 **GOMBE GROUP** ($Tv^G$)

Watkins (1983) assigned near flat-lying tholeiitic basalt flood lavas capping the Suregei and Gombe Plateaus, on the margin of the Koobi Fora Basin, to the Gombe Group. Reconnaissance mapping has traced the flood lavas of this Group southwards and eastwards across the Kinu Sogo Rift, as the Bulal lavas. Wilkinson (1988) extends the same lavas about 100 km southwards to the Kulal shield volcano.

The Gombe lavas typically underlie a featureless plateau in the type-area. In the Kinu Sogo Rift the lavas are disturbed by Pliocene faulting and underlie N-trending valleys and intervening blocks (Chapter 2). The flood basalts are easily distinguished on aerial and satellite photographs by their uniform dark tone. K-Ar



**Plate 10** Columnar jointed basalt of the El Yibo lavas in the El Yibo gorge.



**Plate 11** Fluvial sandstones and conglomerates of the Il Yia Formation, north-western flank of the Suregei Plateau.



**Plate 12** Flow banding in the upper pantelleritic flows of the Hafari rhyolites; western side of the Hafari massif on the Ethiopian border.



**Plate 13** Details of the surge deposit shown in Plate 14.



**Plate 14** Bedded surge deposits defining the rim of Dabacha *maar*.



**Plate 15** Angular unconformity between the Asille Group and the sediments of the Koobi Fora basin, exposed in the Il Biliet valley (27595 97620).

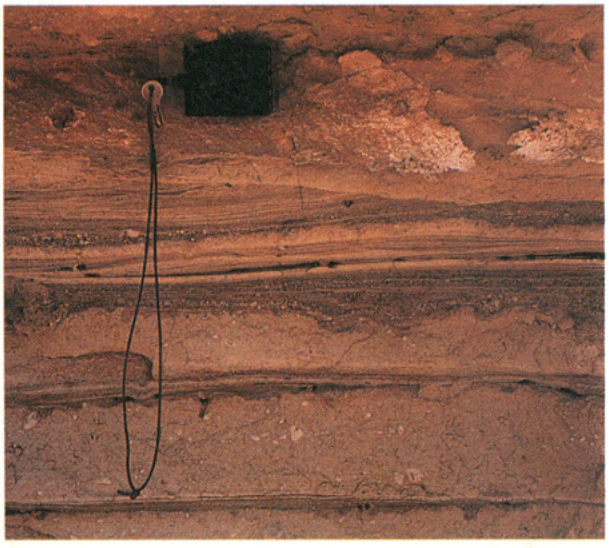


**Plate 16** Sedimentary structures in the pyroclastics of Plate 17.

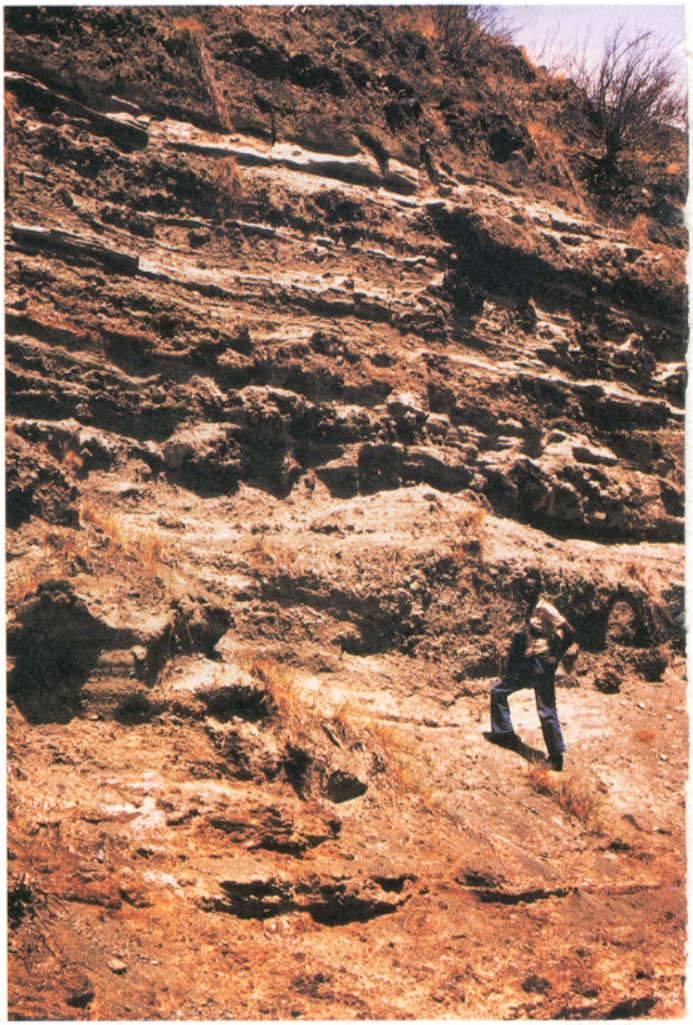


**Plate 17** Stream section into bedded pyroclastics, North Island.



**Plate 18** Plio-Pleistocene sediments exposed in the south bank of the Il Eriet near Ileret Police Post.



**Plate 19** Typical superficial gravel on the summit plateau of Hafari.



**Plate 20** Galana Boi Beds in gully adjacent to the only motorable track up the steep northern escarpment of Kokoi horst. Shell bed in freshwater limestone.



**Plate 21** Lag gravels on the low hills NNE of Ileret.



**Plate 22** Secondary alteration of K-feldspar phenocrysts, porphyritic microsyenite, Jibisa.

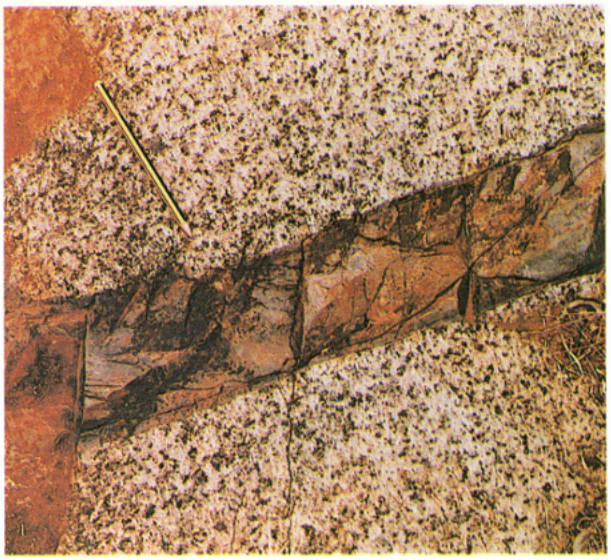


**Plate 23** Microsyenite dyke cutting the central coarse syenite stock, Jibisa.



**Plate 24** Syenite xenolith within the porphyritic microsyenite, Jibisa.



**Plate 25** North-eastern part of Gumuti Dura vent viewed from the south.

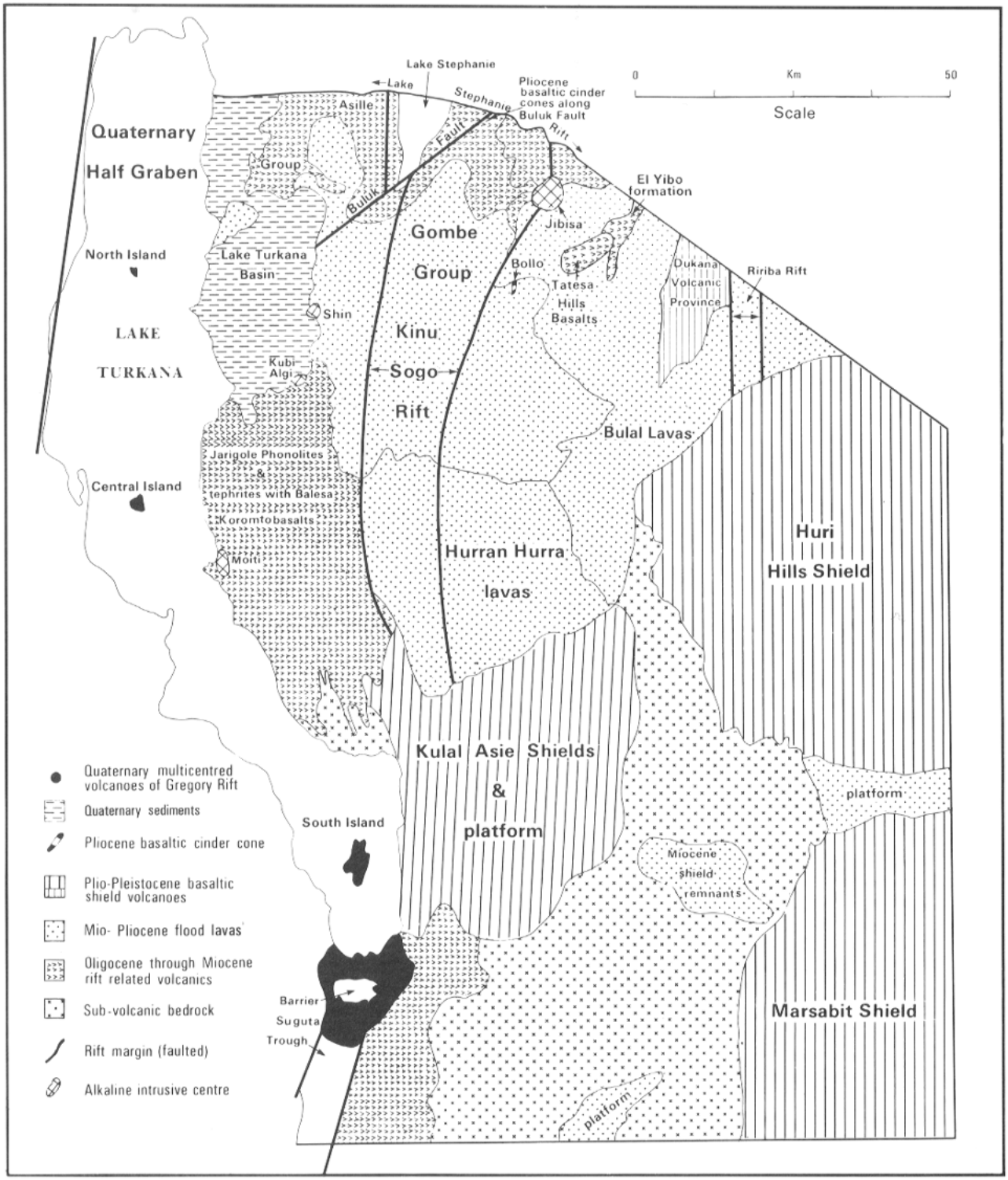



**Figure 7** Regional distribution of the Miocene to Pliocene volcanics

and $^{40}Ar/^{39}Ar$ dating of the lavas has yielded minimum ages of between 6.57 ± 0.35 and 3.73 ± 0.05 Ma, late-Miocene to Pliocene (Chapter 5). The flood lavas represent a recommencing of volcanism after a significant hiatus with an angular unconformity existing where they overlie the faulted dipping strata of the Asille Group. This unconformity is well exposed in the type area and along the scarp slope eastwards from Buluk Police Post. Within the upper Bakate Valley and locally elsewhere the surface of the Asille Group is unconformably succeeded by lacustrine and fluvio-deltaic sediments relating to an early lake. Watkins (1983) has suggested that these little disturbed sediments (Bas Formation) may be laterally equivalent to units of the Kubi Algi and lower Koobi Fora Formations of the Koobi Fora sedimentary basin, or alternatively they may extend back in time the history of sedimentation of Lake Turkana.

Further lacustrine beds, usually a few metres or less in thickness, interdigitate with the lava floods. In his initial study of the Suregei Plateau, Watkins (1983) was able to trace a thin but widespread bioclastic limestone and sandstone bed, the Warata Formation. The flood lavas

beneath this bed were termed the Chen Alia Formation, whilst a thick lava flood above was termed the Harr Formation (Hooker and Watkins, 1980). Type localities for these Formations are as follows:

| | |
|---|---|
| Harr Formation | North side of Suregei Plateau (GR 29083 98347) |
| Warata Formation | East side of Suregei Plateau (GR 28624 97941) |
| Chen Alia Formation | Upper Warata Valley (GR 28875 97735) |

Subsequent investigation of the Gombe Group has revealed the presence of more numerous lava flows which probably have the form of very shallow, overlapping shields: it is evident that only the most painstaking mapping will enable full elucidation of the lava succession. Owing to the small-scale, no attempt is made on the geological map to subdivide the Gombe Group.

As stated in Section 3.8, the Bulal lavas to the east of the Kinu Sogo Rift can be assigned to the Gombe Group, there being an essential contemporaneity between the lava series east and west of the Kinu Sogo Rift. One of the present authors, R. M. K., has interpreted basalt boulder lag deposits forming mounds upon the Bulul lavas near Bollo as remnants of a late Gombe lava that flowed from the west (available isotopic age data is consistent with flood volcanism to the west outlasting that forming the Bulal flows).

At the edge of the Koobi Fora Basin, Gombe Group lava flows alternatively terminate westwards or are overstepped by lacustrine sediments. The capping basalt on the Kokoi horst overlies lacustrine sediments similar to those occurring beneath the flood lavas in the basin margin (Watkins, 1983).

The total thickness of the Gombe Group in the type-area is approximately 180 m with 120 m, 1.33 m and 58.5 m of Chen Alia, Warata and Harr Formations, respectively, in type-section. The sediments preserved beneath the Gombe lavas are up to 40 m in thickness. Along the scarp east of Buluk Police Post the flood basalts are less than 50 m thick. A similar thickness of flood lava probably occurs throughout the Kinu Sogo Rift.

The Chen Alia Formation is composed of several thick flows of aphanitic and aphyric lavas. Individual flows locally utilise valley features to produce an inverted stratigraphy and complicated outcrop pattern. Individual flows may be highly vesicular and flow-banded with inverted *pahoehoe* structures at their bases. Levees and small standing lava pools are preserved (GR 29008 96286). The base of the formation is marked by a chilled black glass, between 1 and 2 cm thick, with an orange brown rind.

The Warata Formation (Watkins and Williamson, 1980) is a complex sequence of lacustrine, carbonate-cemented breccias, carbonate-rich silts, sandstones and bioclastic limestones. It has a freshwater mollusc macrofauna (listed in Watkins, 1983).

The Harr Formation consists of identical lavas to the Chen Alia Formation. Individual flow units have brecciated bases with chill textures and *pahoehoe* structures, and vesicular tops. Possible feeder dykes trend NNE: the dominant Pliocene tectonic trend. Brick red and grey diatomite, with or without limestone, is interbedded in the lava sequence at GR 29484 98111. All the flood basalts are aphyric with microphenocrysts of plagioclase (sodic labradorite or calcic andesine) and augite in a groundmass of plagioclase, magnetite, ophitic clinopyroxene and glass mesostasis. Olivine is present as relatively infrequent, rounded microphenocrysts (cf. the Bulal lavas). Chemically they are quartz tholeiites (Table 2).

The Gombe Group basalts represent renewal of volcanism after a period of about 6 Ma. Feeder dykes trend NNE parallel to contemporaneous rift fractures: the obvious control on magma ascent. Considerable erosion must have taken place during the later Miocene to level off the surface upon which the flow basalts were neatly (for the most part) laid down. The Bas Formation completed the levelling by filling up topographic depressions.

Lava eruption overlapped with the onset of sedimentation in the Lake Turkana Basin, the Bas and Warata Formations representing the easternmost transgression of the early Pliocene lake. An intriguing question is whether the transgression extended far enough eastwards (across a relatively flat area) to enter the Chew Bahir depression.

### 3.10 **DUKANA VOLCANIC PROVINCE AND ISOLATED PLIOCENE BASALTIC CONES** (Qvb, Qvb$^{l}$, and Qvbf)

Pliocene basaltic volcanism in the map-area was dominated by lava eruption (Bulal lavas and Gombe Group) with the cinder cones representing less than 1 per cent of the total volume of magma eruption. Quaternary basaltic volcanism is confined to the eastern Dukana Volcanic Province consisting of two N-trending belts through the village of Dukana (immediately south of the map-area). Part of the westernmost belt straddles the eastern map-area. The Ririba Rift separates the two belts which both extend into Ethiopia with total lengths of about 35 km.

Isolated cinder cones are sited adjacent to Pliocene rift faults in the centre of the map-area: at Bollo on the eastern boundary fault array of the Kinu Sogo Rift and in the Chew Bahir depression along the trace of the Buluk Fault. All these cones are eroded down to low dome-shaped hills.

Three classic volcanic landforms occur within the Dukana Province: steep sided cinder cones with summit craters, *maars* and upstanding blocky lava flows. All show up clearly on aerial and satellite photographs. Gof Wargan, Dibandiba and the flows straddling the narrow track from Dukana into Ethiopia are representative of the three landforms and their volcanic components and define reference sections.

A K-Ar whole rock age of approximately 0.9 Ma has been obtained for a basalt from the southern part of the Dukana Province (Charsley, 1987). The Pliocene age for the central eroded cones is confirmed by the K-Ar whole rock age of about 2.25 Ma (Wilkinson, 1988) from a cinder cone south of Bollo but similarly situated along the eastern margin of the Kinu Sogo Rift.

The Dukana Volcanic Province is founded on the flat lying Bulal lavas. The N–S alignment of the individual belts adjacent to a parallel Ririba Rift suggests that faults peripheral to this rift controlled magma ascent. The Rift faults are probably shallow seated (note their small displacements), which explains the lack of Quaternary volcanism within this rift.

Maximum height of the cinder cones is about 250 m (Dibandiba), whereas the pyroclastic rims to the *maars* are less than 100 m thick (Plate 14). Both types of deposit rapidly taper in thickness radially outwards. Lava flows are narrow and only several metres thick. The eroded Pliocene cones are less than 40 m in elevation.

All cinder cones (i.e. Quaternary and Pliocene) are composed of red-brown weathering, clast-supported

lapilli tuffs. They have radial dips increasing towards cone apices. Thus the eroded Pliocene cones have shallower dipping strata than the Quaternary cones (apical dips of up to 28°). The principal components of the cones are angular scoriaceous basalt lapilli, including many fragmented spindle bombs. Less common are block-size basalt clasts and cow-dung bombs up to 1 m in length. Individual pyroxene and olivine crystals and peridotite nodules are always present in minor amounts. Gneiss xenoliths are confined to specific cones (e.g. at GR 3840 9472). The breccias are poorly consolidated and weather to produce major scree deposits and peripheral bedded, carbonate-cemented, reworked tuffs. These tuffs are matrix supported with lithic and crystal clasts.

