Report 105

Geology of the Maralal area

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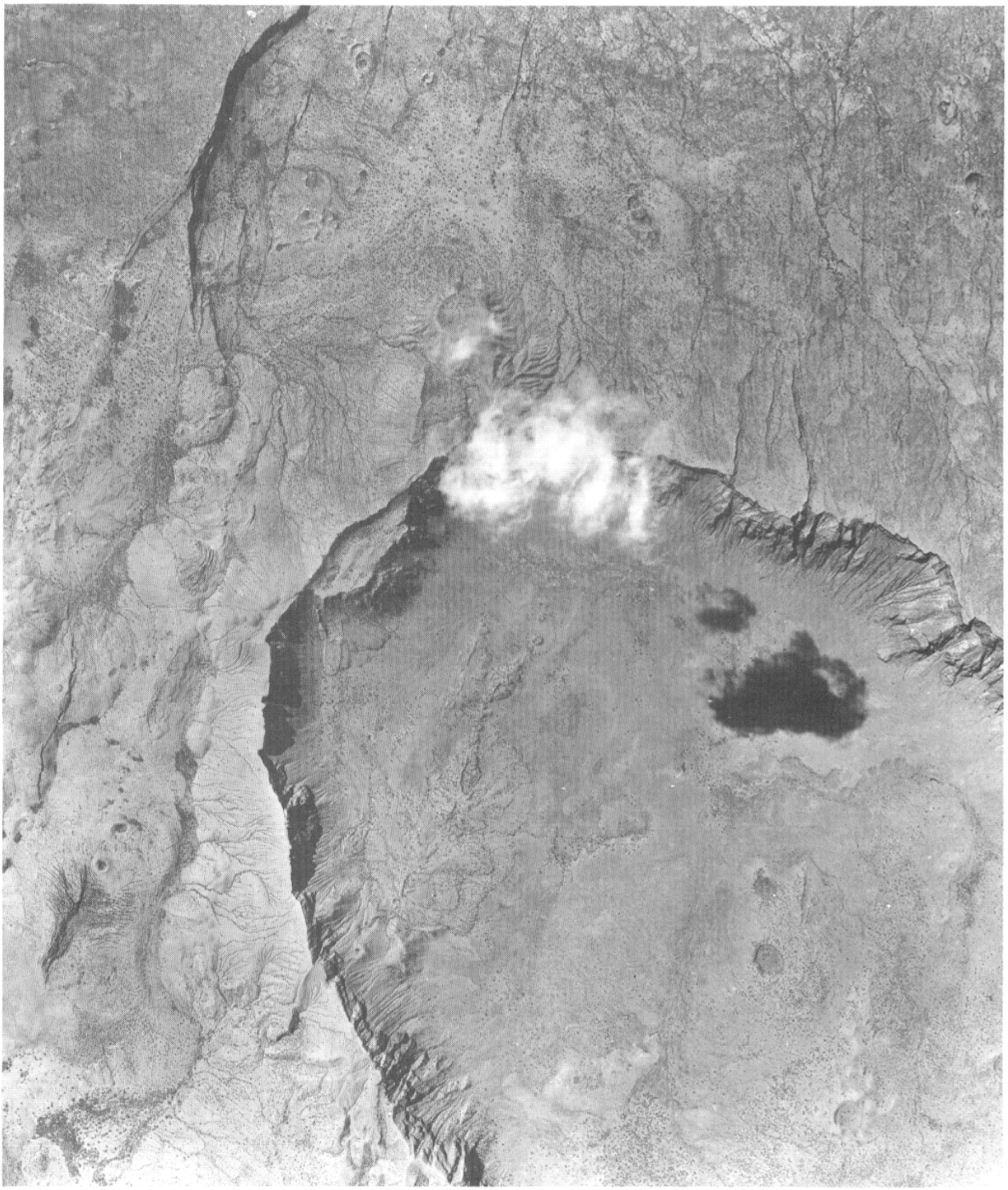


Plate 1 Silali Volcano: part of its summit caldera with sinuous N-trending fault scarps (*Frontispiece*)

MINISTRY OF ENVIRONMENT AND NATURAL RESOURCES
Mines and Geology Department

Report 105

Geology of the Maralal area

Degree Sheet 27
with coloured 1:250 000 geological map and
results of geochemical exploration

R. M. Key

with a contribution by J. Ridgway



NAIROBI 1987

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MAP

1:250 000 geological map of Degree Square 27 (in back pocket)

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PREFACE

This description of a fascinating part of northern Kenya has achieved a new synthesis of the geology of an area which straddles the spectacular northern section of the Gregory Rift Valley, and which has also entailed a detailed revision of the Precambrian Mozambique Orogenic Belt on the eastern shoulder of the Rift, in Samburu District.

The revision of the 'basement' is based on remapping, on the scale of 1:50 000, of the Baragoi area, described by Baker in a Kenya Geological Survey report published in 1963. The names 'Baragoian' and 'Barsaloian', coined in this account for thermotectonic events recognised in the Mozambique Belt, refer to localities in this area: they constitute two important episodes in the orogenic history newly described here on the basis of ground evidence.

For the western half of the area Dr Key has drawn heavily on extensive unpublished descriptions, mostly in the form of post-graduate theses, and two published 1:125 000 maps which resulted from the Kenya Rift Valley Project (East African Geological Research Unit). The Samburu – Marsabit Geological Mapping and Mineral Exploration Project (SMP), a Kenyan-British Technical Co-operative venture responsible for compiling this report and the accompanying 1:250 000 map, is particularly indebted to Dr L. A. J. Williams for providing much helpful advice, and for giving access to his notes on the regional volcanic stratigraphy of the Rift Valley. Dr S. D. Weaver has generously allowed information from his thesis to be published: otherwise it has, regrettably, not been possible to contact all the original research workers who documented this information. SMP must therefore accept full responsibility for any misquotes or untoward bias which may appear in this account.

The relative remoteness of the area notwithstanding, certain aspects of the economic geology deserve mention. The geochemical exploration programme has identified three major anomalous areas in the Precambrian belt, recommended for detailed ground follow up: Siambu Hill (nickel and copper), Marti (zinc, barium and six other metals) and Loroki Forest (zinc, lead, nickel and copper).

Other mineral occurrences of interest are graphite, vermiculite, gemstones and radioactive minerals in the Precambrian rocks, rare earths and fluorite in the Rift Valley. As access improves the geothermal areas of the Rift are likely to be further investigated for their energy potential.

It is anticipated that Dr Key's report will prove a valuable reference work for those concerned with the Mozambique Belt and this part of the East African Rift.

B. D. Hackman
Samburu – Marsabit Project Leader
15 April 1987

Geology of the Maralal area

Abstract This report describes the geology, mineral resources and economic potential of an area 12 182 km² in extent in north-central Kenya bounded by the parallels 1° and 2°N, and by the meridians 36° and 37°E. Six physiographic provinces controlled by bedrock structure are the east and west shoulders and median graben (Suguta Valley) of the Cenozoic Gregory Rift, the Karisia Hills, western foothills of the Musei Range, and an intervening dissected plain underlain by older metamorphic rocks. Drainage is related to four major river systems: the Kerio and Suguta drainages directed north through the western volcanics, and the older Korante Plains and Milgis drainages directed eastwards across the eastern basement. The area is founded on late Proterozoic to early Phanerozoic metamorphosed rocks of the Mozambique Orogenic Belt, although their exposure is essentially confined to the east. A succession of altered sediments and volcanics is cut by a series of minor felsic and mafic intrusives related to a polyphase, Upper Amphibolite Facies, tectonothermal history. The Gregory Rift trends NNE across the western two-thirds of the area with almost continuous early to middle Miocene through Quaternary assorted (basaltic-phonolitic-trachytic) volcanism, subsidiary sedimentation and contemporaneous tensional tectonism. Quaternary volcanism and tectonism are largely confined within the median graben where extensive fissuring may herald plate separation. Significant showings of asbestos, chromite, gemstones (garnet, beryl, sapphire), graphite and vermiculite in the basement area warrant detailed appraisal. Fluorite along faults, and diatomite beds occur in the western volcanics—a potential gold province. Large reserves of river sand and gravel, aggregate and building stone occur throughout the area. An airborne radiometric survey detected a significant uranium anomaly in the east: nearby several other anomalies were discovered on the ground. A regional stream sediment sampling programme over the basement detected anomalies for three areas which have potential for the discovery of significant mineralisation: Siambu Hill (nickel, copper), Loroki Forest (zinc, lead, nickel and copper), and Marti (zinc, barium, bismuth, arsenic, tungsten, niobium, zirconium and uranium). Sparse rural populations rely largely on cold spring water, augmented by seasonal surface or near surface water ponding in sandy river beds. Earth dams and water boreholes are restricted to the few urban centres. Future greater water needs could be met by improving the exploitation of present supplies. Hot springs and fumaroles in the median graben of the Gregory Rift attest to its potential as a major source of geothermal energy.

1 Introduction

Sheet 27 (Maralal) forms the NW quadrant of the phase 1 area (Figure 1) of the Samburu–Marsabit Geological Mapping and Mineral Exploration Project (SMP): a Technical Co-operation project between the British (British Geological Survey) and Kenyan (Mines and Geological Department) Governments. The principal objectives of this phase of the project were to produce geological maps on a scale of 1:250 000 of degree squares and to locate mineral deposits of economic significance. The geological maps incorporate pre-existing data, mostly published at 1:125 000, as well as the new information amassed during fieldwork carried out by project staff between 1981 and 1984.

Mineral exploration involved the field observations of the geologists and a stream sediment sampling programme mostly confined to areas underlain by metamorphic rocks of the Late Precambrian–early Phanerozoic Mozambique Orogenic Belt.

1.1 LOCATION

The Maralal sheet is bounded by longitudes 36° and 37°E and latitudes 1° and 2°N, an area of 12 182 km². The north-east corner is part of the Marsabit District of the Eastern Province and the rest of the sheet is part of three districts of the Rift Valley Province: the Samburu, Turkana and Baringo Districts (Figure 2). Maralal, the district administrative centre of the Samburu District and Lokori in the north-west, are the only sizeable settlements with populations of about 10 230 and 12 465 respectively (1979 census). The rest of the area is largely populated by nomadic ethnic groups: the Rendille, Samburu, Turkana and Pokot in the Marsabit, Samburu, Turkana and Baringo Districts respectively. Locations of small but well-established settlements are shown on Figure 2.

The following topographic maps cover the degree square:

1:50 000
64/2, 64/3, 64/4, 65/1, 65/2, 65/3, 65/4, (all Edition 1, DOS, 1982)
64/1 (Edition 1, DOS, 1983)
77/1, 77/2, 77/3, 77/4, 78/1, 78/2, 78/4, (all Edition 2, DOS, 1982)
78/3 (Edition 2, DOS, 1983).

All are contoured at 20 m intervals and the 16 sheets cover the whole of the Degree Square (Figure 2).

1:100 000 (with form lines only), Series Y633
64 (NW quadrant of degree square 27), 65 (NE quadrant of Degree Square 27).

1:250 000 (contoured at 1000 ft intervals), Series SK60 Edition 2A-SK, Sheet NA-37-9 (Maralal), 1973.

1.2 ACCESS

The Suguta Trough effectively divides the area into two unequal halves, with no vehicular access up its eastern escarpment. A single dirt track skirts the west side of the Trough along the whole length of Degree Square 27 to connect Lokori to Kapedo. Poorly defined tracks

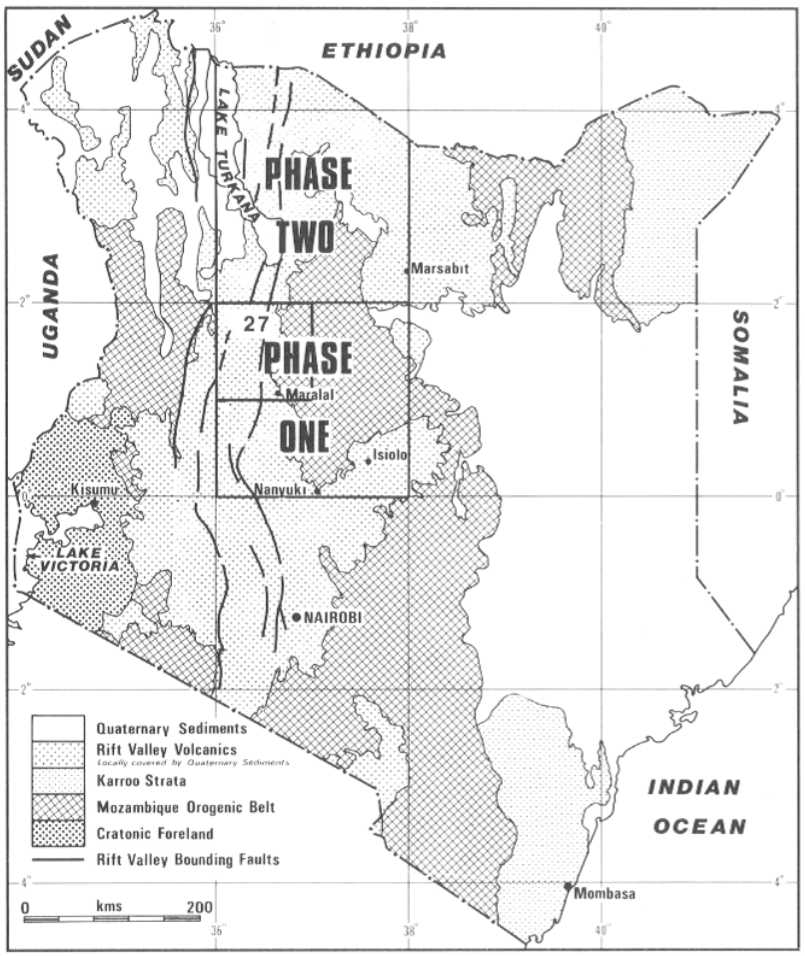


Figure 1 The location of the map-area within Kenya

penetrate eastwards, the best of which is from Lokori to Kamuge.

The main access route from southern Kenya to the east side of Lake Turkana passes through Maralal and Baragoi. It is a poorly-maintained dirt track. A network of small dirt tracks radiates from Maralal to adjacent parts of the Loroki Plateau and Karisia Hills. The track from Maralal to Barsaloi descends the precipitous eastern side of the Karisia Hills. Several tracks extend eastwards from Baragoi including a good dirt track south-east through Barsaloi and Suari to Wamba. The track from South Horr to Laisamis cuts across the NE map corner.

All tracks become impassable during heavy rains and are only suitable for 4-wheel drive, high-clearance, small vehicles or lorries at other times. The present track network is poor with most of the sheet inaccessible to vehicles. Several temporary tracks were constructed by field parties to assist the mapping.

Dirt airstrips suitable for light aircraft are maintained at Maralal, Barsaloi, Baragoi, Poro, Kapedo, Lokori and north of Lomelo.

1.3 PREVIOUS WORK AND OTHER WORK STILL IN PROGRESS

Several late nineteenth century adventurers and explorers traversed Degree Square 27 and included brief notes on its geology in their published travelogues. (Rosiwal, 1891; Cavendish, 1898; Von Höhnel, 1938). Shackleton (1946) and Baker (1963) are excellent accounts with coloured geological maps of the SE and NE quadrants of Degree Square 27. Reference is made to these reports at relevant places in the present account.

The western half of the area was mapped geologically in the late 1960s and early 1970s by members of EAGRU*. Their maps were incorporated into the 1:125 000 geological maps of the Kapedo and Emuruangogolak area (Truckle, 1979a) and of the southern Loriu area (Truckle, 1979b). Much of the work formed the basis of PhD theses and areas of responsibility are shown in Figure 3. Rhemtulla did not complete his thesis but provides a brief account of the geology of his area amongst a collection of scientific papers of the Royal Geographical Society South Turkana Expedition (Rhemtulla, 1970). The geology of the area (Silali Volcano) mapped by McCall is outlined in several papers of which McCall and Hornung (1972) is the most comprehensive. Numerous publications based on the mapping are referred to in relevant parts of the present account, including the review paper (King, 1978) by the EAGRU director.

Khan and Swain (1978) review the geophysical studies (seismic and gravity) on the Gregory Rift to include data from Degree Square 27. Various research projects on the Gregory Rift in central Kenya are in progress. A joint Japanese–Kenyan (National Museum, Nairobi) team is investigating the Miocene rocks of the eastern shoulder for hominid fossils. An international team of geophysicists (KRISP) is conducting deep seismic profiles across the rift in an attempt to understand its deeper structure. The geothermal potential of the rift is being investigated by a UNDP team based in Nairobi.

* East African Geological Research Unit of London University.

An Open University (UK) research student, Seife Berhe, is studying the geochemistry of the eastern metamorphic rocks of Degree Square 27 (specifically the Siambu Complex) as part of a regional study of the Mozambique Orogenic Belt by the University.

1.4 PRESENT WORK

Fieldwork was carried out in each of the eight eastern 1:50 000 sheets during 1982 and 1983. Individual field parties were in the charge of one geologist with a team of drivers, field assistants and locally employed casual labourers. Base camps were established in each sheet with daily foot traverses supported by Landrovers. Stream sediment sampling accompanied the geological mapping in most cases. The time spent mapping each sheet varied from one to three months. Internal reports of the Mines and Geological Department and draft 1:50 000 geological maps were completed by the geologists and these data are incorporated in the present account. The results of the stream sediment sampling programme are discussed at the end of this report by the project geochemist. Four bulk stream samples were also collected from drainage off the Kailimerlim Foyaite in the north-west during helicopter-supported fieldwork in adjacent parts of the phase 2 area of the SMP.

In this report extensive reference is made to the unpublished PhD theses of the EAGRU research students. However by necessity many of the detailed descriptions of individual lithostratigraphic units given in these theses are omitted. There is an urgent need for separate publications of these data so as to formalise many of the (numerous) informal names introduced in the theses.

1.5 SUMMARY OF THE GEOLOGY

The map-area is founded on polyphase metamorphic rocks of the late Proterozoic to early Phanerozoic Mozam-



Figure 2 Place names referred to in the text

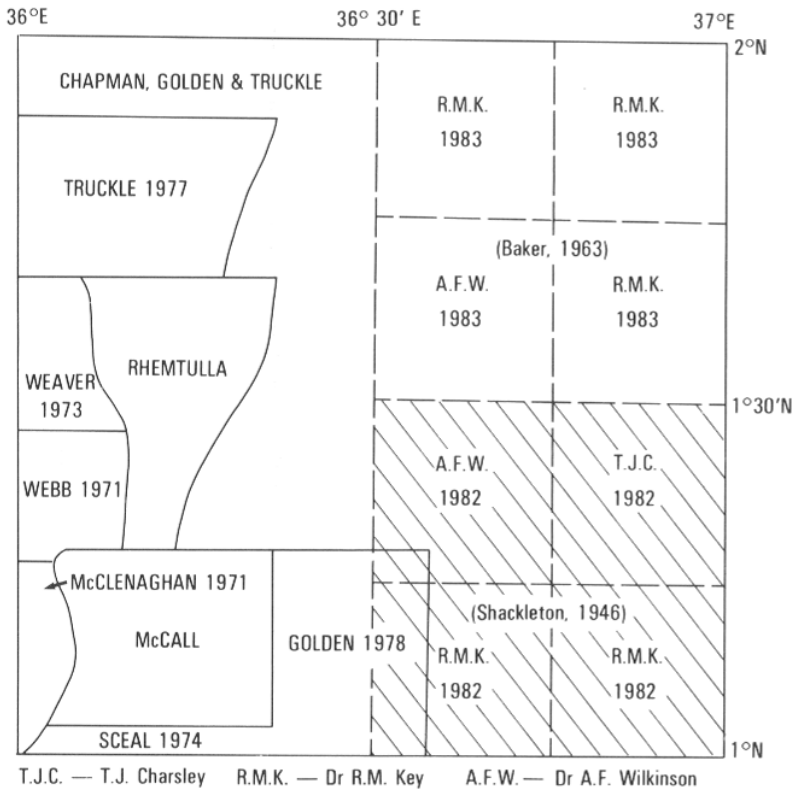


Figure 3 Geological mapping responsibility diagram

bique Orogenic Belt. However in the western two-thirds these rocks are concealed (Figure 4) beneath Miocene through Quaternary volcanic rocks (and subsidiary sediments) of the Gregory Rift. Basement inliers protrude through the eastern shoulder of this rift and a solitary western inlier just extends into the map area at GR*1663 1900.

Four major lithostratigraphic units are recognised in the eastern metamorphic terrain. The original stratigraphic order has been muddled by subsequent tectonothermal events but a prevalent tectonic stacking order is established to include two tectonic units as follows:

6 Augen gneisses (Baragoian tectonites).
5 Siambu Complex (altered basic and ultrabasic igneous rocks and subsidiary sediments).
4 Undifferentiated migmatites (Sabachian sole thrust partial melts).
3 Ol Doinyo Ng'iro Gneisses (altered shelf sediments).
2 Loroki Gneisses (altered arkosic sediments or granites).
1 Ndura Complex (altered arkosic sediments or granites).

An early (c.830 Ma) regional Upper Amphibolite Facies metamorphism (Samburuan) produced coarse gneissic fabrics as well as local partial melt intrusives. Metamorphism was progressive culminating in anatectic melting to form autochthonous migmatites and allochthonous (G2) granites. Continuing tectonism (Sabachian) produced overturned and recumbent folds with thrusting from the north-west with linear fabrics aligned transverse to the regional N–S trend of the orogen. Following a long period of erosion renewed Upper Amphibolite Facies deformation initially (Baragoian) produced upright cylindrical folds plunging gently to the north-west or north-north-west.

Folds limbs were then attenuated in major (c.580 Ma, Barsaloian) N–S shear zones. Tectonic domains related to the tectonothermal episodes have been mapped. Post-tectonic almandine growth in the SE quadrant may be related to a concealed basic intrusive. Warping and brittle fractures are related to post-orogenic crustal adjustments. Several phases of late linear felsic veins and dykes, as well as earlier minor mafic and felsic veins and dykes are recognised.

A long period of erosion followed until the later Tertiary inception of the NNE–trending Gregory Rift. Three volcanic provinces are recognised: eastern and western shoulders and a median graben, each with distinctive histories of extrusive volcanic phases, which, however, have common chemical trends suggesting they shared the same mantle magma source. Eastern shoulder volcanicity started with early to middle Miocene eruptions of basalts to form a series of coalescing shields. This was succeeded by middle to late Miocene phonolitic and basaltic volcanism showing lateral facies changes. The shoulder is capped by late Miocene through Pleistocene (in the north) basalts and trachytes which formed low angle shields.

Quaternary basaltic and trachytic volcanism alternate in four low angle shield volcanoes in the median graben. The volcanoes are separated by fluviolacustrine sediments.

Middle Miocene through Pleistocene volcanism in the western shoulder manifests as flood basalts and trachytic volcanoes with minor phonolitic and trachyphonolitic volcanics on the western margins. In general the trachytic volcanoes are large low-angle multicentred shields or plateaus.

As to be expected there was a fundamental structural control on the locus of volcanism in all three provinces. Periods of normal (synthetic and antithetic) faulting, tilting and fissuring are recognised in the lower, middle and upper Miocene, Pliocene and Quaternary which, at least in part, were contemporaneous with extrusive volcanism. As a result the gross structures of both shoulders are monoclines facing towards the median graben. In Degree Square 27 the median graben is bounded on both

*GR refers to grid reference within the UTM grid zone 37.

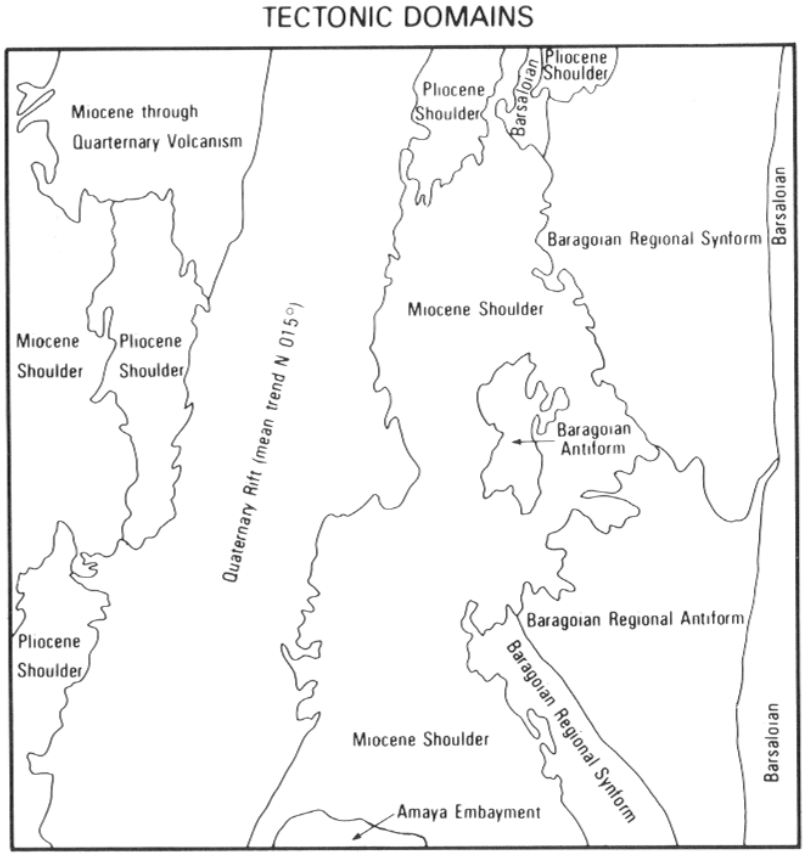


Figure 4 Tectonic domains with three Cenozoic (of the Gregory Rift) and three basement (of the Mozambique Orogenic Belt) components

sides by Quaternary fault arrays with cumulative downthrows of up to 1000 m. Fluviolacustrine and subaerial sediments within the Cenozoic volcanic sequence form laterally restricted, thin lenses. Their deposition was controlled by contemporaneous volcanicity: the volcanics provided the source material and caused temporary ponding of surface water behind volcanic edifices. Subvolcanic sediments infilled hollows in the sub-Miocene land surface and have volcaniclastic and basement-derived debris.

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2 Geomorphology

Six physiographic provinces are identified (Figure 5), coincident with major tectonic domains in bedrock, to emphasise the structural control on landform development. The western and eastern shoulders and median graben of the Cenozoic Gregory Rift have distinctive landforms. The Karisia Hills and adjacent eastern lowlands relate to an older regional (Baragoian) antiform and synform respectively. The eastern map edge is part of the western foot slopes of the Musei Range with pronounced N-trends in a major (Barsaloian) shear zone.

form featureless plateaus (underlain by monotonous, gently dipping lava seqences) cut by NNE river gorges along fault traces.

The northern trachytic volcanoes lack the rugged central areas. Escarpments, up to several hundred metres high, around plateaus are either fault-controlled (adjacent to the median rift) or due to headwater erosion by south bank tributaries of the Kerio River.

Kailimerlim Foyaite (paragraph 6.6) underlies a prominent steep sided hill above the northern plateaus. Mesas

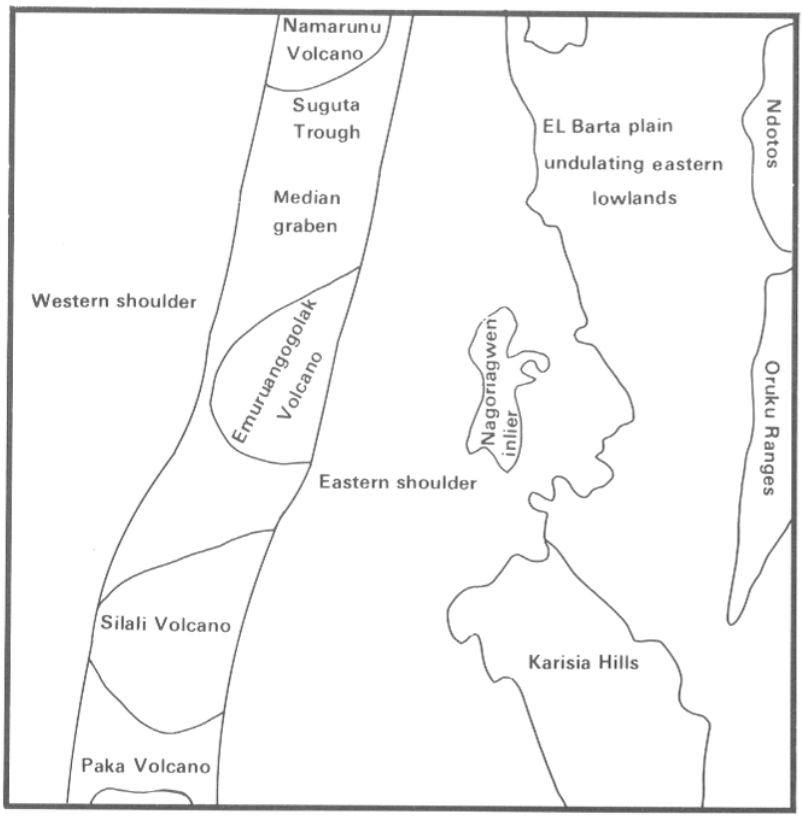


Figure 5 The major physiographic provinces

2.1 WESTERN SHOULDER OF THE GREGORY RIFT

The western shoulder is underlain by middle Miocene through Quaternary volcanic rocks. Contemporaneous erosion, faulting and tilting accompanied intermittent volcanism to fashion the present varied landscape. Coalescing trachytic shield volcanoes and flood basalts dominate the geology. Central zones to the trachytic shields comprise lavas, pyroclastics and minor intrusives. Preferential weathering of the friable pyroclastics has formed rugged terrains: steep pinnacles of resistant trachyte intrusives contrast with landslipped blocks capped by resistant lavas with substantial undercutting of underlying pyroclastics. Back-tilting of faulted blocks in these central areas has also contributed to their chaotic scenery. By contrast the flanks of the trachytic volcanoes and buttes dominate the northern landscape separated by wide U-shaped valleys floored with alluvium and skirted by pebble and boulder pediments (Truckle, 1977a).

Due to the different susceptibilities to weathering of different lithologies there is no direct correlation between the age of bedrock units and their state of preservation. The oldest trachytic lavas are less weathered than younger pyroclastics and coarse basalt lavas. (The relatively young Murgisian basalts in the south are extensively eroded.) Surface water is the main weathering agent, forming radial drainage channels off central volcanic summits, fault controlled linear watercourses and dendritic headwater systems.

The western shoulder dips gently NNE from heights (above sea level) of over 800 m to below 500 m. Central areas of Ribkwo and Kafkandal Volcanoes rise to over 1200 m (above sea level) and the western shoulder is

generally elevated about 200 m above adjacent parts of the median graben. The basement inlier in the extreme west stands at about 800 m (above sea level) which is thus the present level of the sub-Miocene surface on this side of the Gregory Rift.

2.2 MEDIAN GRABEN OF THE GREGORY RIFT

The median graben is bounded by fault-controlled escarpments up to 600 m in height (at Losiolo and below the Loriu Plateau), locally modified by landslipping. It is flat-bottomed with a NNE gentle slope from elevations above sea level of 800 m in the south to less than 300 m in the north (comparable altitude to Lake Turkana further north). Landforms relate to transient unconsolidated sediments-lacustrine flats, shallow braided and meandering river channels, barchan and seif sand dunes, gravel fans. Rising majestically above the floor of the median graben are central trachytic volcanoes: in Degree Square 27, the northern and southern parts of Paka and Namarunu respectively and all of Silali and Emuruangogolak. They have circular plans, which may straddle the whole graben, and rise to 800 m above the adjacent flood plain. These Quaternary edifices have pristine, poorly vegetated landforms: summit calderas, cinder cones, hornitos, discrete lava flows with unweathered surface textures, flat areas of unconsolidated pyroclastic beds. NNE-trending fault scarps and narrow fissures are common features on the slopes of all four volcanoes (Plate 1).

2.3 EASTERN SHOULDER OF THE GREGORY RIFT

The eastern shoulder geology is dominated by middle Miocene through Pliocene stratified volcanics. Along the eastern outer parts of the shoulder these volcanics underlie monotonous, horizontal plateaus (Lopet, Loroki, Tirr Tirr and Emuru Akirim Plateaus). They may have erosional escarpments up to 380 m high facing eastwards towards the metamorphic terrain. Towards the median graben they are increasingly cut by NNE sinuous faults which have controlled erosion. Watercourses have cut steep, narrow gorges (up to several hundred metres deep) along fault traces, and back-tilted fault blocks have exposed basal friable pyroclastic beds. Weathering of this soft material has undermined capping lavas to create numerous landslips in a typical badland topography. Several minor trachyte plugs form impressive pinnacles above the strongly eroded pyroclastics of the inner part of the eastern shoulder in the south. Surface water flow controlled erosion: the steep hydraulic gradients eastwards into the median graben aided torrential, clast-charged flow.

The elevation of the eastern shoulder is controlled by the pre-existing basement landscape, especially the Karisia Hills. The volcanic surfaces dip radially off the Karisia Hills where they are about 2400 m above sea level.

In the extreme north and adjacent to the median graben, mean elevations above sea level drop to below 600 m. The Nagoriagwen inlier (Figure 5) is a continuation of the same basement lithology and structure (antiform) as the Karisia Hills. Contemporaneous (Tertiary), and Quaternary erosion has fashioned the volcanic landscape. The sub-Miocene land surface beneath the volcanic carapace varies in present elevation above sea level from about 2200 m in the south (Karisia Hills) to 1100 m in the north under the Emuru Akirim Plateau. In part this is due to the doming which preceded the Gregory Rift, cf. Baker and Wohlenberg (1971), and also reflects basement lithology and structure.

2.4 KARISIA HILLS

The NW-trending, heavily forested Karisia Hills have asymmetric NE–SW profiles with precipitous NE slopes and gentle SW slopes disappearing beneath horizontal phonolites of the eastern shoulder of the Gregory Rift. The Karisia Hills attain a maximum altitude of 2582 m at Poro; their northern end summit levels generally exceed 2000 m with several flat peaks at about 2200 m above sea level. The constant summit heights led Shackleton (1946) to suggest they may represent a Cretaceous peneplain. The NE slopes have gradients of up to 1:2 to base levels of about 1400 m. This steep declivity is not fault-controlled, but reflects the extreme difference in competence between the quartzofeldspathic gneisses under the Karisia Hills and the friable banded gneisses to the east. The quartzofeldspathic gneisses also form the core of a NW-trending antiform. Summit levels decline in elevation gently to the SSE.

The Karisia Hills and the Ndotos to the east must have formed positive topographic features ever since Palaeozoic uplift and erosion followed the Mozambiquian Orogeny. Shackleton (1946) could well be correct in suggesting that the present summit levels represent the state of erosion by Cretaceous times. A thick forest cover probably existed throughout the Tertiary and Quaternary, protecting the Karisia Hills from any significant erosion.

2.5 UNDULATING EASTERN LOWLAND

Quaternary erosion related to superimposed drainage networks has caused major undulations in the present land surface of the eastern lowlands. Three distinct landforms can be recognised: major hills or inselbergs; undissected (residual) plains; and dissected rugged countryside (Plates 2, 3, 4 and 5). There is a strong lithological control on the positive topographic features as they are invariably underlain by massive quartzofeldspathic gneisses, G2 granites or altered ultrabasics. Their morphology is controlled by bedrock structure: flat-lying rocks form mesa-like features (e.g. Kawab) elongate inselbergs parallel the strike of their dipping bedrock (e.g. Kotikal). Summit heights relate directly to the competence of bedrock, although they may represent remnants of the sub-Miocene, or an older surface (e.g. Siambu Hill).

The El Barta Plain in the north is the largest area not affected by major (Quaternary) river erosion. Headwaters of the major rivers are encroaching into all its marginal areas. The plain has a relatively thick soil/colluvium cover infilling a gentle basin with surface heights decreasing inwards from about 1500 to 1300 m above sea level.

The eastern lowlands are cut by several major drainage systems (see Figure 6 and paragraph 2.7). Active erosion, especially by headwater streams, is stripping the soil/colluvium cover and reducing land levels dramatically (in the extreme NE and SE to below 1000 m above sea level). Much of the area north of northing 153 affected by this erosion lacks a thick soil cover.

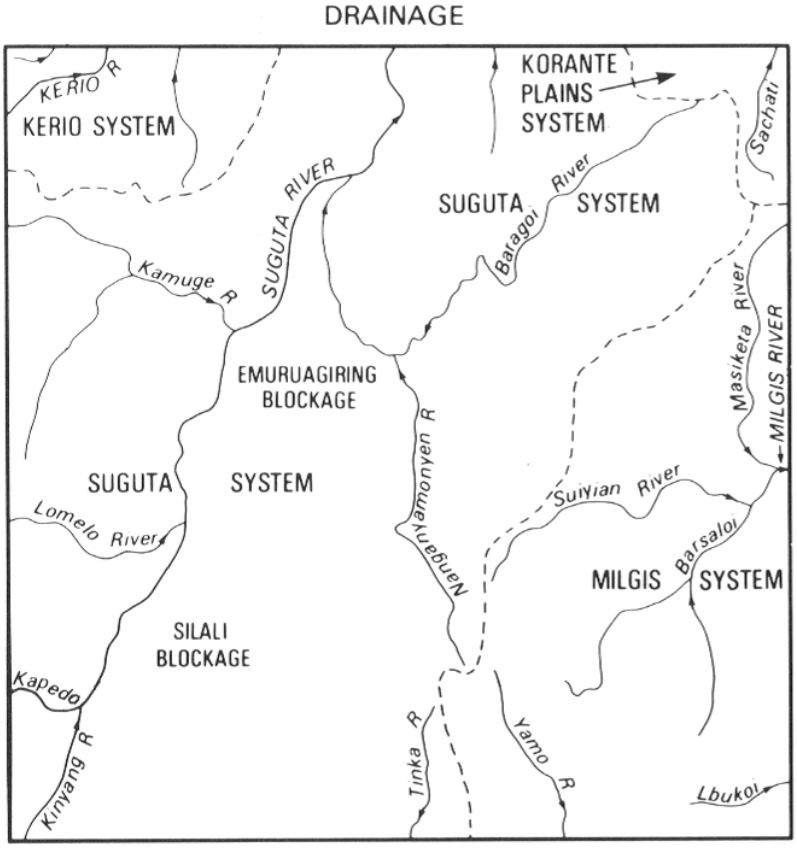


Figure 6 The four major drainage systems

Therefore three stages in the ongoing weathering history of the eastern lowlands can be recognised. Some of the peak heights to the most resistant hills and inselbergs may only reflect pre-Miocene erosion. The residual plains continued to be affected by Tertiary erosion although this actually led to a local thickening of (colluvial) overburden which was transported in from adjacent high ground. Widespread Quaternary headwater erosion is affecting all of the eastern lowlands except for the residual plains (El Barta Plains).

2.6 NDOTOS AND EASTERN HIGHLANDS

The western slopes of the Ndotos just extend into the north-east of Degree Square 27, abutting against the El Barta Plains. Immediately to the east they rise to over 2600 m and are underlain by massive quartzofeldspathic gneisses similar to the Karisia Hills. Further south the eastern ridges of the parallel (N–S) Oruku Ranges define the eastern margin of the eastern lowlands (Plate 3). The quartzofeldspathic gneiss bedrock has a pronounced N–S strike to control the topography. Parallel valleys are eroded in softer banded gneisses and are part of the same Quaternary drainage system described in Section 2.5.

2.7 DRAINAGE

The area is drained by four major river systems (Figure 6): the Kerio and Suguta Rivers in the volcanic terrain, and the Milgis (Barsaloi–Seiya) River and Korante Plain watercourses in the eastern metamorphic area. Drainage patterns have changed since the Miocene because of changing landforms related to the Gregory Rift tectonism and volcanism. For example, the Suguta River, and its NNE-trending watershed with the Milgis directed drainage, originated in the Tertiary.

The Suguta River almost certainly cut right across and captured all east bank tributaries, including the Baragoi River branch of the older Kerio River. The eastern watershed cuts across the grain of the metamorphic basement so that much of the Milgis directed drainage is superimposed. Local examples of river capture, rejuvenation and 'fossilisation' of river, are common. Water courses are controlled by surface slopes and their instability, major structures (e.g. fault scarps and master joints), bedrock, or may be superimposed.

Most watercourses are ephemeral except for the headwater stretches of rivers fed from permanent (hot and cold) springs. These include the Suguta River north of Kapedo and many headwaters on the Karisia Hills and adjacent high parts of the eastern Rift shoulder, e.g. Tinka, Lagolin, Moridjo, Lulu and Ndadapo Rivers.

The Kerio River just cuts across the NW corner of Degree Square 27 on its 350 km long northward journey into Lake Turkana. It has a closely spaced dendritic drainage network on its south-eastern side off the Loriyu Plateau. However none of these tributaries now join the Kerio River, rather they disappear into an extensive flood plain incising older alluvium (Truckle, 1977a). This is because of the present arid climate and also Quaternary tilting eastwards of the western Rift shoulder (see Chapter 8). The tilting has effectively removed the hydraulic gradient in the watercourses, and has caused river capture of the upper parts of many streams by Suguta River directed drainages, (e.g. the Kamuge River System). The watershed ridge between the Kerio and Suguta River drainages is extremely narrow.

The Suguta River meanders along the whole length of the median graben in Degree Square 27 on its northward journey to Lake Logipi. Its course is mainly controlled by



Plate 2 Siambu Hill viewed from the south



Plate 3 The Oruku Ranges viewed across the Barsaloi River

Plate 4 Looking west towards the Lopet Plateau





Plate 5 Looking north towards Kawab



Plate 6 Typical paragneisses of the Ol Doinyo Ng'iro Gneisses



Plate 7 Looking north-west across the median graben from Losiolo with the well-stratified Losiolo phonolites exposed in the foreground escarpment

the gentle NNE surface slope. In the north the land is essentially flat to account for the numerous braids and meanders in an area susceptible to flooding (the catchment area of the Suguta River covers over 10 000 km^2). Intermittent blocking of the river has occurred because of Quaternary outpourings from Silali and Emuruangogolak Volcanoes. Radial drainages off these volcanoes only locally continue into the Suguta River. Most die out in the adjacent flood plains. The Suguta River is presently cutting into these flood-plain deposits. A well developed drainage network joins the west bank of the Suguta River; most tributaries actually join the Suguta River because of the eastward tilting of the west Rift shoulder mentioned previously. Individual watercourses have been modified throughout the Tertiary because of the volcanism and tectonism. Webb (1971) provides good examples for the Ribkwo area.

The east bank drainage of the Suguta River is more complex. Excepting the short watercourses draining off Silali and Emuruangogolak, all the east bank tributaries start on the eastern Rift shoulder or, in the north, from the eastern metamorphic area. In the south only the Amaya River flows into the Suguta River. Others die out in the flood plain of the median graben. Watercourses are strongly controlled by NNE fault traces to produce a rectangular drainage pattern. The major northern tributary is the Baragoi River: this starts on the El Barta Plain where its headwaters are now 'fossilised' and being captured by the Milgis River and Korante Plain drainage channels. In its upper parts the Baragoi River has a dendritic tributary network but in its lower stretches the drainage is rectangular due to the influence of NNE faulting. The prominent U bend in its course where it is directed northwards into the Suguta River may be due to its capture by this river. Previously it must have continued westwards into the Kerio River. The thick accumulations of intravolcanic sediments in the east shoulder near the Baragoi River are, at least in part, due to sporadic blocking of the river by volcanic rocks.

Most of the eastern metamorphic terrain is drained by the Barsaloi River and its tributaries. The Barsaloi River is directed north-east off the Karisia Hills and swings due east at the map-edge along a major fracture zone which takes the Milgis River through the Musei mountain range. The tributaries define a dendritic pattern largely controlled by the regional surface gradients. However the northern tributaries have been reactivated by the late Cenozoic development of the NNE watershed with the Suguta River system: they are superimposed on bedrock and all unconsolidated overburden in their vicinity has been removed. The southern half of the Karisia Hills is drained by watercourses flowing SSE into the Seiya River (which flows north to meet the Barsaloi River just east of Degree Square 27: the confluence marks the start of the Milgis River). The southern Karisia Hills watercourses are remarkably linear, as they are controlled both by regional SSE surface slopes and parallel bedrock structures.

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3 Metamorphic rocks

Metamorphic rocks of the polyphase late Proterozoic to early Phanerozoic Mozambique Orogenic Belt underlie the whole map-area to include the median graben of the Gregory Rift (King, 1978). However, exposure of this metamorphic foundation is essentially confined to the eastern third of the map. Inliers of the metamorphic rocks do protrude through the volcanic carapace of the rift shoulders: mainly along the eastern shoulder in the map-area except for a single western shoulder inlier at GR 1663 1900. The eastern metamorphic terrain has been divided into several lithostratigraphic units based on major (regional) lithological variations (Figure 7). It has to be stressed that no rigid stratigraphic order is elucidated because all lithological contacts possess a tectonic component: present relative orders of the lithostratigraphic units are tectonic. The intimate interlayering of migmatites and banded gneisses and the lateral wedging of lithologies are also tectonically controlled. Lithologies of two or more of the major lithostratigraphic units may be interlayered and, in many cases, it is not possible to ascertain the exact lithostratigraphic setting of individual lithological bands or wedges. Over much of the eastern part of Degree Square 27 the following tectonic stacking order exists (for mapped units):

6 Augen gneisses
5 Siambu Complex
4 Undifferentiated migmatites
3 Ol Doinyo Ng'iro Gneisses
2 Loroki Gneisses
1 Ndura Complex

Local variations are mentioned in the descriptions of individual units. All, except for the augen gneisses, have suffered the same polyphase tectonothermal history and their precursor rocks are hence older than the initial (Samburuan) imposition of coarse metamorphic fabrics, dated at about 830 Ma.

The considerable textural changes imparted by the successive high grade (Upper Amphibolite Facies) metamorphisms and the complete absence of original sedimentary or igneous textures preclude detailed discussions on the original nature of the various precursor rocks, although a general statement on this controversial subject is appended to the descriptions of individual units. Similarly, present lithological thicknesses are controlled by the structural history and bear no relationship to original sedimentary or igneous layering. Lateral tectonic transport has also displaced whole units which again introduce a major element of uncertainty into discussions on palaeogeography.

The Upper Amphibolite Facies metamorphisms have produced small and large scale partial melt phases: quartzofeldspathic veins and pegmatites, mafic clots, and the autochthonous granites described in Chapter 6. The small scale textural variations described below are all typical of high grade metamorphic terrains, i.e. concordant and discordant vein and dyke phases, zones of partial melting, regular gneissic layering and more heterogeneous migmatitic textures, parasitic folds of several generations, brittle and ductile, conformable and cross-cutting shear zones.

The major lithostratigraphic units in this, and the following chapters are described in terms of their nomenclature, distribution, morphology and photogeology, reference sections, age, structural relationships, thickness, lithologies, petrography and petrogenesis.

3.1 NDURA COMPLEX

The Ndura Complex takes its name from the Ndura River in the Kirimun area (Sheet 924), about 50 km south of Maralal. Key (1982) has defined a reference section along the river from the main Kirimun to Ol Doinyo Ng'iro track crossing (GR 2592 0630) downstream to its confluence with the Ewaso Ng'iro River.

The complex is confined to the south-eastern part of Degree Square 27 (Figure 7) where it is variably exposed in the low ground east of the Karisia Hills. The dominant lithology is a banded grey gneiss with a major concordant granite vein phase giving the rocks a highly flaggy character (stromatic migmatites of Mehnert, 1968). The planar fabrics are generally horizontal in Degree Square 27, accounting for the lack of relief in the countryside underlain by the Ndura Complex.

Exceptionally, in the extreme east the migmatites are tectonically interleaved with steeply dipping rocks of the Ol Doinyo Ng'iro Gneisses and form elongate kopjes. The flat areas are devoid of photolineaments apart from major negative lineaments related to master joints or faults. The flaggy nature of the steeply dipping beds is manifested by closely spaced N–S lineaments.

Three easily accessible reference sections are, as follows:

1 Along the tributary of the Barsaloi River between Lorokomonge village (GR 2515 1386) and a point west of Lusien village (GR 2566 1392).
2 That part of the Ol Ketuloni River incising into the north-eastern part of the Karisia Hills (near GR 250 137).
3 Along the stream south of the Wamba to Barsaloi track between GR 2730 1284 and GR 2745 1279 (Lkwasi area).

The Ndura Complex is composed of migmatites consisting of a gneissic palaeosome and several generations of neosome of which the concordant granite veins are volumetrically the most important. The palaeosome gneisses are regarded (at least in part) as contemporaneous with the adjacent major gneiss units (e.g. Ol Doinyo Ng'iro Gneisses in Degree Square 27). The granite veins are folded by the major tectonic episodes, i.e. they are early (Samburuan, Chapter 7 and of late Proterozoic age. No radioisotope analyses have been undertaken on the migmatites.

The Ndura Complex is generally at low tectonic levels, demonstratably basal in Degree Square 27, to the Loroki and Ol Doinyo Ng'iro Gneisses. It is affected by regional Baragoian and Barsaloian folds and older (Sabachian) parasitic folds are ubiquitous to all good outcrops. These

small-scale folds affect both the palaeosome gneisses and the neosome felsic phases.

A present thickness of at least 500 m has been estimated for the whole unit on Degree Square 27. The base of the migmatites is never exposed. Stromatic migmatites composed of banded grey biotite gneisses with from 15 to 50 per cent by volume of foliated *lit par lit* (occasionally discordant) granite veins and sheets are the characteristic lithology. A hornblende-bearing gneiss palaeosome is locally present and in the Barsaloi area a flaggy biotite gneiss with little or no granite neosome is included in the Ndura Complex. In a 1-m section measured at right angles to the gneissosity there were 40 lithological bands (alternating biotite gneiss and granite *lit par lit* veins) with the biotite gneiss bands up to 10 cm thick and the felsic veins up to 5 cm thick. Recumbent parasitic folds and slightly discordant dislocation surfaces are ubiquitous. Particularly mafic-rich gneisses tend to form boudins. A mappable metaquartzite horizon is shown in the east (GR 2740 1290) and form a low ridge of strongly lineated, iron-stained ferruginous metaquartzite. The linear fabric is defined by quartz rods. Locally limonite seams are intercalated with pure quartz seams.