Pyroclastic rims to the *maars* are variably consolidated, finely bedded, matrix-supported ashes and tuff (Plate 13). They are mostly red-brown in colour but grey and pink beds exist. Basaltic lithic lapilli are the most common clasts in a basaltic ash matrix with rarer basalt blocks, olivine and pyroxene xenocrysts, peridotite nodules and gneiss fragments. Secondary carbonate cements the tuffs, which coarsen upwards. At Dabacha the uppermost beds are lapilli tuffs identical to the cone deposits.

The basalt lavas are black-grey, massive aphyric rocks of fresh appearance. They are variably vesicular though never scoriaceous. Rounded olivine megacrysts characterise the field appearance of the lavas.

In thin sections the volcanics are either olivine or ankaramitic basalts with olivine and, in some cases, zoned augite phenocrysts in microcrystalline groundmasses. The phenocrysts are rounded or fragmented anhedral grains varying in size up to 10 mm. Well defined iddingsite rims are present on olivine. Fine plagioclase needles are unorientated in the groundmass where secondary carbonate is ubiquitous.

A single chemical analysis (Table 2) of a Dukana lava plots as an alkali basalt (Figure 5). In fact the Dukana Volcanic Province is the most northerly of a series of alkali basalt multicentred shield volcanoes in northern Kenya. It differs from the others (e.g. Huri and Marsabit Shields) in its relative youth (Quaternary as against Plio-Pleistocene) and the dominance of pyroclastics relative to lavas. More detailed work is needed to ascertain whether there is definite geographical trend with regard to age and composition of these shields. All the shields have a tectonic control (see Charsley, 1987 and Key, 1987a) which has allowed uncontaminated, poorly differentiated alkali basalt magma to reach surface.

### 3.11 NORTH ISLAND TRACHYTES (Qrvt)

As its name implies, North Island is the most northerly of three volcanic islands in Lake Turkana. It is about 60 km north of Central Island and equidistant from the present east and west shorelines. North Island, less than 2 km in diameter, is the summit of a trachytic volcano with two of its centres underlying the main peaks: a breached northern cone and an older southern cone (Figure 4). Preferential erosion of the friable pyroclastic walls of the southern cone has exposed an infilling dioritic plug. The northern cone has a small summit crater mostly infilled by a trachytic flow which has breached the north and west crater walls. These walls are heavily gullied, especially on steep upper slopes which level off on to flat-lying older pyroclastics. The lavas have curved pressure ridges and *aa* textures with upper surfaces disrupted into house-size blocks separated by irregular and deep chasms. All the major morphological elements are clearly visible on aerial photographs (Plate 1).

The distribution of the three principal lithologies: trachytic lava, pyroclastics and diorite plug is shown on the geological map (see also Brown and Carmichael, 1971).

Summit volcanism is Quaternary with fumarolic activity continuing to the present day. The trachytic lava from the northern cone post-dates the bench mark related to the last major high water level of Lake Turkana. The Galana Boi Beds were laid down on the eastern side of the lake during this high water period and have been dated by the radio carbon technique at between 9880 and 4390 years BP (Vondra and others, 1971). Brown and Carmichael (1971) quoted an age of 3250 years BP for the high water level. K-Ar whole rock ages (Appendix 1) tend to confirm the young age of the trachytes, and suggest an age of less than 300 000 years BP for the diorite plug.

**Table 2** New whole rock analyses (Watkins, 1983 provides additional analyses from his thesis area)

| | RK3 53 | RK4 56 | NI 2 | NI 8 | RK3 46 | RK4 13 | RK4 16 |
|---|---|---|---|---|---|---|---|
| $SiO_2$ | 45.60 | 64.00 | 61.40 | 56.40 | 44.80 | 48.60 | 49.70 |
| $Al_2O_3$ | 9.66 | 12.10 | 14.49 | 13.32 | 13.88 | 9.49 | 14.21 |
| CaO | 11.20 | 0.56 | 4.20 | 9.52 | 9.52 | 9.80 | 8.12 |
| MgO | 10.08 | 0.60 | 1.08 | 2.42 | 8.06 | 12.50 | 4.64 |
| FeO | 7.69 | <0.14 | 4.89 | 3.59 | 10.27 | 8.91 | 10.70 |
| $Fe_2O_3$ | 2.73 | 11.07 | 0.64 | 5.51 | 2.16 | 1.71 | 2.11 |
| $Na_2O$ | 5.46 | 5.06 | 8.29 | 4.45 | 3.37 | 4.01 | 4.25 |
| $K_2O$ | 1.08 | 3.68 | 3.62 | 0.54 | 1.81 | 1.02 | 1.08 |
| $P_2O_5$ | 0.71 | 0.10 | 0.02 | 0.52 | 0.28 | 0.44 | 0.38 |
| $TiO_2$ | 2.84 | 0.58 | 0.71 | 1.63 | 2.54 | 1.67 | 2.29 |
| MnO | 0.23 | 0.41 | 0.19 | 0.15 | 0.21 | 0.24 | 0.25 |
| L.O.I. | 2.10 | 2.00 | 0.70 | 1.90 | 2.00 | 1.70 | 1.80 |
| **Total** | 99.38 | 100.16 | 100.23 | 100.23 | 98.90 | 100.09 | 99.53 |

| **Norms** | RK4-56 | RK NI-8 |
|---|---|---|
| Quartz | 13.71 | 8.39 |
| Orthoclase | 21.75 | 3.19 |
| Albite | 41.75 | 37.66 |
| Anorthite | 0.00 | 14.78 |
| Diopside | 1.86 | 24.31 |
| Hypersthene | 12.94 | 2.18 |
| Olivine | 0.00 | 0.00 |
| Aegirine | 0.98 | 0.00 |
| Magnetite | 3.15 | 2.90 |
| Ilmenite | 1.10 | 3.10 |
| Apatite | 0.24 | 1.23 |
| Water | 2.00 | 1.88 |
| Norm Total Error | 0.04 | 0.00 |
| Thornton-Tuttle Index | 77.21 | 49.24 |

| | |
|---|---|
| RK3/53 Olivine basalt, Pliocene | *(Hawaiite) |
| RK4/56 Hematite zone; Sabarei volcanics | (Trachyandesite) |
| NI/2 Trachyte, North Island | (Trachyte) |
| NI/8 Diorite, North Island | (Diorite) |
| RK3/46 Basalt, Bakate Formation | (Alkali basalt) |
| RK4/13 Basalt, Dukana volcanics | (Alkali basalt) |
| RK4/16 Basalt, El Yibo lavas | (Alkali basalt) |

* Name in parenthesis based on Cox and others (1979) alkali vs silica plot.

North Island, like all Quaternary trachytic volcanoes in Kenya is confined to the median trough of the Gregory Rift which implies a tectonic control on their setting. They are also almost equidistantly separated along the Rift and are not situated at the intersection of major discordant fractures with N–S faults.

The island's summit stands approximately 140 m above lake level. The surrounding waters are some 80 m deep giving a total height of the order of 200 m for the volcano.

Volumetrically, the dominant exposed lithologies are well bedded pyroclastics consisting of interlayered lapilli tuffs and ash tuffs (Plates 16 and 17). The beds dip radially off the cones, but are subhorizontal at lower topographic levels. Typical small-scale structures include parallel layering, low angle cross-beds, grading, sets and co-sets defined by variable clast content in the same ash matrix, and bedding sags beneath large clasts. Scoriaceous lava blocks disrupt bedding by initial impact and secondary compaction. Lapilli clast content varies from nothing to 100 per cent in clast-supported lapilli tuffs. The most common lithic clasts are composed of angular trachytic lava.

Solfataras in the centre of the island produce sulphur and gypsum deposits which brightly colour the otherwise drab grey pyroclastics. Individual gypsum crystals are common in natural cavities throughout the pyroclastic rocks. Around the main solfataras, the host tuffs and diorite plug have decomposed to secondary claystones with a paste-like consistency.

The *aa* trachyte flow is composed of a fresh friable, scoriaceous, glassy, black grey lava which is locally porphyritic. Scattered bombs of porphyritic obsidian up to about 40 cm in diameter are part of the initial extrusive phase preceding the lava eruption. In contrast the diorite plug is a massive, coarse-grained, grey rock with abundant feldspar phenocrysts.

Details of the petrography of the volcanics of North Island are given by Brown and Carmichael (1971). Ejecta within the lowermost (flat-lying) pyroclastics consist largely of interlocking zoned plagioclase, pyroxene, and iron oxide grains with or without olivine. Secondary alteration phases related to the solfataras are montmorillonite, serpentine, zoisite, hematite, zeolites, carbonate and chlorite.

The lavas contain bytownite, augite and less common olivine phenocrysts set in a groundmass of feldspar, pyroxene, iron-titanium oxides and interstitial brown glass. The groundmass plagioclase is zoned to anorthoclase. The obsidian bombs contain large bytownite phenocrysts, but, unlike the lavas, contain few olivine and pyroxene phenocrysts. The microlitic feldspar of the groundmass is largely anorthoclase. Small crystals of diopsidic augite and titanomagnetite are also present in a brown glass matrix.

The phenocrysts in the diorite are well twinned, strongly zoned and corroded labradorites. Brown augite forms rare anhedral phenocrysts but dominates a groundmass also containing plagioclase and iron oxide.

Brown and Carmichael (1971) recognised three types of feldspar in the lavas and obsidian bombs: calcic plagioclase phenocrysts, spongy feldspar crystal clusters, and small feldspar laths of the groundmass. The phenocrysts have atypical bytownitic composition which led Brown and Carmichael (1971) to conclude that they are xenocrysts. The sieve-textured crystals range from labradorite to oligoclase, and the groundmass feldspar straddles the albite-anorthoclase compositional range. The only pyroxene phase is weakly zoned diopsidic titanaugite. Zoning is normal in all olivine grains with fayalite content increasing outwards in phenocrysts and still further in any groundmass crystals. Rare richterite and kaersutite were identified and analysed by the same authors. The iron oxide phase proved to be titanomagnetite.

Whole rock analyses (Table 2) confirm the trachytic and dioritic compositions of the lavas and the southern plug respectively. Brown and Carmichael (1971) also quote whole rock analyses of ejecta, lava and obsidian.

## 3.12 UPPER CENOZOIC SEDIMENTARY DEPOSITS OF THE KOOBI FORA BASIN

The discovery of hominid fossils by R. E. Leakey in 1967 (Leakey and others, 1970) along the eastern shoreline of northern Lake Turkana has led to numerous detailed investigations on the host deposits of what subsequently became known as the Koobi Fora Basin. The papers in Part IIIC of 'Geological Background to Fossil Man' (ed. W. W. Bishop, 1978) review much of this work as well as providing comprehensive reference lists (see also Leakey and Leakey, 1978). Thus Vondra and Bowen (1978) described the stratigraphy, sedimentation and palaeoenvironment of the Upper Cenozoic deposits. Findlater (1978) discusses the various tuff marker beds within the sediments. A summary of the geochronological data is given by Fitch and others (1978); the palaeontology documented by Maglio (1972) is reviewed by Behrensmeyer (1978). The following account of the northern part of the Koobi Fora Basin is based on these and later sources. However, significant revision of the stratigraphy and the geochronology has been necessary.

In a recent paper, Brown and Feibel (1986) have revised the stratigraphy to redefine the Koobi Fora Formation as 'sedimentary strata of the Koobi Fora region of Pliocene and Pleistocene age that lie disconformably or unconformably on, or are in fault contact with, Miocene and Pliocene volcanic rocks and/or associated sediments, and are disconformably overlain by the late Pleistocene and Holocene Galana Boi Beds of Bowen and Vondra (1973).' Thus the newly defined unit incorporates the former Kubi Algi, Koobi Fora and Guomde Formations of Bowen and Vondra (1973).

The redefined Koobi Fora Formation is divided into eight members using tuff beds identified by their chemical compositions. The basal five members (Lonyumun, Moiti, Lokochot, Tulu Bor and Burgi) replace the Kubi Algi and Lower Koobi Fora Formations (the 'Lower Koobi Fora Formation' on the accompanying 1:250 000 geological map) and the three upper members (KBS, Okote and Chari) replace the Upper Koobi Fora and Guomde Formations (the 'Upper Koobi Fora Formation' on the accompanying map).

Figure 8 shows the geographical distribution of the Koobi Fora Formation along the east side of Lake Turkana from Allia Bay north to Ethiopia; a strike length of about 80 km. The countryside is broken by low ridges, mesas and cuestas capped by the better consolidated sediments with intervening low ground covered by mantles of Holocene debris (including the Galana Boi Beds). In the map-area, the Ileret ridge and Karari escarpment to an 'Upper' Koobi Fora Formation plateau are the major topographic features, excluding the lava capped Kokoi horst.

Tonal contrasts on aerial photographs between 'Lower' and 'Upper' Koobi Fora Formations, used in conjunction

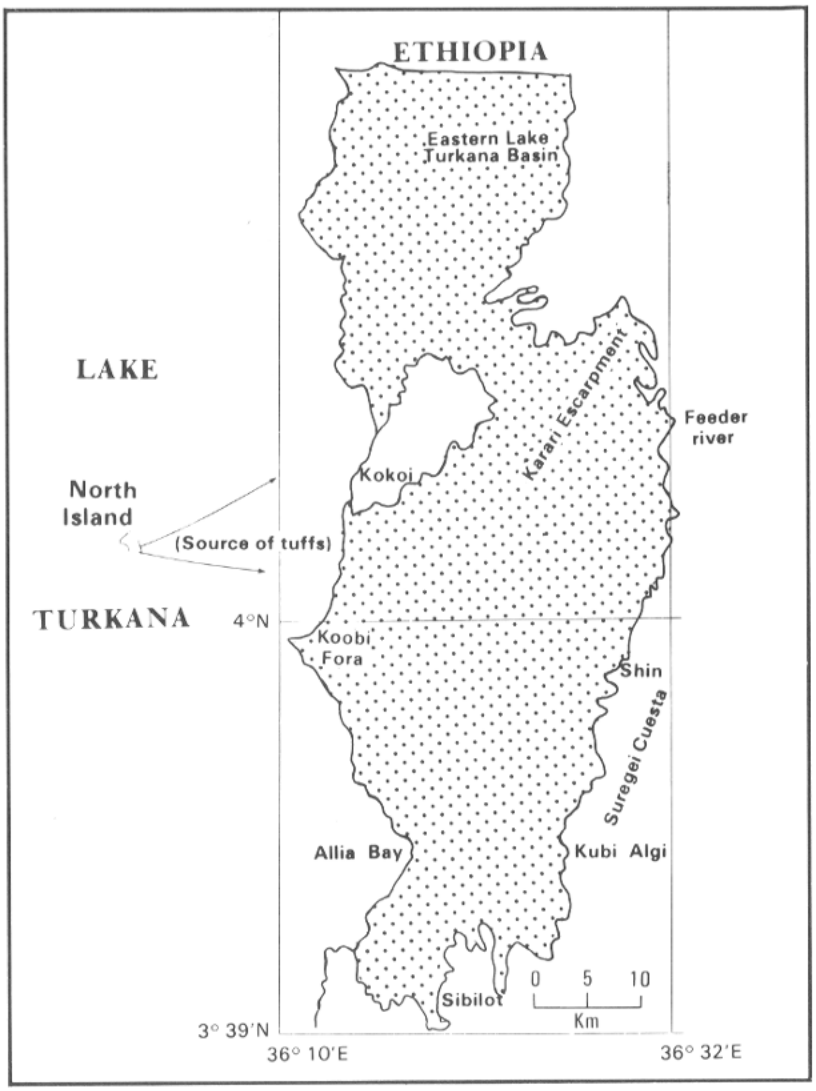



**Figure 8** The full extent of the Koobi Fora Basin in northern Kenya

with the map of Vondra and Bowen (1978) for the area south of Ileret, were used to compile the geology of the Koobi Fora Basin on the accompanying geological map.

Details of type sections are given by Vondra and Bowen (1978) and modified by Brown and Feibel (1986). Type-sections for the uppermost Okote and Chari Members are on the Karari Escarpment and east of Ileret; the lower members are defined to the south of latitude 4°N.

The marker tuff horizons have been dated using various K-Ar isotopic techniques (see Chapter 5) and Figure 9 summarises this data. According to Brown and Feibel (1986) the Koobi Fora Formation spans the time interval from 4.6 to 0.6 Ma.

In the map-area these rocks disconformably overlie both Gombe Group lavas and the Asille Group strata (Plate 15) in the east. However, it has already been noted that some overlap between sedimentation in the Lake Turkana Basin and Gombe Group volcanism took place; (see Section 3.9). The sedimentary rocks generally dip gently westwards except for occasional reversals due to faulting. The principal fault set is NNE, i.e. parallel to the Pliocene faults, with closely spaced ENE fault or tension-fracture related lineaments in the south. A mantle of Holocene debris is widespread on low ground extending 10 km inland from the present lake shore.

The composite stratotype of Brown and Feibel (1986) has a total thickness of 565 m. However, less than 300 m of strata underlies the map area. The Okote Member is 21.6 m thick at its type-area, the Karari Escarpment in a total thickness of about 135 m (according to Vondra and Bowen, 1978). East of Ileret where Vondra and Bowen (1978) record about 210 m of strata, the Chari Member is 42.7 m thick in its type-section.

The Koobi Fora Formation comprises inter-tongued fluvial, deltaic, transitional-lacustrine and lacustrine sediments together with the chemically distinct tuffs. Basal units, extend along the southern margin of the map-area, were laid down on an uneven surface. They consist of coarse-grained, thickly bedded sediments which fine upwards from pebble conglomerates to sandstones to laminated siltstones. Volcanic clasts predominate in the conglomerates. Fossils are rare with ostracods, gastropods, bivalves, abraded mammalian remains and trace fossils. However, basal units exposed along the north-eastern basin margin near the Ethiopian border are mostly fine-grained, lacustrine deposits with diatomites (Watkins, 1986).