All phases of the migmatitic gneisses are leucocratic. The usual banded grey gneissic palaeosome consists of quartz, and K-feldspar (untwinned microcline), plagioclase and biotite with or without minor hornblende, iron ore, apatite, red garnet (present in all lithologies in the south-east). Textures are inequigranular, xenomorphic or xenoblastic with highly variable quartz to feldspar ratios. The hornblende (partly replaced by biotite) is locally a major component of the palaeosome with andesine and quartz. The granite veins are leucocratic with feldspar (major phase) and quartz with trace biotite. Average grain sizes are about 0.4 mm; feldspar and quartz porphyroblasts are up to 2.5 mm in diameter.

The overall leucocratic nature of the Ndura Complex suggests it was originally either a major granite province or a suite of arkosic sediments prior to metamorphism. The granite veins are regarded as partial melts and the stromatic migmatitic texture (relative to the gneissic fabrics of adjacent units) may reflect the deeper tectonic setting of the Ndura Complex. Alternatively the migmatites formed an older cold basement to the precursor sediments to the gneisses. In the type area (Key, 1982) the Ndura Complex appears to define a massive western foreland to the Ol Doinyo Ng'iro Gneisses.

3.2 LOROKI GNEISSES

The Loroki Gneisses take their name from the Loroki Plateau south-east of Marti and from the Loroki (Lorogi) Forest which extends south from the plateau to cover the Karisia Hills massif east of Maralal. A basal more mafic unit, confined to the north and named the Tupa gneisses by Wilkinson (1983a & b) is incorporated into the Loroki Gneisses in this account.

The full geographical extent of the Loroki Gneisses is shown in Figure 7, essentially confined within Degree Square 27. The main outcrop areas are the Karisia Hills and Loroki Plateau. Smaller exposures on the eastern lowlands include Engineji Hill, Rarait (GR 2667 1432), Lolgume (GR 267 147), and Lusien village (GR

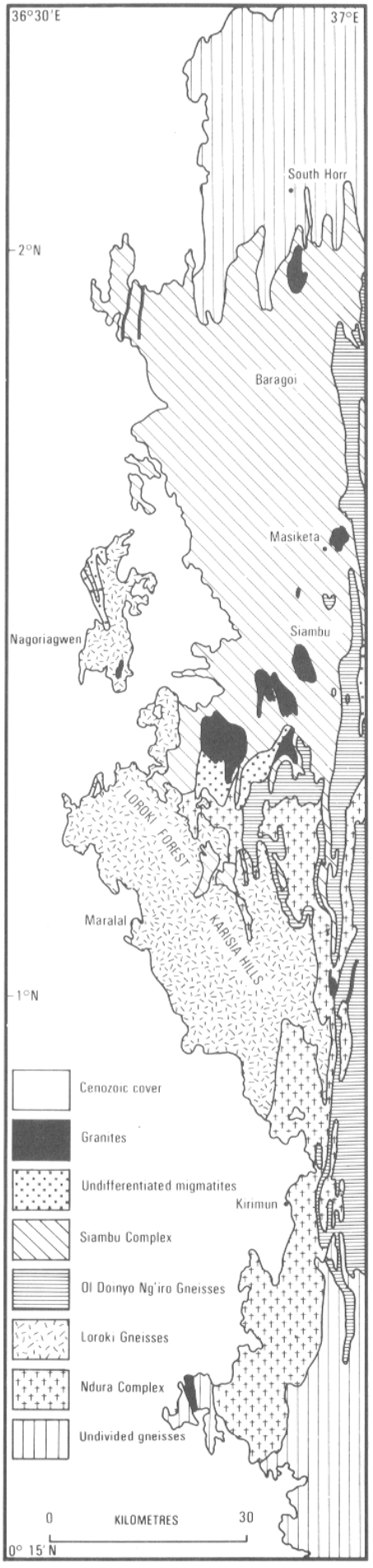


Figure 7 The distribution of the major metamorphic lithostratigraphy associations

2569 1399). Good outcrop also occurs in the Nagoriagwen inlier on the eastern shoulder of the Gregory Rift.

For the most part the Loroki Gneisses underlie high ground (in particular the Karisia Hills massif and the elevated Loroki Plateau). The definitive quartzofeldspathic gneisses form steep craggy tors (or kopjes), elongate ridges (inselbergs) or plateau uplands—due to subhorizontal (flaggy) foliation surfaces. Less competent lithologies mostly underlie lower subdued ground and are exposed in river valleys. Individual bands, e.g. of mafic gneiss, may form rounded ridges parallel to the strike of gneissosity.

Away from the thick Loroki Forest cover the quartzofeldspathic gneisses have pale, highly reflective phototones in contrast to darker tones of associated lithologies. Steeply dipping gneissosity, faults and joints form negative lineaments which are even visible through the thickest forest cover.

No single section or area defines all the lithologies of the Loroki Gneisses. Instead the following localities may be used as reference sections for specific lithologies (all on Degree Square 27) to formalise the unit.

Loole rock (Maralal area, GR 2453 1325) for massive leucocratic quartzofeldspathic gneisses.
Soit Pus rock (Maralal area, GR 2466 1253) for foliated feldspathic fine-grained gneisses.
Water tank hill, Maralal (GR 2424 1196) for porphyroblastic and fine-grained quartzofeldspathic gneisses.
Sasab River south-south-east of Barsaloi (GR 2638 1447); the area north and north-east of Longutukie (GR 2441 1530); eastern edge of the Loroki Plateau for continuous sections mainly through the quartzofeldspathic gneisses.
Quarry east of the Maralal to Rumuruti road and 2.0 km south of the main roundabout by the D.C's. office, Maralal for biotite gneisses.
Hills south of the Maralal to Barsaloi track on the outskirts of Maralal for hornblende plagioclase gneisses.
Tupa River (GR 2432 1427 to GR 2450 1428) and GR 2482 1540 for basal mafic gneisses.
Well exposed ridge south of Maralal (GR 244 120 to GR 2496 1178) for amphibolite.
Lusien village (GR 2569 1399) and the ridge south of the Ndadapo River (GR 248 131) for muscovite-bearing gneisses and schists. The last locality is most conveniently visited by driving along the forestry track to Loole rock. Good outcrops are found in the foot path trending NE off the motorable track east-south-east from Loole rock.
South-east of Maralal (GR 2490 1185) for ultramafic schist, chert and metaquartzite. A motorable track may be followed from Maralal along the silt-clogged valley which extends to the ultramafic schist exposures. The best outcrops are on the northern valley slopes.

The exact age of the precursor rocks subsequently altered to the Loroki Gneisses is not known. They predate all the major tectonothermal events described in Chapter 7, which indicates they were older than c.830 Ma. Cenozoic volcanics, and their basal sedimentary rocks of the eastern rift shoulder, unconformably rest on an elevated weathered surface of Loroki Gneisses northwards from Maralal. Both the Siambu Complex and Ol Doinyo Ng'iro Gneisses structurally overlie the Loroki Gneisses. On the Karisia Hills the Ndura Complex underlies the Loroki Gneisses but this sequence is locally reversed in the Barsaloi area.

The present (tectonic) thickness of the Loroki Gneisses varies from between 1000 and 2000 m under the Karisia Hills, decreasing to over 400 m further north. Individual lithological bands (within the quartzofeldspathic gneisses) may be up to several hundred metres thick, e.g. the amphibolite south of Maralal and the muscovite-gneisses north-east of this town.

The quartzofeldspathic gneiss is generally massive, weakly foliated, pale grey to pale yellowish brown or pink, and leucocratic with accessory biotite. However textural variations include migmatites (adjacent to the Ndura Complex), pegmatitic gneisses, porphyritic gneisses with K-feldspar porphyroblasts up to 10 cm long, and strongly foliated gneisses. Any foliation is defined by impersistent biotite wisps, but may be accentuated by flattened quartz grains and quartzofeldspathic ellipsoidal pods. The feldspar porphyroblasts are flattened by the enhanced foliation. The rocks vary from fine- to coarse-grained: pegmatites are common, as are visible red garnet, muscovite and magnetite grains.

Locally (especially around Maralal) the quartzofeldspathic gneisses are mobilised, with apophyses cutting and enclosing xenoliths of adjacent, more mafic gneisses. The xenoliths retain their original orientation. All other lithologies (described below) form concordant bands or wedges within the quartzofeldspathic gneisses.

Biotite-gneisses are fissile, flaggy, grey-brown, readily weathered (and hence poorly exposed) rocks with regular quartzofeldspathic seams separated by biotitic foliae. A measured 1-m section (in the Maralal area) at right angles to gneissosity recorded about 500 foliae. Discordant pegmatites and concordant, boudinaged hornblende-bearing seams are common. Zones of feldspar porphyroblasts extend from quartzofeldspathic gneisses into adjacent biotite-gneisses.

The hornblende gneisses are banded, rarely speckled, medium-grained rocks composed of plagioclase and hornblende in varying proportions (plagioclase amphibolites to hornblende anorthosites) with accessory quartz, diopside, magnetite and biotite visible in hand specimens. Between 80 and 140 bands were counted in 1 m sections at right angles to gneissosity with white felsic and darker hornblende-plagioclase bands. Individual bands are laterally discontinuous due to wedging and boudinage. Discordant felsic veins are rare and define ptygmatic folds. Hybrid rocks at interfaces between the hornblende gneisses and quartzofeldspathic gneisses contain golden vermiculite plates in a relatively coarse quartzofeldspathic groundmass. South of Maralal is a massive, ridge-forming, black green and white striped amphibolite which weathers to large angular blocks. It has a uniform appearance along its whole strike length with occasional veins of euhedral, clear quartz and pale green, translucent apatite. Hornblende, plagioclase, epidote and diopside grains are readily identified in hand specimen.

Muscovite-quartz gneisses are flaggy (due to a strong muscovite foliation), medium- to coarse-grained, equigranular rocks with silver muscovite flakes embedded in a white quartz groundmass. The ratio of muscovite to quartz varies between extremes of muscovite-schists and almost pure metaquartzites. Sillimanite fibres are common and biotite is present in the north. Muscovite flakes distinguish soils above these gneisses and facilitate mapping. Associated with the muscovite-gneisses are equigranular quartz-hornblende gneisses (calcareous metaquartzites). Thin metaquartzite seams and lenses are relatively common in the quartzofeldspathic gneisses. However calc-silicate pods and marble are confined to the north. Three calc-silicate pods (at GR 2414 1900, GR 2428 1893, and GR 2432 1890) consist of massive, dense, coarse-grained, red and green mottled assemblages of garnet, epidote, quartz and plagioclase. A single marble (at GR 2367 1690) is typically, pale grey-blue to white, massive, homogeneous and coarse-grained.



Plate 8 Typical quartzofeldspathic gneisses of the Loroki Gneisses

Ultramafic rocks are equally rare and confined to two small areas near Maralal. The principal lithology is a pale green rock with a felted texture due to rosettes of actinolite and anthophyllite. Black amphibolite, dark brown ferruginous metaquartzites and pale yellow-brown chert breccias crop out with the ultramafics. Trenches expose amphibole asbestos veins up to 1 m thick with the asbestos fibres orientated at right angles to the vein walls. Minor quartz reefs contain cordierite.

Parasitic folds, ductile and brittle shears, boudinage, and transposition of foliation were observed in most outcrops of all lithologies. All the more felsic lithologies are cut by quartzofeldspathic vein phases, both concordant and discordant to any gneissosity (Plate 8).

The quartzofeldspathic gneisses have equigranular and inequigranular xenomorphic textures in which the combined feldspar content usually exceeds that of quartz. The major feldspar phase is microcline as a late replacement of quartz and oligoclase (which is sericitised). Quartz grains are embayed by microcline but may also be shape orientated to define planar and linear fabrics. Biotite is the only mafic phase as impersistent foliae, locally replaced by chlorite. Accessories include magnetite (altering to martite), apatite, rounded sphene, elongate but subrounded zircon, muscovite and garnet.

The hornblende gneisses have up to 60 per cent of the mode composed of green hornblende with varying amounts of epidote, plagioclase (saussuritised), quartz, and diopside with accessory biotite, zircon, ore, microperthite and microcline. Aligned hornblende prisms may define planar fabrics in inequigranular, sutured mosaics of strained quartz grains. Around Maralal, diopside is the major modal phase as poikilitic plates including plagioclase and hornblende. Over 70 per cent of the mode of the hornblende-quartz gneisses spatially related to the muscovite gneisses is recrystallised quartz.

The muscovite gneisses have xenomorphic-granular groundmasses dominated by quartz, with minor oligoclase and microcline. Hematite dust may coat the felsic grains. Variable amounts of poikilitic muscovite flakes define a strong foliation. Sillimanite as isolated needles or rosettes is common, but biotite and red garnet are only of local significance.

Green hornblende prisms dominate the modes of the

amphibolites. Epidote, plagioclase (oligoclase-andesine range) and diopside preferentially form equigranular mosaics in discrete bands. Quartz and sphene occur as sporadic accessories.

The marbles are essentially monomineralic aggregates of coarse carbonate plates; fibrous tremolite may or may not be present.

In the northern calc-silicate the main phases (identified in hand specimen) form equigranular aggregates. Garnet grains have sieved textures with accessory hornblende, biotite and rutile needles.

Banding in the rare metaquartzites is both due to varying size of constituent quartz grains and the presence or absence of oligoclase. A weak, impersistent foliation is defined by biotite flakes.

Quartz is also the principal phase in the banded biotite gneisses and exceeds the combined modal content of feldspar (oligoclase and microcline). Biotite in foliae and in the felsic groundmass is the major mafic phase with accessory muscovite, epidote, ore and garnet. Microcline is a late replacive phase (as in the quartzofeldspathic gneisses).

Actinolite intergrowths occur in anthophyllite rosettes in the ultramafic schists, which have sericite and talc groundmasses. The associated cherts have limonite stained opaques-rich matrices cut by chalcedony seams and chlorite veins.

Therefore the following mineral assemblages are represented by lithologies within the Loroki Gneisses.

Quartz + K-feldspar + plagioclase + biotite (chlorite) ± magnetite, apatite, sphene, zircon, muscovite, garnet.
Hornblende + epidote + plagioclase + diopside ± quartz, biotite, zircon, K-feldspar, sphene.
Quartz + microcline + plagioclase + muscovite ± hematite, sillimanite, biotite, garnet.
Carbonate ± tremolite.
Epidote + garnet + quartz + plagioclase ± hornblende, biotite, rutile.
Quartz ± oligoclase, muscovite, sillimanite.
Anthophyllite + actinolite + sericite + talc.
Limonite + chlorite + chalcedony.

The Loroki Gneisses are interpreted as a sequence of altered clastic (arkosic) sediments: the minor amphibolite and ultramafic units are possible altered basic-intermediate lava flows/or sills and ultrabasic intrusives respectively. Other lithologies, e.g. hornblende gneisses, may have originally been more calcareous sediments.

They are lithologically similar to the extensive Kotim Gneisses underlying the more easterly high ground (Kotim range northwards to the Ndotos). This would suggest a major phase of continental clastic sedimentation prior to the onset of the Mozambiquian Orogeny in northern Kenya.

3.3 OL DOINYO NG'IRO GNEISSES

This paragneiss unit takes its name from the village of Ol Doinyo Ng'iro, about 60 km SE of Maralal, in the eastern part of the Kirimun sheet (92/4) of Degree Square 35. East to west sections around Ol Doinyo Ng'iro which overlap into the Longopito area (Sheet 93/3) were used by Key (1982a & b) to define formally the Ol Doinyo Ng'iro Gneisses. The present mapping confirmed a northern extension by Shackleton (1946) of these paragneisses towards the area east of Siambu Hill: in fact the whole eastern map margin of Degree Square 27 is underlain by steeply dipping Ol Doinyo Ng'iro Gneisses. This fissile paragneiss unit is utilised by a major (Barsaloian) straightening zone to account for its uniform N–S strike which is reflected by the countryside's morphology. The Kawab Klippe on the northern map margin also consists of Ol Doinyo Ng'iro Gneisses which are extensively developed northwards. In the main eastern area linear N–S strike ridges, mostly composed of massive quartzofeldspathic gneisses, are separated by narrow valleys with good outcrop of fissile biotite gneisses and related metasediments (see below). Watercourses within the N–S valleys form tributaries to the major drainage channels which in fact flow eastwards along the traces of 'late' major fractures. These major rivers incise impressive gorges through the quartzofeldspathic gneiss ridges. The countryside is spectacularly displayed on aerial photographs: pronounced N–S lineaments are related to gneissosity and cut by the valley traces of the discordant E–W fractures.

Closures of major, tight folds shown on the geological map are easily traced on aerial photographs as the major lithologies are tonally different, e.g. marbles are pale grey and homogeneous relative to lighter, harsher quartzofeldspathic gneisses and darker biotite gneisses full of gneiss-related lineaments. The Kawab Klippe forms a major hill due to the presence of competent gently dipping quartzofeldspathic gneisses.

The section along the Lkwasi stream from GR 2773 1273 upstream to GR 2918 1292 is easily accessible from the Barsaloi to Wamba track and is a good reference section for the Ol Doinyo Ng'iro Gneisses.

Charsley (1984) notes that most of the streams draining into the Naupa Guruk and Terengiwe Rivers are dip streams relative to the trend of the gneisses, so that numerous good sections occur along the length of the Oruku Range. The best access is north from the Barsaloi River. A measured section (GR 2758 1562) in biotite schists and marbles within a graphite gneiss unit provided the following sequence from structural top downwards.

1 m +	Compositionally banded medium-grained hornblende-mica gneisses with rare sillimanite – quartz bands.
30 m	Dark green, weathering rusty brown, compositionally banded schist with a few banded amphibolites up to 3 m thick.
6 m	Banded grey and white graphite marble with some discontinuous schist bands up to 10 cm thick.
8 m	Biotite (+ graphite) schist.
2 m	Poorly banded marble.
1 m	Rusty weathering medium-grained biotite gneiss with a 6 cm thick quartzite.
7 m	Banded and boudinaged grey and white graphitic marble.
1 m +	Quartz-sillimanite-mica gneiss.

The age of the original sediments must be older than about 830 Ma—the approximate time of the gneissosity generation. A fuller discussion of the timing of the various tectonothermal events recorded in the Ol Doinyo Ng'iro Gneisses is given in Chapter 7.

The Ol Doinyo Ng'iro Gneisses confined within the main Barsaloian straightening zone have steeply dipping tectonic contacts with adjacent units (*see* map cross sections). They are locally interdigitated with migmatites as well as other gneisses because of horizontal tectonic transport during an early tectonothermal event. The gneisses of the Kawab Klippe overlie Siambu Complex rocks.

Charsley (1984) suggests an unfolded thickness of about 1250 m for the Ol Doinyo Ng'iro Gneisses in the Barsaloi area. Individual lithological bands vary in pre-

sent thickness from less than 1 m for metaquartzite seams to about 250 m for the major quartzofeldspathic and graphitic gneisses.

The main lithologies are banded grey biotite gneisses (and schists), fissile graphitic gneisses, marbles, banded hornblende gneisses, quartzofeldspathic gneisses and metaquartzites with rare metacherts and calc-silicates.

The biotite gneisses are typically, dark brown weathering, grey, fine- to coarse-grained, quartz-oligoclase-biotite rocks with a strong biotite foliation. Invariably there are concordant (to foliation) bands of the other lithologies within a biotite gneiss outcrop (Plate 6). Red garnet is ubiquitous (to all lithologies) in the south-east as well as further north where sillimanite and microcline grains may also co-exist in the biotite gneisses. The graphitic gneisses are basically the biotite gneisses with modal graphite: up to 20 per cent, but mostly less than 5 per cent. The major graphite gneiss member is at Lareodo (GR 2745 2068) where a complete hill section exposes several hundred metres of friable, dark grey, locally black, medium-grained ($\leq$1 mm graphite flakes), finely banded rocks. An interesting feature of these graphite rocks is the presence of various green minerals including green mica, kyanite, and green (tsavorite) garnet as well as an orange, transparent, glassy silicate. Further south between Soit Ng'iro and the Barsaloi area the graphitic gneisses form persistent units with strike lengths of over 20 km. They are more typically, fine-grained, finely laminated, friable dark grey rocks with slivers of quartz aggregates accentuating a micaceous foliation. Slices of marble and metaquartzite are common to all major graphitic gneiss units.

Marbles are common (a typical characteristic of the Ol Doinyo Ng'iro Gneisses) throughout the eastern Barsaloian domain and on Kawap. They always form pods rather than continuous bands. The marbles are pale blue-grey, medium-grained, massive, essentially pure carbonate rocks apart from occasional traces of graphite, forsterite (replaced by serpentine), phlogopite, epidote and actinolite visible in hand specimen.

The hornblende-gneisses are dark grey weathering black-brown, fine- to medium-grained rocks with a pronounced finely spaced banding. Quartz-plagioclase-hornblende bands are separated by monomineralic hornblende seams imparting a strong flaggy character especially in major fold limbs in the east.

Over most of their outcrop area the quartzofeldspathic gneisses are buff, pale yellow to pale greyish orange in colour, thick to massively banded, medium- to coarse-grained, commonly saccharoidal, leucocratic rocks. Accessory biotite and magnetite weather to produce an orange or red spotting. Sillimanite and red garnet are common accessories with rare graphite (GR 2755 1618) adjacent to graphitic gneisses. Local complete recrystallisation and mobilisation of the quartzofeldspathic gneisses has produced allochthonous granites which are described in Chapter 6.

The numerous metaquartzites form thin but persistent, grey to off-white, fine- to medium-grained, essentially monomineralic seams. Flakes of muscovite (GR 2720 1198), sillimanite fibres and garnet porphyroblasts may be present in the equigranular quartz groundmass. The metaquartzite cropping out in the Kileshwa River, according to Baker (1963) is a flaggy and lineated, locally fissile rock. A banding is defined by quartz-rich and quartz-poor (feldspathic) bands. A 30 cm thick, finely banded pale grey metachert is exposed in the Mureiyia River (GR 2713 1248). It is associated with finely banded biotite gneisses within a graded sequence strongly suggestive of a greywacke assemblage.

Calc-silicates are either medium- to coarse-grained, dark green, massive rocks with individual minerals readily visible in hand specimen or friable, yellow-green rocks. These last mentioned rocks are fine-grained and equigranular and form thin concordant seams within biotite or graphite gneisses.

Quartzofeldspathic veins and pegmatites, parasitic folds and related minor structures (Chapter 7) are common to most lithologies (except the marbles, metaquartzites and metachert). Petrographically the biotite gneisses and hornblende gneisses are identical to the similar lithologies described from the Loroki Gneisses. Within the graphite gneisses, graphite flakes either form within biotite foliae or are dispersed at felsic grain boundaries. Sillimanite and muscovite are common accessory minerals. The marbles usually have equigranular textures dominated by equant calcite and dolomite grains. Calc-silicate, graphite and silicate phases form descrete subhedral grains in the carbonate groundmass. Inequigranular xenoblastic and granoblastic textures are described for a marble from the Barsaloi area by Charsley (1984). The quartzofeldspathic gneisses have equigranular to inequigranular xenoblastic textures of quartz, microcline and oligoclase with up to 10 per cent of magnetite, plus accessory biotite, garnet, sillimanite, graphite, zircon and muscovite; combined feldspar content exceeds 50 per cent of the mode. Microcline grains replace both quartz and oligoclase with quartz locally forming elliptical clusters of equigranular grains. Magnetite porphyroblasts are altered to martite.

The metaquartzites are inequigranular aggregates of variably strained quartz grains with minor (large) feldspar, sillimanite, muscovite, magnetite (dust). The metaquartzite of the Kileshwa River is a feldspathic granitised quartzite according to Baker (1963). Feldspar growth is confined to discrete bands enveloping isolated quartz grains in a 'net-texture'. Orientated flakes of biotite and an altered mica define some bands. Charsley (1984) notes the following calc-silicate assemblages:

Quartz + scapolite + garnet + calcite + actinolite + epidote + oligoclase
Garnet + epidote
Actinolite + epidote + calcite + zoisite + garnet

All of the Ol Doinyo Ng'iro Gneisses are regarded as metasedimentary in origin: original facies variations produced the different lithologies (carbonates, calcareous sediments, arkoses, etc). The quartzofeldspathic gneisses are regarded as altered, relatively clean (hence subsequent simple quartz-feldspar-magnetite mineralogy) arkoses. Melting and related upward mobilisation (usually above major fold axes) of the quartzofeldspathic gneisses has formed intrusive foliated granites (or granite gneisses) which are described in Chapter 7.

3.4 **UNDIFFERENTIATED MIGMATITES** (after Charsley, 1984)

Charsley (1984) mapped and described a suite of undifferentiated migmatites from the Barsaloi area. He noted the main outcrop areas and defined reference sections at the following localities:

Doinyo Uasin Hill (GR 2603 1445); the small hill besides the Barsaloi to Maralal track at GR 2606 1434; in the Barsaloi River tributary between GR 2518 1423 and GR 2530 1438; at GR 2762 1583 and GR 2763 1560.

The migmatites are believed to be a sole unit underlying the major thrust which brought the Siambu Complex into tectonic contact with the other major gneiss units. As such it is probably composed of paleosome thrust slices belonging to more than one gneiss group, and including some of the Siambu Complex gneisses.

It is not possible to estimate a thickness for the migmatites. They are intimately interthrust into the Ol Doinyo Ng'iro Gneisses (as tectonic slices too small to show on the 1:250 000 geological map). Elsewhere the migmatites lie between the Siambu Complex and the Ndura Complex or Loroki Gneisses (Plate 9). Locally this relationship is obscured by (G2) granite intrusions.

The migmatites are massive rocks commonly with a gneissic paleosome and a leucocratic (commonly gneissic) neosome phase. In the eastern tectonic slices the neosome forms bands (boudinaged and flow folded) with a homophanous (Mehnert, 1968) phase locally present. At Doinyo Uasin, a leucogranite phase with schlieren and rafts of country rock dominates a banded mafic gneiss paleosome. Elsewhere the migmatites have raft structures and agmatitic textures with several neosome (monzonitic to granite) vein phases.

Ragged xenoblastic, inequigranular textures seen in thin section consist of microcline, quartz, oligoclase and biotite with trace apatite. Two generations of quartz include early rounded intergrowths with feldspar (myrmekite textures were also locally seen).

The tectonic setting of the undifferentiated migmatites explains their polygenetic origin, from adjacent units.

3.5 SIAMBU COMPLEX

The Siambu Complex is named after Siambu Hill (GR 2707 1630) which has a core of ultramafic rocks and is surrounded by rock units typical of the complex. Its full geographical extent (essentially confined to Degree Square 27) is shown in Figure 7. Its fan shaped plan is primarily due to the fact that the Siambu Complex forms the core of the open Morilem Synform plunging gently NNW. Its western extremity is concealed beneath Cenozoic volcanics and the major Barsaloian straightening zone defines the linear eastern boundary. Baragoi is situated almost exactly in the centre of the complex which has a maximum E–W width of about 30 km and a N–S length of over 100 km (in the north it interfingers with the Ol Doinyo Ng'iro Gneisses and it is difficult to demarcate exactly the northernmost extremity, see Dodson, 1963).

The Siambu Complex underlies a dissected plain (mostly the El Barta Plain) separating the Ndoto mountain range from the uplifted eastern shoulder of the Gregory Rift.

However the topography is generally impressively rugged because of (1) deep and extensive dissection, notably in the south and east by the Barsaloi River system, and (2) major hills of altered felsic and mafic intrusives.

Thick residual soil cover is confined to the small undissected remnant of the El Barta Plain. Elsewhere there is good rock outcrop which has contributed towards the distinctive dark tones of the Siambu Complex on aerial and satellite photographic imagery. Subtler detail reflects the contrast between very dark areas underlain by ultramafics (especially hornblendites) and highly reflective white areas of quartzofeldspathic gneisses and altered felsic intrusives. Numerous pegmatites and late fractures define discordant, linear positive and negative lineaments respectively on aerial photographs. On satellite imagery the plan of the Morilem Synform is starkly clear with a

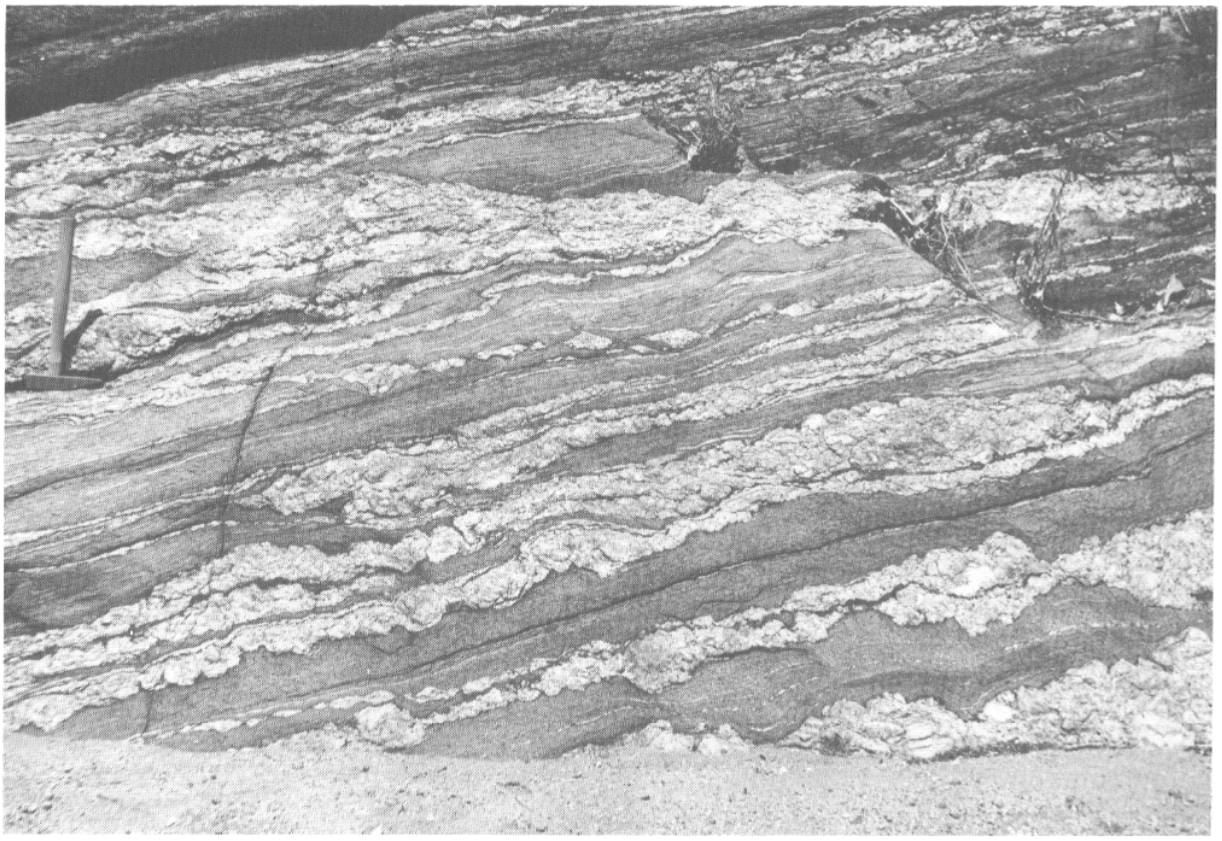


Plate 9 Migmatites exposed in the Nagoriagwen inlier

dark core of the altered mafic and ultramafic rocks around Baragoi enveloped by an almost white zone underlain by mixed felsic and mafic rocks lacking a thick soil cover. Siambu Hill forms a dark area in the southern nose of the synform.

The following, relatively accessible sections provide good exposures of the various lithologies. In the extreme north a traverse eastwards from the Barsaloi to South Horr track immediately north of Sartim along northing 219 (a major river valley) commences with good outcrop of the Sartim Granite (recumbent fold closure) for 1.8 km. This is followed by river exposures of steeply dipping hornblende migmatites (for 0.5 km) and biotite hornblende gneisses (further 0.45 km). A 300 m wide ridge of massive amphibolite deflects the stream northwards and is overlain eastwards by a sequence of plagioclase amphibolites with boudinaged ultramafic pods for about 2.25 km. These form the core of a synform so that the amphibolite (for 0.8 km) and biotite-hornblende gneisses (for 1.3 km) reappear to the east. The traverse then crosses a poorly exposed plain, apart from a granite hill, before meeting the Baragoi to Ilaut track exactly 10.3 km from the start of the traverse.

Still in the north, Kotikal Hill has good outcrop of layered metagabbro (plagioclase amphibolite) and ultramafic rocks. The Baragoi River upstream of Baragoi also provides good outcrops of altered basic intrusives including sheeted dykes in a metagabbro. Further south, in the Masiketa area there are ridges of calc silicate (GR 2650 1825). Upstanding marble ridges are present in the same area by the track from Baragoi to Latakweny. A 14.4 km long traverse towards 244° (true), from Masiketa to the Baragoi–Barsaloi track at Lorregal, provides an excellent section through the Morilem Synform. The traverse starts in amphibolite, well exposed in Masiketa, and then crosses well exposed hornblende gneisses full of pods of ultramafic rock on all scales, until Morilem Hill (serpentinite) is reached after 4 km. The next 4 km through the core of the synform provides good outcrop of hornblende-biotite and biotite gneisses before the hornblende gneiss with ultramafics reappears (for a further 3 km). The last part of the traverse provides good outcrop of biotite and hornblende-biotite gneisses before reaching the Lorregel ultramafics alongside the Barsaloi track.

In the Barsaloi area the best exposures of ultramafic rocks are in the deep valleys on the south side of Siambu Hill. In the stream valley between GR 2714 1610 and GR 2711 1618 there is a long section in actinolite hornblendite with evidence of local alteration to tremolite and talc (GR 2713 1612). Serpentinised dunite occurs in a side stream (GR 2710 1618) overlain by extensive calcrete. A typical mixed-lithology section about 100 m long with tectonic contacts between units is exposed on the west side of Ol Doinyo Narok—a thick amphibolite (GR 2622 1589). The sequence from the structural top downwards is as follows:

Thinly foliated and banded grey hornblende-biotite gneiss
Homogeneous grey biotite-hornblende gneiss
Melagneisses and amphibolite
Actinolite-tremolite-talc rock
Garnet-quartz-feldspar-hornblende gneisses
Hornblende-plagioclase gneiss
Amphibolite

The best section demonstrating the intimate association of the main gneiss units with (their) partial melt products is along the Sitet River before its confluence with the Barsaloi River (GR 2641 1504). Here 87.4 m of leucocratic to melanocratic gneisses with various intrusive phases including coarse-grained granite and pegmatites are well exposed.

The Siambu Complex has suffered the same polyphase tectonothermal deformation as other eastern metamorphic rocks, i.e. the original unaltered rocks were older than c.830 Ma. Three mineral ages of c.585 Ma, 635 Ma and 835 Ma have been obtained from pegmatites in the Nachola area (incompletely documented in an internal Mines and Geological Survey Department geochronology file). The youngest age is from K-Ar dating of feldspars and the other two are Pb ages on monazites.

The Siambu Complex is generally at the top of any structural sequence except where overlain by the Kawap Klippe of Ol Doinyo Ng'iro Gneisses. It is probable that certain metasediments within the complex are tectonic slices from other units, in particular marble and graphitic gneisses from the Ol Doinyo Ng'iro Gneisses.

In excess of 4000 m of Siambu Complex rock is preserved in the core of the Morilem Synform. Individual lithological units do not have constant thicknesses, e.g. the ultramafics are boudinaged on all scales, and vary up to several hundred metres maximum thickness.

The complex consists of a number of distinct metamorphic rocks, including altered sediments and igneous extrusive and intrusive phases, tectonically interleaved and metamorphosed with local partial melting. The tectonic interleaving has occurred on all scales and the geological map is a gross oversimplification of the actual lithological relationships. Nine lithologies are represented on the geological map, although mafic and the early (G1) partial melt products derived during initial metamorphism of the Siambu Complex are also included in this unit together with the altered shonkinite at Baragoi.

The most common rock is a hornblende gneiss, with or without biotite bands, medium grey in colour and well foliated. It invariably contains bands or pods of other lithologies as well as metamorphic reaction zones, e.g. coarse hornblende clots around felsic (quartzofeldspathic with epidote) pods. If biotite is absent the rocks are essentially hornblende-plagioclase gneisses. Garnet and epidote are common in these rocks east of Baragoi where they are banded due to white feldspathic seams. Migmatitic textures are also common, with discordant hornblende-rich veins and diffuse margins to bands.

Baker (1963) provides complete descriptions of the various ultramafic rocks within the main northern part of the Siambu Complex. As he notes, the primary mineral assemblages are rarely preserved in any of these rocks. A black massive metapyroxenite forms low ridges immediately west of Masiketa, and the shonkinite south of Baragoi retains primary bronzite and olivine grains. Mostly the ultramafics are soft, essentially homogeneous unfoliated but felted, melanocratic or hypermelanic rocks in shades of brown or greenish brown. Thin magnesite and amphibole asbestos veins are common and they are variably cut by quartz veins, pegmatites, aplites and intrusive plagioclase amphibolites. Common lithologies are tremolite-anthophyllite rocks variably altered to talc-bearing (locally pure talc) schists, chlorite schists with or without magnetite octahedra, serpentinite (with or without chromite), talc-actinolite rocks with vivid green radiating actinolite needles and talc-carbonate serpentinite. The larger ultramafic bodies are mostly serpentinites. The unit of hornblende gneisses with ultramafic pods on the geological map consists of the dark grey hornblende gneisses with pods of the talc-actinolite/tremolite-anthophyllite rocks on all scales from less than 1 cm up to

mappable bodies. The massive banded amphibolites shown on the geological map include melanocratic and hypermelanic green to black, coarse-grained, commonly monomineralic amphibolites, dominantly hornblendites and banded plagioclase amphibolites regarded as metagabbros. Textural variations in massive amphibolites include the presence of hornblende porphyroblasts, fibrous and schistose rocks. All fabrics are defined by hornblende prisms and are of tectonic origin; in only rare cases is a relict mineral banding preserved between hornblende foliae. Quartz float is common over amphibolite outcrop.

The banded plagioclase amphibolites are variable in texture and composition with fine- to medium-grained, mesocratic to melanocratic poorly foliated rocks grading into coarse-grained, mottled variably banded rocks of obvious gabbroic origin. Individual hornblende and plagioclase grains may define augen in strongly sheared rocks. Baker (1963) notes that these metagabbros become increasingly leucocratic northwards. Kotikal Hill is essentially formed of an altered layered gabbroic anorthosite: anorthosite layers up to 10 cm thick are separated by hornblende-plagioclase seams of similar thicknesses. The banding is an early (metamorphic) fabric cut by a later foliation. Around Baragoi relict igneous textures are preserved in pods within metagabbro. Shear zones tend to reduce the grain size of the various amphibolites with concomitant biotite growth. Further metamorphic alteration of the amphibolites (described by Baker, 1963) produces a crude stratification with granular feldspathic lenticles alternating with schistose hornblende and biotite-rich layers. The development of sieved red garnets as well as replacive plagioclase and biotite are characteristic of these altered rocks. The metagabbro around Baragoi is cut by numerous fine-grained biotite-rich dykes and sills which appear to post-date the metamorphic mineral assemblage of the host rock, i.e. the dykes probably do not represent an original sheeted dyke complex.

The fissile biotite-gneisses are leucocratic, mostly medium-grained well-foliated, frequently flaggy rocks with muscovite flakes, garnet porphyroblasts, sillimanite fibres and minor amounts of graphite flakes often visible in hand specimens. Many of the biotite-gneisses developed from shearing of hornblende-bearing gneisses. Discordant K-feldspar-bearing veins caught up in any deformation form K-feldspar augen. Deep to pale grey-brown metaquartzites form sporadic lenticular bodies commonly with strong tectonic fabrics: foliation, lineation, autobrecciation. Pyrite cubes are visible in a small metaquartzite cropping out about 5 km SW of Baragoi. Rare massive leucocratic quartzofeldspathic gneisses are identical in appearance to the extensive quartzofeldspathic gneisses of the Ol Doinyo Ng'iro Gneisses. Other minor lithologies with modal quartz, too small to be shown on the geological map, include garnet-quartz-feldspar-hornblende gneiss and garnet-quartz-limonite-graphite gneiss (both spatially related to southern ultramafic pods).

Marbles are common throughout the Siambu Complex, most are massive, pale greyish blue coarse-grained rocks of dolomitic appearances. Minor graphite, phlogopite and actinolite may co-exist with the carbonate grains. Pegmatite lenses in any marble are surrounded by soft, pale green sheaths of tremolite and talc (Baker, 1963). The polyphase tectonic history of the Siambu Complex is illustrated on outcrop scale by the interference folds of banding in the calc-silicate rock west of Masiketa.

Migmatitic rocks are widespread in the main northern area of the Siambu Complex and grade into ungranitised rocks at one extreme, and into essentially homogeneous rocks at the other extremes.

The mineralogy and texture of the migmatites (described by Baker, 1963) is determined by the precursor rocks. In the heterogeneous rocks there are relicts of the precursor rocks cut by partial melt products of the migmatisation as well as surrounded by diffuse zones of recrystallised migmatite. The more homogeneous rocks characteristically have diffuse pegmatitic bodies with large hornblende crystals. Shear zones cut and further complicate the migmatite textures, usually producing feldspar augen. Hornblende is the common mafic phase although small areas of biotite migmatite are delineated by Baker (1963). The partial melt phase is generally monzonitic, or trondjhemitic.

The shonkinites which are exposed south of Baragoi are well described by Baker (1963). They are unfoliated dark massive rocks; genetically related to the other surrounding more altered basic and ultrabasic rocks. Baker (1963) states, 'The shonkinite consists largely of pale reddish grey schillerized bronzite in interlocking grains reaching 2 mm in diameter.

Biotite, pleochroic from straw to deep brown occurs as interstitial irregular replacive growths between bronzite grains, and locally there are fine-grained intergrowths of biotite and bronzite between the two minerals. Occurring as scattered inclusions in the bronzite are olivine grains, which are frequently clustered about ore grains. All the larger olivine grains are altered to iddingsite along cracks. They are surrounded by reaction coronas wherever they occur in contact with the alkali feldspar that comprises the bulk of the remainder of the rocks. Next to olivine in these coronas is an aggregate of clear grains which are possibly enstatite. This zone is succeeded by another consisting of minute, bottle-green pleonaste granules in a matrix of fibrous amphibole occasionally associated with biotite flakes. Other intergrowths occur locally between biotite and the alkali feldspar. The feldspar occurs in slightly turbid, untwinned crystals of late crystallisation or in aggregates of grains with sutured margins and patchy polarisation, and appears to be an alkali type. The approximate composition of the rock by volume is as follows:'

Bronzite	40
Biotite	30
Feldspar	20
Olivine	5
Opaque ore	5
Apatite	– '

Minor intermediate intrusive rocks of variable mineralogy are intimately associated with the older rocks of the Siambu Complex and are included within this unit. They are dark greenish grey, pale grey to pale brown; both foliated and massive; fine- to coarse-grained; equigranular, inequigranular and porphyritic. They form concordant (sills) as well as discordant bodies (mostly dykes) on all scales up to the small stock shown on the geological map.

The banded hornblende-biotite gneisses generally have inequigranular, xenoblastic textures with late biotite foliae discordant to any mineral banding. The main mineral phases are hornblende (locally as sieved porphyroblasts), quartz (highly sutured and strained), oligoclase and biotite with accessory muscovite, epidote, garnet, microcline, apatite, carbonate, sphene, zircon, and opaque iron ore. Aligned prismatic hornblende and shape-orientated quartz grains accentuate mineral banding in southern zones. Typical mineral assemblages are as follows.

Quartz + plagioclase + biotite + hornblende
Quartz + plagioclase + biotite + K-feldspar
Quartz + biotite + hornblende + epidote + microcline

Biotite, some quartz and K-feldspar are late crystallisation phases; biotite may breakdown to muscovite.

The ultramafic rocks have xenomorphic or xenoblastic textures with the fibrous rocks showing decussate aggregates in thin section. Petrographic detail of all the different ultramafics (and in fact all the metamorphic rocks) is given by Baker (1963). The major mineral phases are talc, serpentine, olivine, chlorite, tremolite, actinolite, anthophyllite and magnetite with variable amounts of magnesite, plagioclase, carbonate, hornblende, diopside, chromite, biotite, epidote and sphene. Typical mineral assemblages are as follows:

Olivine + diopside + spinel + serpentine
Actinolite
Tremolite + actinolite + talc
Tremolite + carbonate
Anthophyllite
Chlorite + magnetite
Serpentine + chlorite + tremolite
Serpentine + magnetite + tremolite + muscovite + calcite
Serpentine + olivine + magnetite
Serpentine + chlorite + tremolite + magnetite
Serpentine + olivine + tremolite
Talc
Epidote + actinolite + diopside + tremolite + magnetite
Talc + tremolite + carbonate + anthophyllite + actinolite

Chromite is confined to pods in certain serpentinites; Baker (1963) reports lilac mica (?stichtite) in one such pod. Magnesite forms thin veins in serpentinites which also have microscopic networks of chrysotile in brown altered (to serpophite) antigorite groundmasses. Primary igneous textures are not preserved and, except for olivine, no primary mineral phases remain. Charsley (1984) suggests the following sequence of mineral growth.

Olivine → tremolite + magnetite → serpentine + magnetite + chlorite
Olivine (? + pyroxene) → fibrous amphibole + magnetite + chlorite → serpentine + magnetite + chlorite
Olivine → serpentine + carbonate

The hornblende-plagioclase gneisses and amphibolites have inequigranular or equigranular, xenoblastic or granoblastic textures, locally with poikiloblasts of hornblende and rarely plagioclase. The principal minerals are obviously hornblende with or without plagioclase together with variable actinolite, carbonate, magnetite, garnet, diopside, scapolite, quartz, biotite, muscovite, epidote, sphene, apatite and chlorite. The plagioclase composition varies between extremes of oligoclase and bytownite but is mostly in the range andesine to labradorite.

Biotite is a late phase associated with granitisation processes whereas two generations (pre- and syn-hornblende growth) of epidote are recorded. The amphibole is frequently seen to be altering to uralitic hornblende (Baker, 1963). No primary igneous textures are preserved although the more obvious metagabbros have relatively coarse-grain sizes. Relict pale green clinopyroxene cores to poikiloblastic hornblende are rare. Typical mineral assemblages include:

Hornblende
Hornblende + plagioclase + epidote
Hornblende + plagioclase + epidote + garnet
Hornblende + plagioclase + K-feldspar + quartz
Hornblende + plagioclase + quartz + garnet
Hornblende + plagioclase + clinopyroxene

The biotite gneisses have equigranular and inequigranular, xenoblastic and granoblastic textures which may contain feldspar and biotite porphyroblasts. Cataclastic fabrics are localised in major shear zones. Quartz is the major felsic phase and microcline is a late replacive phase of quartz and oligoclase. Biotite is the principal mafic phase with variable, but minor amounts of modal muscovite, epidote, garnet, hornblende, sillimanite. Mineral assemblages include:

Quartz + microcline + muscovite, biotite, sillimanite
Quartz + microcline + oligoclase + muscovite
Quartz + K-feldspar + oligoclase + biotite + garnet + sillimanite
Quartz + K-feldspar + microcline + oligoclase + biotite + muscovite

The metaquartzites are mostly pure quartz rocks with equigranular, inequigranular and cataclastic textures. Secondary oligoclase grains tend to envelope smaller quartz grains. Traces of pyrite, muscovite, epidote, garnet, sillimanite, hornblende, graphite and limonite may occur in the quartz groundmass. Mineral assemblages include:

Quartz + oligoclase
Quartz + muscovite + sillimanite
Quartz + hornblende + magnetite + garnet
Quartz + garnet + graphite

Rare quartzofeldspathic gneisses have quartz and oligoclase lenses set in a quartz-rich equigranular matrix with traces of muscovite, i.e. the assemblage:

Quartz + oligoclase + muscovite

For the most part the marbles are equigranular aggregates of calcite and dolomite with flecks of graphite erratically dispersed in the carbonate groundmass. The calc-silicate west of Masiketa has the following mineral assemblage:

Calcite + dolomite + plagioclase + quartz + sphene + garnet + diopside + phlogopite + epidote

The hornblende migmatites are coarse-grained with inequigranular textures containing poikiloblastic hornblende, feldspar, biotite, sparse subangular quartz, and diffuse trails of small hornblende grains. Biotite is a late overgrowth in discrete foliae or in quartz aggregates. Accessory phases are apatite, sphene, garnet, and clinozoisite.

The intermediate intrusives are composed of varying proportions of hornblende, biotite, plagioclase, K-feldspar, quartz and epidote with accessory apatite, sphene and zircon in xenomorphic, mostly inequigranular textures. Metamorphic mineral assemblages include:

Hornblende + biotite + plagioclase + quartz + epidote
Hornblende + microcline + quartz
Hornblende + biotite + plagioclase + K-feldspar + quartz

Five chemical analyses of representative rocks of the dominant hornblende-plagioclase gneisses (2), an ultramafic (metaperidotite), a hornblende amphibolite (metagabbro) and an early melt phase are given in Table 1. The hornblende gneisses and melt plot in the fields of andesites (diorites) or basaltic andesites on a silica against alkali diagram (Cox and others, 1979). Lower silica and alkali contents confirm the ultramafic and ultrabasic character of the serpentinised peridotite from the Siambu Hill. The plagioclase amphibolite plots in the gabbro field on a silica against alkali diagram. The polyphase metamorphic history of the samples probably involved some chemical redistribution during new mineral growth so the present whole rock chemistry need not relate exactly to original igneous compositions. However the analyses do appear to

substantiate the petrographic evidence that the hornblende plagioclase gneisses represent altered intermediate volcanics or volcaniclastic sediments.

Table 1 Whole rock analyses from the Siambu Complex

	C782/4	C782/46	C782/47	C782/63	27/219*
SiO_2	60.4	55.1	52.8	40.8	50.64
Al_2O_3	13.82	13.08	10.14	1.44	18.75
FeO	6.61	5.17	6.90	6.47	3.70
Fe_2O_3	1.95	1.08	0.69	5.38	1.40
CaO	7.28	8.68	15.12	1.68	10.32
MgO	4.64	9.47	9.27	34.67	9.05
K_2O	0.36	1.63	0.72	0.06	1.05
Na_2O	3.98	2.83	1.82	0.07	3.48
MnO	0.19	0.15	0.25	0.19	—
P_2O_5	0.26	0.39	0.20	0.14	0.12
TiO_2	0.58	1.13	0.46	0.08	0.40
L.O.I	0.7	0.7	0.9	8.2	1.70
Total	100.77	99.41	99.27	99.18	100.61

C782/4 Hornblende-plagioclase gneiss (GR 2571 1645)
C782/46 Early partial melt phase (GR 2633 1538)
C782/47 Hornblende-plagioclase-clinopyroxene gneiss (GR 2645 1534)
C782/47 Ultramafic (serpentinised peridotite GR 2715 1639)
*27/219 Plagioclase amphibolite GR 2445 1950

Analyst: Geosurvey International, Nairobi, except *J. Furst, Geological Survey, Kenya (quoted *in* Baker, 1963).

The Siambu Complex is regarded as an altered mixed igneous and sedimentary sequence. It was thrust as a coherent slab into its present high tectonic setting at an early stage in the tectonothermal history of the Mozambique Orogenic Belt in central Kenya (Chapter 7). Consequently its original geotectonic setting is enigmatic: part of a cratonic (Archaean) greenstone belt thrust southeastwards; part of a Proterozoic ophiolite complex (cf. Vearncombe, 1983, for the West Pokot area), part of a Proterozoic island arc?

More comprehensive chemical data (including radioisotopic analyses) are needed to resolve these questions. The Siambu Complex presently includes altered primary intrusives (the ultramafics and metagabbros) and secondary intrusives (the dioritic suite) due to partial melting during metamorphism. The 'background' banded hornblendic gneisses are regarded as altered lavas; their present chemistry is andesitic. The metasediments are mostly altered shelf deposits, e.g. the numerous disrupted marbles. The juxtaposition of some marbles with ultramafic rocks could suggest that these marbles may, in part, be metasomatic due to release of extensive CO_2 during the metamorphism of the original ultrabasic bodies.

3.6 AUGEN GNEISSES

Mappable wedges of steeply dipping augen gneisses help define the western interface of the major Barsaloian straightening zone. Their strong (Barsaloian) tectonic fabric overprinted and effectively destroyed any pre-existing lithological banding. Units from the Siambu Complex and Ol Doinyo Ng'iro Gneisses, as well as intrusive pegmatites, are incorporated in the augen gneisses. Baker (1963, p.11) provides detailed lithological descriptions. Feldspar augen, up to 4 cm long are flattened (axial ratios of 1:10:10 were measured) by the biotite foliation to accentuate the flaggy character of the augen gneisses. Diffuse and irregular pegmatitic zones are common: locally the breakdown of pegmatite to augen gneiss can be traced in outcrop.

The augen are either coherent microcline prisms or elliptical, granular aggregates of microcline set in a quartz, oligoclase, microcline matrix. The matrix microcline is replacing the other felsic phases. Quartz grains may be shape orientated parallel to the biotite foliae. Epidote is the usual accessory mineral.

3.7 THE WESTERN GNEISSES

Figure 8 taken from Weaver (1973) provides detail of the

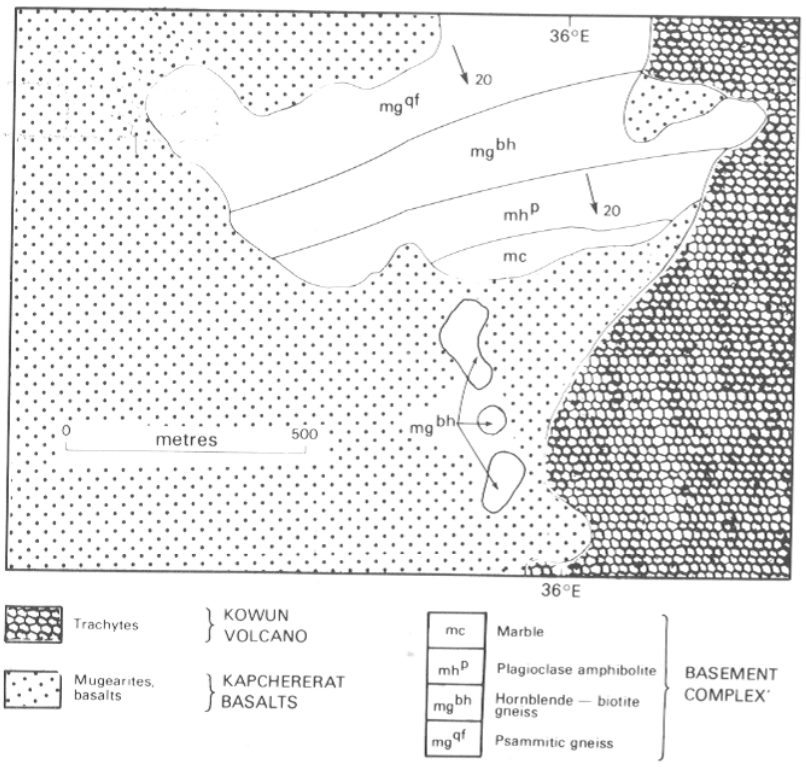


Figure 8 Geological map of the western gneiss inlier (from Weaver, 1973)

small inlier of metamorphic rocks in the western shoulder of the Gregory Rift at GR 1663 1900. Four lithologies are recognised with basal granite gneisses tectonically overlain successively by hornblende with biotite gneisses, plagioclase amphibolite and marble. These rocks strike towards 060° with gentle southerly dips in contrast to the regional strike of 015° and constant moderate easterly dips of the main outcrop area of gneisses further west (see Dodson, 1963; McCall, 1964; Joubert, 1966; Fairburn and Matheson, 1970).

The granite gneisses form ridges above a poorly exposed plain underlain by the other lithologies. An unusual mineralogical feature of these western gneisses is the occurrence of garnet in all lithologies including the plagioclase amphibolite. Otherwise mineralogically and texturally the four lithologies are similar to their counterparts to the east of the Gregory Rift. The lithological layering and any internal banding is sensibly concordant to foliation (biotite with or without muscovite foliae).

Weaver (1973) records the following mineral assemblages:

Quartz + microcline + plagioclase + biotite + garnet with accessory zircon, sphene
Quartz + orthoclase + plagioclase + biotite + garnet with accessory muscovite
Quartz + orthoclase + plagioclase + hornblende + magnetite
Quartz + microcline + sodic plagioclase with accessory biotite, muscovite
Calcite with accessory muscovite, apatite, scapolite
Hornblende + andesine + magnetite + diopside with or without garnet and accessory quartz and apatite

Quartz and microcline pegmatites are common cutting the various gneisses. Weaver (1973) concluded that the granite gneisses are altered psammites, the biotite and hornblende gneisses are also paragneisses, and the plagioclase amphibolites are altered igneous rocks (lavas or sills).

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4 Cenozoic Volcanics (related to the Gregory Rift)

At the time of writing (early 1986) no accounts to formalise the various Gregory Rift volcanic units had been published. Many of these units were introduced by the EAGRU mapping and it is the responsibility of that research unit, and not the present account, to formalise the stratigraphy of the Gregory Rift. Therefore all their stratigraphic terms are left informal pending the publication of the EAGRU data.

4.1 EASTERN SHOULDER OF THE GREGORY RIFT

Lower Miocene through Pleistocene volcanics and subordinate sedimentary rocks wedge eastwards from a maximum vertical thickness of about 4400 m adjacent to the median Rift. The various lithological units (Table 2) can be grouped into four major suites commencing with the basal Samburu basalts: these form eroded shields (overlying a rugged metamorphic rock surface) with a remnant 300 m of assorted basaltic flows, major basaltic pyroclastics, basal and inter-flow sedimentary rocks. It is always in Degree Square 27 overlain by a Miocene phonolite shield of up to 1000 m of widespread basal flows and associated pyroclastics (of the Lopet Phonolites and Rumuruti group) and up to 600 m of more localised upper flows (the Losiolo phonolites). Individual phonolite flows are thin, averaging c.20 m, yet areally extensive from dyke and vent sources. The phonolite shield oversteps eastwards on to the metamorphic basement, but its northern limit near latitude 1°50′N coincides with that of the underlying basalts. A westward facies change introduces up to 2000 m of assorted (Miocene) phonolitic pyroclastics with major basalts, sediments and pumice tuffs, also subsidiary flows of varied composition (basaltic-trachytic-phonolitic). This sequence wedges rapidly eastwards to define a thin unit (Towana Formation) between the eastern basal and upper phonolites but oversteps northwards (Kamolingaran basalts) on to metamorphic basement. Finally, up to 1000 m of Late Miocene through Pleistocene basalts and trachytes were erupted over the western and northern extremities of the eastern shoulder concomitant with identical volcanism further west. Isolated easternmost basalt outliers are of uncertain age.

Samburu basalts (Tvb^{SM})

Shackleton (1946) introduced the name 'Samburu Series' to describe red and purple tuffs with subordinate basalt flows which underlie the extensive phonolites forming the northern Laikipia plains and higher ground north-west of Maralal. The nomenclature is after the Samburu Hills which are named on the military map EAF 1168, Maralal, NW 1:250 000, 1941. Both he and later workers (e.g. Baker, 1963) admitted that the name is inappropriate: the greater part of the type area referred to by Shackleton is underlain by basalts younger than the Samburu basalts. Golden (1978) noted that coeval basaltic rocks were erupted from several centres along the eastern side of the Gregory Rift during the early to middle Miocene. These formed gentle shield volcanoes (including the Chembalao Shield Volcano in the Amaya area) which Golden collectively included in the Samburu 'Series' Basalts. In the present account all basaltic rocks, resting on metamorphic basement and which are older than the extensive phonolite flows, are referred to as the Samburu basalts.

Lithologies of the Samburu basalts are extensively developed at the base of the eastern shoulder's stratigraphy. However exposures are presently confined to the Amaya embayment (good outcrop in the Tinka River) and to the northern parts of the eastern shoulder (especially north-west and west of Marti) along the Samburu monocline.

Table 2 Volcanic stratigraphy of the eastern shoulder of the Gregory Rift

PLIO-PLEISTOCENE	Secumius trachytes ($Tp\text{-}Qpvt^{SC}$) and Emuruangiring trachytes ($Tp\text{-}Qpvt^{EG}$), Tirr Tirr trachytes ($Tpvt^{TT}$) and basalts ($Tpvb^{TT}$), isolated eastern basalts (Qpb) UPPERMOST TRACHYTES and BASALTS	
MIDDLE TO UPPER MIOCENE	Lopogno trachytes ($Tmvt^{LP}$) Nasorut basalt ($Tmvb^{NS}$) Losogol tuffs ($Tmvt^{LT}$) Alengerr tuffs ($Tmvt^{AE}$) Kamolingaran basalts ($Tmvb^{KO}$) Seronut formation ($Tmvt^{SR}$) Komol formation ($Tmvb^{KM}$) Katomuk tuffs ($Tmvt^{KT}$) WESTERN PYROCLASTICS	Losiolo phonolites ($Tmvp^{L}$) Towana Formation ($Tmvp^{T}$) Lopet Phonolites ($Tmvp^{LP}$) Rumuruti group ($Tmvp^{R}$) EASTERN PHONOLITES
LOWER TO MIDDLE MIOCENE	Samburu basalts (Tvb^{SM})	BASAL BASALT

In most cases the Samburu basalts crop out at the base of scarps capped by resistant phonolites. Major agglomerate horizons underlie gently undulating countryside. In the east the rocks are essentially flat lying beneath a subdued phonolite plateau. Towards the west the strata dip at increasing angles westwards to form impressive 'badlands'. Individual resistant lithologies may be traced several kilometres as prominent ribs. These are well displayed on aerial photographs where the agglomerates have distinct homogeneous, pale grey tones.

Good river sections through the Samburu basalts are plentiful, especially in the north. The Tinka River valley due west of Maralal provides good outcrop but is inaccessible within a treacherous, thickly vegetated gorge. Basaltic lavas are well exposed in the Baragoi River between GR 2284 1840 and GR 2297 1876 (north and west of the junction with the Nauorui River) and between GR 2379 1886 and GR 2394 1905. Fluviolacustrine sediments are best seen beneath the phonolites east of the main Maralal–Baragoi track at GR 2465 1750, in the Lomtono River at GR 2386 1876 and the Baragoi River around GR 2394 1907. Good general sections occur in the Nauorui River (c.200 to 250 m of tuffs and lavas), in the Kakipenet River from GR 2263 1574 to GR 2275 1546, in the Narikoyai River from GR 2246 1473 to GR 2260 1447. Baker (1963) observed the following succession on and south-east of Morilem Hill (GR 2400 2045).

	Approximate thickness m
Grey basalt	13
Vesicular purple basalt	17
Cherty limestone with silicified tree trunks	2
Dark grey basalt	5
Coarse sands	1
Dark green tuff	3
Black porphyritic basalt	10
Subvolcanic sediments	
Fine white sand	2
Friable olive-grey clay	1
Conglomeratic grit	1

Silicified tree trunks (cf. Morilem Hill sequence) are locally preserved within the intercalated fluviolacustrine sediments.

Baker and others (1971) record isotopic ages of 18.5 ± 2.9, 19.8 ± 3.6 and 23.0 ± 3.2 Ma for the Samburu basalts north of Maralal. A somewhat younger age of c.16.1 Ma is given by unpublished EAGRU data for the Chembalao Shield to the south. The paper by Chapman and Brook (1978) on the chronostratigraphy of the Baringo Basin quotes ages of 11.8 ± 0.5, 14.2 ± 0.4 and 20.7 ± 0.6 Ma for the Samburu basalts. They note that the youngest age is spurious and conflicts with established older ages for the overlying phonolites.

The Samburu basalts always unconformably rest upon eroded (rugged) surfaces of metamorphic rocks. Locally subvolcanic sediments partly infill major depressions in this surface. In turn, the basalts are overlain by extensive phonolite flows, or in the north-west, unconformably by the Emuruagiring trachytes. Both Shackleton (1946) and Wilkinson (1983b) observed that the (Lopet) phonolites interdigitate with the upper members of the Samburu basalts. However Golden (1978) indicates a period of erosion of the Chembalao Shield prior to its capping by the Lopet phonolites.

The exact thickness of the Samburu basalts is not known because of the absence of completely exposed sections showing basal and upper contacts. Within the Nauorui River basin up to about 130 m of strata are exposed (Baker, 1963).

Wilkinson (1983b) suggests a maximum thickness of about 250 m from the same general area. Further north and eastward the sequence thins rapidly to several tens of metres. In the south Shackleton (1946) indicates a thickness of about 100 m near Amaya. Cross sections in King (1978) indicate several hundred metres of Samburu basalts to the west of Maralal.

Golden (1978) records a maximum measured thickness of 600 m for the Chembalao Shield to the south of Degree Square 27, and suggests that an equal thickness has been removed by extensive erosion.

Within Degree Square 27 the Samburu basalts are principally composed of coarse tuffs with subordinate, but not rare, basaltic lavas. The tuffs are mostly red to purple, with occasional white and rare black layers. They are coarsely, yet regularly, stratified with pumice, lithic and crystal fragments. Lapilli are mostly highly altered, with secondary ferric oxide, carbonate and silica. Crude grading is locally seen in these basaltic fragmental deposits. A lilac to purple unsorted lahar cutting through tuffs is exposed in the Ayena Atesuro River at GR 24081696 (Plate 10). Ash-flow tuffs (Wilkinson, 1983a) are denser (more compact) and finer-grained than other tuffs, varicoloured (reds, yellows, greens) with local streaking. Pumice clasts are flattened and stretched to define strong fluxional fabrics.

The basaltic lavas are mostly plagioclase-phyric, variably vesicular and again reddish to purple (rarely grey) in colour. Individual flows are thin and interspersed throughout the tuffs. Plagioclase phenocrysts, up to 1 cm in size, may be fluxionally orientated. Limonite staining and secondary carbonate are common. Trachytic lavas are rare. Shackleton (1946) records a basal, pinkish buff, streaky flow with glassy feldspar phenocrysts from the Langat Gorge east of the Maralal to Baragoi track. Again the feldspars are fluxionally orientated and the flow has an aphanitic base. Minor intercalated fluviolacustrine sediments are poorly exposed.

The northern extremity of the Chembalao Shield exposed in the Amaya embayment consists of highly porphyritic basaltic lavas. Golden (1978) provides detail for the whole shield, which is composed of basic to intermediate lavas and very rare pyroclastics, with late phonolite and trachytic rocks. Basanites and olivine-analcime hawaiites, mostly porphyritic are dominant in the following succession:

1 Olivine and pyroxene-phyric analcime basanites and hawaiites
2 Plagioclase and pyroxene-phyric hawaiites and mugearites
3 Highly feldsparphyric hawaiites and mugearites

A common feature of all lithologies examined in thin section is strong, locally complete, breakdown of the ferromagnesian phases and all plagioclase phenocrysts.

In the tuffs the phenocrysts are invariably euhedral labradorite laths. Angular clasts of pumice, altered lava, basement mineral fragments (quartz, biotite, microcline) are set in an altered groundmass of plagioclase microlites (locally welded), with subordinate augite, iddingsite, iron ore and golden brown glass.

In the lavas the principal phenocryst is plagioclase. Compositions vary from zoned bytownites and labradorites (basalts) to oligoclase-andesine or anorthoclase (mugearites). They are up to 1 cm long and may be associated with aegirine-augite phenocrysts. Groundmasses are composed of altered aggregates of plagioclase, clinopyroxene and interstitial iron ore with or without

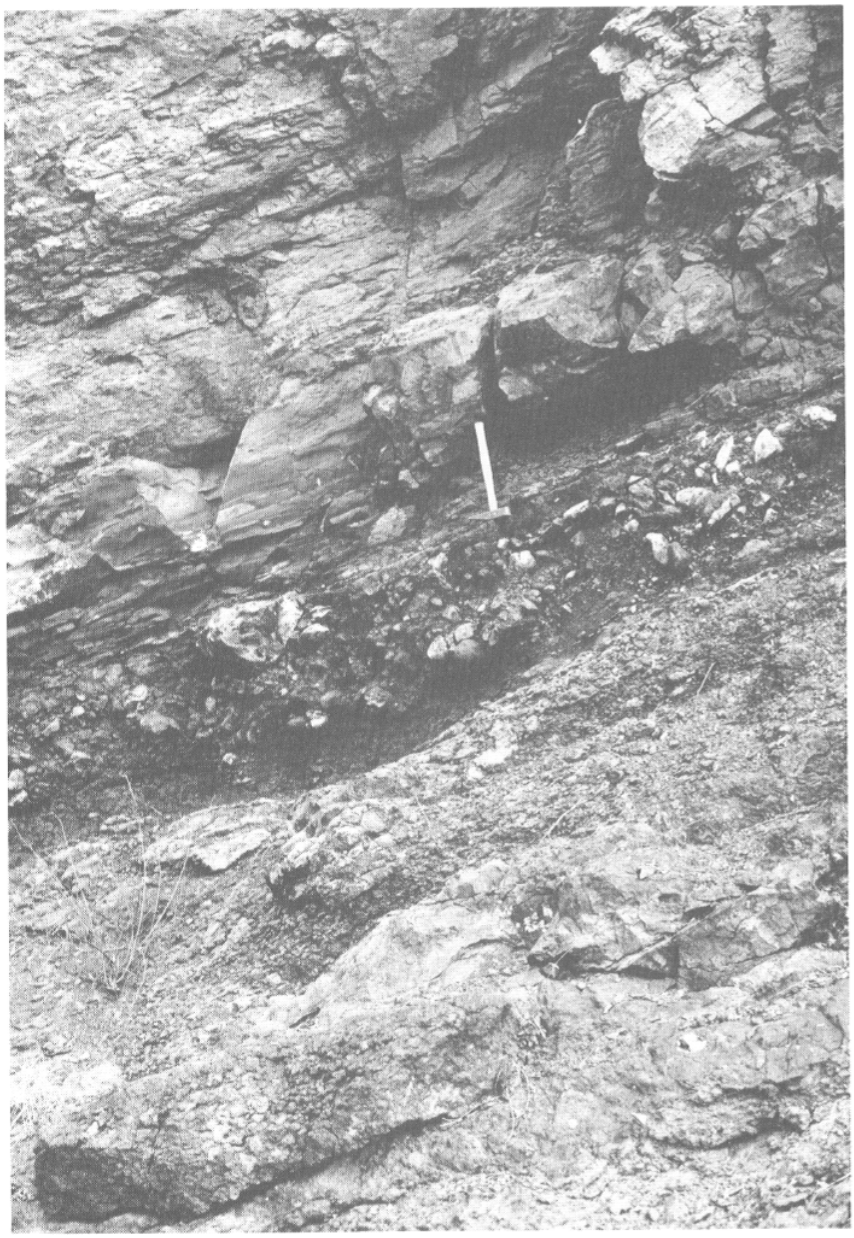


Plate 10 Lavas and lahars of the Samburu basalts in the Marti area

altered olivine, iddingsite, biotite, kataphorite-like amphibole, analcime, chlorite, carbonate, chalcedony and limonite (the last four phases are replacive).

Trachytes may have sanidine, anorthoclase and oligoclase phenocrysts with or without ferroaugite, fayalite, opaques and apatite phenocrysts together with basement derived mineral clasts in altered groundmasses. Labradorite and smaller augite phenocrysts in a groundmass of plagioclase, augite, magnetite and rare interstitial nepheline were noted in a tephrite. Nepheline and augite phenocrysts of variable size are present in basanites in groundmasses of oligoclase, brown glass, iddingsite, magnetite, olivine, augite and nepheline.

Basalts from the Amaya embayment have labradorite phenocrysts within groundmasses of altered olivine, purplish augite, magnetite with or without biotite and analcime.

No chemical analyses are available from the main northern outcrop area. However the petrography clearly indicates a wide range of compositions for the lavas from undersaturated nepheline-bearing rocks through basalts (*sensu lato*) to mugearites. Five whole rock analyses, taken from Golden (1978) are shown in Table 3 to include a wide range of lithologies within the Chembalao Shield. They include pyroxene and olivine ± plagioclase-phyric basalts, labradorite-olivine-phyric basanite and two feldsparphyric trachytes; for the most part strongly undersaturated.

Golden (1978) has demonstrated that the Samburu basalts within the Amaya embayment formed a shield volcano (Chembalao), with a very gentle surface profile and sourced by localised dyke swarms. However the preponderance of coarse tuffs to the north rather suggests numerous vent sources. The coarseness of the tuffs also implies relatively gentle eruptions (MacDonald, 1972). Golden (1978) also suggested quiet eruptions of highly fluid lava with surface degassing to remove the high volatile contents indicated by the presence of modal kaersutitic amphibole. He provides a detailed discussion on kaersutite stability. The ubiquitous alteration of the various volcanic rocks may be related to succeeding cycles of volcanicity. The wide compositional range of the lavas, indicated by their petrography, indicates extensive magma fractionation and/or crustal contamination.

Lopet Phonolites ($Tmvp^{LP}$)

The only new lithostratigraphic terms introduced for the Gregory Rift volcanics are the Lopet Phonolites and Towana Formation. The Lopet Phonolites comprise the major oldest suite of extensive, phonolite flows north of the Laikipia Plateau. Baker (1963) originally referred to these phonolites at their northern limits west of Baragoi as part of the Rumuruti Phonolites. However there they are geographically separated from the type area of the Rumuruti Phonolites. Therefore the new term, Lopet Phonolites is introduced after the Lopet Plateau south-south-west of Baragoi. The full geographical extent of the Lopet Phonolites is shown on the accompanying geological map. Williams and Chapman (1986) refer to these phonolites as the Losuwa Phonolites. Golden (1978) introduced an unnecessary element of confusion by referring to these basal phonolites west of Maralal as the Losiolo Phonolites. Shackleton (1946) had already used this term (retained in this report) for the uppermost phonolites of the same area.

On the Lopet Plateau the present land surface corresponds closely to the original lava surface. Stream erosion (except the Langat River) has made little impression on the flat, hard lava sheets. Edges of lava flows can occasionally be recognised forming low linear marginal levees. To the west and east the edge of the phonolite plateau is a scarp caused by headward erosion of streams draining west to the Suguta Valley and east to the Masiketa River. Phonolite caps protect tilted blocks along the inner sector of the eastern shoulder where at least 12 flows terrace major escarpments.

On aerial photographs the phonolite plateaus have dark tones and their surface although flat, is marked by various manifestations of lava movement. These are frequently concentric ridges, but may also show low mounds of cooled blocky lava with intervening depressions covered by a veneer of fine silt which imparts a paler tone. Variably trending positive lineaments, several kilometres in length, are caused by narrow ridges of phonolite indistinguishable from the adjacent flows and regarded as remnants of eruptive fissures. In the chaotic, dissected western badlands the phonolitic flows cannot easily be distinguished from phonolite pyroclastics except where sufficiently wide remnants of flow sheets survive.

Within the basal pyroclastics in the east the most common lithology is a coarse pumice agglomerate, with angular clasts of pumice and phonolite randomly distributed in a fine tuff matrix. Like other pyroclastics these rocks are pale green, grey, buff or yellow in colour and, rarely, brick red or orange. Welded and non-welded rainfall and ash-fall tuffs are recognised which may contain lithic (phonolite, rarely gneiss) and crystal (sanidine) as well as pumice clasts. Welded tuffs tend to be compact and massive with conchoidal fractures and have fluxional textures due to flattened pumice. The unwelded air-fall tuffs are conspicuously soft, finely bedded, pale buff to white in colour with abundant carbonised plant remains. Waterlain sediments are locally interbedded with the pyroclastics. Occasional bands of dark green agglomerate occur interbedded with the phonolite lavas.

In the east the lavas are uniform in field appearance as medium to dark grey or greenish, massive, compact flows with rare, small phenocrysts of tabular sanidine or glassy anorthoclase. Flow tops may be highly vesicular (GR 2396 1929) with zeolite and carbonate infillings. Trachyphonolites are also recognised as darker, massive aphanitic flows with speckled lustres on weathered surfaces. Individual flows are generally less than 30 m thick.

Further west the EAGRU mapping indicates trachyphonolite and phonolite flows as well as rare trachyte and analcime mugearite flows. The main lithologies form aphanitic, somewhat fissile, dark greenish grey lavas with visible anorthoclase phenocrysts.

The tuffs have microcrystalline to glassy groundmasses. Fragments of altered sanidine, augite, aegirine-augite, magnetite and/or nepheline may be present in the crystal tuffs. In the welded tuffs the brown glass groundmass has radiating clusters of (?) cristobalite and feldspar defining axiolitic or spherulitic textures.

The phonolites have groundmasses of lath-like or equant feldspars (orthoclase, sanidine or anorthoclase) with interstitial granular aggregates of aegirine, augite, cossyrite, an amorphous brown amphibole (kataphorite), nepheline, analcime and rare magnetite. Common microphenocrysts are anorthoclase and sanidine (locally corroded), nepheline (with orthoclase inclusions) and augite. Less common are biotite, analcime, aegirine-augite, sodalite and kataphorite microphenocrysts. Red colours are due to secondary limonite.

Trachyphonolites consist of densely packed, narrow wispy laths of sanidine showing fluxional, trachytic textures. Minor interstitial augite, magnetite, brown amphibole, cossyrite and nepheline may occur.

Representative chemical analyses of three of the various westernmost lithologies of the Lopet Phonolite are shown in Table 3 (taken from Golden, 1978). No whole rock analyses are available from the Lopet Plateau.

The Lopet Phonolites represent a major suite of alkaline under-saturated volcanics whose eruption from fissure and vent sources took place in middle Miocene times. In the north the basal phonolite eruption overlapped with the final outpourings of the older Samburu basalts and the youngest phonolite flows are intercalated with the overlying Kamolingaran basalts. Individual lava flows are typically (for central Kenya) thin, laterally extensive and homogeneous (Lippard, 1973). Golden (1978) suggests that the Lopet Phonolites originally formed a large, low angle shield volcano.

Towana Formation ($Tmvp^{T}$, from Wilkinson, 1983a)

The Towana Formation is defined as the relatively thin unit of very pale phonolitic tuffs and vesicular lavas which occur on top of the Lopet Phonolites and which are overlain by the Losiolo phonolites. It derives its name from Towana Hill (GR 2345 1535) in the centre of the Marti area (Sheet 78/1). The full geographical extent is shown on the accompanying geological map—a group of flat-topped dissected hills around Moridjo and isolated outliers in the Losiolo area. The Towana Formation may be the lateral equivalent of the Katomuk tuffs to the west. On aerial photographs the Towana Formation forms a distinctive unit which can be distinguished with ease from the overlying and underlying formations. The alternation of relatively thin subhorizontal bands of harder phonolite (including Kapiti-type, Shackleton, 1946) with soft, easily weathered tuffs results in a layered-cake topography.

In addition, as a consequence of the softer nature of the pyroclastics, a system of NE–SW and NW–SE-trending fractures has been exploited by weathering agents and erosion. The result is a grid pattern of flat-topped hills, unique in the Marti area. The fractures also affect adjacent Rumuruti group phonolites and basement rocks, but have not been etched out as effectively in these harder rocks.

In outcrop the soft, near-white tuffs have weathered to produce mantles of clay which tend to obscure the phonolites. The latter form thin, flaggy outcrops which collapse when undermined and litter the ground with slabs of lava. The 'Kapiti-type' phonolites are much harder and tend to survive as resistant caps to the hills and projecting ledges at lower levels. Towana Hill is the most prominent local topographic feature. Tuffs and lavas crop out on the flanks of the hill and the summit is capped by 'Kapiti-type' phonolite. Many good sections can be examined west and south-west of Moridjo village, e.g. Merekesh Hill at GR 2345 1453.

The Towana Formation lies between, and where developed, serves to separate, the Lopet Phonolites and Losiolo phonolites, both of Middle Miocene age.

On Towana Hill and east of Moridjo in the Moridjo River valley the Towana Formation is about 160 m thick. It thins slightly to the north-east and thickens to the south to about 200 m at GR 2318 1410.

The tuffs, which are characteristic of the Towana Formation, are pale green, yellowish or pale grey in colour. They may be hard and indurated, but commonly are soft, low density, friable rocks which often show regular banding indicating air-fall origin. They are normally crystal-lithic tuffs of uniform grain size containing a few larger pumice fragments. Common vesicules are carbonate-filled.

Occasionally thin welded tuffs can be observed which have the appearance of convoluted flow-banding due to streaks of multicoloured glass and fiamme resulting from flattened pumice fragments. One such rock from GR 2340 1491 has brightly coloured red, yellow and green bands and secondary axiolitic crystal growth of cristobalite and feldspar. Another specimen from GR 2345 1504 is a massive, compact dark grey rock containing flattened pumice fragments and isolated sanidine phenocrysts.

The lavas within the Towana Formation are typically very pale greenish grey. They are highly vesicular or amygdaloidal, becoming pumiceous with a sharp, cindery texture in some localities. Vesicles are commonly flattened to impart a fissility. Amygdales are filled with secondary zeolite (possibly chabazite). In one amygdaloidal specimen from GR 2363 1488 very large, bladed translucent phenocrysts of anorthoclase are interspersed with spherical vesicles, some partially filled with a black botryoidal mineral provisionally identified as pyrolusite, and others lined with zeolite.

In addition to the vesicular lavas there are other minor lithologies. Several rocks were identified in the field as basalts. They are dark grey, compact and aphanitic with a coarser grain size than normally seen in phonolites, giving the rock a speckly lustre on freshly broken surfaces. However, microscopic examination proved these rocks to be phonolites or trachyphonolites lacking identifiable nepheline.

'Kapiti-type' phonolites, as defined by Shackleton (1946), are dark, mottled, greenish grey fissile rocks of coarser grain size than phonolites typical of the Rumuruti group. They are distinguished above all by very large, translucent anorthoclase phenocrysts up to several centimetres long and rhombic in cross section. Occasionally cruciform twins are observed. Nepheline phenocrysts are inconspicuous or absent, but analcime may occur as small white, spherical phenocrysts.

A singular lithology observed at GR 2350 1526 is a massive, mottled lava with cumulophyric masses of ferromagnesian minerals set in a very pale grey-green, fine-grained groundmass containing scarce soda-orthoclase phenocrysts.

The Towana Formation air-fall tuffs have shards of glass and pumice fragments together with crystals of nepheline and sanidine microlites. In some thin sections, grains of magnetite, augite and brown amphibole (cf. kataphorite) are visible. Amygdales are filled with radiating zeolite fibres.

The welded tuffs have a muddy brown, glassy, banded matrix in which are set globular patches of glass surrounded by radially disposed sanidine and cristobalite (?) microlites with an axiolitic texture. Large euhedral phenocrysts of nepheline and microphenocrysts of sanidine, augite, granular aggregates of aegirine-augite and specks of magnetite may be seen 'floating' in this groundmass. Irregular amygdales may be filled with analcime.

The pale grey-green vesicular and pumiceous lavas display randomly disposed needle-like laths of sanidine in a golden-brown glassy matrix containing partially resorbed nepheline crystals, ferromagnesian grains and specks of magnetite. Amygdales are filled with zeolite (chabazite or natrolite), and in one case a manganese mineral.

The basalt-like phonolites are composed almost entirely of a densely packed decussate groundmass of sanidine laths. Within this are interstitial brown amphibole (cf. kataphorite) with small nepheline grains. There may also be microphenocrysts of pale green augite, diopside, aggregates of aegirine-augite, and specks of magnetite. Minute acicular apatite crystals were observed in one slice whilst amorphous glass made up much of another. Some slides lacked nepheline, and wispy microlites of sanidine displayed a trachytic texture in a brown glassy matrix. These rocks are better termed trachyphonolites.

'Kapiti-type' phonolites have very large anorthoclase phenocrysts which may constitute most of the slide. In addition there may be small phenocrysts of nepheline, analcime, green augite, aegirine-augite and magnetite. Shackleton (1946) also records olivine and apatite from these rocks. The groundmass has a characteristic intersertal texture and consists essentially of sanidine laths, analcime, brown amphibole (cf. arfvedsonite), aegirine-augite, aegirine, diopside, cossyrite, magnetite, nepheline, apatite and olivine.

The distinctively mottled lithology was seen in thin section to consist of granular aggregates of aegirine-augite with magnetite and possibly other ferromagnesian minerals set in a pale groundmass of sanidine microlites with interstitial nepheline and patches of brown glass.

Rumuruti group ($Tmvp^R$)

Shackleton (1946) named (the extensive phonolites of the Laikipian Plateau) the Rumuruti phonolites and, on his map, the phonolites of the Rumuruti Series after the town of Rumuruti situated in the middle of the plateau. Subsequently the EAGRU team working west of Rumuruti identified three separate phonolite formations of what they termed the Rumuruti Group (Truckle, 1977b). This unit has not been formally defined, but the single flow mapped on Degree Square 27 corresponds on the adjacent degree square with the Lower Rumuruti phonolites, cf. Ngelesha (Hackman, in press).

Phonolites of the Rumuruti group can be traced from the type area into the extreme south of Degree Square 27 where they underlie low (c.1900 m elevation above sea level) plateaus south of Maralal. These phonolites have dark grey phototones in contrast to adjacent, paler

plateaus underlain by the younger Losiolo phonolites. Isotopic ages of between 10.3 and 12.0 Ma are quoted (Baker and others, 1972; Chapman and Brooks, 1978) for the Rumuruti group; older ages come from the more northerly flows. The flow exposed south of Maralal is almost certainly the basal unit to the whole group.

South of Maralal the Rumuruti group phonolites rest directly on metamorphic basement. Adjacent plateaus underlain by the Losiolo phonolites are at somewhat higher elevations (to suggest that they are younger). Unfortunately no contact between the two groups of phonolites is exposed.

The Rumuruti group phonolites are less than 30 m thick on Degree Square 27 and comprise a single, badly fractured aphanitic, black-grey amygdaloidal and weakly porphyritic flow. The phenocryst phase is anorthoclase and according to Shackleton (1946) the phonolites classify as the Losugura type of Smith (1931). Lippard (1973) has observed that the various middle Miocene phonolite lava flows of central Kenya are unique in the Cenozoic volcanic record with regard to their large volume and the huge lateral extent of individual flows.

Losiolo phonolites ($Tmvp^{L}$)

Shackleton (1946) named the Losiolo phonolites for the series of thin lava flows exposed in the cliffs at Losiolo (Plate 7) although he did not distinguish them on his map from the underlying phonolites. Shackleton's terminology is retained (albeit informally), as the best section through the lavas is the major scarp feature at Losiolo. The Losiolo phonolites underlie the Loroki Plateau and continue northwards as far as the high ground at Losiolo. Their full geographical extent is shown on the accompanying geological map apart from a southern continuation of the Loroki Plateau into Degree Square 35 (Hackman, in press).

On aerial photographs the Losiolo phonolites are characterised by a smooth, flat texture and uniform dark tone. This is due to the relatively thick soil cover and the resulting vegetation developed on the well-watered elevated plateaus. The constituent thin flows and intervening tuff beds form numerous closely spaced scarps on valley walls and distinguish the Losiolo phonolites from the Rumuruti group to the south. Locally, e.g. the Nashoda River, the valleys expose complete sections down to underlying Towana Formation or metamorphic basement. The water courses on the Loroki Plateau follow curvilinear N–S fault traces. In outcrop the phonolites form rounded boulders with pale reddish brown weathered rinds. The more fissile units have weathered distinctive aluminium grey, greasy surfaces.

The type section for the Losiolo phonolites is the Ngerem precipice down to the floor of the Gregory Rift's median-graben (Mulaso Plain) at Losiolo. Eighteen flows can be identified in a vertical section of about 600 m (Figure 9). Additionally most major river valleys on the Loroki Plateau also provide good outcrop of several flow units. In the north the best sections occur on the outlier of Lonkutukie Hill (GR 2451 1541) and the northern end of Nkabai Plateau (GR 2444 1508).

The exact age of these phonolites is not known. They are obviously younger than c.15 Ma (age of the Lopet Phonolites) and may be c.10 Ma if the regional correlations (see map legend) are correct. Outpourings of the younger extensive phonolite flows have been dated at c.10 to 12 Ma from areas to the south and west (Chapman and Brook, 1978). The Losiolo phonolites conformably overlie

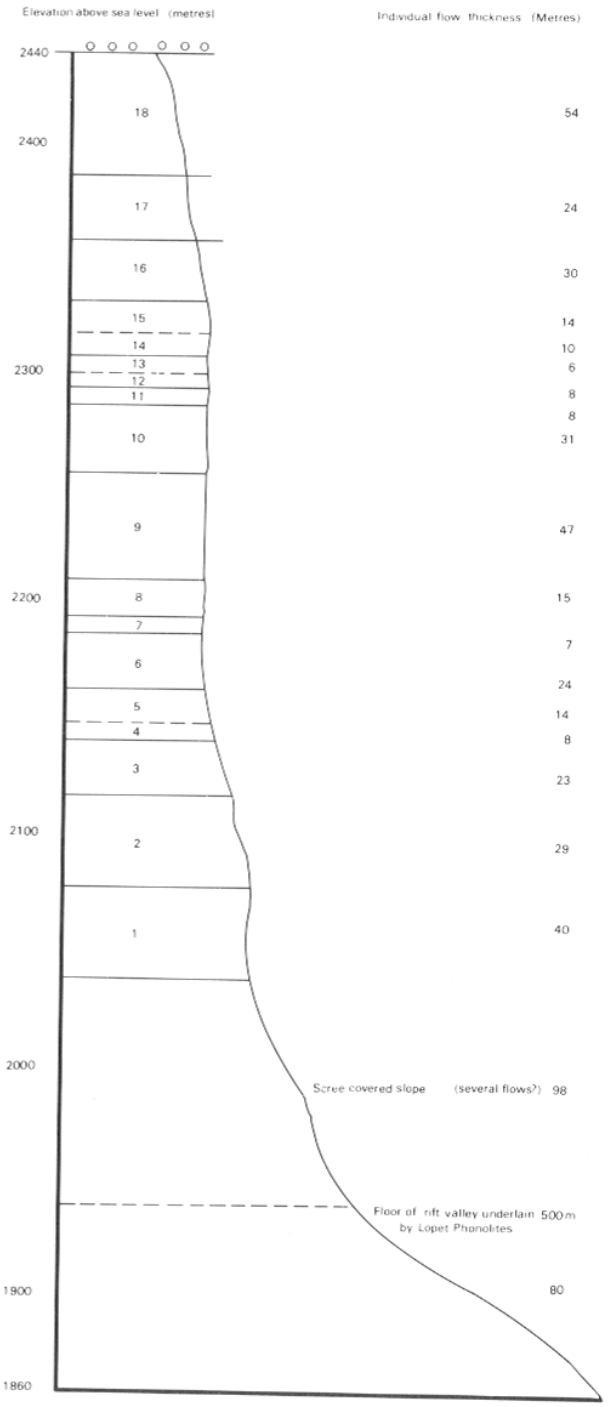


Figure 9 Reference section for the Losiolo phonolites (from Key, 1983b)

the Towana Formation in the west and north but overstep directly on to basement in the east. They are never overlain by younger volcanics.

The thickness of the Losiolo phonolites is variable because of (1) a general thickening westwards towards the Gregory Rift and (2) the rugged nature of the basement depositional floor, especially west of Maralal. In the type area there is about 600 m of phonolite, with individual flows between c.20 and 100 m thick. On the west face of the Nkabai escarpment the unit is about 200 m thick.

The main lithology is an aphanitic, fissile, fine-grained, black, aphyric lava which on high ground weathers to a greasy aluminium-grey powder. Sections through the phonolite sequence reveal clastic bases and scoriaceous

Table 3 Whole rock analyses of the eastern shoulder volcanics (all from Golden, 1978 except 12 and 13)

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
SiO_2	45.59	48.24	48.20	54.48	60.10	58.60	58.53	55.78	51.75	54.30	57.19	55.31	44.46	58.29	63.23	65.17	56.23	45.13
TiO_2	2.96	2.67	2.64	1.09	0.45	1.04	0.73	0.50	0.85	1.46	0.42	0.19	2.67	0.46	0.82	0.62	0.39	2.41
Al_2O_3	12.90	15.33	16.84	18.65	17.85	16.39	16.26	17.91	20.37	16.70	18.65	20.32	16.18	17.32	16.64	13.46	17.72	13.83
Fe_2O_3	5.72	3.79	3.07	3.37	2.33	2.71	5.68	2.84	2.93	3.16	2.04	2.84	4.76	5.03	2.21	4.09	3.54	5.57
FeO	7.10	6.99	8.01	2.85	2.82	3.31	2.42	2.67	2.48	4.56	3.26	1.80	6.90	1.93	2.01	2.68	3.80	6.06
MnO	0.21	0.20	0.20	0.18	0.23	0.26	0.20	0.30	0.18	0.23	0.21	0.31	0.19	0.31	0.19	0.18	0.32	0.16
MgO	8.52	5.19	3.68	1.34	0.60	0.98	0.59	0.48	1.38	1.55	0.42	0.38	5.60	0.74	0.61	0.40	0.36	9.44
CaO	10.28	8.72	8.18	3.99	1.72	1.99	1.65	1.23	4.72	3.94	2.09	1.46	9.69	1.88	0.65	0.40	1.85	10.82
Na_2O	2.75	3.79	3.90	5.77	7.08	5.93	5.27	7.86	5.90	4.98	7.41	8.26	4.48	5.12	6.72	6.34	6.22	1.82
K_2O	1.06	1.91	2.17	4.57	4.59	5.43	5.72	5.56	4.22	4.37	5.18	5.86	1.59	5.51	5.53	4.97	5.53	0.71
P_2O_5	0.52	0.60	0.71	0.24	0.09	0.14	0.04	0.07	0.46	0.41	0.07	0.04	0.76	0.06	0.08	0.03	0.00	0.75
L.I.O.	2.53	1.95	1.55	2.59	1.45	1.28	2.19	3.11	3.77	3.47	1.72	3.69	2.57	2.77	0.64	0.34	2.71	2.12
Total	100.14	99.38	99.14	99.12	99.31	98.06	99.28	98.31	99.01	99.13	98.66	100.46	99.85	99.42	99.33	98.68	98.67	98.79

1 Augite-olivine labradorite phyric basalt
2 Augite-olivine phyric basalt
3 Labradorite-olivine phyric basalt anite
4 Oligoclase-phyric trachyte
5 Anorthoclase-phyric trachyte
— Samburu basalts

6 Aphanitic zeolite mugearite
7 Anorthoclase phyric phonolite
8 Phonolitic ignimbrite
— Lopet Phonolites

9 Oligoclase nepheline phyric phonolite
10 Aphanitic mugearite
11 Oligoclase-kaersutite phyric analcime mugearite
— Seronut formation

12 Phonolite, Seronut formation (Sceal, 1974)
13 Basalt, Komol formation (Sceal, 1974)
14 Mugearite ignimbrite
15 Anorthoclase phyric trachyte
16 Anorthoclase phyric trachyte
17 Aphyric analcime trachyphonolite
— Alengerr tuffs

18 Augite-labradorite phyric basalt
19 Pigeonite-olivine phyric basalt
20 Bytownite-olivine phyric basalt
— Nasorut basalts

tops to individual flows. The clastic bases are matrix-supported with angular clasts, generally less than 20 cm long, of fissile and scoriaceous phonolite. Clastic zones are less than 10 cm thick. In contrast whole flows may be scoriaceous with thicknesses of up to 100 m. In the type area the basal flows are mostly scoriaceous. Vughs are oval shaped, up to 20 cm in length and may either be infilled by a yellow powder (upper flows) or an olive green powder (basal flows). Pumiceous tops are locally present in northern flows which weather to produce thin, intercalated yellowish red soils.

Phenocrysts of sanidine or anorthoclase are rarely seen in hand specimens and never seen to be orientated. Nepheline phenocrysts are absent although microphenocrysts of olivine are locally present. Tuffaceous interlayers are confined to the south.

In the Nkare Narok River a granite boss is exposed which formed a positive topographic feature during the eruption of the phonolite lava. Adjacent to the boss there is a clastic unit (lahar) composed of angular and sub-rounded scoriaceous fissile phonolite clasts up to 1 m long in a fine matrix. It is clast-supported.

As noted above the phonolites are fine-grained and it is impossible in the field to discern individual minerals or groundmass textures. A surprise on examining thin sections was that almost half of all sections showed strong orientation of matrix feldspar laths and the remainder a random feldspar orientation. Re-examination of the rock samples failed to distinguish the two groups which may relate to the orientation of the cut slice. Groundmasses consist of feldspar laths, opaque grains and a brown-stained amorphous mass in which mineral grains are difficult to identify. Wilkinson (1983a) did recognise augite, analcime, aegirine-augite, cossyrite, kataphorite, minute euhedral nepheline and magnetite. Phenocrysts are euhedral and consist of anorthoclase or Carlsbad-twinned sanidine laths and rarely biotite, altered pyroxene, olivine and magnetite.

In the southern tuffs, euhedral untwinned feldspar grains and ragged lithic fragments (in which randomly orientated feldspar laths predominate) are set in a groundmass of fine lithic fragments, aligned feldspar laths and dust. The largest clasts are less than 2 mm long.

The Losiolo phonolites comprise a suite of very regularly interlayered flows with subsidiary tuffs and clastic (lahar) deposits in the south. The frothy nature of many of the flows suggests that they were laid down very quickly and on a relatively stable land surface—there are no large cross-cutting features. Originally the phonolites extended further eastwards and possibly formed a carapace over the whole of the Karisia Hills (Key, 1983b).

Katomuk tuffs ($Tmvt^{KT}$)

This formation was named by Golden (1978) after Katomuk Hill (GR 2090 1290) in the south centre of Degree Square 27. Its main geographical extent is shown on the accompanying geological map although isolated outliers above the Lopet Phonolites continue for about 10 km southwards into Degree Square 35.

There is some confusion about these southern exposures as Sceal (1974) originally regarded them as type areas for the Alengerr tuffs.

The various pyroclastics of the Katomuk tuffs tend to form variably consolidated cappings to fault-controlled low plateaus in the dissected badlands along the eastern margin of the Gregory Rift median graben. For the most part the beds are subhorizontal except for local fault-related tilting. The layer-cake bedding is well displayed on aerial photographs with the unwelded tuffs showing as pale coloured areas.

The reference area is Katomuk Hill itself although thicker successions are exposed to the north. Their whole outcrop area is especially inaccessible. Silicified tree

19	20	21	22	23	24	25	26
43.10	47.36	52.76	62.22	60.33	63.12	61.19	59.05
2.79	2.62	1.73	0.51	0.57	0.57	0.68	0.70
15.02	16.33	16.65	16.96	16.85	14.24	17.21	16.69
4.38	2.71	7.15	2.62	3.84	9.14	2.14	2.60
8.10	8.55	2.48	2.49	1.62	0.00	2.59	3.90
0.20	0.19	0.25	0.16	0.23	0.76	0.23	0.25
7.42	5.64	1.74	0.37	0.33	0.05	0.72	0.61
11.89	8.12	4.71	1.29	1.62	0.26	1.68	2.05
2.85	3.83	5.90	6.31	6.75	4.38	5.80	6.81
1.58	1.34	3.07	5.64	5.11	5.75	5.78	4.76
0.66	1.12	0.52	0.06	0.05	0.08	0.11	0.18
1.03	1.09	1.98	0.78	1.11	1.04	1.15	1.78
99.02	98.09	98.94	99.41	98.46	99.39	99.28	99.38

21 Aphanitic analcime mugearite
22 Sanidine phyric microsyenite
23 Anorthoclase syenite
24 Vesicular correndite
} Nakali trachytes

25 Anorthoclase phyric phonolitic trachyte
26 Anorthoclase phyric trachyte
} Lopogno trachytes

trunks are preserved in soil horizons and *Dicerorhinus* and *Gomphothere* teeth have been located by Sceal (1974) to the south in rocks he originally assigned to the Alengerr tuffs but subsequently (Williams, *pers. comm*) re-assigned to the Katomuk tuffs.

No isotopic ages are available for any lithology in this unit. They overlie with apparent conformity the Lopet Phonolites and are in turn overlain by members of the Komol formation, Alengerr tuffs and Kamolingaran basalts. The Katomuk tuffs are part of the middle Miocene phonolitic tuff belt forming the eastern margin of the Gregory Rift's median graben.

On the southern map margin near easting 2060 the Katomuk tuffs are 150 m thick. They range from over several hundred metres in the north to less than 20 m in their southern extremities. The basal welded tuffs are individually about 30 m thick on average.