Around Ileret, the lower portion of the formation, according to Vondra and Bowen (1978), comprises thick (4 m), loosely consolidated molluscan subarkoses capped by thin (0.1 to 0.5 m) bioclastic carbonates with thin conglomerate and siltstone lenses. These thin towards Karari to interfinger with interbedded thick (3 m) laminated claystones and thin (0.1 to 0.3 m) bioclastic carbonates. Fifty-three metres of intercalated planar cross-bedded

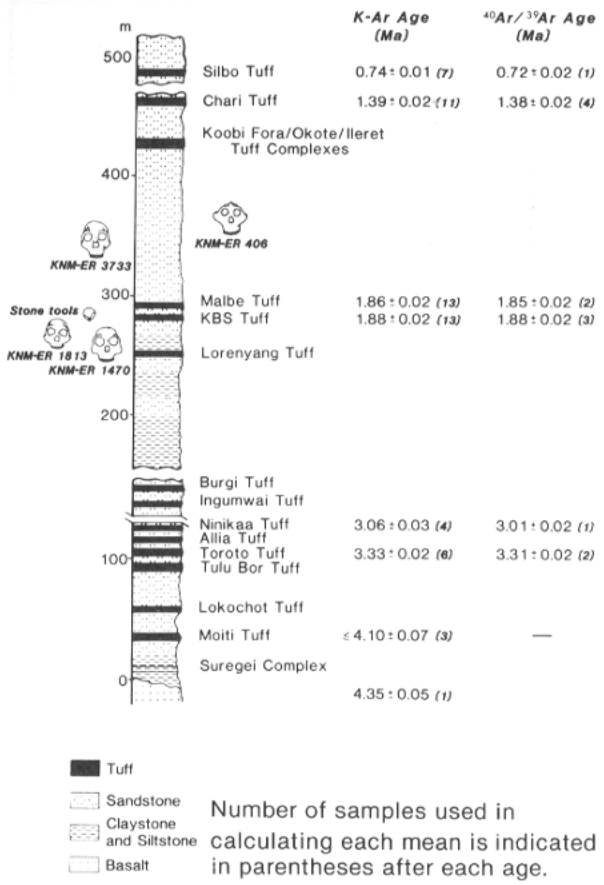



**Figure 9** Generalised tuff stratigraphy in the Koobi Fora region with mean K-Ar and $^{40}Ar/^{39}Ar$ ages shown (from McDougall, 1985)

sublitharenites, and mud-caked siltstones, vertically truncate underlying rocks. The Tulu Bor Tuff separates these rocks from overlying siltstones with sandstone and gastropod lenses succeeded by eight fining upwards cycles of conglomerates and siltstones. Numerous vertebrate fossils include hominid cranial and post-cranial remains.

These lower lithologies are overlain at Ileret (Plate 18) by a sequence of small scale, trough cross-bedded subarkoses with vertebrate fossils overlying thin bioclastic carbonates and molluscan subarkoses. The sequence coarsens upwards into lenticular conglomerates, trough cross-bedded sublitharenites interbedded with tuffaceous siltstones and laminated tuffs. In this coarse upper unit is a laterally continuous, fine-grained, very pale orange subarkose with fish fossils, capped by a thin limestone. The entire sequence contains mammalian fossils with rare hominid fossils and artefacts. At Karari, the upper beds consist of basal channel sandstones and conglomerates which pass upwards into tuffaceous siltstones and tuffs. This sequence contains numerous artefacts and occasional vertebrate fossils. Brown and Feibel (1986) provide detailed successions (their fig. 3).

According to Vondra and Bowen (1978) the topmost beds at Ileret (their Guomde Formation) are characterised by yellowish grey to light olive-brown laminated siltstones and intercalated thin bioclastic carbonates. The equivalent Chari Member of Brown and Feibel (1986) is a major fluvial channel complex dominated by coarse tabular sandstones and claystones and by thin, fining-upward sequences.

Four major lithofacies are recognised by Vondra and Bowen (1978) related to major depositional environments as follows:

1 Laminated siltstone facies (Prodelta or shallow shelf lacustrine).
2 Arenaceous bioclastic carbonate facies (Littoral lacustrine).
3 Lenticular fine-grained sandstone and lenticular-bedded siltstone facies (Delta Plain).
4 Inter-tongued lenticular conglomerate, sandstone and mudstone facies (Fluvial channel and flood plain).

In general the four facies occur in N–S-trending belts which migrated westwards during the overall regression of Lake Turkana. However, three major localised transgressions took place: during the deposition of the Suregei Tuff, after deposition of the Tulu Bor Tuff and, lastly, during the deposition of the Guomde Formation or Chari Member (around Ileret).

Sedimentation rates of between 4 and 47 cm/1000 years have been calculated by Brown and others (1978) within the eastern part of the Lake Turkana Basin. The source area was predominantly the western shoulder of the Lake Stephanie–Kinu Sogo Rifts with sediment transport in several major rivers, i.e. much as today except the rivers are now ephemeral. However, Lake Stephanie may have overflowed through the 'Buluk Gap' into Lake Turkana to have provided a more distant source. Sediment debouching from the Omo River is confined to the northern part of Lake Turkana where it is clearly visible on aerial photographs. Major variations in lake level in the geological past are attributed to tectonic movements rather than climatic conditions although the latter causes short term fluctuations such as the drop in water level of the past few years.

## 3.13 SUPERFICIAL DEPOSITS

Lateral and vertical variations seen in the widespread, derived superficial deposits are caused by the different landforms and their bedrock. For example, alluvium thickness along a single river varies with changing bedrock and channel fill. Similarly, the composition of the alluvium will vary as bedrock changes.

All watercourses are infilled by poorly consolidated *alluvium* which is mostly a stratified sand, silt, gravel mix in which immature coarse deposits are predominant. The larger rivers, e.g. Laga Balal and Il Eriet, have boulder beds and boulder accumulations in coarse uppermost gravels as well as some fine alluvium. These deposits are primarily of volcanic debris (except in the west) reflecting the preponderance of volcanic bedrock. Basalt pebbles are ubiquitous and agate fragments widespread. However, there is also a component derived from sedimentary and/or metamorphic rocks. In the Lake Turkana Basin the majority of watercourses have an upper iron-stained, quartz-rich sand, derived mainly from sedimentary intercalations within the volcanics, the original source having been the metamorphic 'basement' beneath the relatively thin Cenozoic strata. Common 'basement' boulders in the major eastern rivers are washed down from Ethiopia. The marble float in the Garba Merille area also has an Ethiopian source. Thick *colluvial* spreads extend from the lower slopes of the larger hills and mountains, notably Jibisa, Hafari and the Tatesa Hills and along the bases of the Pliocene fault scarps, and the Suregei and Gombe plateaus. These deposits are stratified with boulder beds and lenses in transported soils. They pass laterally into thin residual soils and lag gravels (over plateau basalts). The colluvium around the Tatesa Hills is particularly well

developed and easily distinguished on aerial photographs from the thin superficial debris over the Bulal lavas. The friable bedrock of the Tatesa Hills is easily weathered. Thick basalt scree is ubiquitous at the bases of Pliocene fault scarps but less well developed on hill slopes. All cinder cones have scree of their friable lapilli tuff bedrock. Further breakdown of the breccia scree results in secondary bedded deposits at hill bases (reworked carbonate cemented tuffs). Modern minor *calcrete* deposits are mainly confined within the Sabarei area, particularly in the low ground west of the augite basalt hills. A pale area on the aerial photograph between Jibisa and the Tatesa Hills is thought to be due to surface calcrete and is shown as such on the geological map but was not field checked. Secondary carbonate is common in the Dukana Volcanic Province as a secondary cement to the various pyroclastics. Locally the carbonate cement is the dominant rock component (see above). Clasts of this carbonate-rich material are also found in soil profiles peripheral to cinder cones. Hardpan and friable calcretes with complete and fragmented gastropod shells crop out on the main track south of Sabarei Police Post.

The Kinu Sogo Rift valleys and the eastern *maars* are partly infilled by predominantly argillaceous pan deposits. Evaporation of seasonal surface water gives rise to mud-cracked upper layers. Surface run off from adjacent high ground has introduced a coarser fraction into the pans. Very shallow depressions in the Bulal lava plateau have fine basalt gravel veneers lacking boulders, and relatively thick soils. They occur in the badly drained interfluve areas and are interpreted as pans where surface water has collected in natural depressions. Prevailing ENE winds ensure that all superficial deposits have an *aeolian sand* component. However, significant *aeolian* banks are present only on the windward (SE) side of the Kokoi Horst. The sand is quartz-rich, iron-stained, poorly consolidated and probably less than 5 m thick.

Several types of superficial material mantle the low ground *east of Lake Turkana*. This debris has been produced either through weathering of the underlying sedimentary rocks or by a Holocene transgression (Plate 20) of Lake Turkana. Certain landforms are also made of unconsolidated sediment left by the overall regression of the lake since the Pliocene. Former beaches are marked by standlines parallel to the present shoreline composed of very pale brown sand hummocks up to several metres in height. Adjacent flats are underlain by pale brown clays with a quartz component and a thinly developed surface gravel (intertidal lake bed). Aeolian sand and gravel is banked on windward sides of the low hummocks.

Thin residual, quartz-rich soils cover the higher ground whereas low-lying areas have mantles of argillaceous flood plain deposits and transported, coarser soils. Lag gravels are common (Plates 19 and 21) on upstanding plateau areas but are usually thin, commonly comprising merely scattered pebbles. Basalt pebbles are common over the south-eastern plateaus and quartz pebbles more widespread close to the lake, although Reid and Frostick (1985) show broad basalt spreads issuing from the 'Buluk Gap' to cover former terraces over a wide area.

The present *eastern shore* of Lake Turkana is fringed by a narrow beach composed of fine quartz and heavy mineral sands and clays. The evolution of the beach including the spit adjacent to the rocky headland at Kokoi is described by Reid and Frostick (1985).

The southern extremity of *Chew Bahir* is an ephermeral, desiccated mud flat. The thickness of the lacustrine infill beneath the mudflat is not known, but could be considerable—the median graben defined by the lake has existed since the early Miocene. The low plains which surround and slope gently towards Chew Bahir are covered by various alluvial and colluvial deposits. Most impressive are the thick, coalescing *gravel fans* at the mouths of all drainages from adjacent highlands, especially those entering from the south. The gravels are made up entirely of rounded basalt cobbles and boulders and are clear evidence of former high energy river flow. Towards the salt pan, water-courses lose their individual identity with channels either coalescing or else merging laterally with colluvium. As a consequence, the area peripheral to the pan is underlain by a mixture of arenaceous and argillaceous material, with basalt pebble concentrations. Recent bioclastic deposits on a strand line around the 'lake' indicate late 19th century highwater levels.

*Gravel fans* are found in all major rivers wherever there is a major break of slope. For example, on the south side of Jibisa the steep watercourse in the gully defined by a major fracture spreads into an impressive fan as soon as it reaches the base of the main mountain. All major gravel fans are shown on the geological map.

# 4 Igneous intrusives

## 4.1 JIBISA RING COMPLEX

The igneous complex underlies, and is geographically restricted to, Jibisa Mountain which is about 15 km SE of Sabarei Police Post. Contemporaneous alkaline intrusives are recorded by Davidson (1983) to the north along the eastern shoulder of the Lake Stephanie Rift in Ethiopia. However, Jibisa is unique within northern Kenya. The mountain rises abruptly about 675 m above its own colluvial spread to a height of approximately 1560 m above sea level. It forms a series of arcuate ridges and valleys controlled by concentric bedrock structure. The four principal lithological components have different topographical expressions. The main central ridge is underlain by coarse-grained, oversaturated, peralkaline aegirine syenite or sodic granite with good outcrop including sheer cliff faces on the south and east. Lower, rounded hills flanking the central peaks are underlain by porphyritic microsyenites. An outer ridge of the main mountain is covered by a thick thorn scrub and coarse scree of the underlying diorites and basalts. Erosion of the Sele Gublo Pass has removed this outer suite from the northern flank of the mountain. Relatively low south and west ridges are detached from the main mountain. They are scree-covered and underlain by tholeiitic basalt. Discordant valleys are fault controlled.

Tonal contrasts on aerial photographs distinguish the inner syenites with light tone from the outer diorites and basalts. Similarly it is possible to see tonal differences between the tholeiitic basalts (black) of the detached outermost ridge and the older, more central diorites. Concentric banding within the coarse syenites is visible on aerial photographs, as are ridge-forming microsyenite dykes.

The Jibisa Ring Complex is illustrated by the geological map and section of Figure 10. Exposure is uniformly good and individual lithologies are relatively homogeneous. Consequently any cliff face to the summit ridge is representative of the central syenite. The porphyritic microsyenite is well exposed on the main northern promontory close to the Sabarei to Dukana track. The (chloritised) diorites and intrusive basalts crop out on the detached hill west of the northern salient. The peripheral microsyenite dykes are difficult to access.

K-Ar whole rock ages of 20.7 ± 1.4 Ma and 14.4 ± 0.7 Ma have been obtained from samples of an altered central granite and altered porphyritic microsyenite, respectively (Appendix 1). The outer and older more basic rock types were not dated.

Jibisa is isolated from outcrop of adjacent, mostly younger, rocks by its own colluvial mantle: thus its marginal contacts are concealed. In the north-west an augite basalt lava flow appears to overlie the outermost tholeiitic basalt of the complex. The ring complex is sited on the eastern edge of the Kinu Sogo Rift, and is cut by peripheral rift faults. Its location forms the fulcrum where the faults change orientation from NNE (in the south) to N or NNW (into Ethiopia). On satellite imagery this regional tectonic situation is easily visible: the major bounding fault on the east side of Chew Bahir can be traced into Jibisa. The Miocene age of Jibisa implies that these faults and hence the Lake Stephanie and Kinu Sogo Rifts originated during this epoch. The association of alkaline ring complexes, crustal uplift and rifting is common in Africa (Bowden, 1985).

The temporal relationships of the four major lithologies are shown schematically in Figure 10 based on good field exposures of intrusive contacts (see Plates 23 and 24). The oldest lithology is a chloritised diorite (chlorite is a typical alteration mineral in alkaline ring complexes: Bowden, 1985) which is extensively cut by numerous sills and dykes of relatively fresh basalt. Each diorite outcrop contains sharply discordant basalt intrusions. Both lithologies are truncated by the central syenite-granite complex which is extensively autobrecciated. Late porphyritic microsyenite dykes intrude the ring complex and form a radial dyke swarm. As with the more basic lithologies there is an intimate spatial relationship between the two major felsic intrusive phases: dykes and less well defined zones of the porphyritic microsyenite cut the coarser syenite and sodic granite. The radial dykes intrude all middle Miocene rocks within a radius of about 30 km from Jibisa. The dyke swarm on the southern parts of Hafari and dykes to the south-east at El Yibo radiate from Jibisa. However, the more remote dykes to the north-west are discordant to this pattern and may be related to a different centre or centres in Ethiopia or concealed beneath younger rocks or superficial deposits in northern Kenya. The surface area of the entire Jibisa complex is almost exactly 50 km², made up of 14.6, 12.8, 8.7 and 13.9 km² underlain by the diorite, tholeiite, microsyenite and syenite-granite, respectively. The outer more basic zones are each about 1500 m thick. The central coarse-grained stock has a diameter of about 5 km with the main microsyenite flank about 600 m thick. The high proportion of basic rocks is unusual (cf. Bowden, 1985).

The oldest, chlorite-rich diorite is mostly a massive aphyric, dark grey speckled rock with rounded, black chlorite rosettes several millimetres in diameter. Plagioclase microlites are visible in hand specimen. The rock is locally brecciated into small angular clasts, as well as being cut extensively by the basaltic intrusions. It weathers to a mottled grey-brown colour.

The tholeiitic basalt is a homogeneous aphanitic, dark grey, massive, relatively fresh rock. However, it locally contains secondary chlorite visible in hand specimens as black rosettes.

The central syenite-granite is characteristically a medium- to coarse-grained, strongly-jointed, equigranular, massive pinkish grey rock. Equant, altered feldspar laths, acicular black aegirine crystals and locally, rounded to euhedral, clear and turbid quartz grains are visible in hand specimen. A very wide variety of rock structures and textures are displayed. Brecciated rocks resulting from flow brecciation and subsequent minor

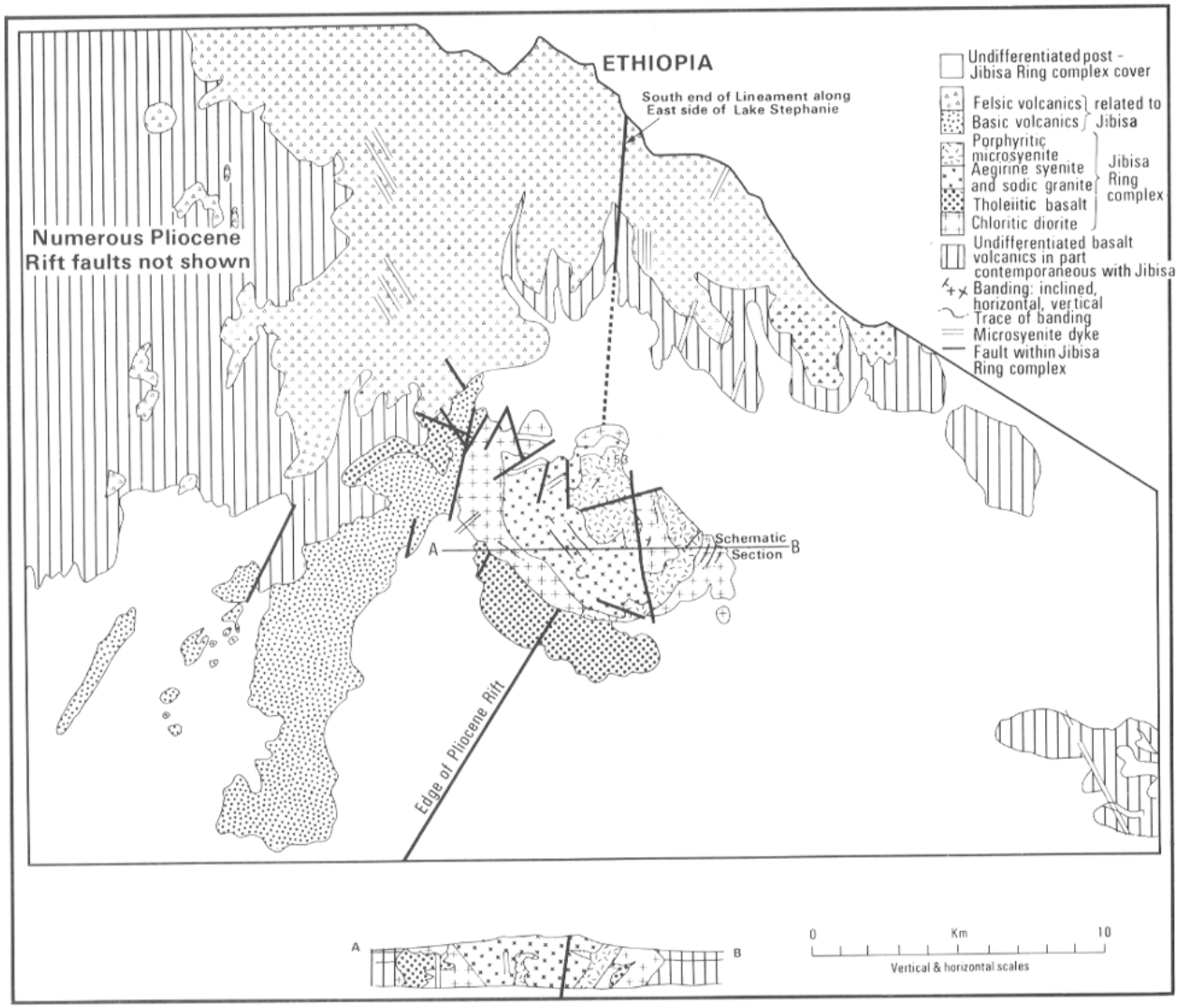



**Figure 10** The geology of the Jibisa Ring Complex

faulting contain angular clasts up to 20 cm across in lesser or greater amounts of a fine-grained pink matrix. Other rocks exhibit banding deformed into asymmetric flow folds. Feldspar occurs as large euhedral crystals in the coarse-grained granites and as euhedral phenocrysts within a fine-grained, grey to pink matrix in the porphyritic rocks. The feldspar crystals have frequently been aligned by flow. The porphyritic rock type grades into the typical equigranular syenite. Rare xenolithic fragments of quartzose rock and basalt of the complex have been incorporated into the autobrecciated syenite-granite, indicating the older age of the basalt. The presence of microsyenite, including porphyritic microsyenite, dykes cutting the syenite-granite stock establishes the relative ages of these lithologies within the complex. A white, friable, easily eroded porphyry is poorly exposed in gullies within the central syenite-granite.

The porphyritic microsyenite (Plate 22) is typically a pale grey, to blue-grey, fine-grained, fissile rock with equant feldspar phenocrysts and minor clots of mafic minerals. Widely spaced feldspar laths are up to 1 cm long and in hand specimens can be seen to be internally altered. Microfractures and closely spaced joints are ubiquitous, and breccia zones are common within the many whaleback and ridge exposures. Rounded xenoliths of the central syenite and of a white, friable rock were noted. The porphyritic microsyenite is locally sheared giving rise to albite-chlorite rich rocks exposed in gullies in the northern outcrop zone.

The microsyenite dykes are homogeneous aphanitic fissile rocks which weather a pink grey colour. Fine feldspar laths are just visible in hand specimens. Individual dykes vary from about 1 to 10 m in thickness.

Thin section examination of all rock types of the ring complex invariably show some alteration to secondary chlorite, carbonate, albite, hematite and unidentified microcrystalline material. The extensive chlorite phase in the diorite appears to be penninite (based on thin section examination). The chlorite rosettes overprint an original phenocryst phase, while individual grains have grown interstitially in the rock matrix. Epidote and/or quartz may form in the rosettes. Strongly altered plagioclase phenocrysts are locally preserved in a matrix of augite, ore and the various secondary minerals.

The basalt is less altered, although secondary carbonate and albitised plagioclase are common. Plagioclase (An40 to An60) and zoned augite or pigeonite microphenocrysts occur in a matrix of clinopyroxene, devitrified glass, opaques, plagioclase, minor quartz, and secondary minerals.

**Table 3** Whole rock analyses of the Jibisa Ring Complex and peripheral volcanics

| | RK4 46 | RK4 89 | RK4 40 | RK3 2 | RK4 74 | RK3 48 | RK4 37 | RK4 25 | RK4 49 | RK4 58 | RK4 70 | RK4 72 | RK4 86 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| $SiO_2$ | 56.00 | 51.80 | 51.00 | 47.90 | 45.70 | 44.30 | 70.20 | 66.20 | 62.90 | 66.80 | 50.40 | 50.50 | 70.00 |
| $Al_2O_3$ | 9.82 | 16.65 | 14.25 | 17.24 | 16.75 | 16.39 | 12.43 | 13.36 | 15.96 | 11.96 | 10.02 | 16.23 | 12.00 |
| CaO | 10.36 | 8.40 | 7.28 | 8.68 | 10.36 | 12.04 | 0.28 | <0.14 | 0.28 | 0.14 | 8.50 | 9.52 | 0.28 |
| MgO | 6.65 | 3.63 | 4.03 | 3.22 | 5.24 | 6.75 | <0.20 | 0.20 | 0.20 | 0.30 | 15.92 | 6.25 | 0.20 |
| FeO | 7.04 | 5.32 | 7.76 | 7.04 | 9.48 | 7.18 | <0.14 | <0.14 | <0.14 | 0.28 | 9.48 | 4.60 | <0.14 |
| $Fe_2O_3$ | 3.61 | 6.16 | 3.56 | 4.78 | 1.53 | 2.91 | 6.79 | 6.71 | 7.14 | 7.26 | 0.71 | 6.57 | 5.75 |
| $Na_2O$ | 2.97 | 2.90 | 2.29 | 4.45 | 4.25 | 3.30 | 7.18 | 7.21 | 7.14 | 5.86 | 1.89 | 2.70 | 6.07 |
| $K_2O$ | 0.48 | 1.39 | 0.78 | 1.57 | 1.45 | 0.78 | 1.27 | 4.16 | 3.68 | 4.46 | 0.42 | 0.84 | 4.24 |
| $P_2O_5$ | 0.27 | 0.34 | 0.42 | 0.64 | 0.76 | 0.70 | <0.01 | 0.16 | 0.17 | 0.20 | 0.16 | 0.27 | <0.01 |
| $TiO_2$ | 1.25 | 2.04 | 1.63 | 2.92 | 2.46 | 3.47 | 0.88 | 0.38 | 0.29 | 0.38 | 1.28 | 1.75 | 0.50 |
| MnO | 0.20 | 0.19 | 0.16 | 0.22 | 0.21 | 0.20 | 0.21 | 0.12 | 0.38 | 0.05 | 0.18 | 0.17 | 0.17 |
| L.O.I. | 1.40 | 2.20 | 6.90 | 0.90 | 1.40 | 1.40 | 1.70 | 1.60 | 1.80 | 2.80 | 0.60 | 1.20 | 1.50 |
| **Total** | 100.05 | 101.02 | 100.06 | 99.56 | 99.59 | 99.42 | 100.94 | 100.10 | 99.94 | 100.49 | 99.56 | 100.60 | 100.71 |

| **Norms** | RK4 46 | RK4 89 | RK4 40 | RK4 74 | RK4 37 | RK4 49 | RK4 70 | RK4 72 | RK4 86 |
|---|---|---|---|---|---|---|---|---|---|
| Quartz | 7.51 | 4.89 | 9.41 | 0.00 | 21.66 | 4.09 | 0.00 | 0.11 | 20.88 |
| Orthoclase | 2.84 | 8.22 | 4.61 | 8.57 | 7.51 | 21.75 | 2.48 | 4.96 | 25.06 |
| Albite | 25.13 | 24.54 | 19.38 | 17.40 | 56.87 | 60.42 | 15.99 | 22.85 | 38.12 |
| Nepheline | 0.00 | 0.00 | 0.00 | 8.92 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 |
| $Na_2SiO_3$ | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 1.36 |
| Anorthite | 12.05 | 28.31 | 26.30 | 22.35 | 0.00 | 0.28 | 17.62 | 29.69 | 0.00 |
| Diopside | 30.89 | 9.26 | 5.92 | 19.83 | 1.20 | 0.00 | 18.83 | 12.98 | 1.19 |
| Hypersthene | 14.85 | 14.75 | 21.09 | 0.00 | 4.81 | 6.58 | 27.46 | 22.18 | 4.99 |
| Olivine | 0.00 | 0.00 | 0.00 | 10.70 | 0.00 | 0.00 | 11.67 | 0.00 | 0.00 |
| Aegirine | 0.00 | 0.00 | 0.00 | 0.00 | 3.57 | 0.00 | 0.00 | 0.00 | 7.55 |
| Magnetite | 2.17 | 2.90 | 2.17 | 2.90 | 1.91 | 3.62 | 2.17 | 2.17 | 2.90 |
| Ilmenite | 2.37 | 3.87 | 3.10 | 4.67 | 1.67 | 0.55 | 2.43 | 3.32 | 0.95 |
| Apatite | 0.64 | 0.81 | 0.99 | 1.80 | 0.01 | 0.40 | 0.38 | 0.64 | 0.01 |
| Corundum | 0.00 | 0.00 | 0.00 | 0.00 | 0.00 | 0.13 | 0.00 | 0.00 | 0.00 |
| Water | 1.39 | 2.19 | 6.88 | 1.37 | 1.70 | 1.79 | 0.59 | 1.19 | 1.50 |
| Norm Total Error | 0.00 | 0.00 | 0.00 | 1.13 | 0.15 | 0.00 | 0.00 | 0.00 | 1.08 |
| Thornton-Tuttle Index | 35.48 | 37.62 | 33.40 | 34.89 | 86.03 | 86.26 | 18.48 | 27.92 | 85.42 |

| | | |
|---|---|---|
| RK4/46 Diorite | Jibisa Ring Complex | *(Diorite) |
| RK4/89 Tholeiitic basalt | Jibisa Ring Complex | (Tholeiitic basalt) |
| RK4/40 Tholeiitic basalt | Jibisa Ring Complex | (Tholeiitic basalt) |
| RK3/2 Augite basalt | | (Hawaiite) |
| RK4/74 Augite basalt | | (Olivine basalt) |
| RK3/48 Augite basalt | | (Olivine basalt) |
| RK4/37 Sodic granite | Jibisa Ring Complex | (Granite) |
| RK4/25 Porphyritic microsyenite | Jibisa Ring Complex | (Syenite) |
| RK4/49 Porphyritic microsyenite | Jibisa Ring Complex | (Syenite) |
| RK4/58 Microsyenite dyke | | (Syenite) |
| RK4/70 Olivine augite basalt, Tatesa Hills | | (Tholeiitic basalt) |
| RK4/72 Basalt, Tatesa Hills | | (Tholeiitic basalt) |
| RK4/86 Pink rhyolite, Hafari rhyolites | | (Rhyolite) |

* Name in parenthesis based on Cox and others (1979) alkali vs $SiO_2$ plot.

The dominant mineral in the central syenite-granite is orthoclase or sodic orthoclase, which forms equant, subhedral laths up to about 3 mm in length. Individual laths are variably replaced by albite (Plate 22). Simple twinning and microperthite are common, but have been largely obliterated by secondary clay alteration. Aegirine is the main mafic phase and hematite the only iron oxide. Quartz crystals form clusters, or occur with orthoclase and twinned albite laths, to define an equigranular texture. Secondary minerals, notably clay together with carbonate and limonite, are common, especially in brecciated syenite.

The friable porphyry exposed in gullies is an albitised porphyritic microsyenite. Corroded orthoclase phenocrysts, partially pseudomorphed by albite, lie in a matrix of altered feldspar, clay, limonite, hematite, minor quartz and rare apatite.

Phenocrysts in the porphyritic microsyenites are orthoclase and plagioclase; in one case the phenocrysts have labradorite cores rimmed by microcline. Albite, orthoclase, hematite and apatite form the fine-grained groundmass with abundant replacive carbonate, chlorite and limonite. Protoclastic albite-rich rocks of shear zones in this unit have felted or aligned equant albite laths with minor hematite as the only other phases. In the albite-chlorite rocks well-twinned albite laths and sparse acicular

aegirine crystals are set in a matrix of acicular chlorite, hematite and albite.

The microsyenite dykes have phenocrysts of variably altered orthoclase and rare nepheline in a groundmass of altered orthoclase, hematite and a mafic phase, usually aegirine. Phlogopite is a common accessory and interstitial quartz was seen in one thin section. Carbonate and limonite are the main secondary minerals. The rocks have trachytic or massive textures.

Seven major element analyses of rocks from the Jibisa Ring Complex are presented in Table 3. A diorite (RK/4/46) was relatively free of secondary minerals. The high calcium content is reflected in the modal augite. Two basalt analyses (RK/4/40 and RK/4/89) plot is the field of tholeiites on a $SiO_2$ against alkali plot (Figure 5 after Cox and others, 1979). Both samples are highly altered with secondary carbonate and chlorite, reflected in their high 'Loss-On-Ignition' values. The high sodium contents of the porphyritic microsyenite (RK/4/49), aegirine syenite (RK/4/25) and a microsyenite dyke (RK/4/58) are reflected by modal albite, aegirine and sodic orthoclase, whereas potasium is confined within orthoclase. Low phosphorus values reflect trace apatite contents, and the high oxidation ratios are in keeping with the presence of hematite as the sole iron oxide.

A single analysis of a quartz bearing sodic granite (RK4/37) plots well within the granite field of Figure 5. Its high $Na_2O$ content is reflected in the dominance of sodic plagioclase laths in the groundmass.

On an A'F'M diagram (Figure 6) the analyses, together with those of related extrusive phases, show an alkali-basalt fractionation path with the oldest diorite intrusive and an olivine-basalt from the adjacent Tatesa Hills being the least evolved, and the various felsic rocks being the most highly differentiated magma. The following facts are pertinent to any discussion on the genesis of the Jibisa Ring Complex.

1 It is the root zone of the volcano from which the voluminous Hafari rhyolites (pantellerites) and the basaltic lavas of west Jibisa were probably erupted.

3 If these extremes of magma types are genetically related, considerable magmatic differentiation must have taken place.

3 The complex has an intracontinental setting intruded into a rigid crust capable of accommodating a radial dyke swarm. Such an anorogenic setting is that of all alkaline granite (A-type) complexes in Africa (Bowden, 1985).

4 The complex has a middle Miocene age, at least for the later felsic intrusions, and is part of a chain of alkaline centres extending north into Ethiopia.

5 Considerable autobrecciation has occurred during intrusion of the felsic rocks and extensive secondary alteration has occurred producing carbonate, albite, chlorite and hematite.

6 The regional tectonic setting is one of Jibisa located at the intersection of two major fault systems, both related to Miocene rifting.

It is believed that the Jibisa Ring Complex is the root of a major volcano whose extrusive products contributed greatly in Miocene times to the volcanic piles on the eastern shoulders of the Lake Stefanie and Kinu Sogo rifts. It is noted that contrasting compositions present in this one complex cover the diverse compositional range represented in the Asille Group volcanic succession, on the western shoulder of the Lake Stefanie Rift. The intersection of two fundamental fracture lines on the eastern shoulders provided a locus for magma ascent at Jibisa.

The considerable economic potential of the alkaline Jibisa Ring Complex and its peripheral extrusives is discussed in Chapter 8.

## 4.2 ANYAKOLIB VENTS

Watkins (1983) described three intrusive pipe-like bodies (Figure 11) situated on the south-west extremity of the Asille Plain close to the steep north-west edge of the Suregei Plateau. The pipes, which are closely spaced in this poorly accessible area, are located on major faults defining an extremely narrow ENE–WSW downfaulted zone within Asille Group strata. They are roughly circular in plan and small in diameter (less than 1 km), except that Gumuti Dura appears to result from the emplacement of two partially overlapping pipes (Figure 11). Gumuti Dura (Plate 25) and Gum Anyakolib are easily recognised features on aerial photographs. The third pipe, Gumuti Alfara, is indistinct having only a small segment of elevated rim.

The pipes are mid-Miocene in age and at their highest levels intrude Il Yia Formation sediments and the lowermost basalt lavas of the Nakwele Formation (Biliet Member). They are no older than 15.8 ± 0.2 Ma since they pierce ignimbrites of the Gum Dura Formation (McDougall and Watkins, 1985; in press). The more precise of two $^{40}Ar/^{39}Ar$ analyses of fresh basaltic glass from tuffisite filling the Gumuti Dura pipe yielded an age of 15.6 ± 0.5 Ma (Fitch and others, 1985). Pipe emplacement was closely contemporaneous with the ENE–WSW faulting.

The pipes are infilled by fragmental tuffisite consisting of abraded juvenile alkaline basaltic glass and xenolithic material. The pipes are interpreted by Watkins (1983) as having penetrated to the surface forming explosion craters, the original mouths of the vents coinciding with the present level of excavation by erosion. Upstanding rims of banded, massive or brecciated siliceous and carbonate rocks are interpreted as hydrothermal sinters and travertines. Well-bedded, often finely laminated, fine-grained carbonates and rarer cherts dip inwards from the rim rocks forming thin carapaces over the tuffisite, and are similarly considered to be the result of precipitation from hot-spring waters. The carbonate rocks are dolomites. Chemical analyses of these and other capping rocks are given in Watkins (1983).

According to Watkins (op. cit.) the pipes were formed by the rapid gas-fluidised emplacement of a volatile and carbonate-rich alkali basalt magma. Adiabatic cooling of the fluid magma on pressure release caused instantaneous chilling to glass. The glass was fragmented and abraded during emplacement, $CO_2$ being a principal fluidising agent. Intersection of shallow chambers of $CO_2$-enriched magmas by deep faults is a most likely control of this rare type of local magmatism. Interaction of groundwater and magma may have also been a significant factor in generation of the fluidised systems.

## 4.3 MINOR INTRUSIONS

The feeder dyke to the Tatesa Hills basalts and the North Island diorite plug have been discussed in Sections 3.6 and 3.11 respectively.

Numerous small intrusions (dykes and plugs) cut Asille Group strata (Watkins, 1983). The largest of these is a columnar jointed, porphyritic olivine basalt stock in the Gum Dura Formation in the area of the Anyakolib Vents

at GR 2824 9832. Basalt dykes have pronounced NNE trends, i.e. parallel to the Pliocene Rift faults, and are interpreted as feeder dykes for the Gombe Group lavas or possibly late Nakwele Formation basalts. A single rhyolitic tuff pipe cuts the Nakwele Formation at GR 2824 9807. In plan it is 380 × 120 m with a ENE long axis and vertical margins.

Fault planes cutting the Asille Group may be coated with chalcedony. Secondary carbonate similarly covers fault and fracture surfaces in the more easterly volcanics.