Basal welded tuffs (phonolitic ignimbrites) are overlain by unwelded pumiceous tuffs, agglomerates, ashes, and tuffaceous sediments with a central trachyphonolite lava. The basal ignimbrites are dark greenish brown, flinty rocks with flattened and angular lithic fragments (trachyphonolite, rarely glass or syenite). The unwelded tuffs are friable and weather to produce prominent earth pillars. They are well exposed near GR 2100 1200 as dazzling white, finely laminated tuffaceous silts and earths (Golden, 1978).

The flows contain abundant phenocrysts of alkali feldspar, nepheline, aegirine-augite, biotite, apatite and opaques with rare plagioclase and fayalitic olivine. Textures are cumulophyric with individual phenocrysts marginally resorbed. Flattened fiamme and analcime vesicles also occur in a groundmass of turbid isotropic, finely comminuted glass. Analcime, interstitial to rare alkali feldspar crystallites, was observed by Golden (1978).

No chemical analyses are available of rocks from this unit.

Komol volcanics

This unit includes the Komol and Seronut formations of Sceal (1974), which clearly form a continuous single shield (Hackman, in press).

Komol formation ($Tmvb^{KM}$)

Sceal (1974) introduced the term 'Komol formation' for a suite of assorted pyroclastics exposed in the Komol valley, locally overlying the Katomuk tuffs. The main development of the Komol formation is in Degree Square 35; on Degree Square 27 they are confined to the summit area of Katomuk Hill (GR 2090 1290). Sceal (1974) identified the following succession.

6 Dark flinty trachytes and phonolites (up to 120 m thick)
5 Porphyritic phonolites (0–100 m)
4 Mugearites (c.80 m)
3 Olivine hawaiite (c.13 m)
2 Porphyritic mugearite (0–23 m)
1 Deeply weathered mugearites (c.60 m)

The olivine hawaiite (of unit 3) caps Katomuk Hill. This is an almost black rock (and referred to as a picrite basalt from its field appearance by Sceal, 1974) which consequently forms a dark patch on aerial photographs.

Three isotopic age determinations on hawaiites from the Komol formation gave 13.6 + 0.2, 14.2 + 0.2 and 15.7 + 0.3 Ma (Golden, 1978). On Katomuk Hill the olivine hawaiite unconformably overlies the Katomuk tuffs and is about 13 m thick. It is distinguished by its almost black groundmass in which olivine and augite phenocrysts are visible to the naked eye. In thin section the olivine phenocrysts are seen to locally enclose biotite laths and microphenocrysts of opaques appear in a groundmass of andesine laths and analcime mesostasis.

A single whole rock chemical analysis (Table 3) by Sceal (1974) confirms the hawaiitic petrography by plotting in this field on an alkali against silica diagram (Cox and others, 1979). The rock is undersaturated with normative nepheline and also has a high Sr content.

Seronut formation ($Tmvt^{SR}$)

This poorly exposed unit unconformably overlying the Komol formation in Degree Square 35 was originally defined by Sceal (1974) after Seronut Hill. Subsequently Golden (1978) referred to these rocks as the Geleamoi trachytes. The original terminology is followed; in Degree Square 27 it is confined to the extreme south centre (near easting 2000) and to a hillside several kilometres to the north-west. It is best developed southwards where reference sections are defined (Sceal, 1974).

Isotopic ages from northern outcrops indicate an age range of c.12.0 to c.13.4 Ma (Golden, 1978) for the Seronut formation. It oversteps the Komol formation and Katomuk tuffs, and is unconformably overlain by the Alengerr tuffs in Degree Square 27. Feeder dykes passing upwards into plugs are identified in the type area. Accordingly to Golden (1978) the Komol and Seronut formation comprise a lava shield to account for external thickness variations. In Degree Square 27 the Seronut formation is about 100 m thick and thins eastwards from a source in the Komol Valley (Degree Square 35).

The dominant lithology is a strongly weathered, friable, greenish brown lava, which has a vesicular top and a

flow-banded base. Sceal (1974) referred to these thin lavas as mugearites and Golden (1978) noted that kaersutite-andesine phyric mugearites are the principal lithology. Porphyritic phonolites and rare basalt form more resistant, and hence more easily identifiable intercalations. Zeolite and calcite amygdales are common in the vesicular flow tops.

In thin section the dominant lavas are seen to be variably porphyritic with phenocrysts of kaersutite and andesine, and microphenocrysts of augite, opaques and apatite in strongly altered groundmasses. Black aphanitic trachyte flows have a high proportion of analcime in their groundmasses. The phonolites have euhedral nepheline, olivine, ferro-augite and kaersutite phenocrysts in a groundmass of euhedral feldspar, aegirine-augite and aenigmatite with minor calcite and analcime. Augite and olivine (altered to serpentine) phenocrysts occur in a plagioclase (An 69), augite, opaques and calcite groundmass in the rare basalt. Four whole rock analyses shown in Table 3 are representative of the principal mugearite flow and of rare phonolite lavas.

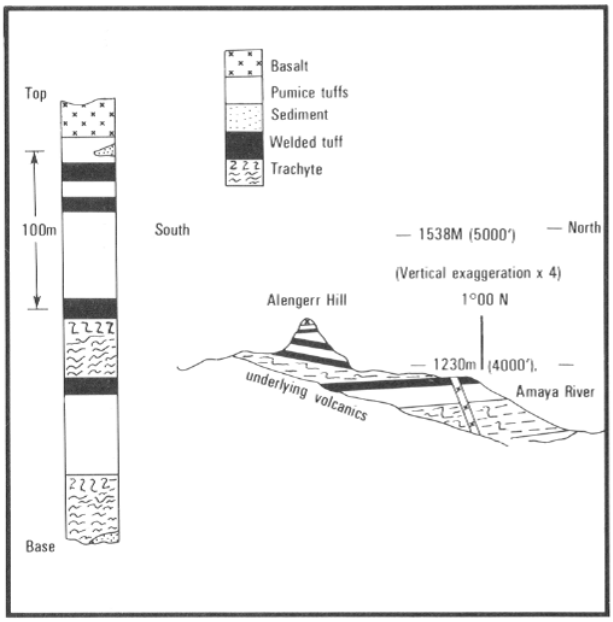


Figure 10 Type area and section of the Alengerr tuffs (from Sceal, 1974)

Alengerr tuffs ($Tmvt^{AE}$)

This predominantly pyroclastic unit was named by Sceal (1974) who defined a type area of Alengerr Hill, whose northern slopes extend into Degree Square 27 at easting 2000. Subsequently Golden (1978) traced the pyroclastics in a 5 km wide faulted NNE belt as far as the southern boundary of the Emuruagiring trachytes: a strike length of about 50 km. Unfortunately Losogol Hill (the type area for the Losogol tuffs) straddles the centre of this NNE belt and is shown as being underlain by Alengerr tuffs in Truckle (1979a). The accompanying geological map to the present report follows Golden (1978) in showing Losogol Hill as underlain by the Losogol tuffs.

The Alengerr tuffs underlie terraced mesas where the tuffs are horizontal (as on Alengerr Hill itself), and are generally over 100 m in thickness. The terracing, caused by alternating resistant welded and soft friable unwelded tuff beds is clearly visible on aerial photographs (to allow accurate estimates of dip).

On Alengerr Hill the basal units are exposed on the northern slopes (cut by the Amaya River) in Degree Square 27. Figure 10 (from Sceal, 1974) shows the complete type succession of 277.2 m of interbedded welded and unwelded pumice tuffs with thin trachytic lavas, rare sediments and an uppermost basalt (peak of Alengerr Hill).

Silicified wood fragments are present in the basal sediments from the type area. A mugearite from the southern flanks of Alengerr Hill was isotopically dated at 10.6 ± 0.8 Ma; Golden (1978) suggested this was too young noting that locally, basal members of the Alengerr tuffs *interdigitate* with Seronut formation lavas dated at c.12 to 13 Ma.

In the type area the Alengerr tuffs locally unconformably overlie the Seronut formation (Sceal, 1974) although, as mentioned above, Golden (1978) suggests this contact is locally conformable. To the north the Alengerr tuffs unconformably overlie the Katomuk tuffs but apparently conformably overlie the Kamolingaran basalts. However an upper unconformity everywhere separates the Alengerr tuffs from overlying Losogol tuffs, Nasorut basalt, Emuruagiring trachytes and the Quaternary basalt lavas of Silali. Numerous trachyte and less common basalt dykes, in part feeder dykes to the intraformational trachyte and basalt lavas, cut the Alengerr tuffs.

The Alengerr tuffs are about 272 m thick in the south and may thicken northwards (the area south of the Emuruagiring Plateau has not been mapped in detail). At Alengerr, over half the succession (Figure 10) comprises unwelded, friable trachytic or phonolitic, off-white pumice tuffs. The welded tuffs are poorly sorted with large trachytic (rarely syenitic) clasts up to 15 cm long in a glassy, devitrified matrix which is variably welded. The thin trachyte lavas contain small feldspar phenocrysts and the sediments are white and diatomaceous. The upper basalt is a pyroxene-phyric basanite. Immediately northeast of Alengerr the unit consists of about 100 m of basal trachytes and phonolites overlain by three or more uniformly welded tuffs. However, unwelded creamy white, friable pumice tuffs and agglomerates reappear in the north together with (basal) phonolitic ignimbrites and various feldsparphyric, flinty, dark greenish grey lavas (phonolitic trachytes and trachyphonolites).

Aphyric, analcime-bearing trachytes with secondary interstitial calcite, and devitrified glassy trachytes with mosaics of anhedral feldspar and ore blades were examined petrographically by Sceal (1974). The welded tuffs have yellow glassy groundmasses with various trachytic clasts and small (1 to 2 mm) crystals of sanidine and nepheline. The basalt has euhedral augite and olivine phenocrysts in an equigranular plagioclase, olivine, augite and opaques groundmass.

According to Golden (1978) the basal phonolitic ignimbrites in the north are petrographically identical to their counterparts in the Katomuk tuffs. He identified phenocrysts of anorthoclase, sanidine, aegirine, ore, biotite and katophorite. The lavas contain turbid microphenocrysts of nepheline and/or sodalite rimmed by sodic mafic minerals. Some contain kaersutite and occasional phenocrysts of ferroaugite, aegirine-augite and, rarely, olivine. Groundmasses haye anorthoclase laths with interstitial ore, aenigmatite, katophorite, arfvedsonite and aegirine-augite. The upper pyroclastics have groundmasses of finely

comminuted glass and pumice with lithic fragments composed of phyric trachyphonolite and phonolite. Rare mugearites have andesine and kaersutite (pseudomorphed by ore and augite) phenocrysts in oligoclase-rich groundmasses. Four representative whole rock analyses, taken from Golden (1978) are shown in Table 3.

Losogol tuffs ($Tmvt^{LS}$)

The map of Golden (1978) shows Losogol tuffs to underlie the Losogol (GR 2040 1240), Nataruk (GR 2068 1280) and Lokapomye (GR 2068 1328) Hills as well as the faulted block immediately south-east of Losogol Hill. Unfortunately the subsequent EAGRU map (Truckle, 1979a) incorrectly depicted the geology of this area. The unit is named after Losogol Hill which marks the centre of the outcrop area (confined to Degree Square 27). The Losogol tuffs are preserved within a faulted 20 km N–S strip which has a maximum width of 6 km. Hill outcrops are separated by alluvial plains. The main hills are several hundred metres high and are underlain by trachyte plugs intruding crumbly tuffs and agglomerates of the Losogol tuffs. Intense dissection of the tuffs has produced a badland scenery.

The pyroclastics have pale phototones.

Losogol Hill is the type area with about 200 m of unwelded air-fall pumice tuffs and subsidiary agglomerates and rare ignimbrites. No isotopic ages are available for the Losogol tuffs. Stratigraphically they are placed between the c.12 Ma Alengerr tuffs and the c.9 Ma Nasorut basalts. According to Golden (1978) the Losogol tuffs were contemporaneous with the upper pyroclastics of the Alengerr tuffs.

Basal contacts of the Losogol tuffs are not exposed although Golden (1978) shows them to be locally unconformably above the Alengerr tuffs. They are unconformably overlain on Lokapomye Hill by the Nasorut basalts and Quaternary basalts from Silali. The tuffs are gently dipping (again incorrectly shown as steep in Truckle, 1979a) but are cut by steeply dipping trachyte sheets and dykes.

The Losogol tuffs form a wedge-shaped mass thickening westwards to about 300 m in vertical thickness. Golden (1978) attributes this thickening to contemporaneous westward tilting of the land surface. Pink, red, green and white air-fall pumice tuffs are the main lithologies with the range of colours due to weathering, especially of iron oxides. Interbedded agglomerates contain phonolitic and trachytic lithic clasts. Finely laminated pumiceous silts with conglomerate channels are localised. Two thin, buff to brick red ignimbrites contain angular and flattened lithic and glass fragments as well as broken crystals of anorthoclase, nepheline, biotite and fayalite. No whole rock chemical analyses of any of the lithologies have been undertaken.

Nasorut basalt ($Tmvb^{NS}$)

The Nasorut basalts are defined by Golden (1978) as the homogeneous basalts which overlie the various phonolitic and trachytic tuff units along the western margin of the Rift's eastern shoulder. They take their name from Nasorut Hill (GR 2064 1303) situated in the middle of the 50 km long by 4 km wide NNE-trending, fault controlled outcrop area south of the Emuruagiring Plateau. Good outcrop is confined to steep-sided hills (up to several hundred metres high) which in the south are capped by trachyte. The basalt outcrop forms relatively dark phototones with respect to the underlying tuffs.

Reference sections are defined by Golden (1978) on Nasorut Hill and on Lopogno Hill where the conformable contact with the overlying Lopogno trachyte is exposed. A basalt from Lopogno Hill gave an isotopic age of 9.41 ± 0.3 Ma. The Nasorut basalts overlie with an angular unconformity both the Alengerr and Losogol tuffs. They are equated stratigraphically by EAGRU with the uppermost (Alengerr) basanite on Alengerr Hill. The basanite also has a slight angular discordance to the underlying tuffs although it is still included in the Alengerr tuffs unit. The Lopogno trachyte conformably overlies the Nasorut basalt although the Emuruagiring trachyte and basalts from Silali locally blanket both units. Trachyte feeder dykes cut the Nasorut basalts which have gentle to moderate (fault controlled) dips.

In the type area the Nasorut basalts are about 250 m thick but may thicken to about 400 m according to Truckle (1979a). Basaltic and hawaiite lavas are the principal lithologies with localised (in north) upper pumice tuffs, agglomerates and capping melabasanites. At Nasorut the unit consists of interlayered plagioclase-augite-phyric basalts and hawaiites. However at Lopogno there are basal olivine basalts with thick intervening vesicular horizons overlain by aphanitic basalts, hawaiites and mugearites capped by 30 m of plagioclase-augite-phyric hawaiites. Feeder dykes to the lavas cut the underlying Losogol tuffs.

According to Golden (1978) phenocryst phases in the basalts and hawaiites are olivine, augite and plagioclase (bytownite and zoned andesine-oligoclase) in varying proportions. The capping basanite has pigeonitic pyroxene, olivine and ilmenite phenocrysts in a matrix of olivine, pigeonite, magnetite, plagioclase (An 76–45) and up to 15 per cent analcime. Three whole rock analyses, two of typical basalts and one of the upper basanite are given in Table 3.

Lopogno trachyte ($Tmvt^{LP}$)

The Lopogno trachyte is always associated with the Nasorut basalts, cropping out in the same area south of the Emuruagiring Plateau. They were named by Golden (1978), after Lopogno Hill (GR 2023 1200) where gently dipping trachytes unconformably overlie basalts. Again they are readily distinguished on photographs from adjacent units which allowed Golden to extrapolate their outcrop area northwards from the type area. This is defined on Lopogno Hill where about 60 m of trachyte or phonolitic trachyte lava is well exposed.

An isotopic age of about 9.0 Ma has been obtained from the trachytes which is in good agreement with the age of about 9.4 Ma for the conformably underlying basalts. In the type area the trachytes cap hilltops. Elsewhere they are unconformably overlain by Silali basalts, the Emuruagiring trachytes or by encroaching alluvium related to a major tributary of the Suguta River. The trachyte flows are fed by dykes cutting the underlying basalts. A northern source is also suggested by Golden (1978) with pumice tuffs and agglomerates appearing to the north of the type area.

The lavas are porphyritic with up to 15 per cent of tabular anorthoclase phenocrysts and microphenocrysts of nepheline (rimmed by sodic mafics), ferro-augite and aegirine-augite in a groundmass with aegirine-augite, aenigmatite and arfvedsonite as well as feldspar. Secondary zeolite locally pseudomorphs nepheline crystals.

Two whole rock analyses shown in Table 3 are of

typical anorthoclase-phyric phonolitic trachyte and trachyte lavas.

The Lopogno trachyte is regarded by Golden (1978) as the dissected remnant of a small trachytic stratovolcano.

Kamolingaran basalts ($Tmvb^{KO}$)

Baker (1963) described 'Porphyritic olivine and augite basalts (Laikipian basalts?)' occurring on top of the Lopet Phonolites in the northern part of the eastern Rift shoulder. These can be traced westwards into the Kamolingaran basalts of Truckle (1979a & b) so the EAGRU terminology is adopted here. These basalts (and related sediments) are confined to the northern parts of Degree Square 27 and have been mapped northwards towards Lake Turkana by Dodson (1963) who referred to them as the 'Porphyritic basalts'. They are well exposed in the Samburu Hills area. The distribution of intercalated sediments concentrated near the base of the basalt sequence is shown on the accompanying geological map (and Figure 11).

The countryside underlain by this unit forms rugged, inhospitable badlands with fault controlled chains of irregular hills covered by *Euphorbia candalabra* (Baker, 1963) and good bouldery outcrops of basalt. Softer sediments and tuffs (in the west) are variably exposed in gullies on hillsides. However, this pale-coloured material is easily recognised on aerial photographs in contrast to the dark basalts: their mapped distribution is essentially based on photo-interpretation. The very good stratification within this unit is also a characteristic photo feature.

The type locality is south of the Emuruagiring Plateau in the Kamolingaran area (Truckle, 1979a). Baker (1963) provides descriptions for the areas to the north-east of the

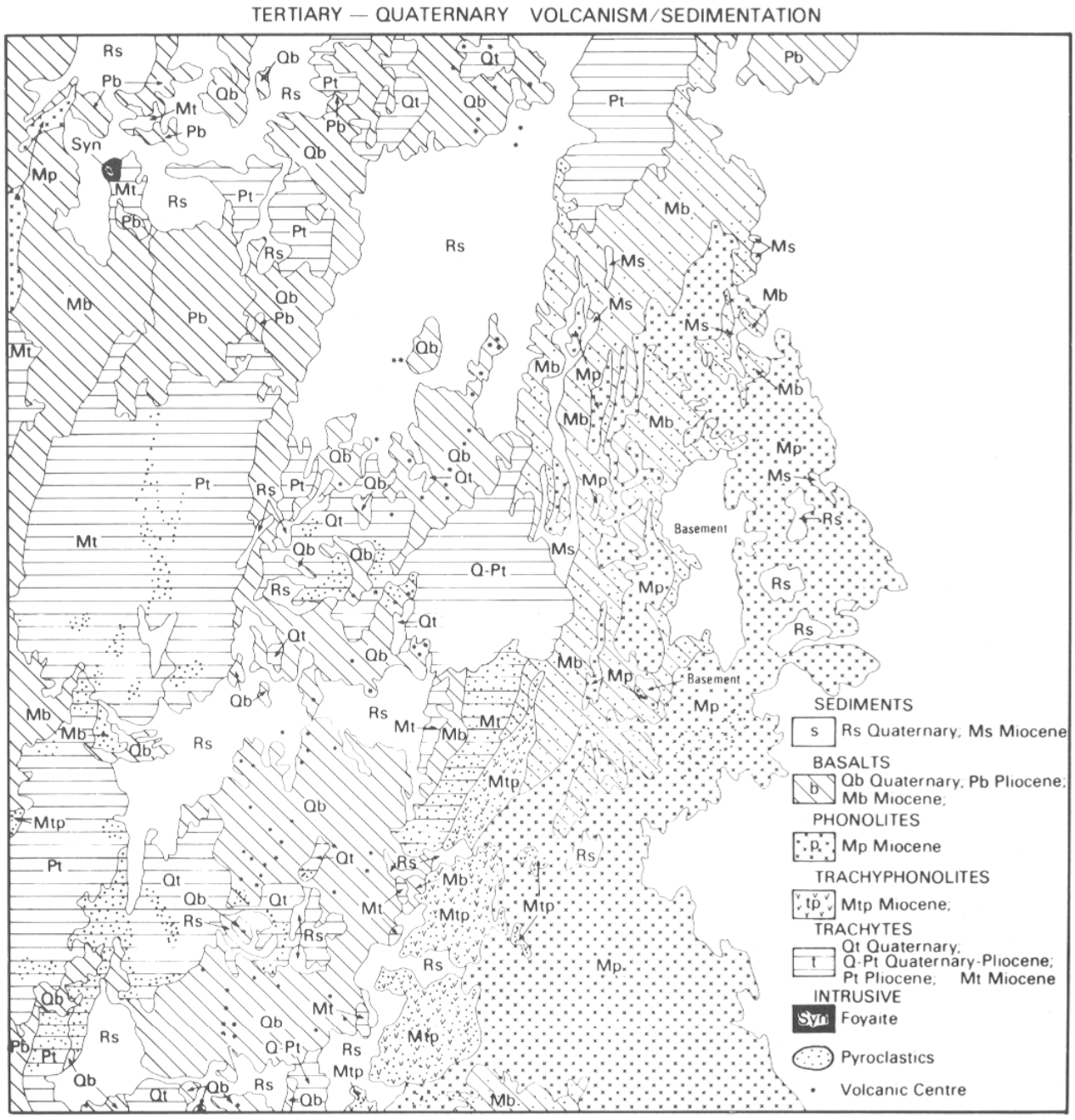


Figure 11 The distribution of the main lithologies within the Gregory Rift

type area. He notes that the sequence is repeated by normal faults and because of the similarity of the various lavas it is impossible to establish a detailed succession. However the lowest lavas are well exposed on Napung ridge (GR 2380 2050). South of Echua Etom (GR 2300 2015) there are good exposures of interbedded tuffs and lavas as well as intrusive dykes and sills. More massive, paler porphyritic basalts occur around Kongia (GR 2235 2000). On the cliffs around the alluvial fan of the Nitelejo River (GR 2223 2042) and also near Kanganya (GR 2223 2023) the top of the Kamolingaran basalts and their contact with the overlying Tirr Tirr trachytes are exposed. The intercalated sediments have a thickness of about 110 m on Waitiakwel Hill (GR 2228 1875) and Baker (1963) gives the following succession north-west of Akangaurak (GR 2253 1860).

(Top)		m
6	Brick red gravels, buff sands, white silts	25
5	Yellow brown volcanic sands with pumice fragments	12
4	Lava conglomerate with gravel beds	8
3	Grey brown ashes with diatomaceous bands	18
2	Grey porphyritic basalt	6
1	Purplish grey gravels and silts	8

The sediments are fossiliferous and are being investigated by a joint Kenya (National Museum, Nairobi)–Japanese team. During their initial fieldwork they found hominoid fossils (remains of the 'Samburu ape') from basal sediments in the Kongia area.

The stratigraphic position of the Kamolingaran basalts indicates that they are between c.9.4 Ma (age of overlying Nasorut basalts) and c.15.0 Ma (youngest age of underlying Lopet phonolites). The joint Kenyan–Japanese team have obtained unpublished isotopic ages of c.13 Ma and c.6 Ma from basal and upper flow units respectively in the Kongia area.

The Kamolingaran basalts generally dip gently to moderately towards the WNW although dip reversals due to faulting are common. The basal contact unconformably oversteps (northwards) the Lopet Phonolites and Samburu basalts on to metamorphic basement, often with basal sedimentary pockets. In the north the unit is, in turn, unconformably overlain by flat-lying Tirr Tirr trachytes and basalts. Westwards they are overlain with apparent conformity by the Nasorut basalts, although the younger Emuruagiring trachytes locally cut across this contact.

According to Baker (1963) the Kamolingaran basalts are the thickest unit within the eastern shoulder. Both his and EAGRU cross sections (Truckle, 1979a & b) indicate thicknesses between 500 and 1000 m. However, as Baker (1963) notes it is difficult accurately to establish the full sequence in the field because of the extensive faulting and lithological uniformity. He records over 100 m of basalts in the basal section at Emuru Etuku and the basal sediments are up to 110 m thick.

The lowest lavas are invariably dark grey, massive basalts with abundant augite phenocrysts, hackly fracture and bouldery outcrop. Minor aphyric basalts locally separate individual porphyritic lava flows. Westwards the lavas become thinner with tuff intercalations. Basalts with olivine phenocrysts are interbedded with the augite basalts (basalts with both olivine and augite phenocrysts are rare). Crystal tuffs form thin beds and south of Echua Etom there are columnar basalt dykes and sills. Many basalts are strongly vesicular, even scoriaceous, and contain many narrow veins and spots of coarse white calcite (locally related to faulting). Around Kongia the basalts contain feldspar phenocrysts which are also characteristic of the uppermost basalts in the west (descriptions all from Baker, 1963).

Baker (1963) also gives details of the various fluviolacustrine sediment intercalations. In addition to the succession north-west of Akangaurak there are buff, porcellaneous clays and olive-green, pale grey-green and brown clays with narrow water-worn pumice pebble beds and interbedded brick red pyroclastics at Waitiakwel Hill.

The basalts are petrographically similar apart from variations (noted in the field) in the size and mineralogy of the phenocrysts. These are commonly small augite, but may include olivine (up to 1 cm long and altered to iddingsite) and basic plagioclase. The groundmass is typically basaltic with plagioclase, augite and iron ore with or without olivine. Baker (1963) tentatively identified analcime to indicate an alkaline trend in the basalts.

The western accumulation of pyroclastics suggests a western source for the basalt wedge (or shield) which is volumetrically important in the northern part of the eastern shoulder. Wilkinson (1983a) noted that a pause in the volcanism prior to the Kamolingaran basalts eruption allowed the accumulation of varied lacustrine and fluviatile deposits in basal shallow basins; these sediments consist almost exclusively of reworked volcanic debris. At the onset of the renewed basaltic activity, thin lava flows were interbedded in the sediments which continued to be deposited. As volcanic activity increased the sediments continued to be locally deposited as thin lenses throughout the volcanic sequence (in a similar manner to sediments within the Samburu basalts).

Tirr Tirr trachytes ($Tpvt^{TT}$) and basalts ($Tpvb^{TT}$)

Originally called the Tirr Tirr Series by Baker (1963) after the Tirr Tirr area north of the Samburu Hills, they underlie the whole of this rugged plateau area with good outcrop on the impressive scarps overlooking the Suguta Valley. A single basalt flow of this unit also underlies the Emuru Akirim Plateau to the east. Dodson (1983) traced the trachytes and basalts for a further 10 km northwards.

The Tirr Tirr area is strongly faulted to produce gently tilted blocks capped by trachytes which weather to rounded boulders. The fault blocks are separated by narrow N–S gorges where ephemeral rivers follow fault traces. The intensity of faulting decreases eastwards, so that the subhorizontal basalt flow capping the Emuru Akirim Plateau forms a featureless flat summit area. The whole unit has a monotonous dark grey phototone with the major western fault traces well defined as dark linear watercourses. The geological boundaries with the underlying steeply dipping metamorphic rocks and Kamolingaran basalts can be drawn with considerable accuracy directly on aerial photographs with minimal field control.

Reference sections may be defined along the western scarps. At Lowoi (GR 2245 2150) the succession, dominated by trachytes, is about 300 m thick and the basal angular unconformity with the Kamolingaran basalts is exposed. The basal basalt is well exposed on the scarp around the Emuru Akirim Plateau.

Isotopic ages of 3.6 ± 0.4, 3.8 ± 0.4 and 3.9 ± 0.4 Ma have been obtained from lavas south-east of Lowoi (Baker and others, 1972). Therefore the angular unconformity at the base of the Tirr Tirr basalts with the Kamolingaran basalts represents a period of erosion of at least several million years. During this period the older unit was tilted westwards. The gentle dips (in the west) of

the Tirr Tirr trachytes and basalts are due to Late Pliocene and younger faulting and warping towards the median graben. Away from the rift shoulder the unit is essentially horizontal.

The thickness of the Tirr Tirr basalt is about 30 m under the Emuru Akirim Plateau. Westwards the basalt is overlain by further basalt and trachyte flows culminating in the 300 m or so succession adjacent to the Suguta Trough.

The basalts are massive, medium grey (paler than the Kamolingaran basalts) porphyritic rocks with a tendency to irregular fracture. Altered reddish brown olivine phenocrysts are more common than augite or feldspar phenocrysts. The trachytes closely resemble the basalts in field appearance (to account for the homogeneous phototone) with greenish tints in fissile outcrops. They form numerous thin flows in the west with intervening tuff horizons and an uppermost massive olivine basalt.

In thin section the basalts are seen to have a groundmass of basic plagioclase, purplish augite and abundant iron ore grains. Augite microphenocrysts are common and olivine is unevenly distributed in this groundmass with secondary iddingsite.

The trachytes have pale green aegirine microphenocrysts, scattered alkali feldspar phenocrysts in groundmasses of oligoclase laths (defining trachytic textures), purplish augite and iron ore dust.

The progressive westward thickening of the Tirr Tirr trachytes and basalts and the appearance of tuff intercalations adjacent to the Suguta Valley indicate a western source area. The present extent of these rocks may constitute the eastern half of a stratovolcano; the western part having been downfaulted and covered by Quaternary volcanics of the Suguta Trough.

Emuruagiring trachytes (Tp-Qpvt^{EG})

There is no published account of this unit which underlies the vast Emuruagiring Plateau straddling the eastern edge of the Gregory Rift's median graben at latitude 1°30′N. The geological boundaries of these trachytes are easily defined both on aerial photographs and landsat imagery because of their uniform grey tones relative to underlying lithologically (and therefore tonally) varied Miocene units. The full geographical extent, corresponding to the Emuruagiring Plateau area, is shown on the geological map of Degree Square 27.

No reference sections are defined: Wilkinson (1983a) notes that between 80 and 100 m of trachytes are well exposed along the plateau's eastern escarpment (e.g. at GR 2234 1618). Williams (1978) quotes an age of c.3.5 Ma for the Emuruagiring trachytes without giving any details of method, etc.

There is a pronounced angular unconformity with underlying more steeply dipping Miocene volcanics. The western limits of the plateau are locally mantled by Quaternary volcanism from Emuruangogolak. The Emuruagiring trachytes are extensively cut by N–S faults, mostly downthrowing westwards, which may gently tilt the trachytes. Cones and flow fronts are shown on Truckle (1979b) towards the SW margin of the outcrop extent.

Thicknesses in excess of 500 m are indicated on cross sections for the Emuruagiring trachytes on Truckle (1979a & b). These sections clearly suggest a pronounced thickening westwards (cf. the contemporaneous Tirr Tirr trachytes to the north).

Shackleton (1946) describes trachyte from the SE corner of the Emuruagiring Plateau as a speckly greenish grey rock with visible flecks of golden biotite and small sanidine crystals. In thin section the groundmass was seen to be composed of mossy cossyrite, aegirine, kataphorite grading into bluish amphibole, laths of sanidine and small poikilitic quartz aggregates.

The Emuruagiring trachytes must originally have formed a quite impressive stratovolcano—a proto-Emuruangogolak, subsequently modified by faulting and erosion prior to the eruption of Emuruangogolak Volcano over its western parts.

Secumius trachytes (Tp-Qpvt^{SC})

This relatively small unit was named by Sceal (1974) after Secumius Hill at GR 1953 1130 near the southern margin of Degree Square 27. It is incorrectly referred to as the Secuminus trachytes on Truckle (1979b). From Secumius Hill the trachytes strike north for about 6 km before disappearing beneath younger lavas of Silali. The trachytes have also been traced southwards for about 9 km into Degree Square 35, where they are shown by Hackman (in press) as part of the Paka volcanic pile.

The Secumius trachytes are confined within a narrow zone of intense N–S faulting, which defines a series of shallow horsts and grabens. To the south, pre-existing fault scarps have confined individual flows and almost certainly explain the linear outcrop pattern. The type area is Secumius Hill, where Sceal (1974) records 74 m of trachyte and mugearite flows. Their stratigraphic setting indicates a Plio-Pleistocene age; no isotopic dates are available.

Around Secumius Hill the trachyte is flanked by Quaternary unconsolidated deposits: alluvium and debris landslipped off older rocks. Further north the unit is overlain by basalt flows of Silali. The trachytes are cut by over 20 small N–S faults (downthrowing westwards) within their 2 km outcrop width. The thickness in the type area is typical for the whole strike length of the Secumius trachyte.

The trachytes form relatively thick flows, some of which are porphyritic with large alkali feldspar phenocrysts in a fine-grained trachytic groundmass. A single mugearite flow is exposed on Secumius Hill and trachyphonolites were noted by Sceal (1974) from basal sections along fault scarps.

Oversaturated and undersaturated trachytes contain modal quartz and sodalite respectively. Individual feldspar phenocrysts may have cores displaying herringbone structures rimmed by anorthoclase. Alkali feldspar laths define groundmass trachytic textures with mossy patches of aegirine-augite and aenigmatite.

Isolated basalt plateaus (Qvb, from Baker, 1963)

About 5 km WNW of Baragoi on the north side of the Baragoi River are several basalt-capped, low hills. The basalts crop out as rounded boulders and large blocks split by curved joints. Their age is unknown and will only satisfactorily be resolved by isotopic dating: lithological correlations with the nearby western lavas may be misleading. They rest on an eroded metamorphic rock surface and are only about 5 m thick.

The basalts are compact, massive, dark brownish grey, faintly mottled rocks with scattered calcite amygdales and minute feldspar phenocrysts. In thin section small cumulophyric groups of labradorite and poorly formed augite grains were noted in a groundmass of purplish augite, olivine, plagioclase microlites and iron-ore.

Table 4 Summary of the volcanic activity forming the eastern shoulder of the Gregory Rift

Stratigraphic unit	King and Chapman (1972) classification	Approximate age (Ma)	Maximum thickness	Shape	Lithology
Secumius trachytes	5	Pleistocene	74	Narrow flows	Trachytic lavas (rare mugearite flow)
Emuruagiring trachytes	3	3.5	500	Tabular prism	Trachytic lavas and (western) tuffs
Tirr Tirr basalts and trachytes	3	3.8 Pliocene	300	Tabular prism	Trachytic lavas and (western) tuffs and minor basalt flows
Kamolingaran basalts	2′	13.0–6.0	1000	Prism	Basaltic lavas and tuffs and related sediments
Lopogno trachytes	2′	9.0	60	Narrow flows	Trachytic lavas
Nasorut basalts	2′	9.4	400	Narrow flows	Basaltic lavas
Losogol tuffs	2′	>Nasorut b.	150	Narrow tabular deposit	Trachytic and phonolitic tuffs
Alengerr tuffs	2′	10.6	270	Narrow tabular deposit	Trachytic and phonolitic tuffs, minor sediments and basalt
Seronut formation	2′	13.4–12.0	100	Narrow tabular deposit	Trachytic-phonolitic—basaltic tuffs and flows
Komol formation	2′	15.7–13.6	13	Single flow	
Katomuk tuffs	2′	>Komol f.	300	Narrow tabular deposit	Phonolitic tuffs and lavas
Losiolo phonolites	2	?12.0–10.0	600	Prism	Phonolitic lavas and (southern) tuffs
Rumuruti group	2	12.0	30	Single flow on 27	Phonolitic lava
Towana Formation	2	12.0	200	Tabular	Phonolitic and trachy phonolitic tuffs
Lopet Phonolites	2	15.0 (older in the north)	800	Tabular prism	Phonolitic and trachy phonolitic lavas and tuffs
Samburu basalts	1	23.0–16.1 Miocene	?300	Several coalescing shields	Basaltic pyroclastics and lavas

Summary

The salient features of the volcanism forming the east shoulder are summarised in Table 4. Over the major part of this shoulder almost continuous volcanism can be demonstrated throughout the Miocene (from c.23 to c.6 Ma). Due to the inherent instability of the environment, widespread erosion almost certainly accompanied volcanism to help account for the numerous erosion surfaces preserved in the stratigraphic column. Only in the south-west did volcanism cease earlier (c.9 Ma). Interestingly both major basaltic units, the Samburu and Kamolingaran basalts, appear to span a period of about 7 Ma. Volcanism recommenced in the middle Pliocene (c.3.8 Ma) as extensive trachyte flows from western sources (prototypes for adjacent median graben volcanoes). The associated basaltic volcanism may have continued into the early Quaternary.

For the initial 7 Ma volcanism was essentially basaltic (with a complete range of basaltic petrochemistry). From c.16 to c.13 Ma the volcanism was mostly phonolitic with important western trachytic flows. The period c.13 to c.9 Ma involved contemporaneous trachytic, phonolitic and basaltic eruptions. Only the basaltic eruptions continued until c.6 Ma. Following the complete standstill in volcanism at the end of the Miocene, the middle Pliocene eruptions were trachytic (with subsidiary basalts) and probably short-lived. Therefore basalts have always formed a part of the volcanic activity and for about 14 Ma were the major eruptive phase. Phonolites were dominant for about 5 Ma (c.15 to c.10 Ma) and the trachytes formed important short-lived flows, possibly for about 1 Ma. Thus in terms of duration of extrusive activity this gives the ratio basalts: phonolites: trachytes of about 14:5:1. This is thought to provide a better indication of the relative importance of lithologies comprising the eastern shoulder than the present (remnant) lithological distribution.

For most units no precise total thicknesses are known and it is impossible to estimate accurately the complete thickness of volcanic and associated sedimentary rock at any given point in the main part of the eastern shoulder. The total succession, in fact, is probably less than 2000 m over most of this shoulder and only significantly increases immediately adjacent to the median graben.

4.2 WESTERN SHOULDER OF THE GREGORY RIFT

This section is based entirely on the documentation, mostly unpublished PhD theses, of the EAGRU mapping.

Middle Miocene through Pleistocene volcanism manifests as cyclic repetitions of flood basalts and trachytic volcanoes with minor phonolitic and trachyphonolitic volcanoes in the extreme west. In general the petrochemistry indicates increased saturation with time, although this variation may occur separately in individual central volcanoes. Basalts are generally alkaline evolving through alkaline basanites to trachytes. Trace element

chemistry of rocks and minerals suggests that all extrusives originated by fractionation of olivine, clinopyroxene and plagioclase from a picritic magma. Most basalts are highly differentiated with at least 20 per cent magma fractionation prior to eruption. The source depth is generally considered to be between 50 and 100 km and progressively shallowing (Truckle, 1977a). Within a cumulative sequence of c.3200 m of flood basalts there is a general decrease in the thickness of individual units (all extensively eroded) with time from c.1100 m, (west of Degree Square 27) for the basal Kapchererat basalts. Olivine basalts predominate from cinder cone and dyke sources with minor intercalations of sedimentary rocks and pyroclastics. The uniform lateral thicknesses of the basalt units primarily results because they formed from coalescing low shields with several source areas.

In general the trachytic volcanoes are very large low-angle multicentred shields (e.g. Ribkwo, Oliyamur) or plateaus (e.g. Nasaken) of peralkaline trachytic lava with up to 30 per cent by volume of trachytic pyroclastics. When preserved a stratiform flank zone made of extensive monotonous lava and pyroclastic flows contrasts sharply with a central zone of eruptive vents (short flows, dykes, plugs, localised pumice tuffs). Calderas and parasitic cones are rare. These volcanoes are strongly eroded; present extrusive thicknesses between 100 and 1000 m in a combined total of c.3600 m (1600 m of Miocene and 2000 m of Pliocene trachytic rocks).

The eastern extremities of the Tiati group and Kasorogol phonolites extend into the map area, where they are each several hundred metres thick. Pyroclastic rocks are important components of both units.

Individual feeder dykes and stocks (apart from the Kailimerlim Foyaite, Section 6.6, p.68) are too small to represent individually on the accompanying geological map. Dyke swarms are represented by single schematic dykes showing the dominant trends.

Kapchererat basalts ($Tmvb^{KP}$)

These basal basalts were originally defined by Martyn (1969) as part of the Tugen Hills Group. His nomenclature was subsequently followed by McClenaghan (1971) and Webb (1971) although Weaver (1973) referred to these basalts as the Turkana Basalts after Baker and others (1970). Truckle (1979b) and the present account stick to the original classification of Martyn (1969).

Three small inliers of the Kapchererat basalts through the Kowun Volcano occur along the western map edge immediately south of the western basement inlier (GR 1663 1900). The upper basaltic lavas are susceptible to weathering so that the inliers have subdued reliefs with dark phototones. Reference sections are given by Weaver (1973) to the west of Degree Square 27 where the formation is thicker and more extensively exposed.

The mugearites which just extend into the map-area adjacent to the western basement inlier have been dated at 15.7 ± 0.6 Ma (Weaver, 1973). A slightly older age of 16.6 ± 0.5 Ma was obtained by Webb (1971) for basal basalts further to the west.

In Degree Square 27 the basal contact of the Kapchererat basalts (with basement gneisses) is not exposed. They are overlain with slight angular unconformity by the extrusives of Kowun Volcano. At the contact the basalts are reddened and strongly weathered. Trachytic sheets and dykes related to Kowun also cut the Kapchererat basalts.

Within the three inliers the basalts have a maximum thickness of 160 m, consisting of 30 m of basal basalt agglomerates overlain by 130 m of basaltic and mugearitic lavas. Seven or eight individual flows have an average thickness of about 20 m. The southernmost inlier consists of about 13 m of deeply weathered basalt. The Kapchererat basalts generally thin eastwards.

The only exposed lithologies (Weaver, 1973) are aphyric mugearites overlain by feldsparphyric basalt flows. The mugearites are silver-grey or purple, somewhat fissile rocks whereas the basalts have vesicular upper surfaces and are also grey to purple with zeolite or calcite filled vesicles. Lath-like, cream-coloured plagioclase phenocrysts, up to 3 cm long, are common although some flows are microporphyritic with the phenocrysts up to 0.5 cm in length. The feldsparphyric basalts are prone to weathering; the exposed surfaces often being crumbly, friable and associated with secondary calcite growth.

According to Weaver (1973) zoned plagioclase laths (An 69 to An 44) are the main phenocrysts with or without olivine, pyroxene and kaersutite phenocrysts. Analcime is common in groundmasses with intergranular or ophitic textures.

Groundmass oligoclase laths have subtrachytic textures. Chemically the porphyritic lavas are undersaturated with normative nepheline. Webb (1971) presents two analyses of these rocks from the west of Degree Square 27.

The exposures on Degree Square 27 represent the eastern extremities of the Kapchererat basalts which had a westerly source: they thicken in this direction towards major tuff units intercalated with basaltic lavas. The Samburu basalts, exposed at the base of the eastern shoulder's succession, were contemporaneous with the Kapchererat basalts. Both are included in 'Group One' of the regional volcanic stratigraphy of King and Chapman (1972).

Kowun Volcano ($Tmvt^{KW}$) from Weaver, 1973

The volcano is named after Kowun Hill (underlain by a small summit caldera) which is situated immediately west of Degree Square 27. The eastern half of the volcano is exposed between northings 177 and 192 along the western map margin. Exposures are controlled by NNE step faults. An exhumed surface of Kowun Volcano forms a low shield composed of rounded hills rising to a maximum elevation of 1185 m (3850 ft) on Kowun Hill. The volcano defines the watershed between the Kerio and Suguta Rivers.

The account of Weaver (1973), if published, could serve to define Kowun Volcano formally. Reliable K/Ar whole rock ages of 15.0 ± 3.4 Ma and 15.2 ± 0.2 Ma have been obtained from trachytic lavas. The volcano always rests upon a weathered planar surface of the Kapchererat basalts, except where it oversteps on to the basement gneiss inlier at GR 1663 1900. In turn it is overlain by various Miocene basalt units (the Tirioko, Lokwanamur and Lomujal basalts); previously covering most, if not all, of the volcano but subsequently eroded to expose the present exhumed surface. Basalt dykes and various forms of trachyte intrusives cut the extrusive trachytic rocks which have quaquaversal dips off Kowun Hill. In the east these primary dips are modified by subsequent tilting caused by the NNE faults (downthrowing to the ESE). Trachytic flows also dip away from the parasitic centre at GR 1709 1860.

The shield has a present maximum (summit) thickness of about 300 m, mainly of trachyte lavas, with the various

trachytic pyroclastics clustered around the few eruptive centres. Truckle (1977a) noted that the pyroclastics increase in the volcano's northern flank where upper pyroclastic units are better preserved. Individual flows and tuff wedges are between 30 and 40 m thick. The various lithologies are highly oxidised with secondary alteration caused by exhumation of the basalt carapace and contemporaneous hydrothermal activity. The lavas form massive, non-fissile, thick flows, which are grey to purple (locally mottled) with common alkali feldspar phenocrysts up to 1 cm long.

Welded tuffs have brick red colours, locally coated blue by opaline silica, with eutaxitic textures due to green-grey fiamme. Pumice tuffs are yellow or green, locally well bedded and are friable when fresh but usually brick-red and porcellaneous (due to baking by the basalt carapace). Air-fall tuffs display graded bedding with crystal, pumice and trachyte clasts.

In thin section the feldspar phenocrysts were seen to be rectangular anorthoclase or, less commonly, 'sanidine-like' laths. The groundmass consists of slender alkali feldspar laths with alkali amphiboles and aenigmatite interstitially, or as subpoikilitic patches. Pyroxene is rare but quartz is usually present as sub-poikilitic to poikilitic aggregates (up to 20 per cent of the mode). Textures are sub-trachytic, mafic minerals are scarce. Therefore the lavas are alkali quartz trachytes. Two whole rock analyses (Table 5) confirm the observed mineralogy and saturated to oversaturated nature of the trachytes. Chemical analyses of the feldspar phenocrysts showed that they are of anorthoclase composition with very low anorthite values. The trend of feldspar crystallisation upward in single flows is of increasing content of the orthoclase molecule and decreasing albite and anorthite.

Lokwanamur basalts ($Tmvb^{LK}$), from Truckle, 1977a

The eastern extremity of these basalts extends into the NW map corner in a faulted strip between northings 2040 and 2105. They are named after the Lokwanamur Hills, the type area, which straddles the western map margin in the middle of the outcrop strip. The basalts underlie low hilly countryside.

Three samples submitted for age determinations gave an unsatisfactory range of dates from 11.4 to 15.4 Ma, best constrained by the more reliable ages of about 15.1 Ma for the underlying Kowun Volcano and about 11.8 Ma for overlying phonolites.

The lavas overstep from metamorphic basement on to the slopes of Kowun Volcano, whose profile controlled the southern thinning of the Lokwanamur basalts. In turn, the lavas are unconformably overlain by the Kasorogol phonolites or faulted against the Lomujal basalts. Trachytic stocks elongated in a NW – SE direction cut the various flows, which dip gently to the SE due to post-deposition tilting.

The formation thins to nothing away from Lokwanamur Hill where it is about 137 m thick. Upper flows overstep southwards on to the flanks of Kowun Volcano.

Eleven flows recognised in the type area are assigned to a lower Phyric Member (8 flows) and an upper Mottled Hawaiite Member. The lower unit is dominated by olivine and salite-phyric lavas with intercalated feldspar-phyric, aphyric and basaltic hawaiites and aphyric basalt flows. For the most part these lower flows are poorly exposed.

Three flows of fine-grained, grey mottled lavas with pink analcime comprise the upper member. Groundmasses of the flows consist of aligned andesine laths with poikilitic patches of intergrown clear plagioclase and dusty brown analcime as well as granular salite and opaques. Glomeroporphyritic zoned plagioclase, rare porphyritic augite and accessory biotite are set in the groundmass.

Truckle (1977a) provides three major and trace element whole rock analyses of analcime-bearing flows of which the analysis of an analcime hawaiite is reproduced in Tables 5 and 6.

Kasorogol phonolites ($Tmvp^{KS}$), from Truckle, 1977a

The faulted eastern part of these plateau phonolites extends along the western map margin between northings 1935 and 2040. The unit was mapped and named by Truckle (1977a). It forms a slightly elevated plateau easily recognisable on aerial photographs. An age of about 11.8 Ma is quoted by Truckle (1979a) for these phonolites, which are a northern extension of the Kamasia Range flood phonolites. They unconformably overlie the Lokwanamur basalts and are faulted against the younger Lomujal basalts in Degree Square 27. The flood phonolites of the Kamasia Range are the thickest (1600 m, according to King and Chapman, 1972) of all the middle Miocene phonolites of central Kenya (Martyn and others, 1968). However, in Degree Square 27 the Kasorogol phonolites are less than 100 m thick and consist of:

4 Pumice tuffs (9 m)
3 Green phonolite (?)
2 Mottled phonolites (main unit, 60 m)
1 Welded tuffs (12 to 15 m)

The upper two units are restricted to the south-eastern extremity of the plateau whereas the basal tuffs are confined to a northern outcrop area. This basal unit comprises a lower impersistent red tuff overlain by a welded tuff and by air-fall tuffs (well-bedded and graded) with an uppermost green-grey porcellaneous ignimbrite. Pumice and lithic clasts as well as altered feldspar (and rare nepheline) phenocrysts occur in these tuffs.

The mottled phonolite comprises a single flow which thickens southwards from 30 m, above the northern tuffs, to at least 60 m. According to Truckle (1977a); 'the rock type is a conspicuous dark olive-green and clots of alkali amphibole impart a marked mottling. Sanidine (up to 5 per cent) and more rarely nepheline phenocrysts (less than 2 per cent) occur but the rock is more usually almost aphyric. The groundmass is holocrystalline, with a weakly defined fluidal texture, and comprised of tightly packed feldspar laths and numerous, small nepheline euhedra set in pools of analcime and sub-poikilitic amphibole clots'. An analysis of this rock (Table 5) shows it to be a phonolite of the plateau type (cf. Lippard, 1973).

The pale green phonolite is commonly flow-banded, with rare folding and boudinage of more competent layers. The fine-grained groundmass contains nepheline (up to 4 per cent) phenocrysts.