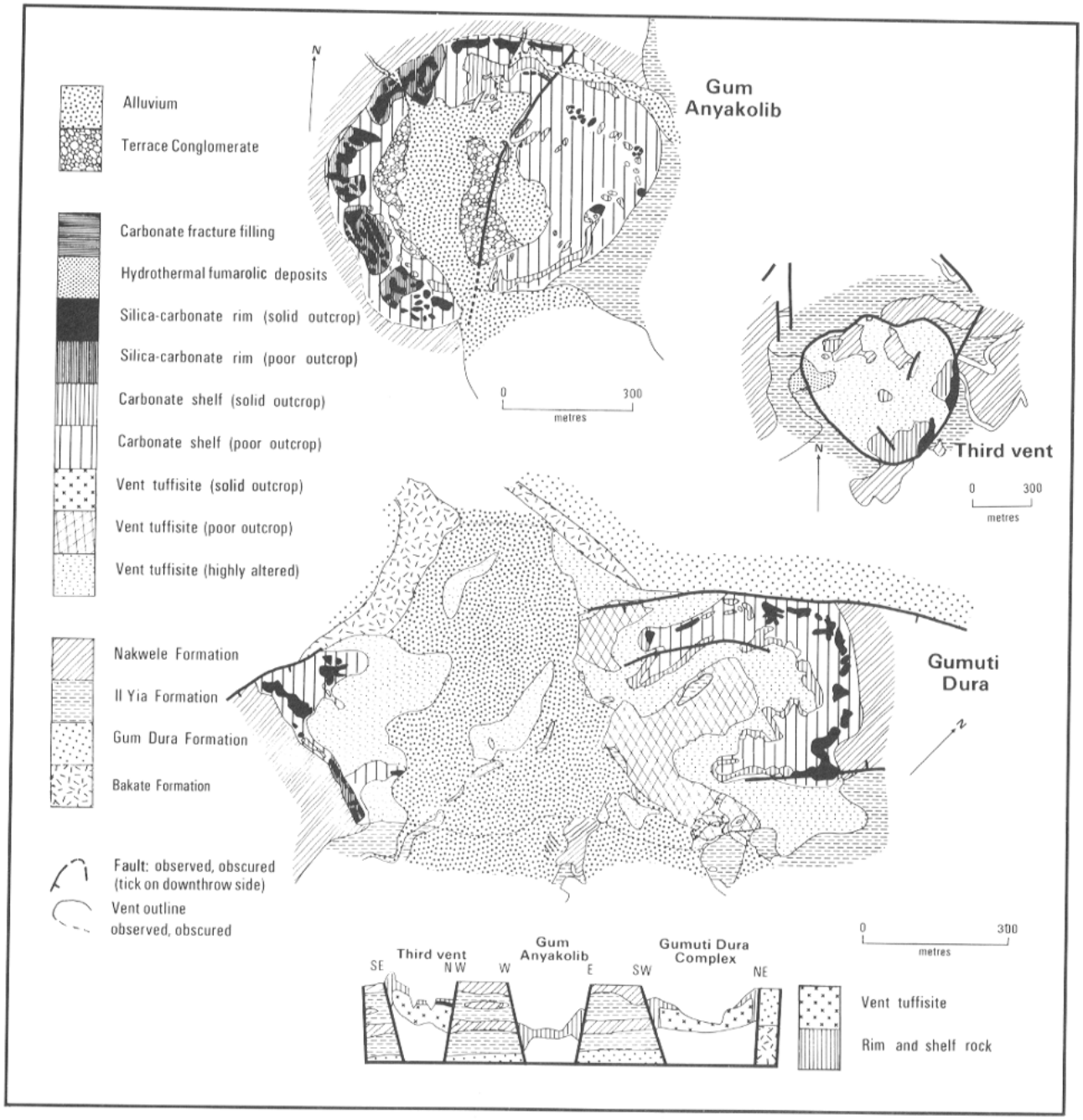



**Figure 11** The geology of the three Anyakolib vents

# 5 Geochronology

The Pliocene–Pleistocene succession of the Koobi Fora Basin has been afforded considerable geochronological study on account of the important vertebrate fossils, including hominids, and archaeological artifacts recovered from the fluvial and deltaic sedimentary rocks. Older volcanic rocks forming the upland margin of the basin to the north-east have also been the subject of major programmes of isotopic age analysis. Results of these studies are summarised in this account. In contrast, no previous radiometric ages are available for units exposed in the central and eastern parts of the map area. Six new conventional K-Ar whole-rock ages on volcanics from this area (Appendix 1) enable broad chronostratigraphic correlation with the well-dated succession to the west.

The first isotopic ages of rocks forming the margin of the Koobi Fora Basin were reported by Fitch and others (1974). These indicated that basalt lavas and pantelleritic rocks subsequently assigned to the Asille Group belong to the Miocene, whereas the extensive flat-lying basaltic lavas of the Gombe Group were Pliocene in age. In 1975, Fitch and others published a $^{40}Ar/^{39}Ar$ isochron age of 16.66 ± 0.10 Ma and a near-isochron age of 16.58 ± 0.13 Ma* for two welded ignimbrites constituting the middle member of the Gum Dura Formation, thus establishing a most useful chronostratigraphic marker. The above results and others totalling 44 samples have recently been reported by Fitch and others (1985) in a summary of a reconnaissance dating programme for the Suregei–Asille district undertaken between 1970–1981 which employed various dating techniques (conventional K-Ar, $^{40}Ar/^{39}Ar$ total fusion, and $^{40}Ar/^{39}Ar$ age spectrum) on both whole-rock samples and crystal separates. The authors proposed a provisional time-scale for the complete succession in this north-east segment of the Koobi Fora Basin margin, according to which, Cenozoic volcanism began in late-Oligocene times, around 28.2 ± 0.7 Ma. It persisted sporadically but without major break until 11 to 12 Ma ago. The unconformity separating the Asille and Gombe Groups forms a chronostratigraphic gap of at least 5 million years. Outbreak of flood lava eruptions is estimated to have taken place around 5.9 Ma ago in the very late Miocene. The reconnaissance nature of the study, in which frequently only single samples of widely spaced strata were analysed, yields only imprecise ages for individual units within the provisional time-scale, as is evidenced by inconsistencies in the age progression of dated samples when compared to stratigraphic sequence. The application of conventional K-Ar whole-rock analysis to basalt lavas, in particular, appears unreliable, partly as a result of the extreme difficulty in collecting samples which are entirely free of minor secondary alteration.

McDougall and Watkins (1985; in press) have used conventional K-Ar analysis of alkali feldspars separated from pantelleritic pyroclastic rocks to calibrate the Asille Group more precisely. Consistency in the results obtained from multiple samples of closely spaced strata from individual pantellerite sequences provides confidence that these are accurate estimates of the ages of crystallisation of the ignimbrites and tuffs. The dated pantellerite sequences, which in each case appear to be the result of short-lived periods of pyroclastic activity, yield ages of 26.9 ± 0.3 Ma (Langaria Formation), 15.8 ± 0.2 Ma (Gum Dura Formation) and 13.0 ± 0.1 Ma (Gurro Member/Sibilot Formation). The age of the Gum Dura ignimbrites is slightly younger than that given by the age-spectra of Fitch and others (1975, 1985).

The Buluk Member and neighbouring strata have been the subject of a detailed K-Ar dating study designed to constrain the age of the hominoid fossils occurring within fluvial sediments at the Buluk site (Harris and Watkins, 1974; McDougall and Watkins, 1985; Leakey and Walker, 1985; Watkins, in press). The fluvial redeposited air-fall tuffs in the upper part of the member yield a K-Ar anorthoclase feldspar age of 16.4 ± 0.2 Ma, whilst two samples of plagioclase feldspar from a minor basalt lava flow interbedded with the fluvial terrigenous sediments give concordant K-Ar ages of 17.2 ± 0.2 Ma. The age of the Buluk Tuffs is supported by fission track studies of zircon which yield a mean crystallisation age of 16.2 ± 0.6* Ma (Hurford and Watkins, 1987).

An interpretation of available age data relating to the rocks of the Asille Group is shown in Figure 12.

Numerous attempts have been made to date tholeiitic basalt flood lavas of the Gombe Group (Fitch and others, 1974, 1985; Hooker and Watkins, 1980; McDougall, 1985; McDougall and Watkins, in press). However, the fine-grained aphyric lavas are generally unsuitable for precise K-Ar age analysis since they invariably contain a glass mesostasis. The interstitial glass which contains a significant proportion of the potassium content of the basalts is prone to hydration and secondary alteration and only very rarely can samples be collected that do not contain at least a small amount of secondary mineraloid or calcite. McDougall (1985) has suggested that even perfectly fresh, isotopic glass in the lavas is an insecure reservoir for radiogenic argon. Consequently, available K-Ar whole-rock ages of Gombe lavas can best be interpreted as minimum cooling ages. $^{40}Ar/^{39}Ar$ step-wise degassing experiments on three samples from the Harr Formation reported by Hooker and Watkins (1980) produced only disturbed age spectra. Available ages on the flood lavas range from a maximum of 6.75 ± 0.35 Ma to 2.25 ± 0.45 Ma, although the latter age, after Fitch and others (1985), is almost certainly 'too young': a date of approximately 3.8 Ma frequently recorded from the flood lavas at the edge of the Koobi Fora Basin is probably a more realistic minimum age for the last products of the Gombe magmatism. Fitch and others (1985) reported wide variation in ages of Gombe lavas, including a number in excess

* These and other quoted ages are corrected using the K decay constants and $^{40}K$ abundance recommended by the IUGS Subcommission on Geochronology (Steiger and Jager, 1977).

* Analytical error calculated as standard error on the mean; cf. that of K-Ar analyses based upon one standard deviation.

of 6 Ma, although they conclude that the 16 dated lava samples may have been produced by two major flood eruptions occurring around 4.9 and 3.9 Ma. However, the recording of similar maximum ages from samples collected from the southern part of the Koobi Fora Basin margin and the Kokoi horst by McDougall (1985) sug-

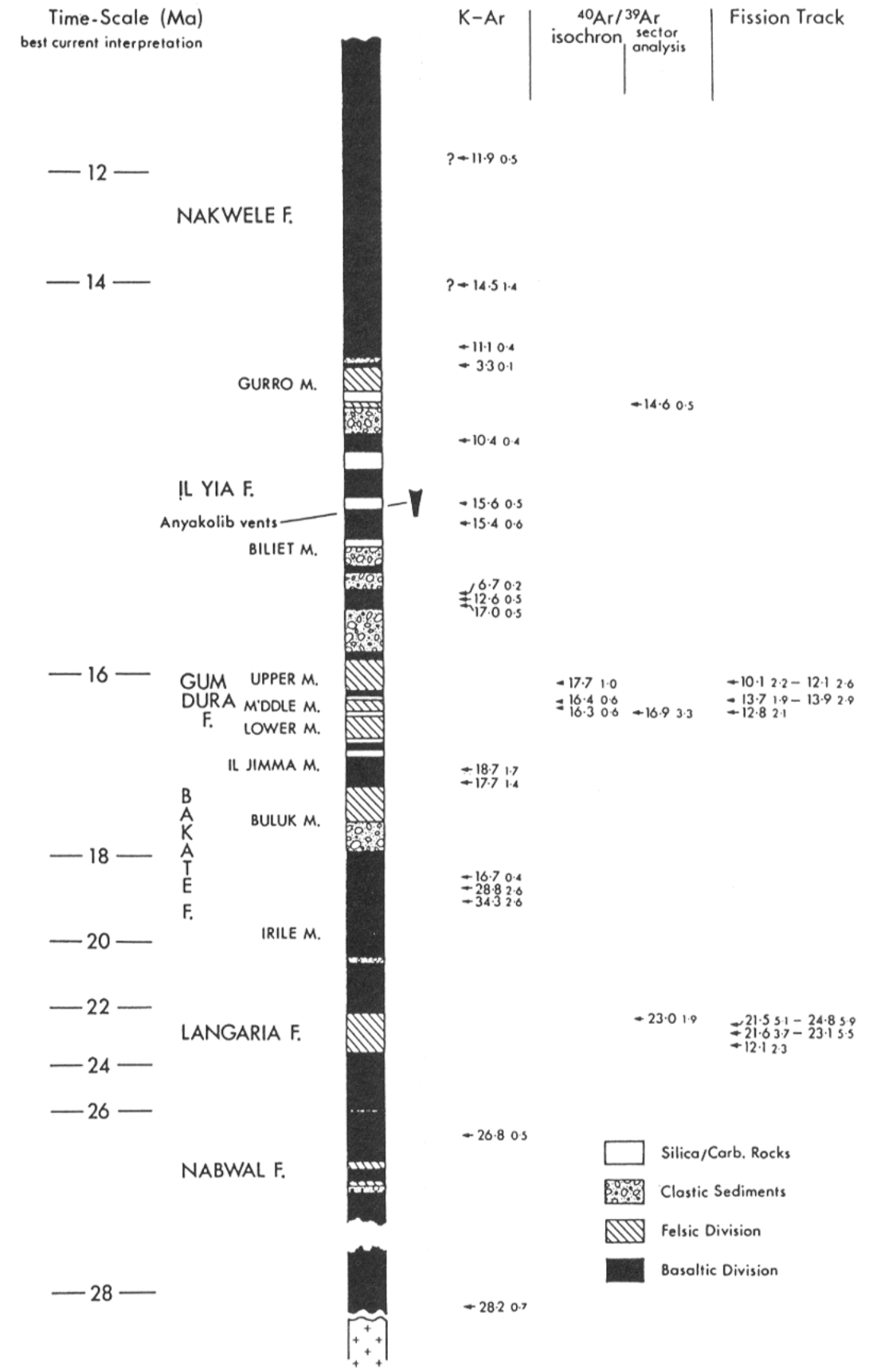



**Figure 12** A summary of the geochronology of the Asille Group

gests that the outbreak of tholeiitic volcanism took place sometime before 6 Ma ago, in the late Miocene.

The sequence of fluvial, deltaic and lacustrine sediments of the Koobi Fora Basin has a thickness in excess of 500 m (Brown and Cerling, 1982). The potential importance to chronological studies of numerous rhyolitic tuff beds, each between a few centimetres and about 2 m thick, present within the sequence was recognised by Behrensmeyer (1970). Whilst the tuffs have been transported in water, this reworking took place soon after the air-fall eruptions (Findlater, 1976, 1978), and in geological terms the age of eruption and final deposition are the same. Bases of the tuff beds are interpreted as isochronous surfaces. Consequently, the age of a fossil or stone tool may be bracketed by the dates recorded from the nearest underlying and overlying tuffs. Curtis and others (1978) and Drake and others (1980) measured ages on volcanic glass of pumice cobbles from the KBS Tuff. However, considerable problems of argon retention arise in the dating of glass (Cerling and others, 1985) and most age analyses have been performed on alkali feldspar crystals contained within pumice clasts in the tuffs, using either convential K-Ar or $^{40}Ar/^{39}Ar$ methods.

Despite the neatness of this approach to calibrating the sedimentary succession, a considerable degree of controversy arose with regard to both stratigraphical and geochronological interpretations of the basin sequence. Difficulties were particularly apparent in the dating of the KBS Tuff (of importance since this unit lies a short distance above the site of hominid skull KNM-ER 1470): the age reported by Fitch and Miller (1970, 1976) of around 2.6 Ma being significantly greater than that assigned to the unit by Curtis and others (1978) of approximately 1.8 Ma or less. Further difficulties became apparent with the inability to equate evolutionary traits of invertebrate faunas (Williamson, 1982) and certain vertebrates, notably the Suidae (Cooke and Maglio, 1972; White and Harris, 1977; Harris and White, 1979), with stratigraphical interpretations and the tuff chronology of Fitch and Miller.

The stratigraphic problems were eventually resolved by geochemical 'fingerprinting' studies of Brown and Cerling (1982) which revealed previous serious miscorrelations of tuff units. It is now known that at least 40 tuffaceous beds are present within the Koobi Fora Basin compared to the much smaller number (amounting to seven major tuff units) recognised by previous workers (Bowen and Vondra, 1973; Bowen, 1974; Findlater, 1976, 1978). The problem of the age of the KBS Tuff was resolved when consistent and concordant ages on anorthoclase from pumice cobbles were obtained by both the K-Ar and $^{40}Ar/^{39}Ar$ dating methods (McDougall, 1985). A detailed programme of analysis of feldspars from major tuff beds recognised from the amended stratigraphy of Brown and Cerling was reported by McDougall (1985). The resulting chronology is reproduced in Figure 9. According to this time-scale, sediments of the Koobi Fora Formation of the basin began accumulating less than 4.3 Ma ago (age of the underlying Gombe lavas) with the first tuff bed, the Moiti Tuff, having an age of about 4.1 ± 0.07 Ma. The uppermost tuff, the Silbo Tuff of the Guomde Formation, yields an age of 0.74 ± 0.01 Ma, early/middle Pleistocene. Interestingly, McDougall points out that as much as half of the time span of the sedimentary succession is unrepresented in the stratigraphic column, there being three hiatuses of sedimentation each of between 0.5 to 0.7 Ma.

The results of early fission track age analysis of volcanic glass from the tuffs (Hurford and others, 1976) have been invalidated by the realisation that very low degrees of heating, such as that developed through surface insolation, are sufficient to anneal the fission tracks (A. Gleadow, *pers. comm.*). However, subsequent analysis of zircon has produced a mean fission track age for the KBS Tuff of 1.87 ± 0.04 Ma (Gleadow, 1980) which is indistinguishable from the K-Ar feldspar age of 1.88 ± 0.02 Ma* of McDougall (1985).

Some efforts have been made to define a palaeomagnetic reversal stratigraphy for the Koobi Fora Basin (Brock and Isaac, 1974, 1976; Hillhouse and others, 1977). However, attempts to match the magnetostratigraphy with the geomagnetic polarity time-scale and thus provide an independent age calibration have not generally been successful, there being a paucity of long sections comprising lithologies suitable for recording reversals especially in the region above the KBS Tuff where the majority of vertebrate fossils are found.

In the light of evidence from other dating studies, the whole-rock conventional K-Ar analyses of six volcanic units sampled during reconnaissance mapping of the map-area must be interpreted as providing only imprecise ages of crystallisation (Appendix 1). They are useful nevertheless in providing broad age comparison of the rocks with those studied in greater details. Sample RK/4/72, a basalt lava from the Tatesa Hills, has a mid-Miocene age equivalent to that of basalt lavas of the Nakwele Formation. The ankaramite (RK/4/63) has an age equivalent to that of the petrographically similar lavas of the Biliet Member (Fitch and others, 1985). The compositionally diverse sequence of the Sabarei volcanics therefore can be tentatively correlated with that part of the succession to the west containing the Gum Dura to lower Nakwele Formations. Neither of the ages from units of the Jibisa Ring Complex, 20.7 ± 1.4 and 14.4 ± 0.7 Ma, matches that of any pyroclastic pantellerite sequence exposed on the margin of the Koobi Fora Basin. However, the ages obtained from single samples of the Nakwele Formation (11.91 ± 1.6 Ma) and Bulal lavas (3.61 ± 0.30 Ma) correspond to existing data. As expected, the North Island volcanics proved too young to be dated accurately by K-Ar isotopes (Appendix 1).

* See previous note on the different methods of calculating the analytical error.