The uppermost tuffs comprise a basal poorly-sorted agglomerate overlain by well-bedded pumice tuffs and welded ash-flow tuffs similar to the lowermost tuff unit. They overlie irregularities in the underlying phonolite surface and the whole succession appears to infill a broad depression between Lokwanamur Hill and Kowun Volcano.

The formation is tilted towards the east, with evidence for a monoclinal axis through the eastern outcrops.

Table 5 Whole rock analyses of volcanic rocks from the western shoulder of the Gregory Rift

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
SiO_2	45.94	45.90	64.38	63.73	48.71	53.67	58.07	47.42	53.14	46.50	61.85	36.97	45.11	49.42	53.65	59.19
TiO_2	2.00	2.22	0.55	0.55	2.43	0.44	1.56	1.74	2.14	2.31	0.45	4.12	3.73	2.30	1.86	0.54
Al_2O_3	16.13	12.83	15.40	16.55	16.90	17.46	17.58	14.17	18.41	16.37	17.38	6.07	15.63	17.31	16.13	12.80
Fe_2O_3	5.89	4.01	3.47	4.67	11.90	6.23	9.01	4.70	9.60	3.60	3.67	8.19	12.90	3.05	8.73	5.46
FeO	5.33	6.72	1.77	0.88	n.d.	1.60	n.d.	5.52	0.33	9.63	n.d.	19.71	2.68	7.14	1.08	3.46
MnO	0.20	0.17	0.31	0.19	0.18	0.40	0.17	0.20	0.10	0.23	0.06	0.37	0.09	0.21	0.31	0.35
MgO	4.63	9.08	0.36	0.16	4.40	0.82	1.38	8.59	0.65	6.02	0.28	14.90	3.45	2.71	1.63	0.55
CaO	8.55	12.66	0.64	0.67	7.61	2.04	4.05	10.84	4.44	8.75	1.76	4.85	9.16	7.67	4.32	1.38
Na_2O	5.39	3.31	7.02	6.84	4.14	6.84	5.62	2.75	4.75	2.97	6.76	1.76	3.09	4.58	5.58	5.08
K_2O	2.39	0.76	5.48	4.70	1.76	5.63	3.44	1.29	2.93	1.58	5.28	0.84	1.10	2.27	3.03	4.41
P_2O_5	0.70	0.35	0.05	0.11	0.58	0.04	0.38	0.33	0.72	0.42	0.03	0.50	0.80	0.85	0.65	0.04
L.O.I.	2.96	2.37	1.21	1.16	n.d.	3.23	n.d.	1.24	1.29	1.88	n.d.	2.01	2.39	2.57	3.09	6.31
Total	100.11	100.38	100.64	100.21	98.60	99.06	101.26	98.79	98.50	100.26	97.52	100.10	100.33	100.08	100.06	99.57

1 Augite-olivine basalt, Kapchererat basalts, Webb (1971)
2 Augite-olivine basalt, Kapchererat basalts, Webb (1971)
3 Trachyte, Kowun Volcano, Weaver (1973)
4 Macrophyric trachyte, Kowun Volcano, Weaver (1973)
5 Analcime hawaiite, Lokwanamur basalts, Truckle (1977a)
6 Phonolite, Kasorogol phonolites, Truckle (1977a)
7 Mugearite, Lomujal basalts (base), Truckle (1977a)
8 Picrite, Lomujal basalts (unit 2), Truckle (1977a)
9 Mugearite, Lomujal basalts (unit 3), Truckle (1977a)
10 Olivine basalt, Lomujal basalts (unit 4), Truckle (1977a)
11 Trachyte, Kailimerlim Volcano, Truckle (1977a)
12 Augite-olivine-magnetite cumulate, Tirioko basalts, Webb (1971)
13 Alkali olivine basalt, Tirioko basalts, Weaver (1973)
14 Hawaiite, Tirioko basalts, Webb (1971)
15 Trachybasalt, Tirioko basalts, Webb (1971)
16 Benmoreite, Tirioko basalts, Weaver (1973)

Lomujal basalts ($Tmvb^{LM}$), from Truckle, 1977a

These basalts are named after Lomujal Hill (GR 1680 1985) situated more or less centrally in the NW quadrant of Degree Square 27. They form a weakly dissected plateau extending from Lokori southwards for over 20 km. To the east they are concealed beneath the Nathelot basalts, and the Kailimerlim and Lokitoe trachyphonolites and unconsolidated debris associated with south bank tributaries of the Kerio River.

Ages from c.13.7 to c.16.1 Ma for lavas from the Lomujal basalts are unreliable in that younger ages were obtained from stratigraphically older flows. This unit is faulted against the c.11.8 Ma Kasorogol phonolites. In the south the basalts bank against the northern slopes of Kowun Volcano and are overlain by the c.6.6 Ma Tirioko basalts which thus provide a minimum age for the Lomujal basalts. Numerous basaltic and trachytic (including syenitic) dykes and stocks cut the Lomujal basalts as well as the Kailimerlim Foyaite stock in the north-east. Lomujal Hill is underlain by one of the largest basalt stocks. NNE-trending faults, downthrowing in an easterly direction, cut the western part of the Lomujal basalts.

According to Truckle (1977a) the Lomujal basalts are up to 600 m thick in the east (beneath the Nathelot basalts). Further west they are thinner, in part caused by extensive erosion.

The formation has easterly dips which increase towards the east from about 5° to 25°. Aberrant dips can be ascribed to vents or faulting. Nine members are recognised based on the dominant lava units and their phenocryst contents:

9 Trachyte
8 Mafic-phyric basalt
7 Aphyric
6 Feldsparphyric
5 Mafic-phyric
4 Feldsparphyric
3 Mafic-phyric
2 Aphyric
1 Feldspar mugearite

Aphyric basalts have less than 10 per cent modal phenocrysts. The mafic phyric lavas have olivine and pyroxene phenocrysts with or without accessory feldspar phenocrysts. According to Truckle (1977a) within the Lomujal basalts, variation in aphyric lithologies is from basalt and basaltic hawaiites through to trachyte. Basalts and hawaiites occur most frequently, whilst mugearites and trachytes are rare.

Individual flows appear to have covered large areas, with aphyric flows thinner (less than 5 m) than the phyric flows (up to 13 m thick). Conspicuous red soils are common. The most complete succession is preserved in the centre of the outcrop area (8 units). North to south stratigraphic variations are, in part, due to lava eruption on to an irregular topography. Unconformable relationships between flows are common. Pyroclastic rocks are conspicuously absent. Truckle (1977a) concluded that the lower part of the sequence was fed from several sources, whilst the upper flows were sourced from linear (N–S) dyke swarms.

Seventeen major (and trace element) whole rock analyses are given by Truckle (1977a) for lavas from the basal four members of the Lomujal basalts. Representative analyses are shown in Tables 5 and 6. The whole-rock analyses indicate the basalts are mildly alkaline and undersaturated; alkali content increases up the succession. The trace element analyses of whole rocks and minerals suggest a genetic link between the Lomujal basalts and the overlying Kailimerlim trachytes. Both formed by mineral fractionation of a picritic magma.

Lokitoe centre ($Tmvp^{LK}$), from Truckle, 1977a

Outcrop is confined to three small fault controlled outliers immediately north of the Lokitoe River near GR 195 180. The Lokitoe centre was originally mapped by Truckle (1977a). The component feldspar trachytes unconformably overlie the Lomujal basalts and are, in the east, overlain by the Nathelot basalts (whose c.4.5 Ma provides a minimum age for the Lokitoe centre). The feldspar trachytes which are less than 10 m thick, are cut by a radial array of trachytic dykes, but with a preferred NE-trend.

17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32	33	34
61.79	43.69	46.12	50.08	53.68	56.80	62.26	45.33	62.04	64.82	64.67	48.05	59.08	59.08	44.48	49.73	56.14	59.11
0.51	2.21	3.45	2.21	1.72	1.52	0.76	2.82	0.78	0.53	0.55	2.05	0.43	0.46	2.42	2.57	1.34	0.98
15.52	14.78	16.07	15.97	16.54	17.14	11.78	14.51	13.91	13.13	11.66	16.77	16.33	15.54	15.06	16.33	16.89	17.22
3.32	4.28	10.30	9.00	8.31	6.70	3.48	3.95	5.42	6.79	7.83	3.33	4.80	4.82	5.29	5.35	5.10	6.99
4.55	8.07	2.28	1.00	1.32	0.71	5.97	7.16	3.02	1.00	1.32	7.18	2.50	3.29	6.14	5.59	3.52	0.66
0.24	0.17	0.21	0.29	0.21	0.25	0.35	0.20	0.20	0.36	0.26	0.23	0.24	0.25	0.13	0.19	0.22	0.13
0.19	9.06	3.51	1.57	1.70	0.87	0.30	7.93	0.30	0.14	0.24	4.79	0.40	0.27	7.15	3.39	1.59	0.58
0.70	11.58	9.53	7.29	4.46	3.75	0.92	10.41	1.18	0.46	0.72	7.96	1.52	1.47	9.61	6.94	4.06	1.82
7.04	2.14	4.11	5.23	5.70	6.12	7.30	2.86	6.28	6.97	6.43	3.99	7.13	6.55	2.48	4.59	6.13	6.87
5.23	1.04	1.48	2.52	2.98	3.14	4.60	1.65	5.05	5.00	5.19	2.51	5.31	4.99	0.91	1.95	3.23	3.74
0.01	0.45	0.75	0.57	0.52	0.37	0.03	0.56	0.08	0.03	0.04	0.48	0.03	0.03	0.40	0.72	0.32	0.26
0.95	2.41	2.62	4.28	2.86	2.72	2.10	2.64	2.14	1.24	1.78	2.96	2.32	2.75	5.85	2.52	1.20	1.62
100.05	99.88	100.43	100.01	100.00	100.09	99.85	100.02	100.40	100.47	100.69	100.30	100.09	100.24	99.92	99.87	99.74	99.98

17 Trachyte, Kafkandal Volcano, Weaver (1973)
18 Olivine basalt, Tumungir basalts, Webb (1971)
19 Olivine basalt, Kachila Volcano, Webb (1971)
20 Mugearite, Kachila Volcano, Webb (1971)
21 Trachybasalt, Kachila Volcano, Webb (1971)
22 Benmoreite, Kachila Volcano, Webb (1971)
23 Trachyte, Kachila Volcano, Webb (1971)
24 Olivine basalt, Nasaken Volcano, Weaver and others (1972)
25 Trachyte, Nasaken Volcano, Weaver and others (1972)
26 Quartz trachyte, Nasaken Volcano, Weaver and others (1972)
27 Pantelleritic trachyte, Nasaken Volcano, Weaver and others (1972)
28 Olivine basalt, Ribkwo Volcano, Weaver and others (1972)
29 Trachyte, Ribkwo Volcano, Weaver and others (1972)
30 Trachyte, Ribkwo Volcano, Weaver and others (1972)
31 Olivine basalt, Lokwaleibit basalts, McClenaghan (1972)
32 Hawaiite, Lokwaleibit basalts, McClenaghan (1971)
33 Mugearite, Jamakana formation, McClenaghan (1971)
34 Mugearite, Jamakana formation, McClenaghan (1971)

A single feldsparphyric trachyte flow is overlain by three strongly weathered welded tuffs with intercalated red, waterlain grits. A single major and trace element whole rock analysis of the trachyte lava is given by Truckle (1977a).

The Lokitoe centre was almost certainly contemporaneous with the lithologically similar but more extensive Kailimerlim Volcano immediately to the north. They are part of the Late Miocene trachyphonolitic volcanism of the western shoulder. However, the Lokitoe centre was largely removed by erosion prior to subsequent basalt eruptions. It was probably never an important topographic feature.

Kailimerlim Volcano ($Tmvp^{KL}$), from Truckle, 1977a and 1979a

Kailimerlim Volcano is confined to the NW quadrant of Degree Square 27 where outcrop is limited by an extensive mantle of unconsolidated debris related to south bank tributaries of the Kerio River.

It is named after Kailimerlim where the best exposures are preserved. Once again the original mapping was done by Truckle (1977a). The volcano forms a gentle shield whose N–S axis extended through Kailimerlim and Ayangyangi. Its constituent lithologies are susceptible to weathering so that they are best exposed on scarp slopes unconformably beneath the more resistant Nathelot basalts. Dates of about 15.4 and 14.3 Ma are quoted by Truckle (1977a) for two trachytic lavas which unconformably overlie the Lomujal basalts and are cut by the Kailimerlim Foyaite stock.

Its central thickness is about 600 m but thins radially (e.g. beneath the Nathelot basalts at Kaliba); in the west the volcano did not extend as far as Nakathuen. The Lokitoe centre may represent the southern margin of Kailimerlim Volcano. The following stratigraphy is given by Truckle (1979a).

4 Phonolites
3 Phonolitic trachyte
2 Welded tuff
1 Feldspar trachyte

The upper 3 units are confined to the north-west between Lokori and Nakathuen, so that the volcano is essentially composed of feldspar trachyte lavas. These lavas dip away from Kailimerlim Foyaite plug to define a half cone structure. Dip values and flow thicknesses increase towards this plug which probably marks a source area for the lavas. South of Kailimerlim, up to 12 m of graded air-fall pyroclastics crop out beneath a single grey trachyte flow. Lithic clasts are mostly derived from underlying basalts: these pyroclastics represent an initial explosive stage of Kailimerlim Volcano. This whole edifice has been tilted eastwards by up to 10° and cut by trachyte dykes (dated at about 14.5 Ma).

Truckle (1977a) provides two major and trace element whole rock analyses of the trachyte lavas. The chemistry suggests a genetic link with the Lomujal basalts. Truckle (*op.cit.*) provides the following summary of the geological history:

1 Lomujal basalts erupted as coalescing low angle shields from several sources.
2 Eastward tilting by up to 10°.
3 Trachyte eruptions with accompanying trachyte dyke, sill and phonolite plug emplacement (see Section 6.6).
4 Further eastward tilting by up to 10°.

Tiati group ($Tmvt^{TA}$). **Chepkoi tuffs and Kameiyun volcanics**

The easternmost tip of this laterally extensive group of mixed trachyphonolitic volcanics just extends into the south-west of Degree Square 27 near northing 140. Webb (1971) and McClenaghan (1971) mapped most of this volcanic sequence and followed Martyn (1969) in referring to it as the 'Tugen Hills Group', subsequently renamed by Truckle (1979a) the Tiati group (after Tiati

Hill which is the main topographic feature in the centre of the volcanics). Reference areas lie to the west of Degree Square 27.

The upper Chepkoi tuffs underlie low, fault-controlled hills: the underlying Kameiyun volcanics are confined to valley bottoms. In the map-area they are unconformably overlain by the (c.6.9 Ma) Tirioko basalts. They dip eastwards at about 15° according to Truckle (1979a), continuing for at least 8 km under the Tirioko basalts towards the median rift.

The Kameiyun volcanics consist of at least 150 m of flow banded (flinty) dark green and grey trachyphonolite lavas with various pyroclastics (mostly to the north-west of the inlier on Degree Square 27), including welded and non-welded tuffs and agglomerates. The welded tuffs have strongly contorted flow-banding with lithic, crystal (sanidine and biotite) clasts and devitrified glass. The non-welded tuffs are white and finely laminated. Sodalite and nepheline occur in the lavas.

The Chepkoi tuffs consist of about 185 m of monotonous pumice lapilli tuffs. They are well-bedded, poorly-sorted with local grading and rare agglomerate lenses.

Williams and Chapman (1986) suggest that the Tiati group and the contemporaneous Alengerr and Katomuk tuffs to the east relate to a proto-Silali Volcano. Its Quaternary edifice is central to the distribution of these Miocene pyroclastics.

Tirioko basalts ($Tmvb^{TR}$)

These basalts were named by Webb (1971) during his initial mapping of the area south of northing 163 where they are particularly well exposed west of longitude 36°E. Subsequently Weaver (1973) and Truckle (1977a) traced the Tirioko basalts northwards into the NW quadrant of Degree Square 27. Truckle (*op.cit.*) refers to this unit as the Napeitom basalts. They have a total NNE strike length of about 54 km north from the Tiati group inlier to Raparap.

The Tirioko basalts underlie strongly dissected hilly countryside with rare cone-shaped hills (GR 1692 1854) and major linear fault escarpments, especially in the north. Basalt and trachyte dykes form low, boulder-covered ridges which, like the fault scarps, are readily visible on aerial photographs. General descriptions, which could serve to define the Tirioko basalts are given by Webb (1971), Weaver (1973) and Truckle (1977a). Exposure is generally good, although many of the basalt outcrops are strongly weathered.

Isotopic ages of 6.6 ± 0.7 Ma and 6.6 ± 0.3 Ma are quoted by Weaver (1973), and 9.3 ± 0.3 Ma and 5.6 ± 0.2 Ma (Truckle, 1977a) have been obtained from basalts and mugearites. Truckle (1979a) decided, from these data, on an age of c.6.9 Ma for the Tirioko basalts, i.e. late Miocene.

Lower and upper contacts are markedly unconformable: the Tirioko basalts rest upon an uneven surface of Kowun Volcano rocks, the Lomujal basalts and the Tiati group. Dyke swarms in the Lomujal basalts are truncated by basal flows of the Tirioko basalts. They are variously overlain by rocks of the Nasaken, Ribkwo and Kafkandal Volcanoes and the Nathelot and Tumungir basalts. Numerous basalt and trachyte dykes and stocks cut the Tirioko basalts. Within Degree Square 27 the Tirioko basalts are from about 150 m (Raparap) to about 300 m thick, but thicken to over 600 m westwards. Overstepping of flows is common, in part due to a rugged depositional landscape.

Porphyritic and non-porphyritic basalts and hawaiites predominate with minor basanites, ankaramites, mugearites and trachytes (including pyroclastics). Dyke and central cone source are identified. The commonest lava is a microphyric or aphyric basalt which is massive, bluish black and sparsely vesicular with or without amygdales.

It weathers (especially adjacent to faults) to reddish yellow hues. Mafic phyric basalts are rare, with olivine and augite phenocrysts in black, fine-grained massive lavas which readily weather to reddish brown earthy material (e.g. at GR 1693 1547). Feldsparphyric basalts and hawaiites are equally scarce (except in the south) and susceptible to weathering. They are grey to purple-grey lavas with glassy tabular plagioclase phenocrysts up to 2 cm long, and ropy surfaces with *pahoehoe* toes.

Mugearites and trachytes form thin (c.15 m thick) rare flows (e.g. GR 1687 1543) which are aphyric (sparse feldspar phenocrysts), buff to brown, fissile and massive with or without vesicular bases and tops. Welded tuffs (GR 1705 1810) have eutaxitic textures with flattened pumice fiamme in red-brown porcellaneous groundmasses. A single diatomite layer, 0.6 m thick is pure white with a shaly texture (GR 1690 1519). The cone at GR 1692 1854 is capped by a basalt with perfect hexagonal columnar jointing.

Weaver (1973) provides excellent detailed petrographic descriptions of the component lithologies. He notes that the lavas of the Tirioko basalts are predominantly alkaline olivine basalts. Hawaiites and mugearites are well represented locally and anorthoclase trachytes confined to the south. The basalts are frequently rich in microphenocrysts with the following common phenocrysts assemblages (first mineral is dominant):

olivine + augite
augite + olivine
augite + plagioclase
plagioclase
plagioclase + augite + olivine

Analcime is present in groundmasses (with intergranular, rarely ophitic textures) of colourless augite, olivine and magnetite interstitially between plagioclase laths.

Hawaiites and mugearites are aphyric and fine-grained. Abundant magnetite is disseminated through the groundmass; olivine is pseudomorphed by iddingsite and the clinopyroxene is pale green. Kataphorite was noted in mugearitic rocks.

Mineral optics and chemistry undertaken by Weaver (1973) indicated the olivines range in composition from Fa 16 to Fa 31, clinopyroxenes are diopsidic augite or augite in basic lavas, ferroaugite in intermediate lavas and hedenbergites (phenocrysts) or aegirine-augite (groundmass) in felsic lavas; plagioclase phenocrysts range from An 80 to An 15 in the basalt-trachymugearite suite.

Thirteen whole rock (major and trace element) analyses were determined by Weaver (1973) to include eight basalts, one hawaiite, two mugearites and two trachytes. A further eight whole rock analyses by Webb (1971) included four basalts, one hawaiite and three trachytes. On an alkali against SiO_2 plot (Cox and others, 1979) all the analyses show a linear trend from picritic basalts through alkali basalts, hawaiites and trachybasalts to benmoreites. They all plot in the nepheline-free range (Hughes, 1982) for lavas of East Africa.

Weaver (1973) noted that most of the basaltic lavas have nepheline in the norm (but no modal nepheline):

only three lavas contained normative quartz. The basalts have high alkali contents with Na_2O always greater than K_2O and compared with alkaline basalts from other provinces the SiO_2 contents are low: from 44.00 wt per cent in an ankaramite to 55.94 wt per cent in a trachymugearite. Ba and Sr contents in basalts are high whereas Zr, Nb, Rb and the light REE increase systematically from the basalts to the trachytes.

The Tirioko basalts are regarded as the eroded remnants of several coalescing low-angle shield volcanoes which had central and dyke sources. Subsequent erosion gives the false impression that they represent a flood basalt sequence. According to Truckle (1977a) the Nasaken trachytic lavas were erupted soon after the upper flows of the Tirioko basalts: the two units representing a single basalt/trachyte cycle. Gentle eastward tilting of the basalts preceded the trachytic volcanism.

Kafkandal Volcano ($Tmvt^{KF}$), after Webb, 1971; Weaver, 1973

Kafkandal Volcano, named after a northern peak, is the oldest of a series of Late Miocene through Pliocene central trachytic volcanoes. It straddles latitude 1°30′N and has a roughly circular plan (10 km diameter), which is a remnant of a more extensive original surface area. The volcano has a classic shield profile rising about 600 m towards central summit heights of about 1540 m. The landscape is impressive with numerous fault scarps, prominent hills related to volcanic centres and intrusive stocks, and a general lack of vegetation.

The northern and southern halves of Kafkandal Volcano were mapped by Weaver (1973) and Webb (1971) respectively. They recognised three stratigraphic units related to three eruptive centres (Figure 12). Reference sections for these units are as follows:

3 K3 (Moru Angitak) GR 1688 1639
2 K2 (Epong) GR 1733 1637
1 K1 (Ngapawoi) GR 1695 1589

Weaver (1973) quotes an isotopic age of 5.9 + 0.3 Ma for the basal trachytes and an age of 5.7 + 0.3 Ma for the middle unit.

The structural relations of Kafkandal are clearly defined: in the west and south it unconformably overlies the Tirioko basalts; to the north and north-east it disappears beneath the southern parts of Nasaken Volcano; similarly to the south-east and south it is overlapped by Kachila and Ribkwo Volcanoes respectively. Three main centres are identified within Kafkandal, related to the three stratigraphic units. Disconformities occur between flows from the different centres and the whole volcano has been tilted eastwards by 5 to 10°. Numerous trachytic plugs and dykes were coeval with, and cut, the extrusive volcanism. NNE-trending normal faults also cut, and locally controlled (the K3 centre), the volcanism.

Kafkandal Volcano presently has a maximum thickness of 610 m of extrusive trachytic material. Individual trachytic lava flows are between 20 and 30 m thick. The thickness of pyroclastic accumulation increases rapidly towards the three main centres; at Moru Angitak there are 300 m of chaotic white trachytic pumice tuffs. The trachytic lavas are usually grey-green, massive porphyritic rocks with thick basal breccias. Phenocrysts are mostly anorthoclase. The lavas are oversaturated with up to 10 per cent modal quartz and without a feldspathoid phase. The trachytic tuffs are friable with extensive clay alteration. Distal parts are well-bedded with respect to proximal chaotic accumulations. Trachyte, rarely syenite, lapilli, volcanic glass and feldspar crystal fragments accompany pumice clasts in the pyroclastics.

The basal Ngapawoi unit underlies most of the volcano and comprises trachytic lavas with pumice tuff lenses, rare welded tuff beds and irregular trachytic intrusions. Initial eruptions were from dispersed centres followed by fissure controlled activity which formed a broad shield (the upper lavas have quaquaversal dips). Glassy bands and streaks parallel flow textures in the lavas. The Epong unit conformably overlies the Ngapowoi Shield with eruptions confined to a more easterly centre. Non-welded, white and yellow, pumice lapilli-tuffs with feldspar fragments dominate this unit, and probably represent a

Table 6 Trace element analyses for the western shoulder volcanic rocks

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
Ba	226	407	571	710	960	1320	1290	818	1	19	715	208	11	32				
Ce	48	55	77	96	156	146	144	118	411	597	106	163	197	518				
La	17	23	33	50	69	79	80	59	231	338	50	94	126	312				
Nb	21	27	45	54	86	92	95	84	298	421	68	127	204	466	75	95	52	174
Rb	12	27	52	50	62	85	86	60	168	211	39	88	162	294	41	89	40	129
Sr	531	944	1227	740	805	788	775	935	27	21	748	14	19	19	817	594	759	11
Zr	64	103	137	180	307	316	308	213	904	1256	195	392	734	1660	172	296	163	452
Nd	19	18	37	36	51	62	62											
Pr	7	7	9	11	13	16	15											
K/Rb	408	335	249	461	359	295	295	418	316	236	423	574	309	177	356	695	327	340

1 Picritic basalt
2 Alkali basalt
3 Alkali basalt
4 Hawaiite
5 Trachybasalt
6 Benmoreite
7 Trachyte
— Tirioko basalts, Weaver (1973)

8 Olivine basalt
9 Trachyte
10 Trachyte
— Ribkwo Volcano, Weaver and others (1972)

11 Olivine basalt
12 Trachyte
13 Quartz trachyte
14 Pantelleritic trachyte
— Nasaken Volcano, Weaver and others (1972)

15 Analcime hawaiite, Lokwanamur basalts, Truckle (1977a)

16 Basal mugearite
17 Olivine basalt
— Lomujal basalts, Truckle (1977a)

18 Trachyte, Kailimerlim Volcano, Truckle (1977a)

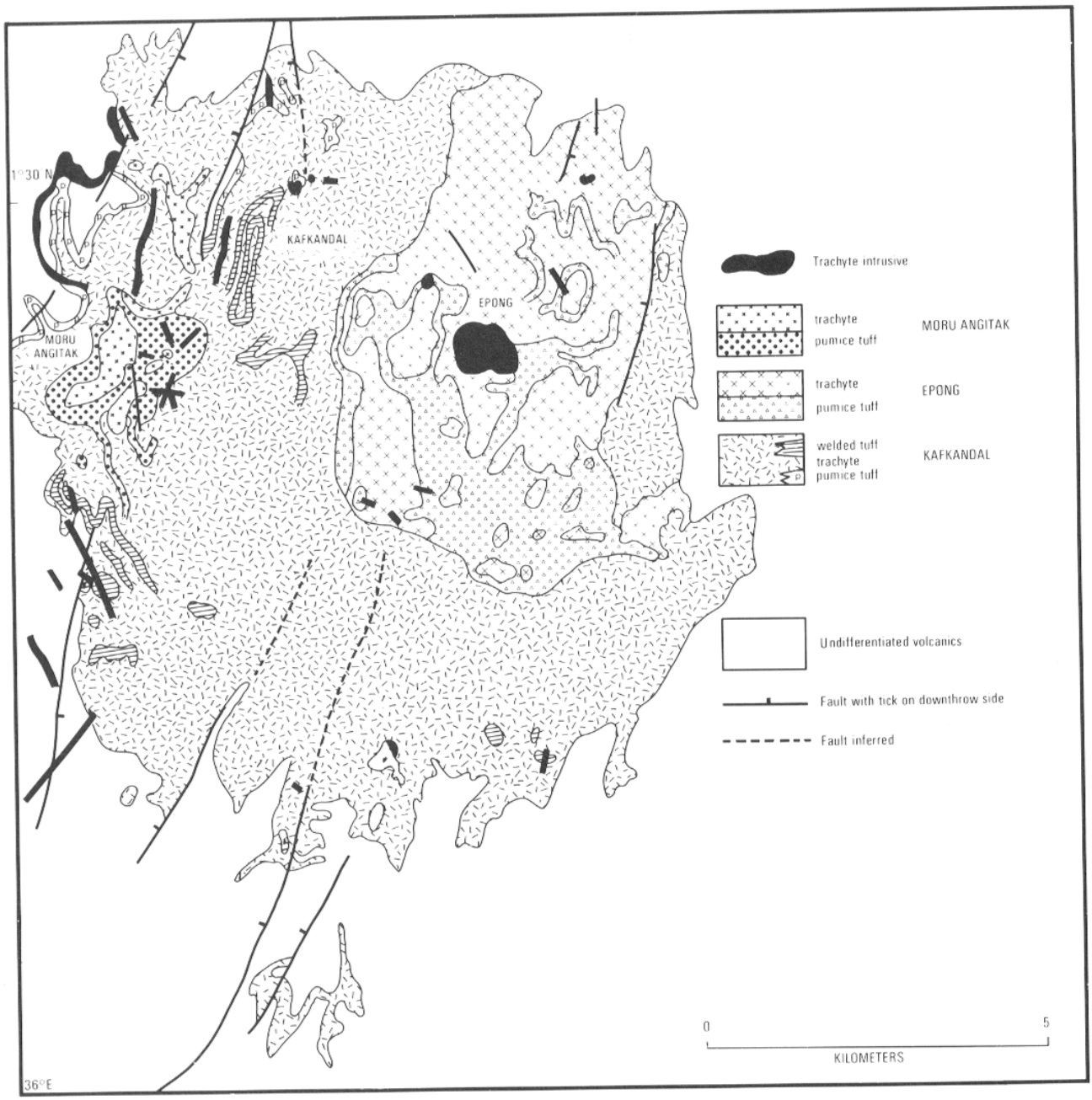


Figure 12 The main units of Kafkandal Volcano (from Truckle, 1979a & b)

collapsed crater infill (moderate inward dips preserved on the west and south walls). Only three thin welded tuff beds are exposed and trachyte lavas are concentrated towards the top of the Epong unit. Finally the Moru Angitak unit erupted through the western part of the Ngapawoi Shield with about 300 m of chaotic tuffs defining the eruptive centre. Again welded tuffs are rare, and trachyte lavas are confined to the upper and distal parts including a 50 m thick uppermost flow (feeder dykes to this flow cut the underlying tuffs.)

Petrographically the lavas vary from soda trachytes, quartz trachytes to quartz-rhyolites (up to 20 per cent modal quartz). The phenocryst phase is invariably alkali feldspar (anorthoclase with Or 12 to Or 38) which forms between 5 and 20 per cent of the mode. Rare phenocrysts of hedenbergite (pseudomorphed by hydrobiotite), magnetite, fayalite, arfvedsonite, aenigmatite, and diopsidic augite were also noted. Groundmasses contain slender alkali feldspar laths with varying amounts of alkali amphibole (usually in mossy poikilitic patches), aenigmatite, quartz with minor pyroxene.

Tumungir basalts (Tmvb^{TM}), from Webb, 1971

These were first mapped by Webb (1971, p.54) on the southern edge of Kafkandal Volcano where they form a series of irregular thin erosional remnants trending NNE for about 10 km. They cap low hills or may be sandwiched, between the Tirioko basalts and the flanks of either Ribkwo or Kachila Volcanoes.

The type locality is Tumungir Kongo (GR 1698 1532) where a maximum thickness of 13.5 m is preserved (comprising three flows each of 4.5 m thickness). At Tumungir Udei (GR 1691 1561 and GR 1700 1558) up to 5 m of pyroclastics are exposed.

No isotopic dates for the Tumungir basalts are available. They unconformably overlie the Tirioko basalts and Kafkandal Volcano to give a maximum age of about

5.7 Ma. As they are overlain by the basal units of Ribkwo Volcano, dated at about 5.2 Ma, the Tumungir basalts are clearly Late Miocene in age. They are cut by NNE faults and the lava flows dip very gently ENE.

In the type area the basalts are black, non-vesicular and fresh: olivine phenocrysts are confined to the basal flow. The pyroclastics at Tumungir Udei are a grey, partly welded tuff and an agglomerate containing spindle bombs in a crudely stratified matrix. All three lava flows have intergranular textures and pale brown augite, and plagioclase phenocrysts were noted in thin section in addition to the olivine visible in hand specimen.

The Tumungir basalts formed a relatively short-lived (less than 0.5 Ma) minor basaltic province with several flows emanating from a single (northern) centre.

Kachila Volcano ($Tmvt^{KC}$), from Webb, 1971

Kachila Volcano was mapped by Webb (1971) who named it after a major eastern hill (GR 1808 1527) close to the Kapedo to Lokori track. Its eastern parts are concealed beneath Suguta River alluvium and younger volcanics of Oliyamur and Kanatim Volcanoes. In the north it apparently interdigitates with the supposedly contemporaneous Nasaken Volcano. Presently Kachila Volcano has an oval plan with a c.25 km long (N–S) axis and a c.7 km short (E–W) axis. Its flat-lying, well bedded strata form low mesas strongly incised by west bank tributaries of the Suguta River, which locally follow NNE fault traces.

The detailed account of Webb (1971) could serve to formalise the stratigraphy, which shows pronounced lateral facies changes, notably on either side of the Aterir River. Kachila Hill (GR 1808 1527) provides good outcrop of northern lithologies and the exposures south of Aterir (GR 176 151) are typical of the southern tuffaceous material.

Webb (1971) quotes five isotopic ages of between 2.7 ± 0.2 Ma and 4.8 ± 0.2 Ma for lavas from Kachila Volcano. However, Truckle (1979a) has not quoted these ages but rather equates Kachila with Nasaken Volcanoes with a time span of 5.7 to 5.4 Ma. The regional chronology is also confusing. Kachila Volcano overlies the c.5.7 Ma Kafkandal Volcano and the c.5.2 to 4.1 Ma Ribkwo Volcano but is overlain by the c.4.3 Ma Kanatim Volcano.

The southern part of Kachila Volcano is not only covered by large areas of alluvium but also by a major flow of the Murgisian basalts.

A maximum thickness of 275 m is given by Webb (1971) on, or west of, Kachila Hill. Generally thicknesses in the north-west and north-east of the volcano are about 200 and 120 m respectively. South of the Aterir River the total thickness decreases to about 60 m.

No source area (dykes or vents) is exposed for Kachila Volcano; it is concealed either beneath overlying flat-lying lavas or (less likely from the exposed facies changes) by the younger eastern volcanoes. Mugearites and basalts form an unusually high proportion of the lavas of this volcano relative to the adjacent trachytic volcanoes. Pumice tuffs are locally very thick; up to 185 m at GR 1766 1550 and GR 1773 1535. The Aterir beds (in the south) consist of tuffaceous sediments.

In the north-west the typical succession starts with about 200 m of variably bedded, locally welded pumice lapilli tuffs with several olivine-basalt flows and one or more trachyte flows. They are overlain by two aphyric olivine-basalt flows, each 10 m thick, intercalated with 3 m of welded tuff. In the north-east the pyroclastics give way to a succession of basalt, mugearite and trachyte flows with minor pumice tuff lavas.

South of the Aterir River, basal basalts are overlain by tuffaceous sediments (Aterir beds) with two thin intercalations of basalt. These may be overlain by fissile trachyte lavas, welded and pumice tuffs with capping mugearites and plagioclase trachytes. Vertebrate fossils have been discovered in the sediments at GR 1756 1525, 1759 1500, 1754 1498, 1757 1498 and 1760 1488. The complete faunal list is:

Reptile
Chelonia
Pelomedusid
Crocodilia
Mammalia
Proboscidea cf.
Anancus sp.
Elephas cf.
subplanifrons
Perissodactyla
Equidae
Hipparion
(*Stylohipparion*) *sp.*
Rhinocerotidae
Artiodactyla
Suidae
cf. *Sus sp.*
cf. *Nyanzachoerus sp.*
Hippopotamidae
Bovidae

All basalts examined in thin section contained olivine (or iddingsite pseudomorphs of olivine) with or without plagioclase and augite phenocrysts or microphenocrysts: the same three minerals constitute the groundmass. Most have intergranular textures, rarely ophitic, or fluxion textures.

The mugearites contain oligoclase phenocrysts (with alkali feldspar mantles) in an indeterminate groundmass. In the trachytes there are plagioclase and sanidine (rarely anorthoclase) phenocrysts in a groundmass of sanidine, altered mafics (aegirine, arfvedsonite and olivine) and magnetite. Possible relict feldspathoid grains were noted by Webb (1971) in only one sample.

Nine rock analyses of Kachila Volcano lavas are listed by Webb (1971) and they follow the typical (western shoulder) trend from alkali olivine-basalt (2), mugearite (1), trachybasalt (4), benmoreite (1) to trachyte (1) on an alkali against silica plot.

Nasaken Volcano ($Tmvt^{NS}$)

This volcano takes its name from a prominent northern hill (GR 1749 1798); the Nasaken River also meanders through the volcano's northern area. Weaver (1973) mapped most of the volcano apart from a southern intercalation with Kachila Volcano which Webb (1971) surveyed. Nasaken Volcano presently underlies an area of about 350 km^2 to the north and east of Kafkandal Volcano. Its eastern parts are concealed beneath younger volcanics. Its western margin was controlled both by NNE fault scarps and the upstanding parts of Kafkandal Volcano.

Morphologically Nasaken and Kachila Volcanics are similar with incised mesas capped by massive trachyte lavas. Uncapped pyroclastics weather to low badlands with gravel and scree veneers. Preferential erosion of basal pyroclastics has undermined many mesas.

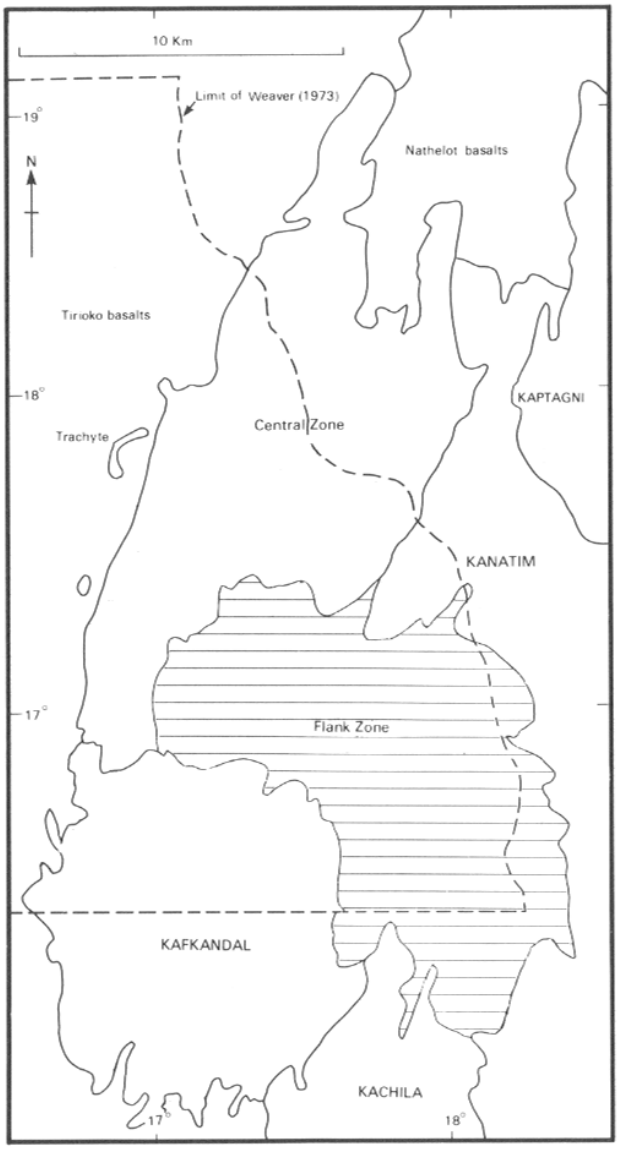


Figure 13 The main units of Nasaken Volcano (from Weaver, 1973)

Weaver (1973) provides enough detail for formal stratigraphic definition. Essentially the volcano consists of a main northern area and a south-easterly flanking unit. Typical northern successions are well-exposed at GR 1726 1788, 1735 1725 and 1749 1798 (Figure 13). An early feeder dyke swarm (trending 035°) is well-exposed at GR 1762 1760. The south-east flank deposits are well-exposed in the numerous river gorges.

Weaver (1973) quotes reliable ages of c.5.7 and c.5.4 Ma for Nasaken Volcano. The regional stratigraphy indicates an age of older then 4.5 Ma (the age of the overlying Nathelot basalts). Kanatim Volcano also overlies the eastern parts of Nasaken Volcano, which itself overlies the Tirioko basalts and Kafkandal Volcano. To the south it intercalates with the northern parts of Kachila Volcano. The major NNE fault forming the north-west margin controlled the early volcanicity of Nasaken and was subsequently reactivated. Thus the early centres are concentrated along the fault and the late downthrowing eastwards partly contributed to the present gentle easterly dip of the whole volcanic complex. Late volcanic centres are also concentrated in the main northern part of Nasaken principally at Nakwamoroi (GR 1733 1755).

Both dyke swarm (trending NNE) and central vent sources are recognised. The volcano is also cut by late trachyte dykes and plugs.

Over 200 m and less than 150 m of extrusive material underlies the northern and southern parts of Nasaken respectively. The maximum measured thickness of 270 m occurs at GR 1735 1725 although a further 130 m of volcanics underlies the base of this succession further to the north. Individual trachyte lava flows are, on average, about 30 m thick but were locally measured at 80 m.

Trachyte lavas dominate the northern succession, with mixed trachytic welded tuffs and lavas within the south-east flank. Basalts, hawaiites, and mugearites make up less than 5 per cent of the volcano. Pyroclastic deposits comprise about 30 per cent of the volcano, concentrated around the main vent sources and in the south-east. Pumice tuffs, pumice-lapilli tuffs, crystal tuffs, ash-flow tuffs (mostly the south-eastern welded tuffs), ashes and agglomerates are exposed as well as small pockets of intra-volcanic epiclastic sediments with thin diatomaceous beds. Explosion breccias help locate vent sources. The trachytic lavas are typically green to grey when fresh, with a pronounced fissility due to the strong alignment of feldspar laths. Individual flows are varied with vesicular tops and complex basal zones comprising lowermost angular breccias, central finer welded breccias and upper massive aphyric (chilled) zones beneath the main fissile trachyte. The variability of the main northern successions is summarised in Figure 14.

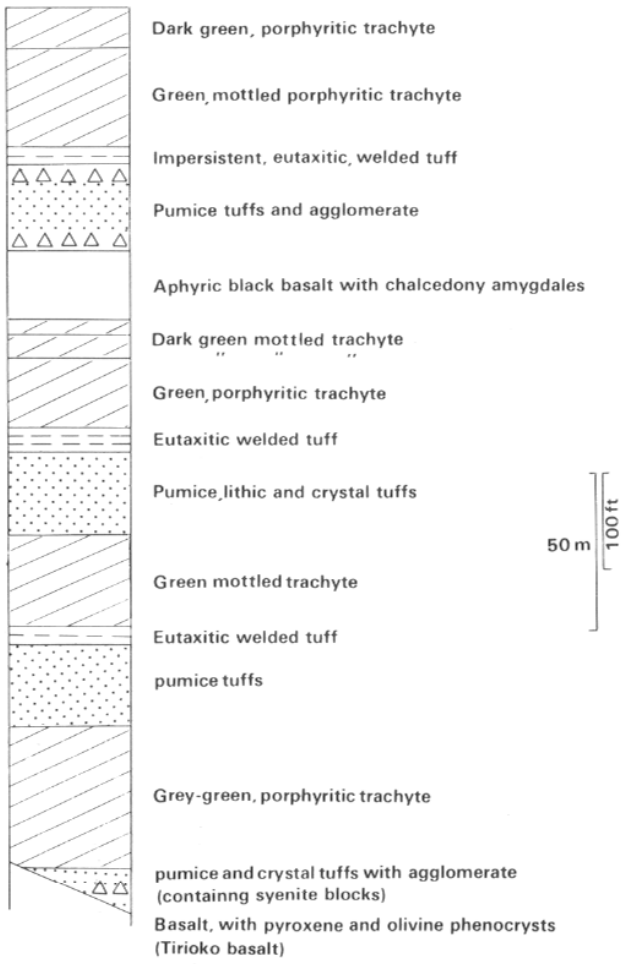


Figure 14 A section through Nasaken Volcano near Emus Ridge (from Weaver, 1973)

The south-east succession consists of three basal olivine basalt flows overlain by a series of alternating fissile trachytes (in flows up to 19 m thick), and welded tuffs with minor pumice tuffs. The trachytic volcanism shows a tripartite cyclicity commencing with pumice tuffs overlain by ash-flow tuffs and final trachyte lava eruption.

The rare basic lavas are either aphyric or have olivine and plagioclase phenocrysts in ophitic or intergranular groundmasses. The trachytes are petrographically identical to the Kachila Volcano trachytes, i.e. they contain anorthoclase phenocrysts with or without minor hedenbergite, magnetite, fayalite, arfvedsonite or aenigmatite phenocrysts in a groundmass of slender alkali feldspar laths, alkali amphibole, interstitial aenigmatite and minor pyroxene (hedenbergite, aegirine-augite or aegirine). No feldspathoid was found in these peralkaline soda- or quartz-trachytes.

Weaver (1973) lists 30 whole rock analyses which show a mildly alkaline and undersaturated trend. Twenty-seven of his analysed rocks have over 59 per cent SiO_2 (i.e. quartz normative); the other three are slightly nepheline normative. The rocks classify as rhyolites (2), trachytes (24), benmoreites (1), trachybasalts (2) and alkali basalt (1) on an alkali against SiO_2 plot. From the trace element contents of these rocks Weaver and others (1972) concluded that Nasaken Volcano originated by fractional crystallisation of an alkali basalt source which initially produced a high level salic cupola. Truckle (1977a) also suggested a genetic link with the underlying Tirioko basalts.

Ribkwo Volcano ($Tpvt^{RB}$)

Ribkwo Volcano is a shallow dome-shaped, multicentred phonolitic trachyte volcano, most of which is exposed in the SW quadrant of Degree Square 27. It is named after Ribkwo Hill and the Ribkwo Plateau (immediately west of the map-area), and has an oval plan with a pronounced (40 km) NNE axis. The volcano has two contrasting morphologies: an eastern central, generally rugged area and a western and north-western, deeply dissected plateau. The eastern terrain comprises a NNE chain of centres with assorted pyroclastics, lavas and minor intrusives. Their different susceptibilities to weathering has caused the rugged topography: the high points relate to trachytic stocks. Erratic dips have also contributed towards the scenery in contrast to the subhorizontal lavas of the plateau area. Deep dissection provides impressive sections through the western lava pile which lacks tuff interlayers (except in the north).

The southern and northern halves of Ribkwo Volcano were mapped by McClenaghan (1971) and Webb (1971) respectively, both of whom provide excellent detail of its geology (Figure 15). Between them they recognised six stratigraphic units (related to separate centres) in the eastern central terrain and three contemporaneous lava units in the plateau area. Good reference sections for the central units are as follows (parenthesised superscripts refer to those in Truckle, 1979a).

6	Kariamangro Centre	(Rc 6)	GR 1770 1400	
5	Lokitet Centre	(Rc 5)	GR 1760 1382	
4	Kukubanga Centre	(Rc 4)	GR 1740 1350	Kablaton
3	Okwabanga Centre	(Rc 3)	GR 1725 1330	
2	Kynmot Centre	(Rc 2)	GR 1685 1280	
1	Kultulpau lavas	(Rc 1)	West of Degree Square 27	

The plateau lavas are well exposed in the various gorges, although the northern tuffs intercalated with the middle

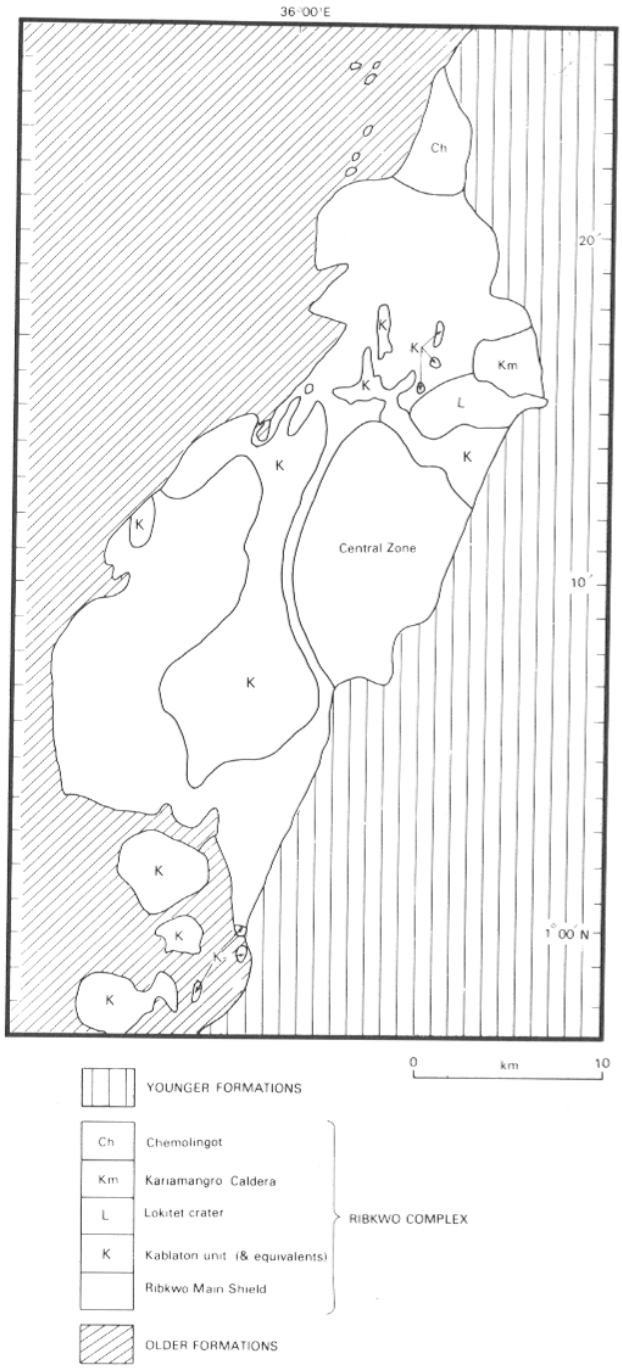


Figure 15 The main units of Ribkwo Volcano

(Rb2) trachyte are best seen between GR 1735 1490 and GR 1742 1527.