# 6 Structure and metamorphism

## 6.1 PRE-CENOZOIC TECTONOTHERMAL HISTORY

Davidson (1983) provides an overview of the tectonothermal history of the Precambrian rocks of southern Ethiopia (Hamar Domain) which extend into the area of Garba Merille. An early Granulite Facies metamorphism accompanied a regional upward-opening fold fan. Subsequently regional retrogression to Amphibolite Facies grade was due to westward-directed overthrusting with related folding and imbrication. Late upright folding produced a regional, steeply dipping, N to NNW trend of planar fabrics, which is prevalent in northern Kenya. Lineations plunge uniformly at about 10° towards 355° parallel to recumbent fold axes generated during the thrusting, i.e. the lineations are normal to the movement directions on the thrust planes. A tight, upright fold is evident in the map-area from aerial photographs. Its surface trace is shown on the geological map.

Regional retrogressive metamorphism isolated zones retaining intact Granulite Facies assemblages. One such zone underlies the general area of Chew Bahir in Ethiopia. Its bounding isograds trend southwards towards Kenya (fig. 6 of Davidson, 1983) with the Garba Merille area on the Amphibolite Facies side. Gneissic xenoliths within the pyroclastic rocks of Dukana contain hornblende as the major mafic mineral. No hypersthene relicts are preserved, and the basement concealed beneath the Dukana Volcanic Province probably is also of Amphibolite Facies.

## 6.2 OLIGOCENE THROUGH MIDDLE MIOCENE TECTONICS

The map-area straddles the major Chalbi–Habaswein topographic depression between the Kenyan and Ethiopian domes delineated by the sub-Miocene erosion surfaces (Saggerson and Baker, 1965). Development of the domal upwarps and the formation of linear segments within the East African Rift System (Baker and others, 1972) were correlated; the Gregory and Ethiopian Rift Valleys dissecting the central portions of the domes along their N–S axes. However, towards the edges of the domes the rifts bifurcate. A lateral offset between the two domes is matched by a similar offset of the zones of rifting (Figure 13). Rift faulting of the Lake Stefanie Rift is considered by Baker and others (1972) and Davidson (1983) to have commenced in the late Oligocene or early Miocene. Evidence is given below to support this and also contemporaneous faulting of the Kinu Sogo Rift which effectively linked the main axial rifts on the Kenyan and Ethiopian domes. It has been suggested that Lake Turkana may similarly have originated as a result of initial, Oligocene rift faulting (Wilkinson, 1988). However, the wide lateral extent of lacustrine sediments preserved beneath the Gombe lavas (Watkins, 1983) indicates that by late Miocene times there was no significantly deep, rifted basin of Lake Turkana.

The western shoulder of the Lake Stephanie Rift is composed of Asille Group strata dipping at about 6°SW except where modified by faults (Watkins, 1983). Monoclinal and open fold axes trend NNW parallel to the rift and reflect vertical fault movement in the underlying metamorphic basement. The monoclines are west-facing (downwards); they appear longitudinally continuous and are reminiscent of similar structures flanking the Gregory Rift, e.g. the Samburu Flexure of Shackleton (1946).

The inner part of the western shoulder is cut by major N-trending faults including the major fault interface with the median graben. The major faults downthrow to the east although outcrop of the Langaria Formation in a peripheral graben is bounded on its east side by a fault downthrowing to the west. The amount of movement on the steeply inclined fault planes is not known. The interface between the western shoulder and median rift is rectilinear or serrated owing to the dextral offset of the N-trending faults by broadly contemporaneous ENE faults (Moore and Davidson, 1978). This is a feature of the whole length of the western side of the Lake Stefanie Rift (Moore and Davidson, 1978) and may be related to dextral strike-slip movement along the length of the rift (cf. Kinu Sogo Rift, below) as well as regional E–W tension.

Small-scale faults are prolific on the outer parts of the western shoulder (Watkins, 1983). All have steep hades; most are normal, but occasionally faults show reversed and/or a degree of strike-slip motion. Early faults ( pre-12 Ma) mostly trend NE and are cut by Pliocene faults trending NNE. There is local 'basement' control on individual fault orientation but the main influence may be an interplay between regional topography (i.e. flank of the Ethiopian dome) and the regional stress system.

The eastern shoulder of the Lake Stephanie Rift is also composed of stratified volcanics of comparable age to the Asille Group. The Hafari rhyolites dip gently ESE; monoclines are absent. Faults are concentrated close to the shoulders' inner margin and are again of several generations with NNE-trending Pliocene faults cutting older faults (N- to NNW-trends). Jibisa Ring Complex is located at the intersection of regional faults of the older trend and faults aligned SSW along the eastern edge of the Kinu Sogo Rift. The early to middle Miocene age for the ring complex indirectly suggests a similar or older age for the controlling structures. This agrees well with the conclusions of Watkins (1983) that much of the faulting on the western shoulder was older than about 12 Ma. Davidson (1983) concluded that the faulting in the Lake Stephanie Rift in Ethiopia occurred from 13 to 5 Ma.

Both shoulders of the Lake Stefanie Rift are approximately 30 km wide whilst the median graben is about 20 km wide. These dimensions are comparable with the Miocene extent of the Gregory Rift suggested by King

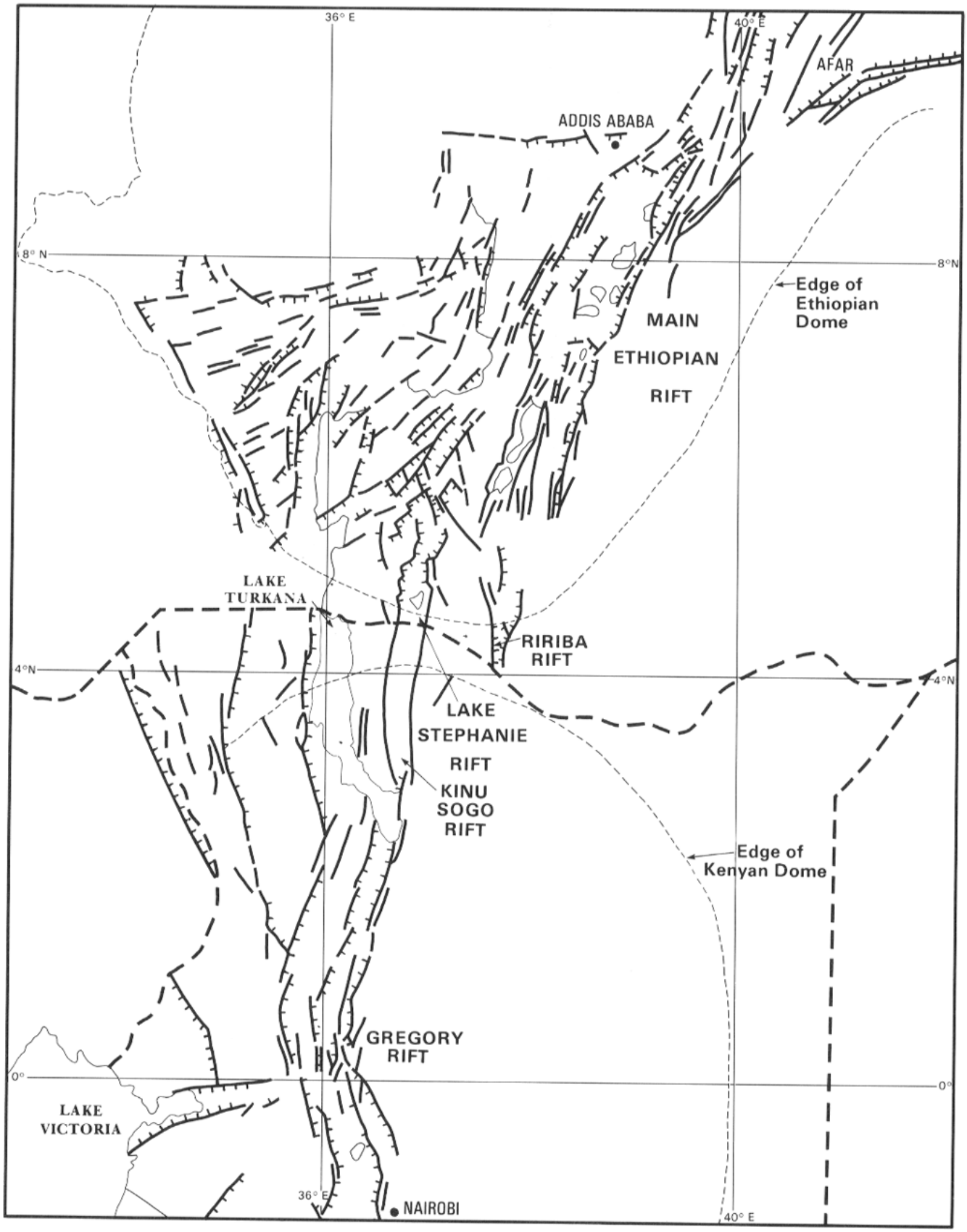



**Figure 13** The regional setting of rift valleys to Ethiopia and Kenya (modified from Davidson, 1983)

(1978). Total vertical displacements on the faults bounding the median graben are not known: topographically the eastern scarp is the more spectacular (Chapter 2). Davidson (1983) reconstructed the elevation of the Cenozoic volcanics-metamorphic basement boundary in southern Ethiopia, confirming the presence of a precursor dome to the Lake Stefanie Rift. He suggested vertical displacement of 1650 and 1100 m on the west and east sides, respectively, of the median graben. Moore and Davidson (1978) noted that throws of individual faults varied longitudinally, i.e. rotation of displacements about poles to fault planes. This caused individual faults to terminate and might explain the southern termination of the median rift itself. The middle Miocene Nakwele Formation straddles the floor of the median rift across the topographic high at the south end of Lake Stephanie. Thus it is clear that no lateral spreading across Lake Stephanie Rift has taken place since the middle Miocene.

The location of Jibisa suggests that the Kinu Sogo rifting started in the early Miocene. However the major tectonic elements presently apparent are Pliocene faults, described in the next section. Similarly, a mantle of post-Miocene basalt lavas covers a large part of the map-area outside of the Lake Stephanie Rift concealing any Miocene structures. The Tatesa Hills currently have a NNE- to NE-trend, subparallel to an exposed feeder dyke to the component basalts. This trend is thought to be a primary feature, the Tatesa Hills basalts forming part of the middle Miocene shoulder of the NNE-trending Kinu Sogo Rift.

## 6.3 PLIOCENE TECTONICS

Rift-related faults are the main Pliocene tectonic elements and are concentrated within the Kinu Sogo–Lake Stephanie and Ririba Rifts. The faults cut Miocene to Pliocene flood basalts of the Gombe Group but predate the Plio-Pleistocene shield volcanoes to the south (Wilkinson, 1988). They are therefore probably of middle

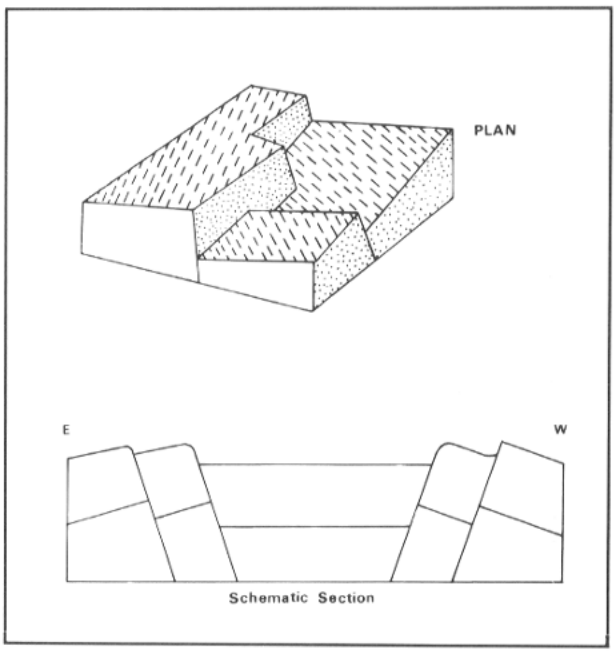



**Figure 14** Trap door faulting in the Kinu Sogo Rift (after the model of Moore and Davidson, 1978)

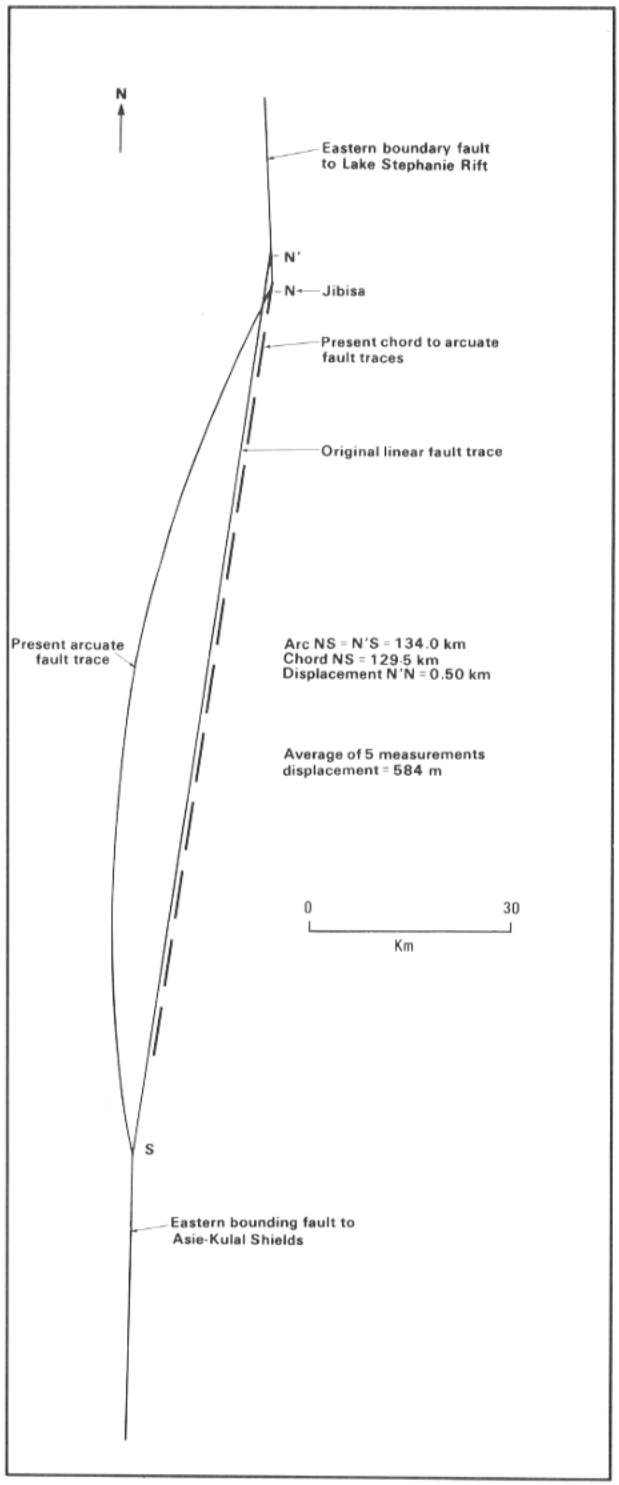



**Figure 15** A solution to the arcuate faults of the Kinu Sogo Rift

Pliocene age and may be contemporaneous with upper units within the younger flood basalts.

The Kinu Sogo Rift is 30 km wide in an ESE direction, and consists of five shallow grabens separated by four horsts. The marginal western and eastern grabens are topographically the most impressive, with well defined and longitudinally continuous scarp margins. The central grabens have low (less than 50 m high) serrated scarp margins, the NNE-trending bounding faults being offset by shorter ENE faults. The central grabens are longi-

tudinally discontinuous suggesting that, in this case, too there are variations in the amount of downthrow along the lengths of the bounding faults (cf. Moore and Davidson, 1978 on Miocene faults in the Lake Stephanie Rift). Intervening horsts are either horizontal plateaus or tilted blocks with the strata rotated towards the steeply inclined fault-planes—the 'trap-door effect' (Figure 14) of Moore and Davidson (1978). Vertical throws on individual faults could not be ascertained precisely but are not large, judging from their gentle topographic expressions. The western bounding fault has produced the highest scarp face, of the order of 100 m, which probably reflects the actual downthrow (eastwards). The faults have steep hades to produce gently sinuous surface traces on a local scale. Individual NNE faults are up to 140 km in length with arcuate traces (Figure 15). Contemporary ESE faults are rarely more than a few km in length. The tilting of horst blocks has produced a keystone effect centred on the grabens. Across the Kinu Sogo Rift there is a close density of about one NNE-trending fault per kilometre. Despite this abundance of faults, elevations of both shoulders and of the central horsts is constant at about 1000 m.

The NE-trending Buluk Fault, which effectively marks the northern limit of the major NNE faults, is delineated by several basaltic centres and must be a fundamental fracture. It downthrows to the NW and is unusual in transecting a Miocene median rift in the African Rift System (see Key, 1987a). The origin of the arcuate rift faults is discussed in Chapter 7.

Downwarping of the northern Lake Turkana (Koobi Fora) Basin commenced in the middle Pliocene, i.e. synchronous with rift faulting 12 km to the east. The eastern margin of the basin was a gentle slope on which accumulated less than 200 m of Pliocene sediments on a surface gradient of less than 0.001. No major contemporaneous faults reached surface on the eastern side, although there was major fault movement on the west side of the basin.

The Bulal lava plateau east of the Kinu Sogo Rift is tectonically undisturbed except for fractures parallel to the NNE rifting. In the extreme east of the map-area the plateau lavas are cut by the gentle western marginal faults to the narrow Ririba Rift. This Rift is about 10 km wide and strikes N–S to disappear beneath the younger Huri Shield (Charsley, 1987).

## 6.4 QUATERNARY TECTONICS

Sporadic downwarping of the Lake Turkana Basin throughout the Quaternary has progressively deepened the axial trough. The present reduced aerial extent of the lake is largely a result of this tectonic process. Nevertheless, despite its linearity, the Lake Turkana Basin is a relatively shallow topographic feature, indicating the possible thick accumulation of sediments under the present lake. Its eastern half currently has westerly gradients of about 0.02. No prominent controlling N-trending faults reach surface to the east although there is a major fault marking the western limit of the basin. Minor N-trending faults on the eastern margin of the basin are confined to the south of the map area. The fault along easting 2937 downthrows eastwards with its scarp controlling Holocene sedimentation.

Faults within the Lake Turkana Basin mostly trend NNE. The largest are the bounding faults to the Kokoi horst with throws of about 120 m (Vondra and Bowen, 1978). There is a ENE-trending belt of closely spaced lineaments in the area south-east of Kòkoi, each lineament being about 1 km long. These lineaments are interpreted as tension fractures related to crustal arching, and possibly heralding the start of an ENE rifting.

The Quaternary Dukana Province has an internal radial network of fractures and small faults (see geological map). Gentle arching of the upper crust in response to the rise of basaltic magma is possibly responsible for this fracture pattern.