According to the reliable isotopic ages quoted by McClenaghan (1971) and Webb (1971) the volcano was active between about 5.2 Ma and c.4.4 Ma. In Degree Square 27 it overlies the Tirioko basalts and the inlier of the Tiati group. It underlies Kachila and (the western parts of) Silali Volcanoes, Murgisian and Lokwaleibit basalts, the Jamakana formation and is partly concealed by unconsolidated debris of the Suguta Trough. It is cut, especially in the east, by numerous trachyte and basalt dykes and stocks, many of which represent feeder dykes to the lavas. The dykes have a preferred NNE-trend parallel to the relatively few faults cutting Ribkwo. Similarly there is a progressive NNE movement of the main centres with time in the central belt.

There is a maximum vertical thickness of over 760 m of extrusive rock under Ribkwo Volcano according to Webb (1971).

A summary of the stratigraphy is given below.

Central Zone (top)		Western Plateau
Kariamangro Centre	trachyte lavas pumice tuffs (2nd period of caldera collapse)	
Lokitet Centre	trachyte lavas pumice tuffs (1st period of caldera collapse, minor intrusive trachyte stocks)	*Rb 3*: trachyte lavas
Unconformity	~~~~~~~~~~~~~~~~~~~~~~~~~~~~	
Kukubanga Centre	trachyte and phonolitic trachyte lava (NNE trachyte feeder dyke swarm)	*Rb 2*: aphyric trachyte lavas
Okwabanga Centre	trachyphonolite and trachyte lavas, pumice tuffs, lahars, trachyte lavas agglomerates	
Kynmot Centre	pumice tuffs agglomerates pumice tuffs	*Rb 1*: trachytic lavas with upper breccia and lower tuffs
Kultulpau lavas	pumice tuff lava breccia trachyte lavas	

Most of the lavas are fissile, porphyritic, dark green-grey, weathering to dark-green or red-brown slabs. They have large phenocrysts of sanidine or anorthoclase. The central area has a high proportion of yellow, white and brown tuffs to include pumice, lithic, lapilli and crystal types associated with agglomerates, lava breccias, lahars and trachytic welded tuffs. Thick tuff beds are rare but numerous thin tuff lenses are separated by thick, but short (less than 10 km length) lava flows. Intercalated tuffs are rare within the plateau lavas except in the north. Mugearite and olivine basalts are locally present in the plateau area.

The basal Kultulpau lavas have a distinctive purple-green colour. They formed a shallow dome which thinned westwards. The Kynmot Centre was produced by a slight southerly shift in the focus of eruptions to form a second shallow tuff and lava dome cut by numerous feeder dykes to late lavas. At this time the plateau lavas started to erupt and a contemporaneous northern pyroclastic centre existed. The Okwabanga Centre represents a slight NNE shift of the eruptive centre to produce alternating purplish trachytes and agglomerates. The upper flows are about 30 m thick; all the rocks have erratically variable dips related to numerous, closely spaced centres, including a NNE dyke swarm. A continuation of the NNE migration of the focus of eruptions produced the Kukubanga Centre with the lavas emanating from a NNE dyke swarm with related sills and plugs. Thin yellow lapilli tuffs are intercalated in the lavas. The contemporaneous (Rb 2) trachyte is an aphyric 60 m thick single flow from a plug source.

Unconformably above the aforementioned central rocks are the basal lapilli and pumice tuffs of the Lokitet Centre, which grade eastwards into agglomerates. Caldera collapse followed the tuff eruptions prior to trachyte lava (extended on to plateau) and welded tuff eruptions. The lavas had NNE-trending feeder dykes. Finally the Kariamangro Centre is essentially a repeat of the volcanism of the Lokitet Centre with a second phase of caldera collapse.

In thin sections the lavas were seen to contain nepheline microphenocrysts as well as the large feldspar phenocrysts. The nephelines are rimmed by aegirine and aenigmatite and altered to cancrinite. Augite and aegirine-augite rarely form phenocrysts. The groundmasses are mainly composed of subparallel sanidine laths with interstitial soda pyroxene and soda amphibole (aegirine, aegirine-augite or aenigmatite). Kataphorite, arfvedsonite, riebeckite, hydrobiotite and magnetite may be present in the groundmass. No quartz was seen, rather the nepheline microphenocrysts constitute up to 10 per cent of the mode.

McClenaghan (1971) presented ten whole rock analyses which plot as trachytes (7), benmoreites (2) and phonolites (1) on an alkali against silica plot. Webb (1971) presented nine whole rock analyses of trachyte, none of which have normative nepheline. He also provided two basanites, one alkali basalt and one benmoreite whole rock analysis.

Nathelot basalts ($Tpvt^{NT}$), mostly from Truckle, 1977a

Initially referred to as 'Basalts of the Lopirapira area' by Rhemtulla (1970), these were later renamed the Nathelot basalts by Truckle (1977a). According to Rhemtulla (1970), Nathelot is the local name given to a major (feeder) dyke amongst a dyke swarm exposed by deep erosion around GR 1855 1900. The account of Truckle (1977a), if published, would serve to formalise the Nathelot basalts.

The Nathelot basalts underlie a large central part of the NW quadrant of Degree Square 27. Around their source area, Nathelot, and northward for 10 km the scenery is impressively hilly but flattens out in the far north to form slightly elevated plateaus isolated by major accumulations of unconsolidated superficial debris. Consequently exposure is better in the south.

Feeder dykes (of at least two major suites) are well exposed at Nathelot. The hills to the north of Nathelot provide good outcrop of the various basaltic lavas: Akichar (GR 1820 1990), Kawalamoro (GR 1890 1950). In the far north the scarp slopes of Kaliba and Nakathuen provide good sections as does the scarp along the east bank of the Kerio River.

Three samples submitted for dating gave ages from about 4.5 to 8.1 Ma (Truckle, 1977a). The older ages are unreliable and the youngest age (taken from a basal lava) is quoted by Truckle (1979b). An upper limit is provided by dates of about 3.5 to 3.0 Ma for overlying trachytes.

The basalts are in contact with many of the lithological units of the NW quadrant; as a rule in the west they unconformably overlie older rocks (Kailimerlim and Nasaken Volcanoes, Lokitoe Centre, Lomujal and Tirioko basalts) and in the east they form the foundation for younger units (Lorikipi and Kamuge basalts, Lomi, Kanatim and Kapatagni Volcanoes, Quaternary sediments). Numerous trachyte dykes and arcuate phonolite dykes as well as basaltic feeder dykes intrude the Nathelot basalts in their main southern outcrop area. They are also cut by two basalt cinder cones (just west of Nathelot) and many NNE curvilinear faults. The whole unit dips very gently towards the E or ESE.

The Nathelot basalts are over 450 m thick in their source area and wedge out rapidly westwards, but less so northwards (to about 45 m beneath Kaliba). Their eastern thickness is not known.

Truckle (1977a) identified five members in the Nathelot area: a major basal unit of aphyric basalts and hawaiites comprises over 80 per cent of the succession. Overlying members contain various thin basaltic and hawaiite flows. A series of eroded domes is recognised in the basal unit, with variable trends to long axes. Feeder dyke swarms show a wide spread of orientations and do not necessarily parallel the elongation direction of the domes. Less than 10 m of bedded arenites are intercalated in the basal member, which also contains a flow rich in harzburgite nodules.

Eight flow units were identified on Kaliba Hill, which is the best northern exposure, where they overlie a very uneven surface of Kailimerlim trachytes. A lithic welded tuff is intercalated within the lavas, which are capped by a red soil and grit horizon. Isolated outliers west of the main outcrop area form steep-sided hills mostly capped by aphyric or poorly feldsparphyric, fine-grained basalts. These flows dip gently eastwards.

No whole rock analyses are available for the Nathelot basalts.

Truckle (1977a) concluded that the constituent members of this unit were either sourced near, or north-east of Nathelot to form a large low-angle shield volcano. The thickness of the volcano was in part controlled by an uneven older topography and by subsidence in the source area during lava eruption. Subsequent faulting and tilting prior to late trachytic volcanism also lowered the lava pile in the Nathelot area, as well as imparting an overall easterly tilt. Steeper easternmost dips are due to a gentle monocline formed by this deformation.

Kanatim Volcano ($Tpvt^{KN}$)

Kanatim Volcano, mapped but not fully documented by Rhemtulla (1970), continues the Late Mio-Pliocene north-north-east migration of central trachyte volcanoes to overlap the NNE flanks of Nasaken and Kachila Volcanoes. Kanatim is the local name for its summit area. The main part of Kanatim Volcano lies immediately east of the central zone of Nasaken Volcano. It is about 15 km long in a NNE direction with a major ash deposit extending a further 15 km N from its north-western extremity. The eastern part of Kanatim Volcano is mantled by younger trachytic volcanoes, so its original full extent is unknown.

A radial drainage off the summit area of the main volcano strongly dissects its flanks to form a series of low mesas. The northern ash is readily weathered to a series of low mounds.

Reference sections have not been defined. An elongated central plug to the main volcano is exposed at GR 1885 1680; the main summit is at GR 1863 1642. The Katatarope River cuts across the northern ash deposits. Weaver (1973) records an age of 4.8 ±0.2 Ma for the basal welded tuff and an age of 3.8 ± 0.2 Ma for an upper trachyte lava.

It overlies a strongly eroded surface of Nasaken Volcano (5.7 to 5.4 Ma), and in the north also overlies the Nathelot basalts (dated at about 4.6 Ma). Minor Quaternary sediments locally mantle its eastern parts not already covered by Kapatagni (about 3.3 Ma) and Oliyamur Volcanoes. Numerous trachyte dykes cut the volcano's summit area which is also traversed by NNE faults. The contact with Oliyamur is, in part, a major NNE fault downthrowing eastwards.

Less than 300 m of trachytic volcanics underlie the summit area and the northern tuffs are considerably thinner. The basal welded tuff is only 6 m thick where it is exposed on the eastern flank of Nasaken Volcano.

Typical pale greenish, fissile trachyte lavas underlie much of Kanatim Volcano. Sanidine-phyric and aphyric lavas are recognised. Brown, non-welded pumice tuffs, as well as forming the extensive northern air-fall deposit also mantle the central plug of the main volcano. Welded tuffs underlie the northern pumice tuffs and expose as green, flaggy eutaxitic ash flows.

In thin section some anorthoclase, albite, aegirine and aegirine-augite phenocrysts may accompany the sanidine phenocrysts. Rhemtulla (1970) also notes the occurrence of kataphorite, arfvedsonite and aenigmatite in the trachyte lavas.

Lokwaleibit basalts ($Tpvb^{LK}$), from McClenaghan, 1971

This formation just extends into the SW corner of Degree Square 27 in a series of N–S fault strips where it overlies Ribkwo Volcano and, in turn, is overlain (eastwards) by the Jamakana formation. The whole unit was originally mapped by McClenaghan (1971) with a type section in the gorge cut by the Lokwaleibit River west of Degree Square 27 (at GR 1330 1114).

Three basalt members were identified by McClenaghan (1971), locally separated by thin beds of light-coloured sediments which form marker beds in the field and on aerial photographs. He also noted that one (porphyritic) basalt member, weathered more easily than the other basalts, and appeared lighter on aerial photographs. Gastropods and plant impressions have been recovered from limestones and tuffaceous sediments.

No isotopic age determinations have been undertaken on these basalts although McClenaghan (1971) equates them with that part of the Ribkwo Volcano dated at about 4.4 Ma. He attributes the N–S fault set cutting the Lokwaleibit basalts as a response to gentle arching on a N–S axis. The faults produced a series of horsts and grabens: in Degree Square 27 a major eastern graben is exposed.

The type succession is about 120 m thick and a similar thickness is present in Degree Square 27. Two basal porpyritic basalt flows are each about 15 m thick with large olivine, labradorite and augite phenocrysts.

They are overlain by from 2 to 6 m of tuffaceous sediments with or without a gastropod-bearing limestone which oversteps northwards on to the Ribkwo Complex. Up to eight aphyric basalt flows, each about 13 m thick, with intercalated tuffaceous sediments overlie (and also thin and overstep northwards) the tuffaceous sediment-limestone unit. Locally, in the east, a fine-grained, grey lapilli tuff may cap the aphyric basalts. A NE-trending feeder dyke is exposed at GR 1685 1206.

The lavas are dark grey to black, weathered to purple-grey with vesicular or scoriaceous tops. Labradorite (two generations) is the commonest and largest phenocryst phase, up to 0.5 cm in length although with almost complete (seriate) gradation into groundmass feldspar. Olivine phenocrysts are variably altered to serpentine and iddingsite, whereas the less common augite phenocrysts are pale brown and fresh. Anhedral grains of magnetite form microphenocrysts.

Lava groundmasses are largely composed of labradorite

(locally shape-orientated) with anhedral augite grains, rare (or absent) olivine grains, ubiquitous magnetite octahedra and fragmented grains, and accessory biotite flakes.

Rare hawaiites have sparse microphenocrysts of andesine, augite and altered olivine in a fine-grained matrix mostly of orientated andesine laths with augite, magnetite and biotite.

Five whole rock analyses given by McClenaghan (1971) plot as alkali basalts (4) or hawaiite on an alkali and silica plot (Cox and others, 1979). The Lokwaleibit basalts had a southern source area. Volcanism was not continuous: the sedimentary intercalations were deposited in ponded water. Red soil horizons are also common, separating individual flow units.

Kalokopon Volcano ($Tpvt^{KL}$), from Truckle, 1977a

Kalokopon Volcano, named by Truckle (1977a) after the Kalokopon Plateau, is the northernmost unit of the NNE-trending chain of trachytic volcanoes which forms such an important part of the west shoulder. Only the southern part of this volcano extends into Degree Square 27; a caldera was identified futher north. Kalokopon Volcano is largely concealed beneath flood lavas of the Lorikipi basalts, unconsolidated Quaternary sediments, and Namarunu lavas.

Truckle (1977a) records a single age determination of 4.4 ± 0.2 Ma for a trachyte lava from Kalokopon. Outcrop is poor, confined mostly to the area west of Namarunu Volcano. Kalokopon Volcano, like the contemporaneous Lomi and Kapatagni Volcanoes, is founded on the Nathelot basalts. The section of Truckle (1979b) indicates a maximum vertical thickness of about 200 m, gradually tapering westwards over a horizontal distance of about 15 km. However Truckle (1977a) indicates a thinner succession with a basal feldsparphyic trachyte flow (21 m thick) overlain by 18 m of welded tuffs in turn capped by more trachytic flows. The whole sequence is tilted and downwarped to the east and north-east.

Lomi Volcano ($Tpvt^{LM}$), from Truckle, 1977a

This was mapped by Truckle (1977a) and named after its hilly summit area at the south end of the Loriu Plateau. Lomi Hill has a peak height of 993 m (3225 ft), which is about 300 m above the western perimeter of the volcano. These western parts are sporadically covered by thin Quaternary sediments. However the northern and southern flanks are wholly covered by the Lorikipi and Kamuge basalts. These lavas clearly overstepped on to the volcano and Truckle (1979b) shows the summit caldera identified by Rhemtulla (1970) to be infilled by the Lorikipi basalts. The eastern margin is defined by the major Western Suguta Fault Swarm. It is not known whether Lomi Volcano extends east of this fault system into the median graben. Rhemtulla (1970) noted that the fault system was active during the eruptive build up of Lomi Volcano. The main outcrop area of the volcano covers an area of about 224 km^2; its original size may have encompassed about 900 km^2. Major river incision of the summit terrain is controlled by NNE fault traces.

A reliable isotopic age of about 3.0 Ma is quoted by Truckle (1977a) for a trachyte flow from Lomi Volcano. It rests on the Nathelot basalts (locally, in the north-west, overstepping on to Kailimerlim Volcano) and is overlain by flood basalts, and Quaternary sediments. It is cut by NNE faults and numerous trachyte and basalt dykes and stocks, notably in its summit area (including NE-, WNW- and WSW-trending trachyte dyke swarms). Many of the trachyte intrusives represent feeders to the feldspar trachyte lavas. The lavas had quaquaversal dips away from the summit caldera which have been modified by eastward tilting of the eastern part of the volcano by 10–15°.

The eastern and western flanks of Lomi Volcano are composed of several trachytic flows up to 20 m thick with subservient intercalations of ignimbrite and impersistent yellow tuff lenses. Total thicknesses of about 45 m are recorded by Truckle (1977a) for western flank successions.

A summit caldera is well defined by steep west and north-western walls but its eastern and southern limits have been obscured by a combination of erosion, tilting and local complexitites in rock distribution. However the caldera probably covered an area of about 4 to 6 km^2 now blanketed by three horizontal aphyric and ankaramitic basalt flows (of the Lorikipi basalts). In this area there are complex relationships between feldsparphyric trachyte flows and assorted pyroclastic wedges.

A cyclicity to the pyroclastics was noted as follows:

3 Ignimbrites
2 Poorly welded pumice tuffs
1 Agglomeratic pumice tuffs

The coarser units are polymict with syenite and basalt clasts besides feldsparphyric trachyte clasts. Sedimentary breccias are related to landslipping following movement in the caldera fault zone. Good evidence for this zone is given by: drag of flow units, breccias, secondary silicification, veining and reddening of lavas. Several periods of movement along the fault zone are indicated by increased dips to older flows relative to younger flows.

Kapatagni Volcano ($Tpvt^{KP}$)

Kapatagni, unlike the adjacent trachyte volcanoes, is unique in underlying a flat featureless (and therefore distinctive) plateau, after which it is named, in the NW quadrant of Degree Square 27. Well-defined scarps bound the northern and eastern limits of the plateau, which is no more than about 90 m above adjacent low ground. To the south-west it overlaps on to the flanks of Kanatim Volcano.

The eastern scarp is controlled by the Western Suguta Fault Swarm (Mugor fault of Rhemtulla, 1970) with a graben to the east preserving Kamuge basalts. A small faulted inlier of Kapatagni Volcano reappears east of this graben before being obscured by the western side of Emuruangogolak Volcano. NNE-trending faults and master joints locally influence a poor drainage system on Kapatagni Plateau.

The scarp slopes afford good outcrop of the two main trachyte lava units (see below). Tuffaceous units are well exposed in the SW corner and on Lopugut (GR 1910 1875) in the NE corner. Kapatagni Volcano was initially surveyed by Rhemtulla (1970) and remapped by Truckle (1977a). The legend of Truckle (1979b) shows an age of 3.3 Ma for the contact between the two trachyte lava units.

In the west the basal contact of Kapatagni Volcano is exposed with either Kanatim Volcano or the Nathelot basalts. In the east the upper contact with the Kamuge and Lorikipi basalts or Emuruangogolak Volcano is exposed. Major NNE faults are confined to the eastern half of Kapatagni.

Rhemtulla (1970) suggests a total thickness of about

86 m, to include 60 m of trachyte lavas and a maximum thickness of 26 m for welded tuffs or bedded grits and reworked tuffs (possibly thicker in the south-west). The pyroclastics tend to form a wedge between two major trachyte units. However a crescent-shaped pyroclastic unit mapped by Truckle (1977a) near Lopugut in the north-east, marks the centre from which the basal phyric trachyte unit was erupted. Up to 36 m of intercalated pyroclastics, trachytic and basaltic lavas, and sediments are poorly exposed. Up to 40 per cent of this sequence is made of pyroclastic cycles fining upwards from polymict agglomerates into crystal and pumice tuffs. Radial dips, modified by eastward tilting and subsidence, define a cone structure. Trachyte dykes trending ENE and N cut the northern succession.

On the main plateau the lower trachyte unit consists of feldsparphyric lavas which are only exposed on the lower scarp slopes or around the present margin of the volcano. They are overlain by aphyric lavas which are extensively exposed on the plateau itself.

Kajyamamuk Volcano ($Tpvt^{KJ}$)

The summit of a trachytic volcano underlies the NNE-trending Kayjamamuk Ridge on the south-east side of Oliyamur Volcano. The two volcanoes are separated by a NNE-trending valley although locally Oliyamur can be demonstrated to overlie Kajyamamuk Volcano. Kajyamamuk Ridge is only 2.5 km NW of Lomelo. It has a maximum height of 898 m to rise about 300 m above the Suguta River which passes close to its northern end.

The volcano is in the area mapped by Rhemtulla: this mapping has not been fully documented but the detail of Kajyamamuk Volcano shown on Truckle (1979a) is based on this unpublished work. A traverse up the middle of the central ridge commencing at the end of the track north-west from Lomelo passes through the three main lithological units shown on Truckle (1979a), and could serve as a reference section.

No isotopic age determinations have been undertaken on Kajyamamuk Volcano, which according to Truckle (1979a) was contemporaneous with Kanatim Volcano. These two, early to middle Pliocene complexes are geographically separated by the younger Oliyamur Volcano. Basalt lava flows from Emuruangogolak Volcano and Suguta Valley alluvium, mantle the eastern parts of Kajyamamuk Volcano. Its NNE elongation together with a parallel trachyte (feeder) dyke swarm indicate a strong tectonic control: the volcano in fact straddles the western boundary of the Suguta Trough which is invariably fault-controlled. No fault is shown in this area by Truckle (1979a) although the pronounced NNE linearity of the valley separating Oliyamur and Kajyamamuk Volcanoes strongly suggests an underlying fault.

The cross section of Truckle (1979a) shows an unknown thickness of basal trachyte lavas under Kajyamamuk Ridge. The upper mugearites are about 100 m thick and the summit pyroclastics form a wedge thickening to the SSW. Both the mugearites and pyroclastics (pumice tuffs) are confined to the SSW half of the outcrop area where they overlie the trachyte lavas.

Jamakana formation ($Tpvt^{JM}$), from McClenaghan, 1971

This formation just extends into the SW corner of Degree Square 27, where it is exposed on the SE flank of Ribkwo Volcano in the area mapped by McClenaghan (1971). It underlies a hilly landscape with prominent mesas—a reflection of its varied lithological content.

McClenaghan (1971) identified seven members and describes suitable outcrops to serve as type areas. The lower six members are exposed in Degree Square 27 on and around Karmosit Hill (GR 1685 1138). A conglomerate exposed on Karmosit Hill and a nearby butte contains vertebrate fossils, including remains of elephant, rhinoceros, pig, hippopotamus, giraffe and bovides. An isotopic age of 3.4 ± 0.2 Ma has been obtained from a lava within the Jamakana formation. In Degree Square 27 it is faulted against the older Lokwaleibit basalts in the west and is overlain unconformably to the east by the Murgisian basalts, Silali pyroclastics and Suguta Valley alluvium.

McClenaghan (1971) suggests a thickness of about 113 m for the main part (members 3 to 6 inclusive) of the Jamakana formation and notes rapid lateral thickness variations for individual units. He identified the following stratigraphy:

7 Cheptonowa trachytes (south of Degree Square 27)
6 Fine-grained, green welded tuff and trachytic breccias (5 m)
5 Coarse yellow pumice and lapilli tuff (34 m)
4 Mugearite (13 m)
3 Coarse yellow and white pumice tuff with upper conglomerate (61 m)
2 Mugearite
1 Subaerial poorly bedded pale brown and yellow lapilli tuff

There are no intrusions associated with the formation. The higher tuffs are thicker south of Degree Square 27 where a possible source is identified.

The mugearites are fine-grained, purple-grey with moderate fissility. They are sparsely porphyritic and frequently vesicular. The phenocrysts are anorthoclase, rare anhedral augite, olivine pseudomorphed by chlorite and iddingsite, and sparse kataphorite microphenocrysts. Partly aligned oligoclase laths predominate in the groundmass with interstitial mafic grains (magnetite and augite with accessory biotite, kataphorite and pseudomorphed olivine). McClenaghan (1971) provides two whole rock analyses of mugearite lavas (Table 5) which in fact plot as benmoreites on an alkali against silica diagram (rather high total alkalis for mugearites).

Oliyamur Volcano ($Tpvt^{OL}$)

Oliyamur Volcano is situated within the NNE-trending fault complex separating the median graben and western shoulder to the east of Kanatim and Kachila Volcanoes. Consequently it has a rectangular plan with a c.15 km NNE length and c.4 km width. It was mapped and named by Rhemtulla in the early 1970s. The volcano rises to a central peak (Lomelo Hill) with an elevation of 1102 m (3579 ft) compared to perimeter altitudes of about 600 m. Due to its relative youth it is not as deeply incised as adjacent trachyte volcanoes. Shallow valleys follow NNE fault traces or follow the surface gradient from the central peak.

The three major lithostratigraphic units (trachyte lavas, pyroclastics, and dykes) are all exposed on Lomelo Hill which could therefore serve as a type area. Weaver (1973) records isotopic ages of 2.7 ± 0.2 and 2.4 ± 0.1 Ma for Oliyamur Volcano. It overlies (in the west) Kanatim and Kachila Volcanoes and (in the west) Kajyamamuk Volcano. In the NNE it is partly obscured by basalt flows from Emuruangogolak Volcano and is locally covered by alluvium. Pyroclastic rocks are confined to the SSW half

of Oliyamur with a major centre under Lomelo Hill. A NNE (feeder) trachyte dyke swarm cuts through the hill.

The topography suggests about 500 m of trachytic rock under Lomelo Hill, which according to the section of Truckle (1979a) is dominated by lavas.

Murgisian basalts ($Qptb^{MR}$)

These strongly eroded basalts were mapped and named (after Murgisian Hill at GR 170 122) by McClenaghan (1971) in the SW corner of Degree Square 27. They form two main low (about 100 m) dome-shaped hills, Murgisian and Namatono: several erosional remnants crop out further to the SE. The hills are covered by rounded basalt boulders with a thick scrub cover. Murgisian Hill is the type area.

The basalts have not been isotopically dated but unconformably overlie the Jamakana formation (c.3.4 Ma). Individual lavas flowed around the flanks of Ribkwo Volcano and down valleys cut into the Jamakana formation. They dip at about 6° to the SE, an original slope. NNE-trending feeder dykes are recognised on the west side of Murgisian Hill. A mantle of Silali tuffs covers the eastern parts of the Murgisian basalts which are also covered by alluvium of the Suguta River.

In the type area the Murgisian basalts have a thickness of 86 m. The lavas are susceptible to weathering, so this figure probably does not represent an original thickness despite their relative youth.

The lavas are dark grey, porphyritic and highly vesicular basalts with olivine and feldspar phenocrysts. Both phenocryst phases are cumulophyric and up to 3 mm in length. Olivine grains have marginal iddingsite alteration. Augite occurs as sparse phenocrysts, but is a major groundmass constituent together with labradorite and magnetite. Calcite amygdales are recorded by McClenaghan (1971). The basalts have not been chemically analysed.

Summary

An incomplete record of volcanicity on the western shoulder is obtained by conventional mapping because of the extensive development of Pliocene units which effectively conceal their bedrock. No major vertical sections are preserved and it is possible that the apparent fundamental differences in the geology of the eastern and western shoulders may be somewhat misleading. It must also be stressed that basaltic units weather far more easily than the trachytic units (compare the present morphology of the Murgisian basalts with morphologies of any Pliocene trachyte volcano). Therefore there may have been considerably more basaltic material erupted on the western shoulder than is suggested by the preserved stratigraphy (Table 7).

Volcanism commenced (in Degree Square 27) at about 16.6 Ma with the Kapchererat basalts from a westerly source. Basalts were subsequently regularly erupted until the Quaternary (Kamuge and Lorikipi basalts, described in Section 4.3). Generally the basalts were laterally restricted and erupted over short periods of time relative to the eastern shoulder. Phonolitic eruptions were of minor significance on Degree Square 27 and essentially confined to the c.11.6 Ma Kasorogol phonolites. However trachytes were impressively developed, initially with the c.15.1 Ma Kowun Volcano and, later the almost continuous Late Miocene through Pliocene series of central volcanoes.

These volcanoes migrated NNE with time along the inner part of the western shoulder. Their unique morphology has been commented on by Webb and Weaver (1975). In contrast the extrusion of basalts appears to have migrated eastwards towards the median graben with time although this trend is not as obvious as the trachytic migration.

The stratigraphy can be divided into three major stages on the basis of the main locus of volcanism and on the lithologies. Initially during the period from about 16.6 to c.11.7 Ma, volcanism appears to have concentrated to the west of Degree Square 27, with source areas on the western side of the west shoulder. Volcanism was compositionally varied with major basalt, phonolite and trachyte extrusions. The basal Kapchererat basalts were contemporaneous with the Samburu basalts to the east. Subsequent contemporaneous units on the west and east shoulders are generally lithologically disparate. During the (middle) period from about 11.7 to c.6.6 Ma the volcanism included, and must have been dominated by, the two major basaltic units of Lomujal and Tirioko. Despite the susceptibility of the basalts to erosion, both these units still retain thick successions of basaltic rock.

Finally, from c.5.9 Ma to the present the volcanicity has been dominated by the central trachytic volcanoes accompanied in the north by major basalt eruptions. A case could be made for grouping all the trachytes into a single unit (cf. the Samburu basalts composed of small coalescing shields).

It would be impossible to ascertain exactly the vertical thickness of volcanics and related sediments within the main part of the western shoulder, without drilling a series of vertical cored boreholes. Most units display pronounced lateral thickness variations so it is not known how far any one unit continues beneath an overlapping upper unit.

The bimodality of the volcanism clearly shown by the stratigraphy has been commented on by, amongst others, Weaver and others (1972) and McCall and Hornung (1972). Basaltic and trachytic rocks dominate the extrusive volcanic phases although the whole rock analyses do show that the intermediate lithologies are present in *minor* amounts. Both Weaver and others (1972) and McCall and Hornung (1972) demonstrate that all effusive phases could have originated from a mantle-derived basalt magma. A cupola of a salic fraction above, but possibly in contact with the basalt source, provided the trachyte surface material. As these authors note, the salic cupola must have been of quite considerable size, which by inference requires an even larger basaltic magma chamber, also posited by the various geophysical investigations (e.g. Baker and Wohlenberg, 1971; Maguire and Khan, 1980).

Relatively deep fractures are required to tap the deeper basaltic magma chamber, possibly explaining the parallel migration of basaltic lavas and major faults towards the median graben with time. By contrast the salic magma chamber is nearer surface, and, in part, its eruptive phases are controlled by different tectonic processes (cf. Weaver and others, 1972). However Oliyamur and Kajyamamuk Volcanoes were probably controlled by the deep faults marginal to the Suguta Trough, perhaps explaining why they do not follow the usual NNE migration with time of the other trachytic centres. A possible control on this NNE migration could be progressive tensional opening on a slightly arcuate fracture system with a dextral strikeslip component (Hackman, pers. comm). The large scale structure of the western shoulder in Degree Square 27 indicates arcuate fractures (see Section 8.2).

Table 7 Summary of the volcanic activity forming the western shoulder of the Gregory Rift in Degree Square 27

Stratigraphic unit	Ages (approximate) Ma	Maximum thickness	Shape	Classification of King and Chapman (1972)	Lithology
Kamuge basalts		150	plateau	5	basalt
Lorikipi basalts		250	wedge	5	basalt
Murgisian basalts	Pleistocene	86	plateau	5	olivine basalt
Oliyamur Volcano	2.7–2.4	500	elongate shield	4	trachyte
Jamakana formation	3.4	113	wedge	4	trachyte, mugearite
Kajyamamuk Volcano		300	elongate shield	4	trachyte, mugearite
Lokwaleibit basalts		120	wedge	4	alkali basalt, hawaiite
Kapatagni Volcano	3.3	86	shield	4	feldspar trachyte
Lomi Volcano	3.0	300	shield	4	trachyte
Kalokopon Volcano	4.4	200	shield	4	feldspar trachyte
Kanatim Volcano	4.8–3.8	300	shield	4	feldspar trachyte
Nathelot basalts	4.5 Pliocene	600	wedge	3	basalt
Ribkwo Volcano	5.2–4.4	760	shield	2	feldspar trachyte, phonolite (minor)
Nasaken Volcano	5.7, 5.4	400	shield	2	trachyte
Kachila Volcano	5.7, 5.4?	275	shield	2	trachyte, mugearite, basalt
Tumungir basalts		<20	minor plateau	2	alkali olivine basalt
Kafkandal Volcano	(5.9 base, 5.7)	610	shield	2	quartz trachyte
Tirioko basalts	6.9, 6.6	300	coalescing shields	2	basalt, hawaiite
Tiati Group		335	easternmost tip	2	trachyphonolite
Kailimerlim Volcano		200	shield	2	phonolite, trachyphonolite, feldspar trachyte (major)
Lokitoe Centre		<10	minor centre	2	feldspar trachyte
Lomujal basalts		600	wedge	2	basalt, mugearite
Kasorogol phonolites	11.6	200	eastern extremity	2	phonolite
Lokwanamur basalts	13.7	200	eastern extremity	2	basalt, hawaiite
Kowun Volcano	15.2, 15.0	300	shield	2	quartz trachyte
Kapchererat basalts	16.6–base, 15.7 Miocene	160	wedge (considerably thicker to west of Degree Square 27)	1	mugearite, basalt

4.3 MEDIAN GRABEN OF THE GREGORY RIFT

This section, like the preceding section, is based entirely on the EAGRU mapping.

Slightly undersaturated Quaternary basaltic and trachytic volcanism alternate in four low angle shield volcanoes (Figure 5). Locally flows from adjacent volcanoes overlap but mostly the volcanoes are separated by superficial sediments of the Rift floor. Silali and Emuruangogolak Volcanoes are completely represented. The north edge of Paka and south edge of Namarunu straddle the south and north map margins respectively. All, except Namarunu, have summit calderas (variably filled and or breached by the uppermost flows). Apart from the youngest flows which form narrow tongues emanating from well-defined vents, both trachytes and basalts comprise extensive and uniform lava flows. Units mapped along the western margin of the Suguta Trough by Truckle (1977a) can be related to the shield volcanoes. The Lorikipi basalts may represent an early phase of Namarunu; the Kamuge trachytes and basalts, early phases of Silali.

Total thickness of extrusives on Silali approaches 1000 m and, on Emuruangogolak 700 m. Pyroclastics, mainly pumice tuffs, are commonest around summit calderas but also form persistent interlayers in the lava piles. Basaltic and trachytic parasitic cones are ubiquitous and controlled by NNE-trending faults; parasitic craters are less common and dykes are rare. Minor sedimentary intercalations are confined to fluvio-lacustrine deposits, including diatomites.

The western and eastern extremities of the volcanoes overlap on to older volcanics of the Rift shoulders: to what extent the older rocks continue under the median Rift is not known. However, the Plio-Pleistocene Emuruagiring trachytes straddle the eastern edge of the median Rift and dip beneath Emuruangogolak Volcano.

Individual accounts of the major shield volcanoes all conclude that the constituent basaltic and trachytic extrusives are genetically related. Fractionation of a source basaltic magma initially produced a zoned chamber of batholithic dimensions with basalt overlain by trachyte. Both parts were subsequently tapped by intersecting deep fractures with further mineral fractionation during ascent through the crust. Therefore even the basaltic effusives are relatively fractionated.

Lorikipi basalts ($Qpvb^{LR}$), from Truckle, 1977a

The Lorikipi basalts, first named in Truckle (1977a) form a narrow (up to 2 km wide) outcrop strip along the western edge of the northern Suguta Trough. Upper flows overspilled westwards to cover large parts of the Loriu Plateau capping the Kalokopon, Riaten and Kaliba Plateaus as well as partially infilling Lomi Caldera. The plateau lavas are essentially flat-lying but are tilted eastwards near the edge of the Suguta Trough. The Lorikipi basalts have dark phototones.

Truckle (1977a) records good sections at Kamuge, east of Lomi Volcano, and near Kalokopon which could serve as reference areas. The age of the Lorikipi basalts is bracketed between c.3.3 Ma and c.0.8 Ma: the ages of the underlying Lomi trachytes and overlying Kamuge trachytes respectively. An isotopic age of 4.0 ± 0.1 Ma from a basal lava overlying Lomi trachytes is clearly spurious. The Lorikipi basalts also unconformably overlie parts of Kapatagni and Kalokopon Volcanoes and are also unconformably overlain by the Kamuge basalts and flows from Namarunu as well as unconsolidated sediments.

Lavas along the western interface of the Suguta Trough are strongly faulted (NNE-trends) and tilted by up to 25° eastwards. The Lorikipi basalts infilled surface irregularities of the Loriu Plateau with flows overstepping on to higher parts of the plateau, e.g. the flanks of Lomi. As a result there are rapid lateral thickness variations for the plateau basalts to a maximum thickness of about 180 m east of Kalokopon Hill. At Kamuge and east of Lomi the full succession is not exposed: only the upper 100 to 120 m of the Lorikipi basalts are seen.

Truckle (1977a) notes that the Lorikipi basalts, unlike older basaltic units, are unusual in lacking any evolved lavas. Basal units, partially exposed at Kamuge, are mafic phyric basalts and are overlain by aphyric basalts. In turn, on the Loriu Plateau, the aphyric lavas are capped by poorly olivine- and pyroxene-phyric basalts. An eastern source is indicated by the lateral thickness variations and stratigraphy. Thin interbeds of tuffaceous sediments together with basaltic scoria are exposed in the Kamuge area. Rare agglutinate cones, locally with pyroxenite nodules are exposed east of Lomi. In this same area are intercalated volcaniclastic sediments derived from weathering of Lomi Volcano.

No chemical data are available for the Lorikipi basalts which mark the start of a Quaternary cycle of basaltic/trachytic volcanism within the Suguta Trough.

Kamuge trachytes ($Qpvt^{KM}$), from Truckle, 1977a

Fault-controlled outcrops of the Kamuge trachytes are confined to the small area north of the waterhole in the Kamuge River centred on GR 197 189. They were named by Truckle (1977a) and cap low hills above a faulted plateau underlain by the Lorikipi basalts. The hills define the reference sections for this geographically restricted unit. A sample of a basal ignimbrite has been dated at 0.8 ± 0.1 Ma.

At the south end of the Emoruamus Plateau, the Kamuge trachytes are sandwiched between the underlying Lorikipi basalts and overlying Kamuge basalts.

The succession dips at varying angles eastwards with maximum dips of 30° adjacent to the Suguta Trough. Concomitant with increasing dip is an increase in the thickness of the Kamuge trachytes: the Kamuge River sequence is about 15 m thick whilst Truckle (1979b) shows a maximum thickness of about 100 m.

A strongly welded, yellow/brown ignimbrite and a poorly welded pumice-obsidian tuff may form the base of the Kamuge trachytes. They are overlain by glassy or feldspar trachyte lava.

Tilting and erosion of the Kamuge trachytes preceded eruption of the Kamuge basalts. The trachytic sequence may have erupted from Emuruangogolak Volcano.

Kamuge basalts ($Qpvb^{KM}$), from Truckle, 1977a

This unit was named by Truckle (1977a) after the Kamuge River which bisects the outcrop area to the north-west of Silali. The basalts underlie the Emoruamus Plateau and a more southerly, fault-dissected plateau—a total outcrop (N–S) length of about 30 km and (E–W) width of up to 5 km. The lavas are flat-lying or gently tilted eastwards and have dark monotonous phototones. It is the youngest volcanic unit on the Loriu Plateau and predates the latest development of the Suguta Trough.

The scarps surrounding the Emoruamus Plateau provide good exposures of the basalts. Along the northern scarp, in the Amaries area, the basal contact with underlying (Amaries) sediments is exposed. The main outcrop area is between the Kamuge River and Lomi Volcano: basalt boulders mantle exposures.

The Kamuge basalts, in the south, overlie the Kamuge trachytes, and are therefore less than 0.8 Ma in age. A single sample selected for dating had no measurable radiogenic argon confirming its youth. Locally they also unconformably overlie with slight angular discordance Lomi and Kapatagni Volcanoes and the Lorikipi and Nathelot basalts. The Amaries sediments are locally wedged between the upper flows of Lomi Volcano and the Kamuge basalts. The NW lavas of Silali mantle the southernmost parts of the Kamuge basalts. The unit is strongly faulted and locally (GR 1953 1892) cut by basalt dykes.

The Kamuge basalts have a total thickness of less than 100 m comprising several lavas up to 10 m thick. Locally a wedge of brown basaltic tuff underlies a lower aphyric basalt capped by olivine-phyric basalt. Plugs and dykes protrude above the present basalt surface indicating eroded sources. The basalt lavas on the southern flanks of Lomi Volcano are elevated by about 266 m relative to the main plateau outcrop area to suggest a source on Lomi. The gentle eastward tilting of the unit is due to monoclinal warping along the margin of the Suguta Trough.

Western scarp basalts and isolated basalt flows (Qrvb), from Truckle, 1977a

Truckle (1977a) mapped a single flow of aphyric basalt unconformably above the Lorikipi basalts. The flow crops out over a length of about 18 km and is up to 12 m thick. It is locally altered adjacent to syenite veins. Three flows of moderately phyric basalts may cap the aphyric basalt: all dip gently eastwards. Fresh looking basalt lava overlies the lacustrine sediments of the 'Suguta Beds' in the northern part of the Suguta Trough. The lava can be traced

back to a cinder cone source in a similar manner to the youngest flows on Namarunu. It is possible that these lavas are only a few hundred years old. Representative analyses of the five major element whole rock analyses of these basalts by Truckle (1977a) are shown in Table 8.

Paka Volcano ($Qpvt^P$, $Qpvb^P$, $Qrvt^P$, $Qrvb^P$)

Paka is the Pokot name for the volcanic mountain centred on 36°10′E and 0°50′N, i.e. immediately south of Degree Square 27. The northernmost extremities of a representative, but attenuated stratigraphic sequence of trachytic and basaltic lavas just extend into the map-area. Tuffaceous horizons are extremely rare, being confined to the main summit area. Basal mugearites (and a parasitic cone) are exposed at Sekur; uppermost basalt flows reach Gulungul; an older basalt, and flows related to the two main trachytic episodes underlie Chemogogoch. The geology of the whole volcano is described by Sceal (1974).

Outlines of individual basalt and trachyte flows, cinder cones, and fault scarps are well-defined on aerial photographs where flow fronts and internal surface flow features, especially pressure ridges, stand out clearly. Trachytic flows are short, and can often be completely delineated on single 1:40 000 aerial photographs.

The whole volcano has a sparse vegetation, so that outcrop is uniformly excellent. Therefore the locations listed in the opening paragraph serve as reference sections for their underlying bedrock. Paka is a Quaternary volcano with extrusive activity continuing into historical times. Steam jets are visible on the flanks on the main mountain.

Within Degree Square 27 the eastern side of Paka is in fault contact with the Secumius trachytes, treated as part of Paka Volcano by Hackman (in press). The northern perimeter of Paka is surrounded by alluvium. The thickest flow unit underlies Gulungul where a borehole reached a depth of about 100 m and was still in the same basalt. Individual trachyte flows are thinner.

All the basalts are black to blue-grey, usually vesicular with quite coarse grain sizes (individual grains are just visible to the naked eye). *Aa* and *pahoehoe* textures, collapsed lava tunnels, hornitos, trachytic glass rafts and cinder cones sources are described by Sceal (1974).

The trachytes are dark green to black, aphanitic, with streaky textures and usually aphyric, although alkali feldspar phenocrysts locally constitute up to 30 per cent of the rock volume. They may weather to bright red soils. Syenite bombs are recorded in source cones by Sceal (1974). Pumice tuffs are rare in Degree Square 27.

Plagioclase, olivine and augite phenocrysts were noted in the basalts; they rarely exceed 0.5 cm in diameter but may exceed (collectively) 30 per cent of the mode. Plagioclase is the major phenocryst phase although locally exceeded by both augite or olivine (in picritic lavas) phenocrysts. Groundmasses are made up of augite, subsidiary plagioclase and opaques. Secondary minerals are uncommon. The plagioclase composition varies from An 90 to An 20 (in mugearites). Ophitic augite grains have mauve colours, and are generally zoned and titaniferous. Optical examination of the olivine phenocrysts indicated that they are forsteritic. The basalts mostly have intergranular texture grading into ophitic and, in mugearites, pilotaxitic textures.

In the trachytes, Sceal (1974) determined the alkali feldspar phenocryst phase to be anorthoclase (Or 38). Green pyroxene phenocrysts are rare in groundmasses made up of alkali feldspar laths and anhedral grains with zoned aegirine-augite and aenigmatite grains. Arfvedsonite and nepheline are rare but sodalite crystals are common. Both trachytic and intergranular textures are recorded by Sceal (1974).

Faulting occurred throughout the eruptive life of Paka: periods of faulting followed both main periods of trachytic volcanism as well as accompanying the final basalt eruptions.

Sceal (1974) undertook a detailed geochemical study of Paka with 50 major and trace elements, whole rock analyses (Tables 8 and 9). He concluded that both basaltic and trachytic extrusives had a common basaltic magma source of batholithic dimensions. Initial mineral fractionation produced upper trachytic and lower basaltic parts subsequently tapped via deep fractures.

He concluded, 'Repeated sampling of the top of the magma chamber, replenished by a continuous fractional crystallisation process beneath, could account for the preponderance of trachytic lavas in the volcanic edifice, the lack of an evolution with time of the trachytic lavas and the distribution of trace elements observed within individual flows.' (Compare also the McCall and Hornung (1972) evolutionary model for Silali.)

Silali Volcano ($Qpvt^S$, $Qpvb^S$, $Qrvt^S$, $Qrvb^S$), mostly based on McCall and Hornung, 1972

Silali Mountain with its impressive summit caldera is undoubtedly the major topographic feature of the median graben in Degree Square 27. It completely straddles the rift valley with an early tuffaceous unit extending on to the western rift shoulder. The main volcano has a NNE length of about 40 km and a width of about 28 km, i.e. a surface area of the order of 1120 km^2. The summit peak stands more than 700 m above the rift valley floor: the volcano is a large shield with relatively gentle profiles. Interestingly the summit caldera is elongated at right angles to the volcano's long axis: its ESE width is 8 km and its NNE diameter is only 5 km. The caldera walls have inner vertical drops of about 300 m; they remain unbreached and the caldera is not infilled by a lava pool. Other details of the morphology reflect the volcano's youthfulness: parasitic cones retain classic profiles, individual basalt and trachyte lavas are demarcated by unweathered flow fronts and surface (pressure ridges, *aa* and *pahoehoe*) textures. Fault scarps, including the caldera walls, are sharply delineated. All these features are clearly visible on aerial photographs; the volcano and its caldera are both large enough to provide landmarks on satellite imagery (cf. the front cover).

A detailed stratigraphy has been elucidated (McCall and Hornung, 1972; Truckle, 1979a) and the following localities provide reference sections for the major lithological units.

7 *Upper* (*Gulungul*) *basalts*: north-eastern limits with flows from the parasitic cone at Lokoyamama (GR 2088 1350).
6 *Upper* (*Black Hills*) *trachytes*: N-trending chain of craters (Black Hills) to east of the summit caldera (GR 195 132).
5 *Pumice tuffs*: east and west of caldera (GR 194 129 and GR 1867 1282).
4 (*Katenmening*) *basalts*: presently exposed over most of volcano except for western side.
3 (*Summit*) *trachytes*: Silali peak (GR 1850 1288) and whole of western side.
2 (*Kapedo*) *trachytic tuffs and* (*Missun*) *basalts*: around Kapedo (GR 1765 1290).
1 *Basal* (*Jebubbin*) *trachytes*: northern summit area (GR 1928 1323).

Table 8 Whole rock analyses of the median graben volcanics

	SILALI								PAKA		
	1	2	3	4	5	6	7	8	9	10	11
SiO_2	51.10	53.30	47.97	59.84	59.36	59.30	59.10	56.00	46.34	59.45	59.18
TiO_2	2.05	1.72	2.70	0.72	0.43	0.36	0.52	0.66	2.92	0.83	0.53
Al_2O_3	16.30	16.20	16.22	17.63	14.66	20.80	17.10	15.40	14.98	14.89	15.55
Fe_2O_3	0.70	2.65	4.48	5.87	8.87	0.84	4.93	5.62	3.28	4.14	5.46
FeO	8.99	6.39	7.15	3.13	0.41	3.13	4.48	0.44	10.35	3.78	2.45
MnO	0.19	0.20	0.21	0.29	0.30	0.18	0.29	0.25	0.19	0.30	0.32
MgO	4.54	3.84	5.26	0.43	0.33	0.46	0.40	0.57	5.71	0.92	0.60
CaO	8.63	7.24	10.33	1.70	1.50	0.95	1.53	1.30	11.03	2.32	1.79
Na_2O	3.42	3.15	2.27	6.14	5.00	5.87	7.43	10.40	3.10	6.00	6.71
K_2O	1.96	2.68	1.12	4.83	5.30	6.77	5.18	5.12	0.73	4.84	4.78
P_2O_5	0.64	0.51	0.57	0.17	0.16	0.14	0.14	0.22	0.38	0.09	0.05
L.O.I.	0.52	1.20	1.58	1.64	3.14	0.93	1.00	3.68	0.71	2.42	2.47
Total	99.04	98.68	99.86	102.39	99.46	99.73	102.10	99.66	99.72	99.89	99.89

(n.d. not determined)

1 Hawaiite, Silali
2 Hawaiite, Silali
3 Average 8 gabbro inclusions, Silali
4 Average 29 trachyte lavas, Silali
5 Average 3 trachyte tuff lavas, Silali
6 Syenite enclave, Silali
7 Phonolitic trachytic, Silali
8 Glassy phonolite, Silali

— McCall and Hornung (1972)

9 Basalt from Gulungul, Paka
10 Trachyte lava, Paka
11 Trachyte lava, Paka

— Sceal (1974)

No isotopic whole rock determinations have been undertaken. McCall (1968) suggests an early Pleistocene age for the initial eruptions with the caldera forming some time in the late Pleistocene. Fumaroles remain active, notably on the western flanks.

According to McCall and Hornung (1972) the foundation upon which Silali rests comprises Lower Pleistocene and older intermediate/acid volcanics—trachytes, comendites, rare phonolitic trachytes and phonolites. However, it is apparent from the geological map of Degree Square 27 that the only basal contacts now exposed are on the extreme eastern and western flanks. There is no evidence of what actually underlies the main volcanic edifice. In the east the outermost lavas rest directly on various Miocene units of the eastern Rift shoulder. To the west the Kapedo tuffs mantle parts of Ribkwo Volcano, the Murgisian basalts and the Jamakana formation. A 15 km wide swath of normal faults extends along the whole NNE summit axis of Silali. Summit depressions, notably the main caldera, have an alluvial infill and the whole volcano has an enveloping alluvial apron.