The angular discordance between the fundamental (deep) N-rifting of the Lake Turkana Basin and the apparently more shallow NNE-rifting may reflect different tectonic regimes. Epicentral data from Khan (1985) indicate that central Lake Turkana is currently seismically active, suggesting that this is a northern, albeit deflected, continuation of the Suguta Trough structure. The Kinu Sogo Rift zone is not active at present, but clearly was in the Pliocene: it now remains as an *en echelon* link between the Gregory and the Ethiopian Rift Valleys.

# 7 Geological history

The map-area is founded on metamorphic rocks of the Mozambique Orogenic Belt*, although these have only limited outcrop in the study area. The geological history of the northern extension of the Garba Merille outcrop into Ethiopia, established by Davidson (1983) is essentially identical to the geological history of the Mozambique Orogenic Belt in north-central Kenya (Charsley and others, 1984). In this general region the orogenic belt consists of upper Proterozoic supracrustals deposited within an ensialic setting with subsequent high grade deformation continuing into the early Palaeozoic. An initial metamorphism, increasing in grade northwards from upper Amphibolite to Granulite Facies, imparted a ubiquitous gneissic fabric (defining an initial upward-opening fold fan in Ethiopia). Regional west-directed (in Ethiopia) and east-directed (in Kenya) thrusting and related folding under Amphibolite Facies conditions juxtaposed different lithologies including slices of possible oceanic crust (Key, 1987a). Later more localised, upright folding and shearing established a series of tectonic domains. The Garba Merille area is part of the Hamar Domain (Davidson, 1983) equivalent to the Barsaloian of Charsley and others (1984). Crustal melt granites of minor aerial extent were generated during the early high grade metamorphic events.

Cambro-Ordovician mineral ages (Cahen and others, 1983) time the crustal uplift marking the end of the Mozambiquian Orogeny. The next geological event (Table 4) preserved in the map-area is the late Oligocene inception of Cenozoic rift-related volcanism. Little is known about the geological record of the intervening 400 Ma. Regional geophysics (Ministry of Energy confidential data) indicates that the map-area flanked the major Mesozoic marine-sedimentary basins in north-east Kenya.

Three distinct periods of Cenozoic volcanism can be recognised. The oldest, late Oligocene to middle Miocene, phase is associated with contemporaneous rifting (Lake Stephanie and Kinu Sogo Rifts). The volcanics define the rift shoulders. Volcanism recommenced at the end of the Miocene to continue into the middle Pliocene forming extensive flood basalts east of the Lake Turkana Basin. Finally, Quaternary volcanism was confined to specific centres: North Island (trachytic) and Dukana volcanic provinces (basaltic), again associated with rift tectonism.

The oldest Cenozoic volcanics are probably the basal (Nabwal Formation) basalts to the Asille Group forming the western shoulder of the Lake Stephanie Rift. The oldest basalts on the east shoulder, may be of comparable, late Oligocene age. Volcanism with intermittent fluvial sedimentation continued in the areas of both shoulders until about 12 Ma BP. The western shoulder is composed largely of basaltic rock whilst the eastern shoulder is dominantly felsic (pantelleritic) in character. A total thickness of less than 1500 m of Cenozoic rock underlies both shoulders.

Away from the Lake Stephanie Rift the extensive Mio-Pliocene lavas conceal older strata. However, in the Laga Balal gorge there are exposures of basal hawaiites (El Yibo lavas) which underlie the c.12 Ma Tatesa Hills basalts. The lateral extent of the El Yibo lavas is unknown; they may represent a basal unit of the eastern shoulder of the Kinu Sogo Rift.

Volcanism ceased around 12 Ma, and there was a volcanic hiatus until approximately 6 Ma when the flood basalts of the Bulal lavas and Gombe Group started to mantle the general area east of the developing Lake Turkana Basin. Contemporaneous basaltic pyroclastic centres were confined to major fault zones, notably the Buluk Fault. The plateau lavas had feeder dyke sources with dominant NNE-trends parallel to the Pliocene rifting.

Quaternary volcanism in the Dukana Volcanic Province is now extinct: the products are entirely basaltic and this province is similar, albeit on a smaller scale to the several basaltic shield volcanoes east of the Gregory Rift Valley in northern Kenya (Key, 1987b). The still-active North Island trachytic volcano is part of a chain of multicentred trachytic volcanoes confined to the Quaternary median graben of the Rift Valley (Key, 1987a). The association of basaltic/diorite rock of the southern plug with the trachytes is typical.

Accumulation of lacustrine and fluvio-deltaic sediments associated with 'proto' Lake Turkana commenced prior to the major outpourings of Gombe flood lavas. Repeated regression and transgression of this early lake, indicated by the interdigitation of thin lacustrine beds and the subaerial lava flows, may be explained by volcanotumescence and deflation accompanying the voluminous tholeiitic basalt eruptions (Watkins, 1986). Subsequently, in excess of 500 m of fossiliferous lacustrine, deltaic and fluvial sediments have been deposited in the Koobi Fora Basin over a period of around 4.3 Ma (Brown and Cerling, 1982; McDougall, 1985). Controversy exists over the origin of the numerous tuff beds within the sedimentary pile. Most recent authors, including Watkins, prefer a distal source based on tuff chemistry (cf. Brown and Feibel, 1986). However, Key notes that pyroclastic interclations are common in the sedimentary infill of the Quaternary median graben of the Gregory Rift Valley to the south which are derived from local large trachytic/basaltic centres such as Paka and Silali (Key, 1987a). Therefore it is tempting to suggest that the numerous tuffs of the Koobi Fora Formation were similarly derived from the central trachytic/basalt volcanoes in the lake. Against this idea is the pantelleritic composition of the tuffs which do not match that of rocks exposed on the islands.

Watkins (1983) preferred separate geneses for the three contrasting magma types erupted in the Suregei–Asille

* The median graben of the Lake Stephanie Rift may be underlain by young crust; the fault grid on its western side in Ethiopia could, in part, be due to E–W tension which may have 'opened' the rift to form new crust.

**Table 4** Summary of the geological history of the Sabarei area

| Sedimentation | Volcanism | Intrusion | Tectonics/ metamorphism | Regional setting | Group/ Formation | Age | Economic content |
|---|---|---|---|---|---|---|---|
| Unconsolidated superficials | | | NNE- to N-trending faults; reactivation of older faults | Generally widespread | | Holocene | Clay, sand, gravel, water concentrations |
| Reworked volcanic debris, calcrete | Basaltic lavas and pyroclastics around cones and *maars* | | Radial fracture system | Basaltic shield peripheral to rifts | Dukana Volcanic Province | Quaternary | Aggregate, surface water concentrations |
| Fluviatile, deltaic and lacustrine sediments | Multicentred trachytic volcano, thin tuffs | Dolerite | Normal fault control on W side, NNE- and N-trending faults in E. ENE tension joints | Lake Turkana Basin | Koobi Fora F North Island trachytes | Pliocene to Recent | Fossils inc. hominids, solfataras, water, gravels, clay, minor evaporites, gold |
| Minor lacustrine sediments | Flood basalts and hawaiites, pyroclastic centres | Feeder dykes | Faults within Kinu Sogo and Ririba Rifts | Plateau basalts to east of Quaternary Lake Turkana Basin | Gombe Group<br>Bulal lavas | Late Miocene to middle Pliocene | Aggregate, agate, groundwater, fossil wood |
| Minor volcaniclastic sediments, evaporites | Interlayered basic and acid volcanics from alkali basalt, mantle source | Feeder dykes, vent tuffisites, minor basaltic cones, Jibisa Ring Complex | Faults within Kinu Sogo and Lake Stephanie Rifts | Shoulders of the Kinu Sogo and Lake Stephanie Rifts | Tatesa Hills basalts, Hafari rhyolites, Sabarei volcanics, West Jibisa augite basalt, Asille Group, El Yibo lavas | Late Oligocene to middle Miocene | Fossils, groundwater, springwater, iron, agate, gold, aggregate. Potential for uranium in Hafari rhyolites and apatite etc. in the alkaline Jibisa Ring Complex |
| Ensialic sediments, including limestones | | Minor felsic crustal melts | Polyphase high grade tectonothermal history | Mozambique Orogenic Belt (Hamar Domain) | | Upper Proterozoic to Lower Palaeozoic | Marble |

district. Alkali basalt magma formed by relatively small-scale melting of the upper mantle incident upon a rising geothermal cell. Slow ascent of the magma with frequent ponding enabled differentiation and resulted in the characteristic eruption of small batches of varied alkali basaltic lavas. Larger degrees of melting at shallower depths in the upper mantle produced the source magma for the voluminous tholeiitic lavas of the Nabwal Formation. After subsequent differentiation tholeiite magma rose rapidly from large reservoirs utilising major rift fractures. The rapid ascent precluded significant shallow crystal fractionation. The lavas are characteristically of near-uniform composition and represent a very high rate of extrusion. Flood lavas of the late-Miocene to Pliocene Gombe Group are of equivalent composition and are interpreted as forming due to a renewed increase in geothermal gradient after a volcanic hiatus which terminated the first magmatic episode. A probable origin of felsic magmas involved partial melting of lower crustal rocks: pyroclastic pantellerites of closely similar compositions have erupted at intervals between 26 (Langaria Formation) and 13 Ma ago (Sibelot Formation).

The limited chemical data (Table 3) on the volcanic rocks of the eastern shoulder of the Lake Stephanie and Kinu Sogo Rifts also suggest an alkali basalt magma source. The various analyses, including those from the Jibisa Ring Complex, define an alkali basalt differentiation path on a A'F'M diagram (Figure 6). Early basalts are again tholeiitic flows. Jibisa Ring Complex is thought to be the eroded root zone of one source volcano for the east shoulder material. Basaltic lavas flowed south-west (augite-basalts) presumably down a regional slope: aerial rhyolitic fragmental deposits were wind borne northwards.

Extensive secondary alteration is general in all rocks of the alkaline Jibisa Ring Complex as well as including

regional hematite growth.

Tectonism responsible for Cenozoic rifting can also be separated into three phases contemporaneous with the three periods of volcanic activity. There is also a spatial link between tectonism and volcanism. The Lake Stephanie and Kinu Sogo Rifts developed in the late Oligocene or early Miocene with faulting continuing until the middle Miocene. Middle Pliocene rifting or downwarping took place in (from west to east) the Lake Turkana Basin, Kinu Sogo Rift and the Ririba Rift. Vertical movements in all three cases were less than in the older rifting. However, an important dextral strike-slip component is indicated from the geometry of the Kinu Sogo Rift (Figure 15). Quaternary rifting is confined to the Lake Turkana Basin (asymmetric graben with major faulting confined to its western edge): the major, older faults of the Lake Stephanie Rift in Ethiopia remain seismically active (Gouin, 1979).

Uncertainty exists regarding the exact timing of faulting within the Lake Stephanie and Kinu Sogo Rifts. The location of the early Miocene Jibisa Ring Complex at the intersection of the eastern bounding faults to these rifts suggests a comparable or older age for the faults. Similarly the fundamental contrast between the volcanics of opposite shoulders of the Lake Stephanie Rift might imply an early existence of a separating median graben. Watkins (1983) concluded that a majority of the pre-Pliocene faults affecting the Asille Group in the Suregei–Asille district are older than 12 Ma. The post-16 Ma Hafari rhyolites on the west shoulder are weakly faulted again implying an older age for the extensive fault network in the underlying Sabarei volcanics.

The subsequent developments of the Lake Stephanie and Kinu Sogo Rifts have been described in Chapter 6. They are an integral part of the East African Rift System (Figure 16) and during the Miocene represented the northern extremity of the Gregory Rift. The Miocene morphology of the Kinu Sogo Rift is concealed by late Miocene–Pliocene flood lavas. However, the Lake Stephanie Rift and the Gregory Rift of central Kenya are morphologically similar: c.20 km wide median graben flanked by c.30 km wide shoulders with steep, normal faults locally defining major scarps between the shoulders and median graben. Volcanic and tectonic activity effectively ceased in the Lake Stephanie Rift in the middle Miocene (after the median rift was infilled by basalts of the Nakwele Formation). The Pliocene Buluk Fault transects the southern end of the Lake Stephanie Rift.

By contrast the Kinu Sogo Rift, after being infilled and mantled by late Miocene–Pliocene flood basalts, was reactivated in the middle Pliocene. A series of shallow horsts and grabens developed in the basaltic carapace and the regional arcuate-curvature of the fault traces is interpreted as due to oblique right-lateral strike-slip displacement along the rift. Originally the faults are thought to have had linear traces trending NNE parallel to the overall trend of the Kinu Sogo Rift. The morphology of the Pliocene Kinu Sogo Rift may be likened to a sigmoidal mega-tension gash. Warping of fault traces by the strike-slip motion (Figure 15) suggests a total displacement of c.580 m since the middle Pliocene (a movement rate of 0.16 mm per year). The western bounding faults of the

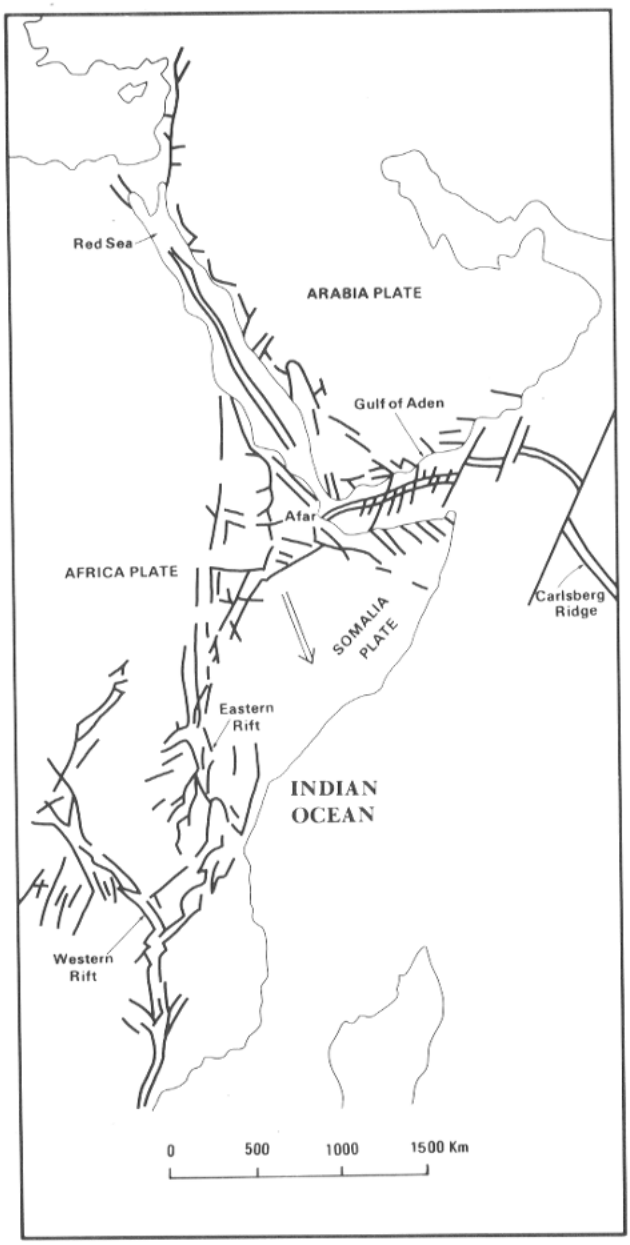



**Figure 16** Structural pattern of the Afro-Arabian rift system, showing the eastern and western rifts of East Africa, and the mean movement vector of the Somalia Plate

Ririba Rift just cut the eastern extremity of the map-area. They downthrow to the east and are also shallow-seated.

In all three Pliocene rifts the westernmost faults seem to be dominant with downthrows to the east. Certainly the Lake Turkana Basin is not marked by major surface faults on its eastern side: the downwarping on this side appears to be a passive response to normal faulting along the western shoreline of Lake Turkana.

A radial fracture pattern within the Dukana Volcanic Province is related to localised crustal uplift caused by ascent of basaltic magma within a confined area of the crust.

# 8 Economic geology

The principal objective of Phase 2 of the Samburu Marsabit Project was to produce 1:250 000 geological maps of previously unsurveyed areas of North Kenya. Detailed investigations for mineral deposits were not part of the brief. However, the mapping has delineated 'targets' which warrant follow-up investigations. In Sheets 3 and 4 the major 'target' is the Jibisa Ring Complex and its related extrusive volcanic rocks.

## 8.1 LOCAL NEEDS

The nomadic lifestyle of the local Gabbra has no need of direct exploitation of surface minerals or rocks. Plentiful gravel spreads and stream sediment deposits readily meet the construction requirements of the residents (mostly police) at Sabarei, Buluk, Kokoi and Ileret. Clay deposits (for bricks or ceramics) are widespread at the south end of Chew Bahir and within the Lake Turkana Basin. A small fishing community is gradually establishing itself on the northern beach-head of North Island. The solfataras on this island could meet a local demand for ochre and, in the long term, the area might support small-scale exploitation of geothermal energy.

## 8.2 NATIONAL RESOURCES

Worldwide, one in twenty alkaline ring complexes supports some form of mining (R. L. Johnson, *pers. comm.*), so these bodies statistically are most attractive exploration targets. Base, precious and rare-earth element concentrations and various industrial minerals (notably phosphates and magnetite) are presently being mined from such ring complexes. Therefore, despite its remoteness, the Jibisa Ring Complex warrants detailed ground investigations.

**Table 5** Gold values of selected whole rock samples

| Sample number | Gold value (ppm) | Comment |
|---|---|---|
| RK/NI/8 | 0.17 | Diorite plug on North Island |
| RK/4/89 | 1.12 | Tholeiitic basalt, Jibisa |
| RK/4/37 | 0.30 | Sodic granite, Jibisa |
| RK/4/70 | 0.03 | Olivine basalt, Tatesa Hills |
| RK/4/72 | 0.03 | Hawaiite, Tatesa Hills |
| RK/4/74 | 0.28 | Augite basalt, Jibisa |
| RK/4/89 | 1.04 | Pink rhomb porphyry, Hafari rhyolites |
| RK/4/49 | — | Porphyritic microsyenite, Jibisa |
| RK/4/25 | — | Aegirine syenite, Jibisa |
| RK/4/46 | — | Porphyritic microdiorite Jibisa |
| RK/4/58 | — | Microsyenite dyke, Jibisa |

Note: detection limit is 0.02 ppm. Analysis is by Geosurvey International Ltd, Nairobi.