A maximum (summit) thickness of about 700 m is indicated by the present shape of Silali. The cross sections of Truckle (1979a) indicate that 500 m of this thickness comprise the Summit trachytes. Other units are not more than 150 m thick although the thickness of the basal trachytes is unknown.

No detailed descriptions of the various lithologies (and their petrography) have been published. The trachytic lavas are volumetrically the major component of Silali as massive, coalescing 'button' flows. The youngest flows show signs of incipient inhomogeneity (autoliths and patchiness) and the trachytic cones to the east of the summit caldera (the Black Hills) are almost pure black glass. Syenite enclaves are common; welded tuffs only form impersistent thin veneers to the trachytic flows (notably in small spills down satellite cone flanks).

Aa and *pahoehoe* textures have formed in the youngest basalt lavas (cf. Paka) which also contain blocks, several metres long, of gabbro and dolerite. These blocks have basalt veneers, vesicles and in some cases, interstitial glass. They, like their host rocks, are alkalic olivine basalts in composition and are interpreted as cognate xenoliths.

The western subaerial Kapedo tuffs grade westwards into a fluviatile series in which primary sedimentary structures abound and piedmontine torrent wash becomes increasingly intercalated in the mixed subaerial/water-laid tuff sequences.

Agpaitic mineral sequences occur in the trachytes and related, subsidiary phonolites. McCall and Hornung (1972) present major and trace element analyses of the various lithological units of Silali. Their results are reproduced in Tables 8 and 9 and they summarise 'In the trachytes relative to the basalts there is an increase in Si, Na, K and depletion of Mg, Ca, Ti; total iron shows slight depletion and the iron shows oxidation to the ferric oxide. Of the minor elements, Rb, Zn, Y, La, Zr, Pb and Nb are increased; Cu, Ba and Sr depleted; and Cr, Co and Ni show slight depletion. The trends seen in trachytes, etc. are incipiently evident in hawaiites and mugearites. The consistent bimodality, peralkaline character of the intermediate to acid rocks, major and minor element contents, and intimate intermingling of the two suites in space and time can only be interpreted in terms of complex differentiation processes affecting a mantle-derived basalt parent'.

Silali is a composite volcano, a dome built from clustered vents: the first phase of trachytic effusion was followed by emission of trachyte pyroclasts, which, in turn, was followed by a return to quiet effusion. Later, the volcano suffered (Holocene) sagging along a median,

EMURUANGOGOLAK				SUGUTA TROUGH	
12	13	14	15	16	17
45.26	48.27	62.14	62.61	48.89	49.05
3.14	3.57	1.10	0.81	2.44	2.43
14.85	14.30	15.15	11.01	15.76	15.73
2.59	13.28	3.25	11.06	13.04	12.90
10.20		2.88		n.d.	n.d.
0.22	0.28	0.28	0.44	0.20	0.20
6.95	3.91	0.84	0.15	5.62	5.71
10.86	8.29	1.65	0.90	10.28	10.26
3.00	5.40	6.63	7.84	3.99	3.90
0.70	1.43	4.04	4.11	0.90	0.90
0.80	1.47	0.21	0.05	0.43	0.45
0.46	n.d.	1.06	n.d.	n.d.	n.d.
99.12	100.20	99.23	99.01	101.54	101.51

12 High Ti-basalt, Emuruangogolak
13 Andesine-basalt, Emuruangogolak
14 Peralkaline quartz trachyte, Emuruangogolak
15 Obsidian, Emuruangogolak
— Weaver (1978)

16 Hawaiite, Suguta Floor
17 Hawaiite, Suguta Floor
— Truckle (1977a)

meridional zone, in which a fine grid of normal faults developed immediately after extensive emission of thin basalts flows from many small vents situated along these fracture lines. The basalts erupted obscured the trachytes of the main phase, but further trachyte eruption then occurred just prior to the formation, by subsidence, of an oval, 8 × 5 km caldera: in the last events, following the formation of the caldera, basalt and trachyte have been erupted from minor scattered centres, mainly situated outside the caldera'.

The caldera is a Glencoe-type structure; the walls show a dropped fault block and some scalloping, and there is negligible mantling of pumice associated with the actual caldera formation (i.e. not a Krakatoan type phenomenon).

McCall and Hornung (1972) conclude that high and low magma reservoirs existed beneath Silali, with the basalts tapping a deeper chamber and the trachytes emanating from high-standing (near surface) cupola reservoirs.

Emuruangogolak Volcano ($Qpvt^{Eu}$, $Qpvb^{Eu}$, $Qrvt^{Eu}$, $Qrvb^{Eu}$)

Weaver (1977) gives a detailed account of the geology and petrochemistry of Emuruangogolak from which the following brief summary is paraphrased. Following eruption of an early trachytic shield volcano, two episodes of caldera collapse occurred, each preceded by explosive pyroclastic activity. Post-caldera volcanism consisted of alternating phases of basalt and trachyte eruption. The basalt lavas are high-Ti ferrobasalts of mildly alkaline 'transitional' composition; the trachytes are peralkaline and oversaturated. A genetic relationship is established between all the volcanics based on their major trace element contents. Intermediate volcanics are conspicuously absent.

The volcano covers an area of about 600 km² with summit heights of over 700 m above the Suguta Valley which it completely straddles. The two episodes of summit collapse produced shallow nested calderas of 4 km and 8 km average diameter with maximum wall heights of 75 m. The late formation of an incomplete circular fracture may herald a third caldera collapse event. Parasitic cones stand up to several hundred metres with steep (40°) slopes and well-preserved central, circular craters. Cliff features on the northern side relate to old shore lines of Lake Suguta. Youthful features such as flow fronts, hornitos, driblet spires and levees are well preserved. The summit calderas, as on Silali, are orientated at right angles to the trend of the median graben.

Trachyte lavas from the oldest and youngest exposed part of the main shield give K/Ar ages of 0.9 ± 0.1 Ma and 0.5 ± 0.1 Ma respectively. Shells collected from upper, intercalated sediments were dated by their 14C contents as 11 000 to 9000 years BP. Charcoal preserved beneath the most recent flow gives a 14C age of 250 ± 100 years. Skinner and others (1975) used three youngest flows for secular magnetic variation curve constructions which suggest that these flows are from 900 to 700 years old. The youngest flow maybe only about 70 years old: elderly Turkana tribesmen recall earth tremors and volcanic activity.

The early trachyte shield had a summit thickness of 600 m (with 3 to 5° slopes): individual trachyte lava flows vary from 10 to 30 m thick. The overlying pyroclastics are up to 150 m thick whereas the associated trachyte lavas are highly viscous, short (button) flows up to 100 m thick. Numerous lava tubes and tunnels in the basalt lavas have been measured at 8 m in diameter and traced for about 300 m.

The volcanic products of Emuruangogolak are of basaltic (20 per cent by volume) and trachytic (80 per cent) compositions including the plutonic and pyroclastic equivalents of the lavas. Following collapse of an initial caldera, basalt and trachyte were erupted in about equal proportions and in a remarkably consistent alternating sequence. Four cycles of basalt-trachyte volcanism may be inferred, three of which occurred after the appearance of a second caldera. In the post-caldera history, trachyte lavas were erupted only in the summit area from vents located on the three ring fractures. By contrast basalt volcanism is far less central in character and the fracture systems from which basalts were erupted appear to have a more regional significance. This implies that trachytes had a high level cupola source whilst the basalts tapped a deeper magma (see discussion on the western shoulder trachyte volcanoes). Truckle (1979a & b) summarised the stratigraphy:

9 Upper trachyte
8 Basalt with sedimentary intercalations
7 'Arc trachyte' (flows arcuate in plan)
6 Basalt
5 Lower trachyte (formation of the second caldera)
4 Trachytic tuffs
3 Basalts with basal sediments (formation of the first caldera)
2 Pumice tuffs and agglomerates
1 Main shield trachyte

The main shield trachyte and the overlying pumice tuffs are exposed on the western flank of the volcano. The pumice tuffs are concentrated on the upper slopes with the agglomerates adjacent to the source cone forming Emuruangogolak summit peak. Similarly the oldest basalts now crop out on the western side; basal sediments are exposed on a scarp slope south of Kamuge at the northern end of the Kangemonginolei (Kagimoging'olei) Plateau. The trachytic tuffs are confined within the outermost caldera. The upper basalts form a carapace over much of the northern and southern parts of the volcano, whereas the various upper trachytes are confined to discrete flows. The lower trachyte units breached the summit calderas, to

Table 9 Trace element analyses of volcanic rocks from the median graben of the Gregory Rift

SILALI (from McCall and Hornung, 1972)

Element	Trachytes and syenites (72 analyses)		Basalts and gabbros (49 analyses)	
	average (ppm)	range (ppm)	average (ppm)	range (ppm)
Rb	142	55–260	16	0–55
Cu	20	0–50	73	0–160
Sr	42	0–535	617	265–388
Ba	200	0–2250	884	225–5100
Zn	208	20–330	85	45–200
Y	96	8–230	34	20–70
La	103	20–210	16	0–65
Zr	916	70–1920	237	75–1450
Pb	15	0–45	2	0–15
Nb	335	80–410	27	1–85
Cr	54	10–190	155	50–480
Co	44	0–155	71	0–180
Ni	321	65–1270	509	165–915

K/Rb ratio: basalts = 580; trachytes etc. = 495

PAKA (from Sceal, 1974)

	Basalt 9	Basalt 10	Trachytes 11 (from Table 8)
	(ppm)		
Rb	11	110	223
Sr	491	42	74
Ba	411	174	291
Zn	82	161	252
Y	33	81	205
La	34	102	258
Zr	150	663	1634
Nb	26	155	349
Ce	61	202	379
Nd	10	96	185
Pr	16	24	44
K/Rb ratio:	6700	440	214

be confined to the uppermost summit slopes, although one lava flowed most of the way down the northern slope. Contemporaneous trachytes erupted from the parasitic cone at Losetom. The 'arc trachytes' are concentrated in the central caldera and along the SE (faulted) margin where they erupted from a parasitic vent. The youngest trachyte forms a single flow on the upper parts of the south summit area.

The trachyte lavas are dark green, almost black, porphyritic rocks with alkali feldspar phenocrysts. Thin trachytic ignimbrites and vent-related agglomerates are intercalated in the lower shield flows. Uppermost trachytic flows are locally arcuate in plan, as a consequence of their feeder dykes utilising caldera faults.

The trachytic pyroclastics are yellow, brown or green pumice tuffs, locally with trachyte lapilli and (rare) rounded blocks of trachyte and syenite up to 0.5 m in diameter. Agglomerates are proximal to cones, as are basaltic cinder beds.

A black obsidian flow locally caps the second caldera and has up to 25 per cent by volume of alkali feldspar phenocrysts.

The basalt lavas are mostly porphyritic with olivine and/or plagioclase phenocrysts visible in the field. *Pahoehoe* and *aa* flows occur; the most recent flows were fed from NNE fissures. Faulting along the fissures was contemporaneous with lava outpouring. Small gabbro nodules are locally present and, together with associated olivine nodules, may make up 30 per cent of the lava volume.

The basalts contain up to 15 per cent of olivine (Fo 87–80) and up to 5 per cent of zoned plagioclase (An 82–62) phenocrysts in aphanitic groundmasses. Pale coloured clinopyroxene phenocrysts are rare. The groundmasses have fine intergranular textures containing abundant small clinopyroxene granules between plagioclase laths and are particularly rich in skeletal ilmenite crystals and cubes and octahedra of magnetite. Olivine is present as small sparsely distributed grains. A few hawaiitic lavas are aphyric with rare phenocrysts of labradorite; the groundmass plagioclase is in the andesine range. Two successive basalt flows in the north, interbedded with lake sediments, are highly cumulophyric picrites. They have up to 40 per cent large phenocrysts, typically 15 per cent olivine, 15 per cent plagioclase and 10 per cent clinopyroxene. The groundmasses have intergranular textures with labradorite laths, clinopyroxene granules, ilmenite and magnetite blebs and rare olivine grains.

The trachytes are feldspathic with 15 to 30 per cent mafic minerals. They have 5 to 30 per cent anorthoclase (Or 10–38) phenocrysts with minor clinopyroxene and fayalite (Fa 92–96) phenocrysts. In some, magnetite occurs as microphenocrysts and rims early fayalite. Groundmasses are composed of alkali feldspar laths or microlites and subpoikilitic patches of alkali amphiboles (ferrorichterite-arfvedsonite), aegirine pyroxenes and aenigmatite. Interstitial quartz constitutes up to about 5 per cent of modes.

The obsidians are composed of pale brown-green glass containing phenocrysts of anorthoclase, green clinopyroxene and fayalite.

The syenite clasts consist largely of microantiperthitic feldspar plates, a little interstitial quartz and mafic clots (intergrowths of zoned green clinopyroxene, ferrorichterite-arfvedsonite and aenigmatite). A few discrete fayalite crystals are rimmed by magnetite.

Weaver (1977) analysed over 80 rock samples which showed a pronounced bimodality in their compositions: no rocks had silica contents intermediate between basalts and trachytes. He showed the basalts have mildly alkaline 'transitional' compositions with 1 to 4 per cent normative Ne, distinctly high Ti and Fe contents and low Fe_2O_3/FeO ratios. Variations in the basalts from olivine basalt through 'high Ti' ferrobasalt to 'andesine-ferrobasalt' is explicable in terms of the fractionation of the phenocryst phases. The determined trace element abundances agree well with this scheme. A magma source at 40 to 60 km depth is estimated.

The trachyte lavas range from subaluminous to strongly peralkaline compositions with up to 20 per cent normative *ac* and *qz* from 1 to 12 per cent. In general the total iron oxide increases as alumina decreases and trachytes of both comenditic and pantelleritic affinity are represented. There is no simple variation with time of the composition of trachytes erupted. The syenite bombs are chemically similar to the trachyte lavas, so probably represent

crystallisation of trachytic liquids at deeper levels. In broad terms the chemical variations can be explained by crystal fractionation dominated by alkali feldspar approaching the Or 38 composition and to a lesser extent by fractionation of small quantities of the mafic phases.

A combination of the field and chemical evidence suggests that the trachytes were generated by crystal fractionation of basaltic magma; trachytic magma initially collecting in an upper cupola above the primary chamber (cf. previous sections on Silali and Paka).

Namarunu Volcano ($Qpvt^N$, $Qpvb^N$, $Qrvt^N$, $Qrvb^N$), from Truckle, 1977a

Namarunu Volcano straddles latitude 2°N; its southern portion covers an area of about 10 by 10 km² in the extreme north of Sheet 27. Its summit area, after which it is named, is over 620 m above sea level compared to perimeter levels of about 300 m. Outcrop is essentially confined to Namarunu Hill, Nakitoekirion Hill and adjacent parts of the Loriu Plateau.

There is no summit caldera but Namarunu Hill approximates to the culmination of a broad volcano. The western outcrops correspond to the western flank of the volcano; the eastern flank has been downfaulted into the Suguta Trough and mantled by superficial sediments. Numerous parasitic cones are mainly concentrated along NNE-trending normal faults. These faults define sharp scarp features, and upper basalt and trachyte lavas are outlined by unweathered flow fronts and surface textures. They can be traced back to source vents. The lavas are clearly defined on aerial photographs and satellite imagery: summit algal limestones form a distinctive pale toned area.

A Quaternary age for Namarunu is indicated by its setting on the edge of the Suguta Trough and because it is founded on the Lorikipi basalts. K/Ar whole rock dating of a trachyte gave an anomalously high age of 6.8 ± 0.5 Ma.

The maximum thickness of volcanic rock under Namarunu Hill is about 600 m according to Truckle (1979b). This consists of about 300 m of the basal main shield trachyte and up to 200 m of overlying basalts (which thicken westwards). The full stratigraphy is:

4 Basaltic pyroclastic cones (various ages)
3 Basalts and intercalated sediments
2 Basaltic pillow lavas and intercalated sediments
1 Main shield trachytes and intercalated tuffs

On Namarunu Hill the trachytes are dominated by feldsparphyric trachyte lavas, individually up to 30 m thick. Pyroclastic flows and air-fall pumice lapilli tuffs and agglomerates are common to include an uppermost, 30 m thick, poorly sorted, unbedded trachyte agglomerate. The lavas dip radially off cones and there is an overall quaquaversal arrangement of dips off Namarunu summit: the main source area to the volcano. Secondary silicification and reddening of the lavas is due to hot spring activity.

At least 365 m of thick trachytic lavas with lenses of air-fall and ash-flow tuffs are well exposed on Nakitoekirion Hill. Basal units are concealed by younger basalts abutting against fault scarps. The trachytes dip westward. Up to 25 per cent of the succession comprises pyroclastics with a basal unit of poorly welded pumice lapilli-tuffs and ignimbrites over 20 m thick. This basal unit is exposed on the Loriu Plateau where it thins northwards. Here a green feldsparphyric trachyte overlies the pyroclastics and is over 10 m thick.

The overlying sediments and intercalated basalts are assigned by Truckle (1976; 1977a) to the 'Suguta Beds' and briefly described in Section 5.4. The basalts capping Namarunu Volcano are poorly exposed aphyric and porphyritic (plagioclase; plagioclase with pyroxene and minor olivine) lavas. They, like source cinder cones, are variably weathered to suggest different ages of eruption. The cones comprise basaltic scoria with bombs and spindle-shaped lava blobs.

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5 Cenozoic sedimentary rocks

Cenozoic sedimentary rocks, confined within the Gregory Rift, either infill subvolcanic hollows in a weathered basement surface or define intravolcanic lenses. The basal sediments comprise basement-derived clastic debris, whereas the intravolcanic sediments embrace a wide lithological range in which the amount of basement-derived material decreases rapidly up the succesion. Primary pyroclastics (air-fall tuffs, lapilli tuffs and agglomerates) and reworked waterlain deposits are recognised. The primary deposits are directly related to volcanic centres where they are cut by crystalline intrusives (see last chapter and Williams and Chapman, 1986).

Sedimentation in the Gregory Rift was, and is, being controlled by contemporaneous volcanism and tectonism. A primary volcanic control has already been cited. Secondary factors include breaks in volcanism, concomitant with the presence of surface water ponded by volcanic edifices, also earth movements related to volcanism producing lahars. Cenozoic tectonism also influenced sedimentation by altering the landscape to define troughs in which lacustrine or landslip material accumulated. Major fault scarps are the principal tectonic elements in these processes.

5.1 SUBVOLCANIC SEDIMENTS (KIRIMUN FORMATION)

Key (1982) formally assigned all subvolcanic sediments within the eastern shoulder of the Gregory Rift to the Kirimun Formation. The type area, around Kirimun, is about 45 km SSE of Maralal where Shackleton (1946) first mapped these sediments. In Degree Square 27 outcrop of subvolcanic sediments is confined to the NE margin of the east shoulder: in river valleys or lower slopes of escarpments capped by the more resistant overlying lavas.

There are no subvolcanic sediments exposed beneath the western shoulder in Degree Square 27, but further west Weaver (1973) records subvolcanic sedimentary wedges up to 450 m thick related to major fault scarps. Sediments of the Kirimun Formation are everywhere overlain by volcanic rock and have minimal morphological expression. Quartz gravel spreads delineate areas once covered by the Kirimun Formation and show as pale areas on aerial photographs.

No fossils have been found since the original discoveries of Shackleton (1946) from the type area. This fossil assemblage indicates a Miocene age, confirmed by the middle Miocene age of the overlying and locally interdigitating Samburu basalts and Lopet Phonolites. Sedimentation and subsequent partial erosion probably occurred intermittently throughout the early to middle Miocene.

The Kirimun Formation infilled hollows in the sub-Miocene basement surface, and is preserved beneath either the Samburu basalts or Lopet Phonolites. Locally the onset of both these volcanic episodes overlapped with sedimentation; manifested by discrete volcanic interlayers or a volcaniclastic component to polymict sediments. The thickest sedimentary sequence in Degree Square 27 occurs on the escarpment capped by the Lopet Phonolites west of Sikira at GR 2483 1750. Here the Kirimun Formation is about 63 m thick. Elsewhere it is between 1 and 45 m thick.

Widely ranging lithologies have been recorded by Baker (1963), Wilkinson (1983a & b) and Charsley (1984) between extremes of variably bedded and sorted coarse arenites and finely laminated argillites. Chemical sediments are not present but baked (porcellanite) units are common beneath lavas. The following measured sections are typical:

West of Sikira

	m	
Rounded quartz pebble conglomeratic grits	60	63 m
Coarse conglomerate with basement clasts up to 15 cm long	3	

GR 2465 1750

Fissile Lopet Phonolites	2	
Coarse, green, well-bedded conglomerate	12	22 m
Coarse volcaniclastic grit with pumice fragments	3	
Coarse pumiceous agglomerate	3	
Coarse volcaniclastic grit	2	
Finely-laminated buff-white siltstones	2	

GR 2465 1750

Fissile Lopet Phonolites	3	
Coarse, green pumiceous agglomerate and tuff	10	
Clean, silica-cemented quartz gravels and conglomerate	15	43 m
Finely-laminated, pale-yellow siltstones	25	
Localised basal conglomerate	3	

Thin sections showed 60 to 80 per cent of the arenaceous rocks to consist mostly of angular to subrounded quartz grains with microcline, oligoclase and lithic clasts. Quartz and calcite cement these mineral and lithic fragments. Polymict rocks have assorted basement and volcanic, mineral and lithic fragments (Plate 11).

The Kirimun Formation represents deposition in isolated natural surface depressions (occasional shallow lakes or pans) fed by fast (torrential floodwash) and shallow rivers, partly as *in situ* soils and in part as air-fall tuffaceous deposits and re-worked volcaniclastic sediments.

5.2 MIOCENE INTRAVOLCANIC SEDIMENTS

Primary air-fall pyroclastics, described in the previous chapter, are relatively common within most Miocene volcanic units. However epiclastic and chemical sediments are far less common. Epiclastic sediments are recognised by their small scale structures of laminar, graded, current and cross bedding, slumping, channeling, flame structures and argillaceous seams. Within the eastern shoulder they are either located within the major Samburu and Kamolingaran basalts (Plate 12) or the

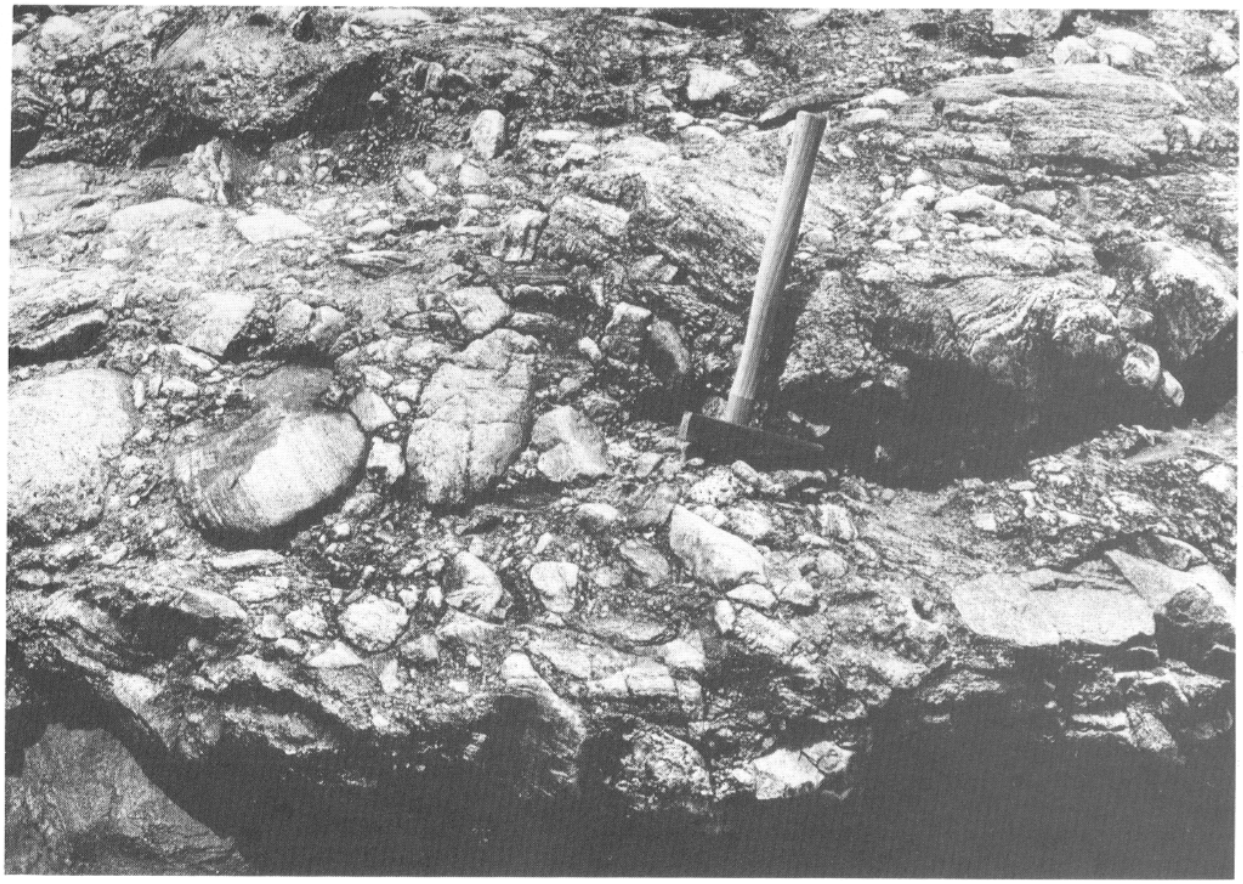


Plate 11 Clastic, basement-sourced sediments of the Kirimun Formation, Marti area



Plate 12 Volcaniclastic sediments within the Samburu basalts, west of Marti

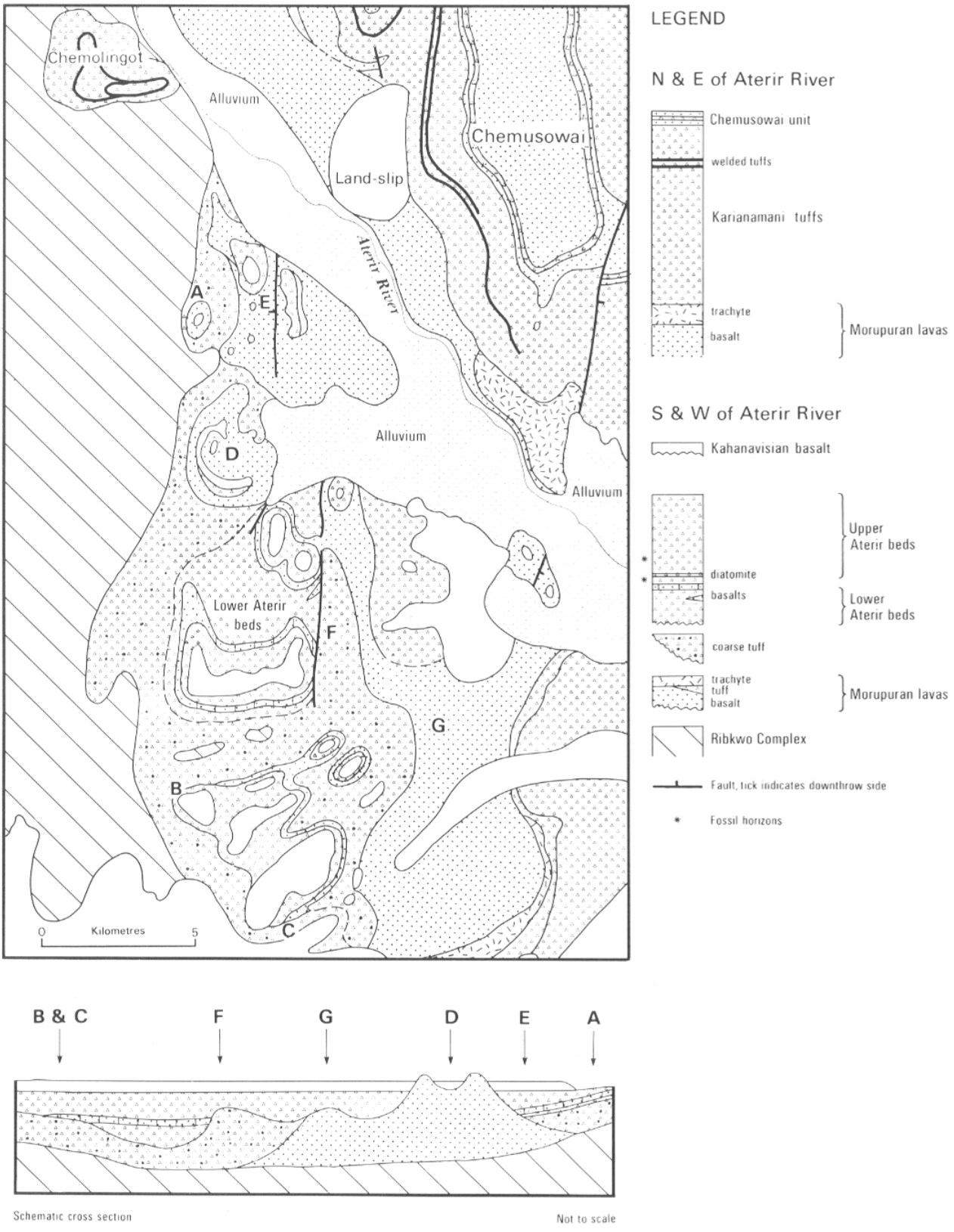


Figure 16 The Aterir Basin (from Webb, 1971)

western tuff belt. Their important vertebrate fossil assemblages have been documented (last chapter) and silicified wood remains are widespread. Such sediments only form an important component in the Kamolingaran basalts (Section 4.1, p. 34).

Two thin and aerially insignificant dolomitic limestones occur in the western shoulder volcanics: in the basal Kapchererat basalts west of Degree Square 27 and in the Tirioko basalts at GR 1715 1856 (Napeitom Hill). The younger limestone is about 5 m thick with basal algal structures and gastropod fossils in its upper parts. Epiclastic sediments are not well developed in the western shoulder with the exception of the Aterir beds of Kachila Volcano. These vertebrate fossil-bearing tuffaceous sediments (see last chapter) have a basal (0 to 18 m thick) subaerial tuff unit overlain by 0 to 9 m of fluviatile, well-laminated, fine-grained white tuffaceous sediments, which partly scour into the older tuffs. They in turn were overlain by a thin brown earth prior to the fluviatile and lacustrine deposition of further (3 to 18 m) tuffaceous sediments and diatomites. The stratigraphy and details of the geology are shown in Figure 16 (from Webb, 1971). Leaf impressions are preserved in argillites within Nasaken Volcano.

5.3 PLIOCENE SEDIMENTS

Primary air-fall pyroclastics are recognised in all the Pliocene volcanic units, whereas epiclastic and other sediments are of minor significance. Thin hard limestones with gastropod fossils and algal laminations occur in the Lokwaleibit basalts together with cherts and epiclastic tuffaceous sediments. At the base of the Nathelot basalts (demarcated on the accompanying map) are up to 30 m of white massive limestones with or without grit, polymict conglomerate and pisolitic beds. Truckle (1977a) shows these sediments represent lacustrine infill of a valley blocked by the earliest Nathelot basalt flows. Metamorphic basement clasts in the conglomerates indicate a source area to the north-west. Vertebrate fossil remains from the Jamakana formation sediments were listed in the last chapter. Red soils or reddened water-laid gravel horizons indicate minor breaks in lava pile sequences in several of the trachyte volcanoes.

5.4 QUATERNARY SEDIMENTS

Eastern basement

Quaternary deposits mantling the eastern metamorphic rocks include alluvial infill and overspill to most watercourses, residual soils, colluvium passing laterally into talus at the bases of major hills, and minor calcrete spreads.

Meandering and braided, shallow, poorly-defined channels related to sporadic flow after the last heavy rains are traced in the yellow-brown, medium-grained sands with gravel bars which top the alluvium of most watercourses. Rare sections through this alluvium reveal alternating bands of unconsolidated silts, clays and sands with rarer gravel lenses. Streams on the Karisia Hills and the fossil drainage network in the El Barta Plains are unusual for their dark grey silt-rich alluvium, possibly related to former volcanic covers. Total thicknesses of alluvium are rarely known but most watercourses are effectively filled, with shallow banks. Almost certainly the widest rivers have the thickest alluvium. Shallow flood deposits mantle low interfluves upstream of major confluences.

Residual soils, although common, are rarely more than 2 m thick and are being eroded by the more active drainage systems (e.g. around Barsaloi and Masiketa). Soils are mostly red-brown sandy loams with local variations related to underlying bed-rock. Mafic-rich gneisses weather to dark brown soils on the El Barta Plains. Vermiculite-bearing soils overlie much of the Karisia Hills although the soils above the muscovite-quartz gneisses near Maralal are full of silvery muscovite flakes. Quartz pebble seams and kankar nodules are locally seen in soil profiles.

All colluvium is a crudely stratified, poorly-sorted mixture of lithic clasts, sand, silt and clay which is thickest around the peripheries of major hills. Thicknesses of about 12 m (GR 2727 1390) are revealed in gullies cut by torrential run-off from kopjes. Substantial erosion of colluvium is taking place by the major drainage systems. Colluvial spreads represent transitional deposits linking talus with residual soils. The extent of colluvium is directly related to slope steepness and the extent of source hills. The colluvium was formed by exfoliation, hillcreep, sheetwash and to a lesser extent by wind action prior to Quaternary uplift. The El Barta Plains in the east are underlain by thick colluvium mainly derived from the Ndotos. This thins rapidly westwards until residual soils are the dominant cover south of Baragoi.

Fresh talus in varying states of consolidation is ubiquitous around all major hills and kopjes, but is never very thick.

Calcrete sheets are rare and of limited lateral extent. Around Barsaloi they are associated with small patches of 'black cotton soil'—dark grey, structureless silty clays. Both deposits are thought to be related to pools of stagnant water formed by flood overspill on to impervious bedrock. Minor ferricrete occurs at GR 2441 1530.

Eastern shoulder of the Gregory Rift

Quaternary unconsolidated mantles generally reflect (Pleistocene) periods of strong weathering which involved massive transport, often over considerable distances.

A noticeable feature of alluvial infill to watercourses draining volcanic terrains is the abundance of boulders which vary in size up to several metres in diameter. Close to the median graben these rounded boulders are essentially the sole alluvial component and reflect torrential flow. Upstream a coarse volcaniclastic gravel forms a matrix to the transported boulders.

Chaotic boulder accumulations due to landslip are a common feature of the precipitous interface between the east shoulder and median graben. They are either due to normal faulting, fissuring or to preferential erosion of softer tuffaceous rocks underneath plateaus or mesas capped by resistant lavas. Talus, in part related to landslipping, is especially well developed at the base of major fault scarps and erosional escarpments to phonolite plateaus.

Another feature of the eastern shoulder of the median graben are numerous terrace deposits related to former high lake levels in the graben (Baker, 1963). These deposits are up to 25 m thick and elevated by over 100 m above the present graben floor. They consist of mixtures of silt, gravel and boulder beds and are readily distinguished on aerial photographs.

Shallow, flat-bottomed depressions (pans) on the phonolite plateaus adjacent to cooling ridges in underlying flows, are partly infilled by less than 2 m of waterlain silt, clay and volcanic pebble gravels.

Again these laterally extensive deposits are readily visible on aerial photographs. Shallow soils overlying the volcanics are characterised by uppermost lava pebble and cobble gravel spreads in subservient dark volcaniclastic silt matrices. Coarse boulder spreads cover interfluves of the more dissected western terrains related to torrential overwash. Ferricrete, in which phonolite clasts are cemented by limonite, forms an insignificant deposit at GR 2406 1562.

Median graben of the Gregory Rift

Truckle (1976; 1977a) documents evidence for Quaternary lakes within the Suguta Trough. Consequently Pleistocene lacustrine sediments are an important part of the geology of the Suguta Trough, especially north of Emuruangogolak (which probably defined the southern shoreline of an extended Lake Turkana prior to the eruption of the Barrier Volcano at the south end of that lake). The trough was first initiated about 3 Ma, and subsequently filled by volcanics. It was the reforming of the trough at about 0.75 Ma that led to the eventual development of Lake Suguta. Several ages of lacustrine sediments are preserved as intercalations between trachytic lavas on Namarunu and (the northern side of) Emuruangogolak Volcano. These, like the later mud flat which presently separates the two volcanoes, consist of laminated reddish brown clays, up to several tens of metres thick.

An intimate mixing of lacustrine and fluviatile sediments has resulted on the floor of the median graben from the erratic meandering and braiding of the Suguta River across the mudflats separating the central volcanoes. Truckle (1976; 1977a) provides great detail on sediments and an intercalated basalt which accumulated during the Quaternary in the northern part of the Suguta Trough. He refers to this succession as the 'Suguta Beds' in which four member are recognised (Figure 17). The sediments of members A, B and D were deposited during high water level of a Lake Suguta. Strand lines along the flanks of the Suguta Trough suggest the lake surface reached the 600 m contour during deposition of member D. Truckle (*op. cit.*) elegantly demonstrates a connection via a western spillway along the Kerio River between Lake Suguta and Lake Turkana. The youngest sediments are dated by 14C measurements on contained gastropods at about 9600 years BP. Pillow structures occur in the member C basalt at Gwasinyono Arrolio.

South (i.e. upstream) of Emuruangogolak and Silali, extensive flood plain deposits of the Suguta River have developed from intermittent blocking of the river by lava flows from these volcanoes. Due to the contemporaneous volcanicity, it is likely that the thickness of Quaternary sediment under the median graben is variable but nowhere considerable. Both flanks of the median graben have extensive spreads of coarse volcaniclastic debris, caused either by landslip off the shoulders, or by outwash fans—especially where the source rivers cut across basaltic units. Typically the fans consist of chaotic, unsorted loose accumulations of rounded to subrounded basalt lava clasts ranging in size up to boulders several metres in diameter.

Seif and Barchan dunes identified (Truckle, 1979b) south of Namarunu on the western side of the median graben indicate winds from the WNW. The dunes rest on a mudflat surface and are Holocene accumulations. The thin algal limestone locally capping Namarunu Volcano is related to the 9600 years BP high lake level, contemporaneous with the eastern terrace deposits (Truckle, 1976). Remains of *Crocodylus*, *Lates sp.*, *Clarias sp.* and rare gastropods (*Melanoides tuberculata*) and bivalves occur in the 'Suguta Beds'.

Primary air-fall pyroclastic layers related to the Quaternary volcanoes have been described in the last chapter. South of Kapedo one such deposit from Silali is intercalated with fluviatile sediments of the Suguta River.

Western shoulder

The most extensive Quaternary sediments occur in the north-west, related to south bank tributaries of the Kerio River. These alluvial deposits consist of interdigitating flood plain spreads and outwash fans (again concentrated where rivers debouch off exposed basalt plateaus). The Amaries sediments are exposed in headwater gullies of several of these south bank tributaries. The Kerio River is unique in having a sand-rich infill with clasts of basement rocks; other western rivers are clogged by volcanic boulders.

The Amaries sediments were located and described by Truckle (1977a) to the south-west of Lomi Volcano, where they infilled a fault-controlled trough prior to eruption of the overlying Kamuge basalts. 10.6 m of clastic sediments comprise a basal wedge (up to 0.9 m thick) of

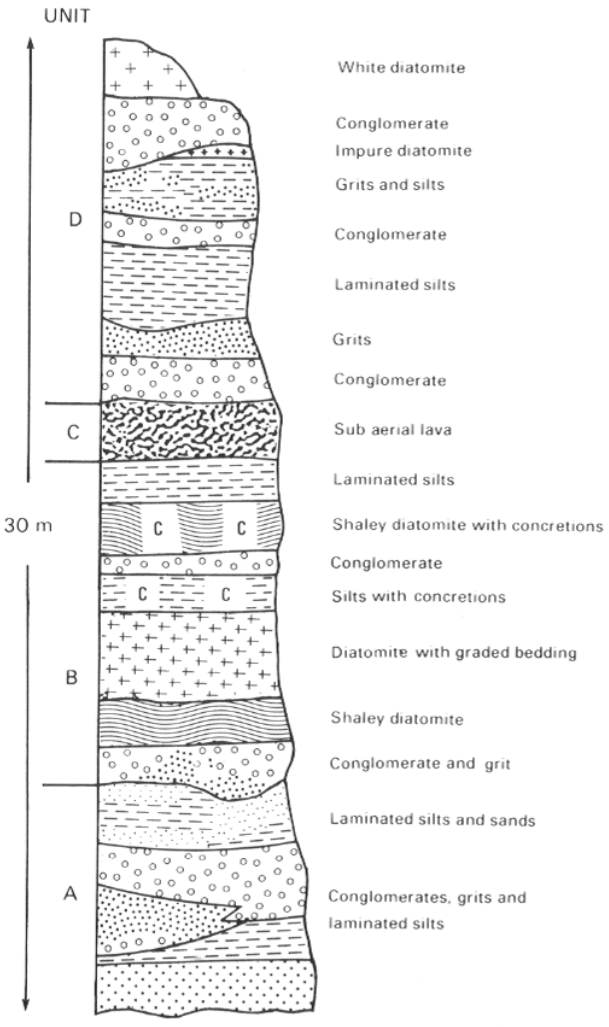


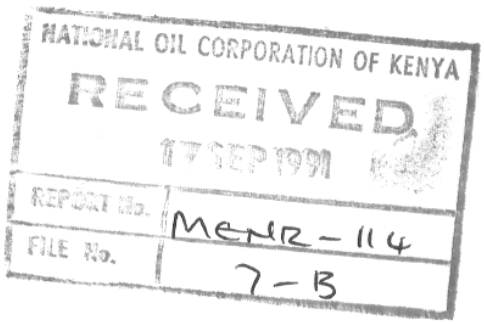
Figure 17 Type section of the 'Suguta Beds' (from Truckle, 1976)

reworked pumice-rich tuffs overlain by 4.5 m of red-brown, porous, poorly-bedded grits with upper well-bedded silts and graded grits. An upper unit 4.5 m thick has conglomerate lenses in grey grits and sands. Lateral facies changes and erosional unconformities in the sediments point to torrential deposition.

Thin and discontinuous, flat-lying diatomite beds noted by Truckle (1977a) overlying plateau lavas, are attributed to the periodic high water levels of Lake Suguta manifested by deposition of members A, B and D of the 'Suguta Beds'.

Quaternary erosion of the western shoulder volcanics has preferentially scoured the softer pyroclastic beds and coarse basaltic lavas in preference to the more resistant trachytic and aphyric basalt lava flows. Therefore central zones of the trachyte volcanoes, with their large accumulations of pyroclastics, are more eroded, and erroneously look older, than the surrounding flatter terrains. Landslipped lava blocks caused by undermining of softer basal pyroclastics are common in these central zones. Local redistribution of pyroclastic beds by wind and surface water is also common, e.g. south of Nasaken Volcano. In contrast the monotonous flood lavas are covered by thin gravel spreads (coated by desert varnish) and are essentially stable surfaces. Talus is well-developed, both along major fault scarps and plateau escarpments.

Unusual Quaternary sedimentary units include small mud volcanoes in the Nasaken River at Nasaken (Weaver, 1973); evaporite concentrations around hot springs (Webb, 1971): 1 m thick pure white diatomaceous layers in an alluvial sequence at GR 1688 1155 (Sceal, 1974); finely bedded siltstones over the Jamakana formation with very gentle northerly dips at GR 1718 1115 (Sceal, 1974); 'black cotton soil' over pumice tuffs (Weaver, 1973).



6 Igneous intrusives

Minor felsic veins ubiquitous in all metamorphic rocks are not considered in this chapter. They are briefly described in the relevant parts of Chapter 3.

6.1 METAMORPHOSED ULTRABASIC BODIES (EXCLUDING THE SIAMBU COMPLEX)

Metamorphosed ultrabasic intrusives are essentially confined to the Siambu Complex. Brief descriptions of these important rocks have been given in Chapter 3, and Baker (1963) provides details of those intrusives within the NE quadrant. Two minor metamorphosed ultrabasic pods were discovered in the south-east (Key, 1983a) within the Loroki Gneisses. Exposed in a footpath on the west side of the Ol Ketuloni headwater valley (GR 2516 1315) is a massive, pale green serpentinite containing white mica flakes and thin white magnesite veins. This ultrabasic appears to form a concordant (probably boudinaged) pod in hornblende gneisses; it is about 20 m thick, several hundred metres long and steeply dipping. The serpentinite weathers to a typical orange-brown colour.

At Opiroi there is small hill (GR 2560 1277) west of the main track to Maralal, with sporadic outcrop of a pale green, massive felted rock entirely composed of talc rosettes. This also forms a pod in the same hornblende gneisses.

Both of these ultrabasic bodies probably represent parts of thin slices of the Siambu Complex thrust into the Loroki Gneisses at an early stage in the tectonothermal history of the area.

6.2 METABASIC INTRUSIVES (EXCLUDING THE SIAMBU COMPLEX)

Early, i.e. pre-gneissosity, basic intrusives are again essentially confined to the Siambu Complex and have already been described (see also Plate 15). Two further occurrences, described below, were probably also part of this complex and subsequently tectonically emplaced into adjacent lithostratigraphic units. Both occur in the same general area as the two ultrabasic pods mentioned in Section 6.1.

In the valley east of Samu Devr Hill there is a low rise (GR 2598 1292) underlain by a massive, black metagabbro. The hill is isolated by surrounding alluvium so that its contacts with the adjacent Ndura Complex are unexposed.

At GR 2606 1315 there is a line of rounded boulders trending 010° for about 100 m of the same metagabbro. Adjacent to this elongate pod or dyke, the host Loroki Gneisses contain up to about 20 per cent by volume of red garnets (about 0.5 cm in diameter).

Baker (1963) describes several types of metasomatised basic dykes cutting gneissic fabrics in the Siambu Complex. These include a group of four or five narrow (less than 1 m thick) 'epidiorite' dykes exposed north-west of Lamerok Hill (GR 2503 1875), which resemble metadolerites in the field. In thin section hornblende, biotite and epidote grains occur in a plagioclase and quartz matrix. However the main suite of metabasic dykes comprises the foliated biotite-rich, rather light-coloured rocks exposed in the Lamerok, Baragoi, Lebanyuki and Kolowatoni rivers. The foliation is variably developed. Hornblende, quartz, plagioclase, sphene and microcline occur with biotite which forms small and large plates. Slightly less metasomatised and more massive dykes are exposed in the Baragoi River about 5 km SW of Nachola. They are essentially fine-grained, greenish leucocratic rocks with small hornblende phenocrysts in a groundmass of alkali feldspar, quartz and epidote.

All these dykes were emplaced during the time interval following the generation of the gneissic fabric, but before the (Baragoian) emplacement of minor felsic veins: pegmatites and quartz veins cut the metabasic dykes. These dykes are confined to the Siambu Complex, suggesting a genetic link with their host mafic-rich rocks.

6.3 G1 SODIC INTRUSIVES

Early, pre- or syn-gneissosity, felsic intrusives are confined to small, partial melt products of their host rock. No large intrusives of this age are recognised. This may be partly due to the difficulty of positively identifying such highly altered bodies; large parts of the Loroki Gneisses may be orthogneisses.

Within the Siambu Complex there are irregularly shaped bodies of dark greenish grey, light grey (mottled) or grey-brown rocks. They are massive in their larger outcrop areas but foliated or gneissic when interlayered with the host gneisses (Charsley, 1984). The common mineral assemblages consist of varying proportions of oligoclase, K-feldspar (microcline and/or orthoclase), biotite, hornblende and quartz. They formed as allochthonous bodies during the initial metamorphism which generated the gneissic fabric of the Siambu Complex. The *lit par lit* felsic veins of the Ndura Complex are of comparable age.

Ledeiro Hill, due south of Maralal (GR 247 114) is underlain by a flaggy, medium-grained melanocratic granite which is stained red-brown by the weathering of numerous magnetite grains. The granite has a gneissic fabric, and also contains xenoliths of hornblende gneiss, suggesting a late kinematic origin with regard to the gneissosity. Biotite, hornblende, red garnet and feldspar porphyroblasts occur in a groundmass of plagioclase, orthoclase, quartz, magnetite and sericite.

6.4 G2 GRANITES

The most impressive felsic intrusives are the G2 granite stocks, partial melt products of quartzofeldspathic gneisses which moved upwards to form autochthonous intrusives into the Siambu Complex. The contem-

7 Tectonothermal history of the eastern metamorphic province

The metamorphic rocks have been affected by a series of tectonic episodes (Table 10) individually identified in the field by their small-scale structures, e.g. planar and linear fabrics, folds and related minor intrusions. The superimposition of these small-scale structures provided the field evidence for the temporal relationships of the tectonic episodes. Early events were homogeneous over the whole map-area, whereas later episodes were variable with respect to the regional distribution of their related tectonic fabrics. Therefore tectonic domains (Hepworth, 1967) can be identified (Figure 4) where one, usually late, tectonic episode is dominant. Individual tectonic domains are characterised by a consistent trend of structural elements, which produce a pronounced surface 'grain', allowing domain boundaries to be defined on aerial photographs and satellite imagery.

7.1 PRE-METAMORPHIC EPISODES

The polyphase tectonothermal history has effectively obliterated all early, non-metamorphic fabrics. Possible sedimentary bedding is preserved in calc-silicates at GR 2650 1825 and igneous layering is present in an altered ultrabasic at GR 2690 1806. In both cases the primary fabrics are at a high angle to the metamorphic fabrics. Discrete lithologies form wedges rather than continuous bands and there is no evidence to suggest that gneissosity exactly parallels primary fabrics. In other words, very little can be stated about the palaeogeography of the initial sedimentary and volcanic manifestations. Similarly there is no sign of the 'cold' basement which floored these supracrustals. All the metamorphic rocks presently exposed appear to have undergone the same essential tectonothermal history.

7.2 SAMBURUAN EPISODE

The progressive, Upper Amphibolite Facies, regional metamorphism which produced the coarse metamorphic fabrics (gneissosity, early migmatites) is referred to as the Samburuan (Hackman and others, in press). Initially the event was probably static producing subhorizontal gneissic fabrics with physical separation of unorientated felsic grains and shape-orientated mafic minerals.