The phosphate-bearing rocks in particular, are susceptible to weathering and invariably underlie low, poorly exposed terrain. Preliminary reconnaissance mapping commonly misses these deposits. In terms of its size, Jibisa is comparable to Phalaborwa: a major producer of copper and phosphate, with subsidiary gold, silver, platinoids, nickel and iron (Coetzee, 1976). The best, and most practical exploration technique for Jibisa would be soil geochemistry based on a detailed grid over the whole complex, together with surface geophysics and rock chemistry. Bowden (1985) notes that, in West Africa, economic uranium deposits are located in extrusive felsic volcanics peripheral to alkaline ring complexes. Therefore the Hafari rhyolites should be investigated for radioactive mineralisation as well as for gold*. Finally, the secondary iron horizon peripheral to Jibisa is laterally extensive but highly variable with regard to its hematite content. Detailed mapping and extensive sampling is required to assess the economic importance of the iron mineralisation. Kenya presently lacks an internal source of iron.

The mineralogy of the tuffisites cutting the Asille Group is described by Watkins (1983). There is no indication of deep-source mineralogy, such as garnet or diamond.

The sedimentary infill of the Lake Turkana Basin is famous for its vertebrate fossil content including hominids. The scientific importance of the fossils has long been recognised (Leakey and others, 1970). There is also an economic aspect through the sale of replicate fossils (casts), as well as the attraction of tourism. Evaporite and gypsiferous beds are too thin to be of economic importance.

Agate, including 'blue lace' and petrified wood float are common in the area underlain by Asille Group rocks and the Sabarei volcanics. However, these deposits and concentrations of baryte crystals (in the Asille Area) are probably too small to be of economic importance.

Regional geophysical studies (Ministry of Energy, confidential data) indicate that no Mesozoic sedimentary basins underlie the map-area. Therefore the potential for oil concentrations beneath the Cenozoic volcanic cover is not promising. No oil seepages have been recorded along any of the major faults in the map-area. The only area worthy of oil exploration is under Lake Turkana where considerable thicknesses of sediment may have accumulated.

## 8.3 BULK STREAM SEDIMENT SAMPLES

Thirteen bulk stream sediment samples were collected:

* Gold is present in detectable quantities in seven of ten assayed whole rock samples from Jibisa and related extrusive phases (Table 5). Values of up to 1.12 ppm (grams/tonne) Au emphasise the potential economic importance of these alkaline rocks and the necessity for detailed mineral exploration. A value of 0.17 ppm Au was obtained from a whole rock sample of the North Island diorite plug. Therefore this plug and the area of solfatara alteration also deserve further exploration.

five from rivers within the Lake Turkana Basin, two from rivers within the Bulal lavas, four from drainage channels off Jibisa, and two from drainages off Hafari. The samples were collected to provide regional geochemical information and their analyses are presented in Table 6. No discussion of the data is attempted because of the small number of samples involved.

**Table 6** Analytical data on stream sediment samples

| Element | Ag | Ba | Co | Cu | Fe | Li | Mn | Mo | Ni | Pb | Sr | Zn |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Highest value | n.d. | 350 | 47 | 53 | 9.0 | 6 | 1640 | n.d. | 138 | 11 | 84 | 169 |
| Lowest value | | 80 | 14 | 6 | 4.1 | 1 | 460 | | 10 | <1 | 32 | 47 |

1 Values in ppm except Fe (in percentage)
2 n.d. not detected.

# 9 Water resources

Despite the low mean annual rainfall (less than 400 mm/year, Figure 17) and the lack of perennial surface water away from Lake Turkana, there is still sufficient spring and well water to satisfy present demands of migrant Gabbra pastoralists and the small resident communities in the map-area (see Keith-Jones, 1984).

## 9.1 SURFACE WATER

The only permanent surface supply is the alkaline water (Walsh and Dodson, 1969) of Lake Turkana, with a relatively high fluoride content, but suitable for livestock and domestic human needs. The southern end of Chew Bahir is a mud flat which holds water only after heavy rains. Similarly all watercourses have seasonal flow following rains. The Laga Balal and its tributaries have particularly impressive gorges testifying to former strong perennial flows from the Ethiopian highlands. Local Gabbra remember such flows during the early years of this century.

Pans within the Kinu Sogo Rift grabens and *maars* of the Dukana Province retain shallow ponded water for relatively long periods after rains.

## 9.2 WELLS, SPRINGS AND WATER BOREHOLES

Wells and springs are concentrated within three geological provinces: the Lake Turkana Basin, the central rifts and the Bulal lava plateau (Laga Balal). The wells within the Lake Turkana Basin shown on the 1:250 000 topographic map at Bani, Ilgimirr, Kokoi, Gambur and Karari, are all seasonal and dry up in times of drought. The water is thought to be of good quality: they were all dry during the fieldwork. The small community at Ileret relies on near-surface water in the sandy sediment of the Il Eriet River. Numerous hand-dug, shallow wells pepper the wide river bed in the vicinity of the police post.

Perennial springs are confined to the central rifts at Garba Merille, Buluk, Kersa-Bis, Arabtirs and Sabarei. At Sabarei a main outlet is on the river bank. Elsewhere the water is exploited from shallow, hand-dug wells within river sediment. The water is of good quality and these springs are relied on by the Gabbra during droughts.

Numerous, shallow wells have been sunk into the alluvium of the Laga Balal, notably in the El Yibo area. Interestingly the water quality is variable, with saline water at Saru and El Sardu within a fossil west-bank tributary parallel to the Dukana track. Fresh water is extracted from the wells in the main river. All the wells are thought to retain almost permanent supplies of water although rest levels will vary according to climatic conditions. Records of just two, dry, water boreholes are kept in the Ministry of Water's headquarters in Nairobi. The driller's logs are reproduced in Appendix 2. Both holes, at Sabarei and Illeret, were stopped at shallow depths. The mission at Ileret has a piped supply of water which comes from a shallow borehole on the Il Eriet River.

## 9.3 POTENTIAL WATER SUPPLIES

The narrow steep-sided gorges of the Laga Balal and its tributaries form ideal dam sites. However, because of the importance of the wells in this river system to the Gabbra pastoralists, the full impact of damming seasonal flows must be assessed before any dams are constructed. All wells downstream of any dam would necessarily retain water only for short periods after rainfall. Therefore it would be sensible to construct dams as far downstream as possible, i.e. south of Dukana.

Groundwater potential is good in the same three provinces which contain near surface spring-fed water. Within the Lake Turkana Basin groundwater will collect at the base of the Pliocene to Recent sediments (base of the Kubi Fora Formation) due to the absence of major intraformational impermeable units, and the extensive faulting in the upper sediments. Therefore boreholes should be relatively deep, i.e. as much as 300 m to the lava or metamorphic basement contact. Supplies could be large especially if former drainage channels are intersected.

Within the central rifts, the obvious sites for boreholes are the major faults, especially intersection zones of several faults. Due to the relatively high concentration of faults it is unlikely that high yields will be obtained.

Water boreholes in the east should be sited using geophysical methods to intersect former river beds in the metamorphic basement surface beneath the plateau lavas. Otherwise yields will be low. All boreholes should penetrate through the capping plateau lavas.

The Hafari Plateau is capped by the impermeable, massive porphyries which prevent groundwater build-up within this massif. This is reflected in the absence of springs on the lower slopes and the lack of major spring-fed drainage channels. Therefore, incident rainfall is retained in the thin soil cover of the plateau which consequently has the best grazing in the whole area.

## 9.4 GEOTHERMAL ENERGY

Solfataras in the centre of North Island are surface manifestations of a geothermal energy system within the trachytic volcano. Its isolation in the middle of Lake Turkana means that the hot springs will probably never be utilised unless there is a significant development on the island itself.

## Monthly Rainfall Totals 1977-1984
## (figures in millimetres)

### Ileret Police Post Elevation 588m

| Year | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sep | Oct | Nov | Dec | Total |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 1977 | 69.5 | 31.3 | 7.5 | 65.9 | 67.5 | 4.4 | 1.1 | 0.0 | 1.7 | 248.6 | 109.9 | 6.9 | 61.43 |
| 1978 | 5.0 | 36.0 | 104.2 | 23.5 | 4.1 | 0.0 | 15.0 | 0.0 | — | 20.8 | 80.7 | 29.0 | 318.3 |
| 1979 | 49.6 | 25.0 | 68.4 | 0.0 | 25.4 | 47.8 | 7.0 | 0.0 | 0.0 | 15.2 | 43.1 | 45.3 | 326.8 |
| 1980 | 4.8 | 1.5 | 13.3 | 116.0 | 88.5 | 0.0 | 0.0 | 0.0 | 0.0 | 17.0 | 50.4 | 0.0 | 291.5 |
| 1981 | 0.0 | 4.3 | 240.6 | 58.6 | 8.5 | 0.0 | 10.3 | 9.7 | 3.4 | 10.2 | 24.0 | 0.0 | 369.6 |
| 1982 | 8.6 | 10.7 | 44.4 | 49.9 | 61.4 | 12.0 | 4.5 | 3.5 | 0.8 | 84.3 | 90.8 | 54.6 | 425.5 |
| 1983 | 2.8 | 74.2 | 6.6 | 64.3 | 9.9 | 6.0 | 2.8 | 7.6 | 24.6 | 6.7 | 46.6 | 6.6 | 258.7 |
| 1984 | 5.0 | 6.3 | 9.1 | 78.0 | 15.2 | 0.0 | 0.0 | — | 3.7 | 0.0 | 90.4 | 67.4 | 275.1 |

### Sabarei Police Post Elevation 762m

| Year | Jan | Feb | Mar | Apr | May | Jun | Jul | Aug | Sep | Oct | Nov | Dec | Total |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 1977 | 23.6 | 30.7 | 8.4 | 174.4 | 83.1 | 0.0 | 0.0 | 0.0 | 0.0 | 80.3 | 249.8 | 1.6 | 651.9 |
| 1978 | 35.3 | 18.4 | 91.3 | 33.9 | 0.0 | 0.0 | 2.0 | 0.0 | 15.0 | 57.7 | 13.0 | 113.5 | 380.1 |
| 1979 | 26.0 | 92.4 | 88.6 | 124.0 | 6.6 | 33.0 | 10.5 | 0.9 | 9.0 | 32.0 | 58.0 | 73.4 | 552.4 |
| 1980 | 0.0 | 0.0 | 22.9 | 86.4 | 62.1 | 0.0 | 0.0 | 0.0 | 0.0 | 23.8 | 68.7 | 8.1 | 272.0 |
| 1981 | 0.0 | 56.0 | 153.5 | 262.5 | 27.5 | 27.5 | 4.8 | 2.8 | 1.2 | 19.5 | 18.1 | 6.1 | 578.5 |
| 1982 | 10.4 | 15.2 | 45.1 | 83.2 | 56.0 | 0.0 | 0.8 | 0.0 | 1.2 | 91.4 | 127.9 | 36.2 | 467.4 |
| 1983 | 2.0 | 19.0 | 22.5 | 15.0 | 10.8 | 6.0 | 12.8 | 24.5 | 11.8 | 14.3 | 35.0 | 7.0 | 180.7 |
| 1984 | 0.5 | 13.1 | 10.4 | 71.3 | 7.8 | 0.0 | 0.0 | 0.0 | 27.9 | 26.3 | 97.2 | 24.0 | 278.4 |

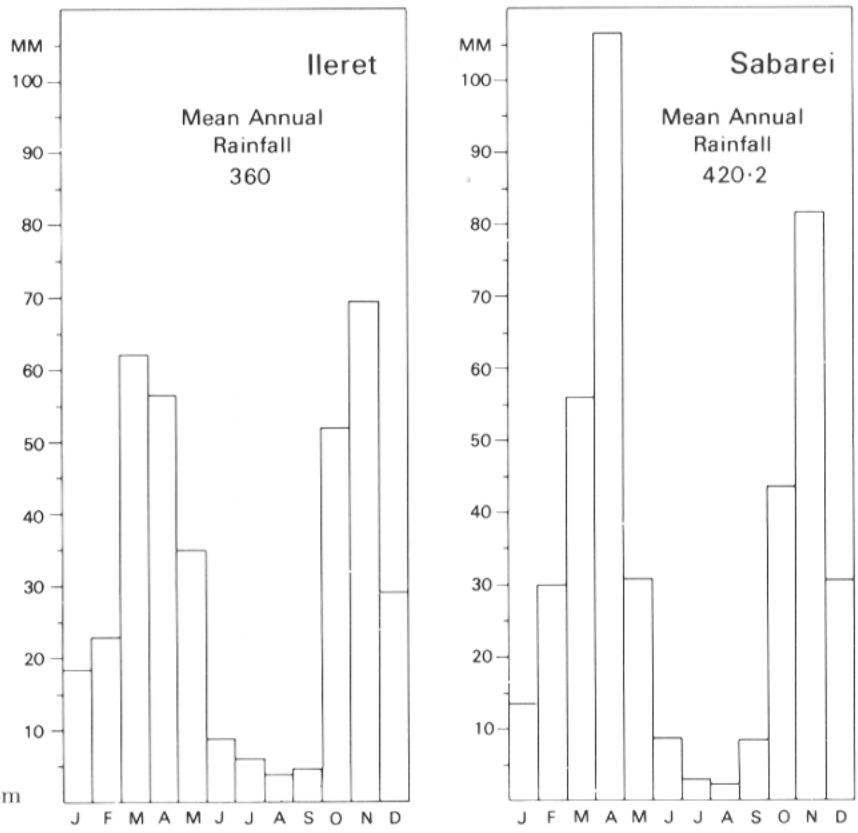



**Figure 17** Rainfall data from Ileret and Sabarei Stations

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# APPENDIX 1 Radio-isotopic age data

**Table 7** Results of conventional K-Ar whole rock analysis by Teledyne Isotopes

| Sample | Method | %K | 40 Ar* (sec/gm × $10^{-5}$) | %$^{40}$Ar | Isotopic Age (Ma) | Note |
|---|---|---|---|---|---|---|
| RK/4/37 | | | | | | |
| Sodic granite, | Whole rock | 2.66 | .213 | 31.6 | 20.7 ± 1.4 | |
| Jibisa | | 2.66 | .217 | 32.6 | | |
| RK/4/49 | | | | | | |
| Porphyritic microsyenite, | Whole rock | 4.32 | .241 | 77.0 | 14.4 ± .7 | |
| Jibisa | | 4.32 | .243 | 83.3 | | |
| RK/4/63 | | | | | | |
| Ankaramitic basalt, | Whole rock | .67 | .042 | 33.0 | 16.2 ± 1.0 | |
| Sabarei volcanics | | .67 | .043 | 45.7 | | |
| RK/4/72 | | | | | | |
| Basalt, | Whole rock | .69 | .029 | 48.8 | 12.1 ± 1.9 | *a* |
| Tatesa Hills | | | .034 | 54.9 | | |
| | | | .039 | 60.2 | | |

Note *a* Heterogeneous argon content in sample.

**Table 8** K-Ar whole rock ages by BGS, London

| Sample | %K | Sample wt for Argon | Rad$^{40}$Ar ± S.D. (nl/g) | Atmos (%) | $^{40}$Ar (nl/g) | Age + 2 sigma (Ma) |
|---|---|---|---|---|---|---|
| RK/NI/2a | 3.180 | 1.92256 | n.d. | 100 | 0.56 | n.d. |
| | | 4.9517 | n.d. | 100 | 0.62 | n.d. |
| RK/NI/2b | 3.290 | 3.7474 | n.d. | 100 | 0.55 | n.d. |
| North Island trachyte | | | | | | |
| RK/NI/8a | 0.397 | 1.4324 | 0.0036 ± 133 | 99.2 | 0.43 | n.d. |
| | | 4.5292 | 0.0021 ± 112 | 99.0 | 0.21 | n.d. |
| RK/NI/8b | 0.422 | 4.5171 | 0.0025 ± 364 | 99.6 | 0.56 | n.d. |
| North Island diorite | | 8.5183 | 0.0172 ± 53 | 97.7 | 0.73 | 1.05 ± 1.10* |
| | | 10.7697 | 0.0040 ± 115 | 98.9 | 0.35 | n.d. |
| Average: | 0.410 | | 0.0031 ± 29% | | | 0.19 ± 0.11 |
| i.e. probably less than 300 000 years old | | | | | | |
| RK/4/16a | 0.874 | 2.5633 | 0.3198 ± 26.5 | 95.1 | 6.26 | 9.39 ± 4.96* |
| | | 3.1784 | 0.1179 ± 2.5 | 66.0 | 0.23 | 3.47 ± 0.19 |
| RK/4/16b | 0.803 | 2.5477 | 0.1179 ± 2.7 | 70.7 | 0.29 | 3.77 ± 0.21 |
| | | | | | mean = | 3.61 ± 0.30 |
| Bulal lavas | | | | | | |
| RK/3/46a | 1.704 | 1.3934 | 0.7623 ± 1.2 | 41.1 | 0.53 | 11.47 ± 0.36 |
| RK/3/46b | 1.603 | 1.7010 | 0.8049 ± 1.8 | 55.5 | 1.00 | 12.87 ± 0.53 |
| Nakwele Formation | | | | | | |
| | | | | | mean = | 11.91 ± 1.6 |

* Anomalously high age associated with high atmospheric argon contamination—probably due to fractionation of residual argon. Not used in mean.

n.d. not determined—radiogenic argon content below limit of detection.

# APPENDIX 2 Water borehole inventory

**Borehole C5001, Sabarei Police Post**

Comments: DRY

Drillers log: (metres)

| | |
|---|---|
| 0– 4 | Yellow brown to grey silty sand and gravel |
| 4– 12 | Light grey clay |
| 12– 20 | Grey tuff |
| 20– 36 | Light grey felsic lava |
| 36– 42 | Light grey tuff |
| 42– 56 | Light grey lava |
| 56– 62 | Light grey tuff |
| 62– 68 | Dark grey lava |
| 68– 82 | Light grey to brown tuff |
| 82– 94 | Yellowish brown to grey tuff |
| 94–100 | Light grey lava |

**Borehole Ileret Police Post**

Comments: DRY

Drillers log: (metres)

| | |
|---|---|
| 0– 6 | Silty sand |
| 6– 10 | Silty sand/ash |
| 10– 22 | Sandy silt |
| 22– 28 | Silty sand |
| 28– 38 | Greenish clay |
| 38– 52 | Pale brown clay |
| 52– 54 | White sand |
| 54– 64 | Mottled clay |
| 64– 76 | Greenish brown clay |
| 76– 82 | Green sandy clay |
| 82– 84 | Dark green to grey clay |
| 84– 94 | Pale green to grey clay |
| 94–100 | Pale green clay sand |

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## MINES AND GEOLOGICAL DEPARTMENT

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Chief Geologist: Mr F. G. Theuri

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