The metamorphic grade in the several kilometres thick vertical section presently exposed (from the Siambu Complex down to the Ndura Complex) is uniformly in the Upper Amphibolite Facies with an increase in K-metasomatism downwards (more microcline in basal units). Immediately to the north there are contem-

Table 10 Summary of the tectonothermal history of the metamorphic rocks

Episode	Approximate age (from Rundle, 1983)	Tectonic elements	Metamorphic grade	Associated Igneous activity
	Ma			
'Late'	580–450	Brittle fractures and shears of variable orientation with a pronounced ENE set; open folds and asymmetric warps with NE and E trends to axial traces	Greenschist Facies (post tectonic almandine growth in SE)	
Barsaloian	580	Major shear zones trending N with internal parallel structures: tight folds plunging gently N; mylonitic S surfaces. Open folds outside shear zones plunge gently N	Amphibolite Facies	Local remelting to produce minor veining
Baragoian	630	Regional synforms and antiforms with NW axial traces and moderate plunges to the NW (and rarely to the SE). Parasitic folds. New gneissic fabrics in cores of regional folds	Amphibolite Facies	Local melting to produce various vein phases
Sabachian	830 (culmination of Samburuan episode)	Subhorizontal fabrics: recumbent to overturned folds (including nappes), thrust complexes. Axial traces trended and plunged gently SW. Parasitic folds	Amphibolite Facies	G2 granites
Samburuan		Coarse gneissic fabric—subhorizontal. Minor folding. Earlier fabrics obliterated by Samburuan structure	Upper Amphibolite Facies	G1 sodic intrusives, minor vein phases and partial melt concentrations

poraneous Granulite Facies assemblages in the felsic gneisses (Dodson, 1963; Key, 1987). Typical mineral assemblages defining the gneissic fabrics are as follows (see Chapter 3).

Sillimanite + K-feldspar + quartz + oligoclase + biotite + garnet
Sillimanite + muscovite + quartz + K-feldspar + epidote
Hornblende + biotite + quartz + K-feldspar + epidote
Hornblende + plagioclase + quartz + garnet
Hornblende + epidote + plagioclase + diopside + quartz
Carbonate + quartz

No metamorphic orthopyroxene was seen and clinopyroxene is rare. Secondary mineral growth related to the late metamorphisms is ubiquitous, e.g. K-feldspar overprints, recrystallisation of the common fabric-forming minerals (quartz, plagioclase, biotite and hornblende).

Load pressure must have played an important controlling role in the formation of the mineral assemblages: it is suggested later that the present exposure levels represent depths of over 12 km (pressure of over 4 Kbars).

The role of shearing stress is not known, but was critical in the Sabachian episode which directly succeeded the Samburuan. No contemporaneous major plutons are recognised at present (root zone) exposure levels. The Samburuan episode must be related to a major mantle thermal anomaly. The mineral assemblages suggest temperatures of about 650°C at pressures of about 5 Kbars (Pattison and Harte, 1985).

7.3 SABACHIAN EPISODE

The episode which produced subhorizontal tectonic fabrics in the gneisses and migmatites, and which appears to have formed immediately after the generation of the Samburuan fabrics, is referred to as the Sabachian (Hackman and others, in press). Again, it was a regional event although its structures are generally modified or completely masked by later events.

Overturned or recumbent folds (including nappes) related to low angle thrusts were generated (Plates 13 and 17), with tectonic transport from the north-west. Still recognisable major folds include the Masiketa dome (GR 271 169), recumbent folds east of Kawab, Kotikal basin, folded southern tip of the Siambu Complex (GR 2715 1200) and the recumbent fold closure in the western limb of the Morilem Synform. Rotation of fold axes to compensate for the effects of subsequent folding indicates that they plunged at about 20° towards the SW. A distinctive feature of the Sabachian linear fabrics is their variable plunge trend (due to subsequent rotation) compared to the very regular plunge directions of later lineations.

Parasitic folds are ubiquitous in all good outcrops and are prominently displayed by the *lit par lit* felsic veins in the Ndura Complex. They are tight to isoclinal, similar folds with one or both limbs strongly attenuated. In the stromatic migmatites the fold limbs may completely disappear to leave rootless folds outlined by mafic pods.

The major thrusts (shown on the geological map and sections) are rarely exposed, but minor dislocation planes are visible in most outcrops. They cause the wedging of lithological beds and individual gneissic layers as well as attenuation of fold limbs.

On a large scale the thrusting juxtaposes massive migmatites and muscovite-bearing gneisses (e.g. at GR 2766 1591). Any flagginess in the major quartzo-feldspathic gneisses is due to Sabachian thrusting: the lateral continuity of the graphitic gneisses may result from their preferential utilisation as movement zones (the graphite acting as a lubricant).

Sabachian planar (S) fabrics are parallel to gneissosity which is therefore always a composite fabric. Linear fabrics are strongly developed in minor and major fold hinge zones and down dip on most dislocation surfaces. They are stretch lineations parallel to the movement direction of the thrust plates (cf. Shackleton and Ries, 1984). A typical Sabachian SL fabric had axes of 1:6:8.

The metamorphic grade remained in the Amphibolite Facies throughout the Sabachian deformation: the G2 granites were emplaced towards the end of the episode. They are crustal melt intrusives whose upward ascent was controlled by the Sabachian structures. They locally have a Sabachian foliation and their age of 828 ± 33 Ma (Rundle, 1983) provides a minimum data for the Samburuan and Sabachian episodes.

The Sabachian episode was not a thin-skinned tectonic

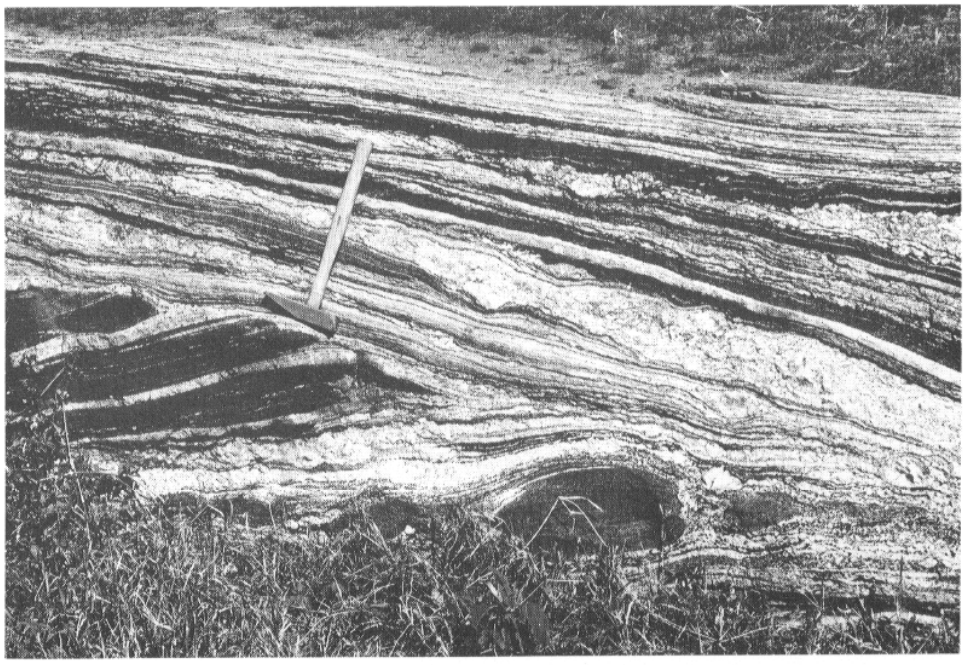


Plate 13 Sabachian thrust, Marti area

poraneous partial melting seen in the quartzofeldspathic gneisses of the Loroki Gneisses (described in Chapter 3), represent exhumed allochthonous root zones to the G2 granites.

G2 granites were initially recognised and named during mapping of the first Degree Square (36) of the SMP (Hackman and others, in press). Within the group, individual plutons may be named: the Sartim Granite (GR 266 220), Nakefenda Granites (GR 240 212), Luwamara Granite (GR 2725 1790), Suyan Granite (GR 2663 1591), Lawai Granite (GR 2615 1562), Lekima Granite (GR 254 148) within Degree Square 27. All these plutons intrude the Siambu Complex, although elsewhere they are recognised in other gneiss and migmatite units.

The decisive factor in the location of the G2 granite plutons appears to be their tectonic setting: many occur at high tectonic levels in the crests of antiforms. At lower crustal levels, e.g. in the Ndura Complex (at GR 2410 1117) they form sheets concordant with the gneissosity.

The G2 granites form major inselbergs and hills. A single massive positive landform may encompass the whole of one pluton and rise several hundred metres above the surrounding countryside. Considerable internal variations in relief are caused by preferential weathering along master joints. The granites have pale phototones in contrast to the mafic gneisses of the Siambu Complex: granite contacts are clearly visible on aerial photographs. The large plutons are also outlined on satellite imagery.

Exposures are ubiquitously good; reference sections may be defined in any of the six named plutons. For ease of access the Luwamara Granite would be the best reference unit, as its outcrops straddle the track from Masiketa to Latakweny.

Rundle (1983) obtained a Rb-Sr whole rock isochron age of 828 ± 33 Ma for a G2 granite from the type area in Degree Square 36. This dates the initial emplacement, although several of the G2 granites in Degree Square 27 have undergone subsequent mobilisation. The high initial Rb/Sr ratio derived from the isochron confirms its inferred crustal source.

G2 granite contacts are always sharp, with no sign of chilling of either the granite or country rock. Granite offshoots into the country rock cut gneissosity and early isoclinal folds of gneissosity (Plate 16). In turn the granites are deformed by the Baragoian and Barsaloian tectonic events to produce a wavy foliation. The Nakefenda Granites are deformed to augen gneisses within a Barsaloian straightening zone. The Sartim Granite has a migmatitic envelope around a central para-autochthonous granite stock. Gneiss xenoliths are confined to certain parts of individual plutons, where they are locally extremely abundant.

The G2 granites vary in size from less than 5 m thick for the small intrusion at GR 2455 1415 to the Lekima Granite whose oval plan covers an area of about 70 km^2.

Freshly broken granite surfaces are white, but most outcrops are pale yellow, orange or dirty brown-pink in colour. The granites are medium- to coarse-grained, massively homogeneous, although locally gneissic with streaks, wisps or flecks of dark minerals (usually biotite). As noted above if a pluton is subsequently deformed within a Barsaloian shear zone, cataclastic fabrics are developed. The granites are leucocratic rocks, locally with a patchy development of secondary alkali feldspar and give high, locally anomalous, scintillometer counts. The Suyan Granite or its remobilised offshoots appear to cause an airborne uranium anomaly (Chapter 9). Granitic (and aphanitic) offshoots cut the country rock and pegmatite veins are clearly defined within individual plutons. The principal minerals (quartz, K-feldspar, plagioclase, biotite, magnetite), are readily visible in hand specimen with occasional plane crystal faces to the felsic grains (Shackleton, 1946). Dark (biotite and hornblende) schlieren accompany gneiss xenoliths.

Most specimens examined in thin section had xenomorphic, inequigranular textures. Quartz, as lobate grains, and ragged anhedral oligoclase laths are variably replaced by large microcline plates (which also forms primary intergranular grains). Myrmekitic zones form at the interface of oligoclase and replacive microcline. Plagioclase is also locally altered to sericite or muscovite plates. Individual biotite flakes may show alteration to chlorite. Not all contacts are replacive, cf. the plane crystal faces noted in outcrop.

The relative proportions of the three principal felsic minerals vary greatly; any of the three may be dominant. Biotite is the usual mafic phase and magnetite the iron oxide phase. Hornblende, apatite, sphene, zircon, epidote and muscovite occur in accessory amounts. Red garnet is extremely localised but may be present in significant quantities as small cracked anhedral grains.

The G2 granites are crustal partial melt products sourced in quartzofeldspathic gneisses: their emplacement was, in most cases, structurally controlled, at about 828 Ma, following the imposition of gneissic fabrics on the country rocks.

6.5 'LATE' FELSIC INTRUSIVES

Linear aplite, granite, pegmatite and quartz dykes and veins are well developed in the NE quadrant. They do not have a constant orientation, as they utilise all available open planar fabrics. However many trend roughly NW to follow a Baragoian fracture cleavage related to the Morilem Synform.

The aplites are fine-grained, off-white, mottled by flecks of opaques, and occur in swarms of narrow (about 1 m wide) but continuous (up to 2 km strike length) dykes. They form low ridges which are readily visible in the field and on aerial photographs, as they cut across the regional gneissosity trend. They may have a faint foliation parallel to their steeply dipping walls. Silvery muscovite flakes are ubiquitous in small amounts in a groundmass of quartz, sodic plagioclase and microcline with minor biotite and secondary sericite, calcite, chlorite and epidote. Quartz is the principal modal phase and may form euhedral phenocrysts (the quartz porphyries of Shackleton, 1946). Plagioclase (partly replaced by microcline) is the principal feldspar phase.

Late granite sheets are relatively rare although Baker (1963) records both irregular 'hornblende-granite' sills and discordant leucocratic biotite-granite dykes. They may have a weak internal foliation parallel to their margins. The 'hornblende-granites' are mainly composed of microcline grains with interstitial sodic plagioclase, quartz, hornblende, sphene and rare biotite and epidote. Leucocratic granite dykes contain modal muscovite in a felsic groundmass of roughly equal amounts of quartz and K-feldspar (orthoclase and microcline) and lesser oligoclase and biotite.

Several generations of pegmatites are recognised, including late linear pegmatites, as well as earlier deformed veins related to G2 granites or migmatisation processes. The permeation and segregation pegmatites of

Baker (1963) are examples of the earlier intrusives. Late intrusive pegmatites (Baker, 1963) are the most abundant and vary in size up to widths of about 30 m, with strike lengths of several kilometres. They are sheet-like with parallel walls and have sharp contacts with country rocks. The larger pegmatites are zoned (Baker, 1963): cores of pure, massive quartz are flanked by coarse feldspathic zones. Outermost wallrocks are of fine-grained quartz and feldspar with mica and/or garnet. Numerous other mineral phases have been identified, but always in minor amounts. These include radioactive minerals (euxenite, columbite, pitchblende, samarskite, monazite), notably in the Nachola area, in pegmatites that utilise Baragoian fractures; green beryl, aquamarine and schorl as fractured euhedral crystals embedded in the massive quartz core zone; and magnetite.

An important swarm of pegmatites trends SSE from Nachola through Nagiss and Lorregel as far as the Masiketa River. Large pegmatites are also present further south towards Marti at GR 2383 1587; GR 2423 1513; GR 2398 1764; GR 2415 1868; GR 2460 1930; GR 2423 1693; GR 2438 1680.

Quartz veins are poorly exposed, and generally only identified by a relict milky quartz float. For the most part they are entirely composed of quartz although tourmaline, red garnet and epidote occur in veins at GR 2680 1530, GR 2594 1467, and GR 2618 1553 respectively. A siliceous breccia infills a fault zone at GR 2269 1571 which trends towards the NW. It is about 2 m thick with country rock (mafic gneiss) clasts embedded in a fine siliceous matrix.

6.6 CENOZOIC INTRUSIVES

Basalts

Basalt dykes, sills and plugs are recognised as feeders to the various basaltic lithostratigraphic units of both Rift shoulders. However similar intrusives are rare in the median graben: a single basalt plug cuts the south end of Namarunu Volcano. Two major dyke swarms are recognised (Truckle, 1977a) in the NW quadrant, cutting the Lomujal and Nathelot basalts. The older Lomujal swarm has a pronounced NNE trend with associated plugs, cone sheets and two sills. The dykes can be traced for about 9 km in a NNW direction (their southern extent is masked by younger Tirioko basalts).

The younger Nathelot swarm is only exposed in basal lavas of the Nathelot basalts in the confined area north of Nathelot: short dykes trend NNE, NNW and ESE. Truckle (1977a) regards the dykes as feeders to the upper lavas of the Nathelot basalts. There was a clearly defined tectonic control on basalt dyke emplacement; most follow NNE-trending faults. Lithologically, petrographically and chemically the minor basalt intrusives are identical to their extrusive lava phases.

Trachytes

Trachyte dykes and irregular stocks (rarely cone sheets and sills) are prominent components of the central zones of the trachytic volcanoes of both shoulders, but are apparently absent from the median graben. These intrusives locally cut basaltic units (Lomujal, Nathelot and Tirioko basalts) as dyke swarms radiating from adjacent trachyte centres.

Within the trachytic volcanoes the intrusive trachyte plugs and dykes fed several generations of extrusive lavas to span the life of the individual volcanoes. The string of isolated plugs south-east of Silali represents cores of small eroded volcanoes (Golden, 1978) and are of late Miocene age: Nakali (7.4 ± 0.2 Ma), Lokapomye (6.9 ± 0.3 Ma). The irregular shapes of many of the trachyte plugs largely reflect variations in wall rock: plugs penetrate further into softer pyroclastics, often forming elongate apophyses but are severely restricted by massive lava flows. The stocks are mostly several tens of metres in diameter, although an 800 m diameter stock is exposed near the summit of Kafkandal Volcano. The dykes may have a flow related fissility and any sediments are baked immediately adjacent to the trachyte intrusives. Dykes were intruded along fractures—either NNE faults (e.g. Oliyamur and Kajyamamuk Volcanoes) or summit ring fractures.

The trachyte intrusives are identical mineralogically and chemically to their related extrusive phases.

Phonolites

Phonolite intrusives are rare and are either confined (as scarce feeder dykes and plugs) to the eastern phonolite units or in the NW quadrant as several small plugs and a single circular dyke.

Lithologically the eastern intrusives are identical to their host lavas, which makes them difficult to identify in the field. Within the Lopet Phonolites the dykes locally form ridges up to 2 km in strike length. The plug at GR 2220 1260 in the Losiolo phonolites has a slightly coarser grain size than adjacent lavas which gives it a rougher weathered surface to produce a very distinctive phototone.

The minor plugs in the north-west occur at Morualem (GR 1855 2110), with three plugs immediately south-west of Kailimerlim. They all form low rises surrounded by unconsolidated Quaternary debris, and (together with the adjacent foyaite) probably formed the final phase of Kailimerlim Volcano.

Foyaite (from Truckle, 1977a)

Within a NNE line of circular foyaite intrusives in the NW map corner the southernmost Kailimerlim plug (GR 177 208) is the largest. It underlies a major hill, rising 210 m above the adjacent Akapangalei plain, to cover an area of 3 to 4 km^2. The Turkana vernacular name refers to its mottled appearance with alkali feldspar and nepheline phenocrysts.

The intrusion is lithologically varied, expressed as differing proportions of phenocrysts, types of phenocryst, absence of phenocrysts, groundmass grain size. Altogether Truckle (1977a) identified six syenitic rock types within the plug. Generally the foyaite is coarsely crystalline with randomly orientated textures, although layers of aligned feldspars may impart a strong fluxion banding (at GR 1758 2085 and GR 1772 2089). Angular phonolite blocks are confined to the oldest and outermost phase where phonolite veins up to 15 cm wide also cut the same syenite.

The Kailimerlim Foyaite intrudes the north-west part of the largest exposed part of Kailimerlim Volcano. The other two, small, foyaite intrusives occur at Kachangal (GR 1775 2160) and at GR 1773 2135. They are surrounded by alluvium but are within the northern area of Kailimerlim Volcano.

process in Degree Square 27, as it incorporated the lower migmatites as well as the various gneisses. The sub-horizontal attitude of its structures may be compared with root zone structures of other major thrust belts outlined by geophysical profiling. In particular the work of Brewer and Smythe (1984) on the Moine Thrust suggests depths in excess of 12 km for these flat zones.

A major erosional break followed before the end-Precambrian on-set of vertical tectonic regimes recorded by the structures of the Baragoian and Barsaloian episodes.

7.4 BARAGOIAN EPISODE

This areally restricted tectonothermal episode was first recognised in the south-east part of the SMP area (Hackman and others, in press). However, from the satellite imagery it was quickly established that this episode's structures are best preserved in the Baragoi area (notably the Morilem Synform of Baker, 1963) and was consequently named the Baragoian. Figure 4 shows the distribution of the Baragoian domains in Degree Square 27. Along the eastern map edge and adjacent to the eastern Rift shoulder these domains are overprinted by Barsaloian structures.

Major, upright, cylindrical folds whose axial traces trend and plunge NW to NNW are the principal Baragoian structures: the domains of Figure 4 correspond to the Morilem Synform, Karisia Hills Antiform and the adjacent (to the south-west) synform. Large parasitic folds were identified and named by Baker (1963), all of which refold recumbent Sabachian folds. Similar parasitic folds are also recognised within the Karisia Hills Antiform, with wavelengths of up to 10 km. Locally these parasitic folds are replaced by axial planar shear zones.

Enveloping surfaces to the folds are simple but confine complex internal structures (cf. Baker, 1963 and see Plate 14). This is because the Baragoian episode folded essentially horizontal structures on a large scale but smaller heterogeneities, such as parasitic Sabachian fold closures were refolded to produce small-scale interference structures.

The small-scale parasitic Baragoian folds have variable styles from open to tight, symmetrical to asymmetrical, with or without an axial planar fabric. In major fold closures, local melt products were injected along axial planar foliations to define new gneissic fabrics. In the Masiketa River outcrops, at the core of the Morilem Synform, it is possible to see this new coarse fabric cutting the original gneissosity. Other planar fabrics include the NNW- to NW-trending brittle fractures infilled by the various felsic veins and dykes (Chapter 5). Lineations (rods) are confined to fold closures (M zones) and plunge at up to 25° towards the NNW or NW. Plunge reversals, e.g. in the north-west part of the Morilem synform are due to later warping. Typical Baragoian SL fabrics have axial ratios of 1:7:20. Minor slip along Baragoian planar fabrics is demonstrated by displacement of discordant felsic veins (Baker, 1963).

The tectonic events were accompanied by an Amphibolite Facies metamorphism in rocks now exposed with recrystallisation of quartz, plagioclase, hornblende and biotite to define the new fabrics. Local melting formed the gneissosity mentioned earlier and small areas of migmatite (discussed by Baker, 1963, for the Baragoi area and around GR 2698 1605 near Barsaloi). There was a strong structural control on the melting (confined to major fold closures).

The pre-Rift topography was strongly influenced by the Baragoian structures with the Karisia Hills and Nagoriagwen inlier forming major positive features along the axis of a Baragoian antiform.

The Baragoian episode was controlled by regional horizontal compressive stress trending roughly ENE to produce the major folds of Degree Square 27.

7.5 BARSALOIAN EPISODE

Satellite imagery of central Kenya reveals a major N–S zone of closely spaced parallel lineaments, the north-west part of which just extends into the eastern side of Degree Square 27 (Figure 4). Because of good exposures in the Barsaloi River where it cuts right through this zone the deformation which produced the N–S structures is

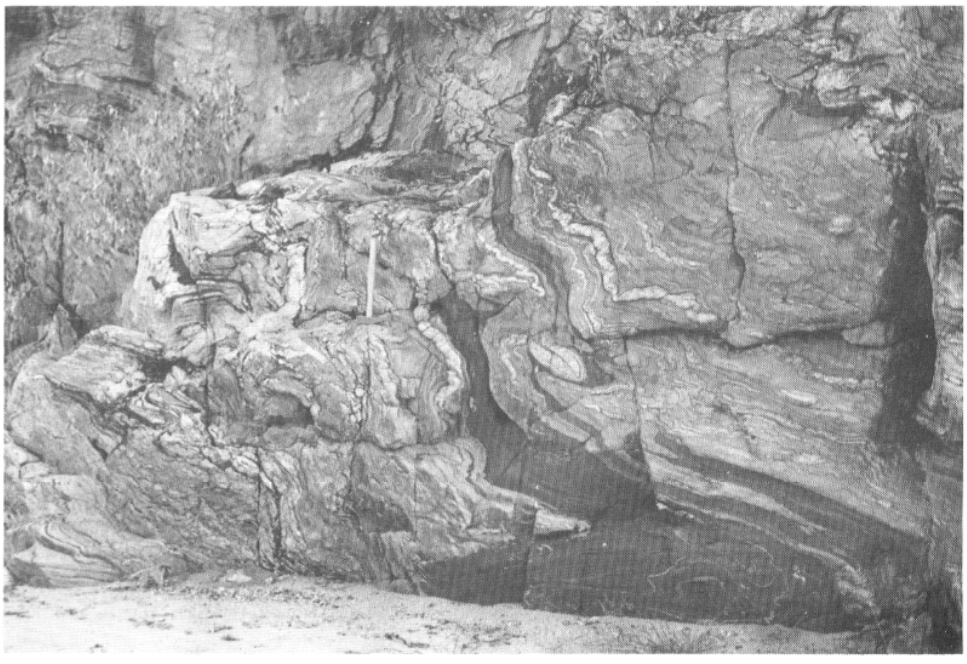


Plate 14 Interference folds in migmatites exposed in the Nagoriagwen inlier

referred to as the Barsaloian (Hackman and others, in press). A smaller, parallel shear or straightening zone (Hepworth, 1967) also cuts Baragoian structures near the eastern Rift shoulder and extends into the Nagoriagwen inlier.

Within the shear zones (the eastern zone is about 20 km wide) all pre-existing planar fabrics are transposed into vertical N–S orientations. Mylonitic fabrics are common with numerous discrete shear surfaces, pods of augen gneiss, intrafolial folds and near perfect isoclinal folds with wavelengths of up to 4 km. Typical axial ratios of S fabrics are 1:10:10. Rodding lineations are also common with very gentle plunges to the N and less commonly to the S. Interference folds are absent in the shear zones because of the transposition of pre-existing fabrics into N–S orientations.

Outside of the shear zones the Barsaloian deformation produced minor open folds which tighten towards the shear zones. The folds retain their N–S axial traces and a very gentle plunge (to N and S). During the Barsaloian, flattening of planar fabrics against the Luwamara Granite took place (adjacent to the major shear zone).

However the Baragoian felsic veins and dykes in this area were not deformed. Further south a Sabachian fold closure is preserved at the south end of the Siambu Complex which was refolded during the Barsaloian to define a major interference fold (section EF on the geological map).

There is no evidence for strike-slip movement within or adjacent to the Barsaloian shear zones. Baragoian S surfaces can be traced into the main Barsaloian shear zone at GR 277 190 with no lateral offset. Therefore it is concluded that shear zones reflect lateral compression (pure shear), i.e. the N–S lineations do not reflect movement direction (compared to the earlier Sabachian lineations which were parallel to movement directions). It is thought (see also Hackman and others, in press) that the same stress field controlled both the Baragoian and Barsaloian episodes: initially producing the upright folds, with subsequent shearing along attenuated fold limbs. Otherwise it is difficult to envisage how the discrete Barsaloian shear zones, separated by gently folded terrains, could have developed.

An Amphibolite Facies metamorphism again accompanied the tectonism in rocks presently exposed, locally with new sillimanite growth defining the tectonic fabrics. Further south granite injection accompanied local melting with the intrusives dated at about 570 Ma (Rundle, 1983).

7.6 LATE EPISODES

A post tectonic almandine growth in all lithologies in the south-east corner may be related to a major basic intrusion concealed beneath the surface. Similar almandine growth was noted next to the metagabbro at GR 2608 1315. Otherwise the basement rocks have suffered no further metamorphic mineral changes since the Barsaloian.

Post-Barsaloian open folds and fractures of several generations reflect the waning and post-orogenic period of erosion and crustal adjustment. They were of considerable importance in controlling landscape development.

An open warp with an axial trace extending north-east across the El Barta Plain caused the plunge reversal in the Morilem Synform. The resultant depression preserved the (remnant) El Barta Plain from Quaternary headwater erosion. Parallel open warps with sheared out crests in the Nagoriagwen inlier (GR 2406 1695) are on strike with the El Barta Plain depression. At the south end of the Karisia Hills there are upright, open folds (two synforms separated by an antiform) which plunge at about 30° to the SW. These folds are cylindrical with wavelengths of about 6 km. Other late folds include the low amplitude flexures with E–W axial traces in the Nagoriagwen inlier (GR 2404 1402) and parallel warps further south at GR 2333 1429 and GR 2419 1510.

Late brittle fractures (and rare shears) show a complete orientation range. Large ENE fractures are important controls of drainage systems—the Barsaloi River follows one such fracture in the extreme east. Many fractures also cut the Tertiary volcanics and are related to the Rift Valley tectonic regime.

Uplift and erosion continued well into the Palaeozoic (indicated by blocking temperature mineral dates, Cahen and others, 1984) and possibly until the Mesozoic marine transgression across East Africa.

7.7 DISCUSSION

The Samburuan and Sabachian episodes

The main effect of these two episodes was to impart a subhorizontal planar fabric within the Mozambique Orogenic Belt. Folding involved both gneisses and basal migmatites: tectonic interleaving of gneiss and migmatite resulted from transport of the migmatites along low angle dislocation planes.

Axes of Sabachian folds (including nappes) were orientated roughly SW, i.e. *transverse* to the assumed N–S trend of the orogen. In this respect north-central Kenya is typical of much of the Mozambique Orogenic Belt further to the south where early transverse structures are the main tectonic imprint (see Cannon and others, 1969; Hepworth, 1967; Hepworth and Kennerley, 1970; Johnson, 1968; MacFarlane, 1969; Sanders, 1965; Weiss, 1959; and the summary in Shackleton and Ries, 1984).

Hepworth (*op. cit.*), in the most comprehensive analysis of the structural evolution of the Mozambique Orogenic Belt, suggests that these early transverse structures (Kondoan) were formed during a period of extension across the E–W width in Tanzania of the orogen. Comparisons may be made between the Kondoan and Sabachian: both produced recumbent fold structures on parallel axes. Recent work on thrusts and nappes (Dahlstrom, 1969, 1970; Elliot, 1976a & b) emphasises the role of gravity as a major stress component.

However, in the Samburu region the role of gravity still has to be considered in the context of assumptions about orogenic mechanism and the thermotectonic sequence: the involvement of a sequence with ophiolitic affinities (the Siambu Complex) implies an early phase of crustal separation.

The temporal sequence of gneisses followed by syntectonic migmatites and granites for the Samburu–Sabachian episodes has been recorded in many high-grade terrains, and can be explained in terms of the role of the incipient fluid phase (see e.g. Brown, 1983; Powell, 1983; Thompson, 1983; Holt and Wightman, 1983; Touret and Dietvorst, 1983).

Rocks undergoing Amphibolite Facies metamorphism contain a CO_2-rich fluid with an ability to remove 'low field strength' elements (K, Rb) to deposit them (metasomatically) at higher crustal levels. Partial melting of the precursor gneisses at the prograding front produced

migmatites, providing a mechanism for the post-gneissification generation of the G2 granites and migmatites.

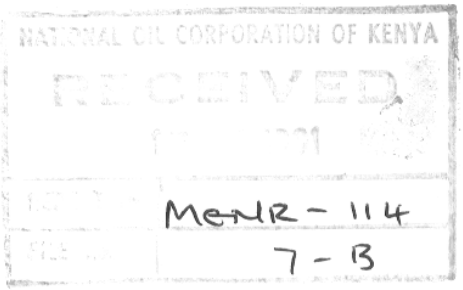
Streaming of the (devolatilisation) fluid may be linked to the generation of large scale shear zones in Amphibolite Facies rocks (cf. Thompson, *op. cit.*), controlling the evolution of the Sabachian structures. The graphite gneisses commonly form movement zones during the Sabachian and these rocks, and the ultrabasics of the Siambu Complex, were a possible source of CO_2-rich fluids. The G2 granite age of about 828 Ma agrees well with dates of about 850 Ma for the early metamorphisms recorded in the Mozambique Orogenic Belt throughout east and south-east Africa (Cahen and others, 1984).

The Baragoian and Barsaloian episodes

During the Baragoian episode upright regional folds plunging to the NNW were produced, presumably due to ENE–WSW directed compression: the Baragoian represents a period in which the Mozambique Orogenic Belt closed in northern Kenya.

During the Barsaloian episode the vertical 'straightening zones' appear to have preferentially developed on the limbs of the major Baragoian folds. It is suggested that the Barsaloian domains formed at the end of the Baragoian episode due to the inability of the rocks to continue to react to the stress field by forming open folds with axes plunging NW.

Shackleton and Ries (1984) argue that recorded lineations, from their map essentially associated with Baragoian and Barsaloian events, define the stretch direction: the implication is that they are signatures of rigid plate obduction towards the south, i.e. *parallel* to the assumed orogenic trend. The dynamics might be integrated geometrically with the concept of a Samburu–Sabachian E–W-trending orogen: however, if the structures in question are in fact Baragoian–Barsaloian, they could be much younger, a response to an E–W stress regime constraining the N–S orogen as it is conventionally conceived. The apparent orthogonal change in orogenic trend still begs an explanation, requiring clearer statements on the tectonic level at which different parts of the 'stacked sequence' were mobilised, and better definition, in terms of age determinations, of the earlier phases of the orogenic sequence.


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8 Cenozoic tectonics

8.1 EASTERN SHOULDER OF THE GREGORY RIFT

The principal structural elements of the eastern shoulder are monoclinal downwarps facing westwards towards the median graben and high-angle normal faults. Several generations of downwarping are distinguished by angular discordance between various stratigraphic units (documented in Chaper 4). Of these, the Samburu Monocline, first identified by Shackleton (1946), and Quaternary downwarping along the margin with the Suguta Trough, are the most important. The Samburu Monocline is principally defined by dip variations in the various Miocene flood phonolites and basalts. Its NNE-trending axial zone passes just west of Losiolo and continues south-westwards into the Putere–Kamuga Lineament (Golden, 1978). South of 1°N this lineament strikes into a series of fault blocks. As a result of the downwarping the phonolites and basalts adjacent to the median graben have dips of up to 45° westwards.

Quaternary downwarping is defined by the northern trachytes. These are flat-lying except in the marginal zone to the median graben where they dip gently westwards. Therefore the Quaternary monoclinal axis is further west than the Miocene fold axis (Figure 18).

Several ages of normal faults can be distinguished, although the middle to late Miocene faults predominate and tend to conceal earlier faults. Shackleton (1978) and Golden (1978) suggest an early to middle Miocene fault set confined to the Samburu basalts, without indicating the fault trend. Major NW- and WSW-trending lineaments recognised on satellite images of the eastern shoulder (Figure 18) can be related to (buried) faults, which, in part, controlled the deposition of the flood phonolites. Hackman (in press) has recognised these faults to the south: locally they show a sinistral strikeslip component. They pass laterally into late (Loldaikan) basement shears which were clearly reactivated in Miocene times.

All the middle to late Miocene faults are high-angle, normal faults, mostly trending in the arc N to NNE with shorter complementary faults trending roughly ENE. They have sinuous traces, in part due to subsidiary strikeslip movement (King, 1978). Individual faults can be traced for about 10 km along strike, although *en echelon* faults are commonly linked over strike distances of up to 40 km in an anastomosing network. The density of these faults increases markedly towards the inner margin of the eastern shoulder. Isolated, antithetic, eastern faults have minor downthrows eastwards, whereas synthetic western faults downthrow up to several hundred metres towards the median graben. Back-tilting (cf. Griffiths, 1980) of

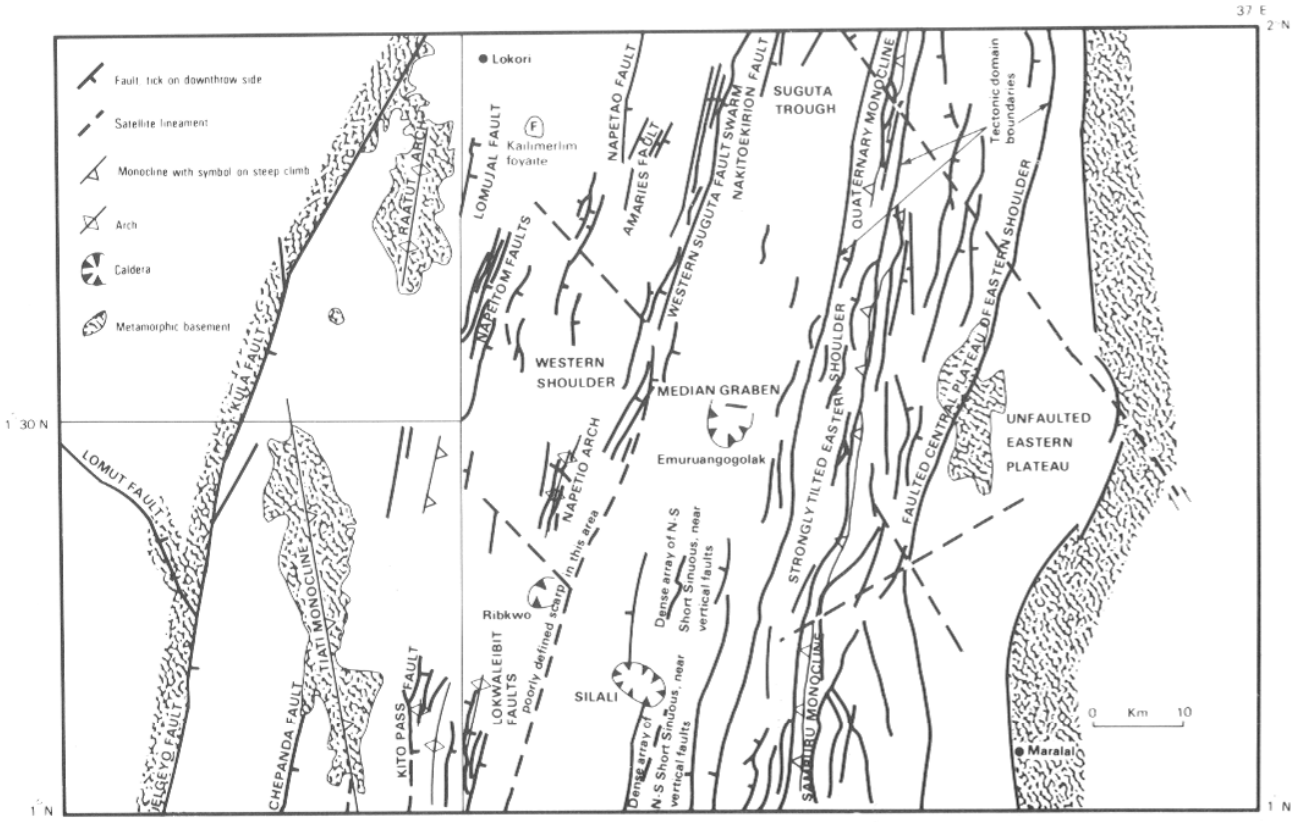


Figure 18 The structure of the Gregory Rift between latitudes 1° and 2°N

fault blocks was common and formed a series of horsts and grabens across the eastern shoulder (see map cross sections).

Quaternary faulting is confined to the immediate vicinity of the median graben. These faults trend NNE: most downthrow to the west. The larger faults define a series of steps which drop the eastern shoulder into the median graben. Several minor antithetic faults are shown by Baker (1963). Major landslipping is presently taking place into the Suguta Trough along curved (concave upward) slip planes utilising the Quaternary faults. Golden (1978) notes that this process, which is akin to normal faulting (King, 1978) has probably continued throughout the structural evolution of the eastern shoulder.

Most joints in the volcanic rocks are primary, cooling-fractures although larger NW, NE and E sets are related to the downwarping and faulting (Wilkinson, 1983a & b).

The structural history of the eastern shoulder is summarised in Table 11. Three tectonic domains can be distinguished (cf. Golden, 1978): an essentially undeformed outermost zone of flat-lying volcanics; a central plateau area, about 18 km wide, of tilted fault-blocks defining shallow horsts and grabens; and an inner zone, about 7 km wide, of densely faulted and steeply-downwarped, westward-dipping volcanics. The Nagoriagwen inlier of metamorphic rocks is due to a 'basement' tectonic episode; it forms the core of a Baragoian antiform, and is not due to Cenozoic arching.

8.2 WESTERN SHOULDER OF THE GREGORY RIFT

Between longitudes 1°N and 2°N, the western shoulder of the Gregory Rift has a constant width of about 48 km between the NNE-trending Elgeyo–Kula Fault set and the fault-controlled scarps bounding the median graben (Figure 18). At 1°N, the Elgeyo Fault lies about 40 km to the west of Degree Square 27; its northern extension, the Kula Fault, approaches to within 4 km of the NW map corner.

The main structural elements of the western shoulder are steeply dipping normal faults, monoclinal downwarps facing eastwards and more symmetrical arches together with major dyke swarms (McClenaghan, 1971; Webb, 1971; Weaver, 1973; Truckle, 1977a).

The normal faults have gently sinuous trends, with a regional arcuate pattern: they trend NNE in the northern half of Degree Square 27 and N–S in the south. Most faults are synthetic, to downthrow eastwards, although antithetic faults locally help define horsts (or arches), e.g. the Lokwaleibit Faults. The faults tended to migrate eastwards with time (see aforementioned references): Quaternary faulting is restricted to, and defines, the boundary with the median graben. However reactivation of the older, western faults occurred, e.g. for the Lomujal and Napeitom Faults. Throws on individual faults vary from the cumulative displacement of about 3500 m on the Elgeyo Fault (King, 1978) to less than 100 m on many of the smaller faults in Degree Square 27. Cumulative downthrows eastwards on the major Miocene faults in Degree Square 27 are as follows: 750 m (Napeitom Faults); 245 m (faults related to the Napetio Arch); 250 m (Lomujal Fault).

Vertical movements on Pliocene faults were less spectacular, although up to 100 m displacement is proved on faults cutting Ribkwo Volcano. Two periods of minor eastward downthrow are recognised on the Amaries Faults (Truckle, 1977a). A dextral strikeslip component within the arcuate network of faults of the western shoulder, with progressive location of extension zones northward, could account for the NNE migration of the trachytic volcanoes throughout the Pliocene. The Quaternary faults are described in Section 8.3 as they relate directly to the median graben.

Truckle (1977a) provides enough detail of the structural history of the northern part of the western shoulder to prove an almost continuous eastward tilting ('creep') since the early Miocene. However the amount of tilting was not consistent: Quaternary downwarping, confined to the faulted scarp zone adjacent to the median graben, was the most intense (dips of over 30°). This imparted a gross monoclinal structure to the western shoulder. The major Miocene monoclines (consistently facing eastwards) are to the west of Degree Square 27. However, there was a progressive eastward migration of the downwarp axes: a proto-Suguta Trough was formed at about 3 Ma by gentle downwarping of Loriu Plateau trachytes. Local westward dips are due to fault-controlled back-tilting: west of the Napeitom Faults and on the Loriu Plateau west of the Western Suguta Fault Swarm.

The major dyke-swarms either trend NNE, commonly along crests of arches: or form radial arrays related to domal uplift accompanying volcanism.

NW-trending lineaments, recognised on satellite imagery, may relate to concealed Miocene faults (cf. Section 8.1). The two lineaments, shown on Figure 18, strike into the median graben towards Emuruangogolak and Silali, to suggest that deep faults controlled the location of these central volcanoes. (Note the NW trend to the long axes of their summit calderas.)

Landslipping has occurred on many fault scarps. The structural history of the western shoulder is summarised in Table 11.

8.3 MEDIAN GRABEN OF THE GREGORY RIFT

The median graben has a constant width of between 20 and 30 km and a persistent NNE orientation in Degree Square 27. Its present floor slopes north-north-eastwards from about 800 m (at 1°N) to about 300 m (at 2°N). This could either reflect an underlying tectonic control related to the gross structure of the Gregory Rift (Kenya Dome; King, 1978) and/or more substantial infilling by volcanics and sediments in the south. The present morphology of the median graben can therefore be likened to a major ramp (Griffiths, 1980). It is a Quaternary domain (Truckle, 1977a and Table 11) comprising major boundary faults and downwarps with more central fissures, normal near-vertical faults and collapse structures (calderas).

The NNE-trending boundary faults are best seen in the north where they comprise narrow zones of stepped, steep escarpments facing the median graben. Up to several hundred metres downthrow westwards can be proved along the boundary faults with the eastern shoulder, by displacement of Plio-Pleistocene trachyte (Baker, 1963). This fault zone is less than 5 km wide.

The north-western boundary of the median graben is defined by the Western Suguta Fault Swarm (Figure 18). According to Truckle (1977a) these faults coincide with the site of the late Pliocene downwarping which created a proto-Suguta Tough. Up to several hundred metres downthrow eastwards can be demonstrated for individual

Table 11 A summary of the tectonic history of the Gregory Rift in Degree Square 27

Age	Western shoulder	Median graben	Eastern shoulder
Recent–Pleistocene	Major NNE-trending normal faulting, downthrowing eastwards and confined to the boundary with the median graben. Major eastward downwarping in the same zone. Minor westward backtilting related to the faults. Landslip	Major normal faulting along marginal zones to define graben; fissuring; near vertical N- to NNE-trending fault swarms; collapse structures (caldera); landslip; NW fracturing to control central volcanoes	Major NNE-trending normal faulting, downthrowing westwards and confined to boundary with median graben; gentle westward downwarping in same zone; landslip
Pliocene	Minor downthrowing, mostly eastwards on new or reactivated faults (e.g. Lomujal F. Ribkwo faults; Napetio Arch faults.) Dextral strikeslip movement of the western shoulder. Several periods of eastward tilting by up to 20°. Local crustal arching related to trachytic vulcanism	Proto-Suguta Trough formed by gentle downwarping of western shoulder towards the east	
Upper Miocene	Major N- to NNE-trending normal faulting, mostly on reactivated faults; related arching (Napetio Arch); gentle eastward tilting; landslip	Graben must have existed because of major faulting on both shoulders	Major N- to NNE-trending normal faulting, eastern faults downthrow gently east and western faults downthrow several hundred metres west; westward tilting in west with back tilting related to faults; landslip
Middle Miocene	Several periods of tilting eastwards (angular discordances of up to 15° between lithostratigraphic units); faulting (Napeitom F, Lomujal F) on N to NNE and possibly NW trends		Downwarping westwards to produce Samburu Monocline; normal faulting trending NNE and NW
Lower Miocene	Early downthrows eastwards on major western faults (Elgeyo F, Kula F,) tilting eastwards locally (Tiati Monocline) on NNW axes		Faulting and gentle warping

faults, e.g. on the Nakitoekirion Fault. Truckle (1977a) suggests that the total downthrow amounted to about 1000 m, by cumulative movements on faults in a zone about 20 km wide encompassing Namarunu Volcano in the extreme north. South-south-westwards the fault zone narrows to less than 5 km and individual fault scarps become less impressive: at Lomelo the fault scarps are from 20 to 30 m high compared to about 300 m on Namarunu. Marginal downwarps, facing the median graben, and contemporaneous with the boundary faults were more impressive on the western side (dips of over 30°) compared to the eastern side (dip of less than 20°). Strata on the floor of the median graben remain horizontal. Closely spaced, near-vertical normal faults with sinuous NNE to N traces, up to several kilometres in length, in the median graben cut the central volcanoes and intervening sediments. Individual faults have minor downthrows although aggregate displacements of over 200 m have been recorded across several adjacent faults (Sceal, 1974). Downthrows to the east and west occur (see map cross sections). Spasmodic fault movements have been interspersed with basalt eruptions along parallel fissures (Sceal, 1974). Rotation in the fault plane due to differential vertical displacement along the strike has tilted fault blocks to form ramps (Griffiths, 1980) along the long axis of the median graben. Landslip has occurred on the older fault scarps. Individual fault traces are deflected towards the summits of the central volcanoes: this may be due to a topographic control on steeply dipping (westwards) fault planes.

Recent faulting is widespread: dilation (pull apart) cracks, locally concomitant with normal faulting, are recognised on Emuruangogolak (Rhemtulla, 1970), Silali (McCall and Hornung, 1972), and west of Secumius (GR 1953 1135). Individual fissures, trending NNE, may be several hundred metres long and deep, and tens of metres wide. On Emuruangogolak the latest basalt eruptions utilise such fissures.

The summit calderas of the various Quaternary volcanoes are regarded as collapse structures of the 'Glencoe type' (McCall and Hornung, 1972). They were controlled by near-vertical, circular faults with downthrows of up to several hundred metres: calderas, e.g. Silali, are manifestly elliptical with the long axes trending WNW–NW, to indicate a fundamental tectonic grain across the long axis of the rift. Parallel magnetic anomalies (shown by the EEC-funded 1982, aerial geophysical survey, confidential data) reinforce this suggestion of deep control by transverse faults.

The structural elements of the median graben can be integrated into a tensional regime with limited horizontal separation across its long NNE axis.

8.4 TECTONICS OF THE GREGORY RIFT IN RELATION TO VULCANICITY

Theories on the relationship between tectonics and vulcanicity in the Gregory Rift beneath Degree Square 27 must take cognizance of the following evidence:

1 The chemistry of the volcanic rocks suggests that the various extrusive phases within a discrete volcanic unit had a common source. This is particularly well documented for the central trachyte-basalt volcanoes (Chapter 4). An alkali basalt (picrite) primary source is generally indicated by the chemistry.

2 Despite gross lithological differences between the eastern and western shoulders and median graben, the chemical trends displayed by these rocks are similar (Figure 19). They all follow an alkaline basalt differentiation path, already indicated by whole-rock chemistry of individual volcanic units. Almost all extrusives are quite highly fractionated. Studies of the central volcanoes of the median graben have all suggested initial fractionation of the source magma to produce trachyte cupolas derived from the underlying alkali basalt magma. The layered magma chamber is generally estimated (Chapter 4) to be at depths of about 50 km.

3 The regional geophysical evidence, summarised by Baker and Wohlenberg (1971), agrees with the petrochemical evidence for a large layered magmatic body beneath the Gregory Rift. In particular the gravity data suggest a basic body to within about 40 km of the earth's surface overlain by a less dense cupola. These data also show an east to west asymmetry to the deep structure, across the rift.

4 Tectonism and volcanism migrated towards the median graben with time; both are presently confined within, or adjacent to the median graben. Throughout the history of the rift there has been a strong tectonic control on the locus of volcanism. The north-north-east migration of the Mio-Pliocene trachytic volcanoes of the west shoulder has been explained by a component of strikeslip movement along a grossly arcuate fault system. The Quaternary central volcanoes appear to be sited where transverse faults cut across the median graben. Therefore it would appear that the volcanic rocks were erupted along deep-seated faults. The inward migration of these faults with time could, in part, be explained by infill of early faults by solidified magma and/or lateral pressure from an expanding median graben. Truckle (1977a), considered that, for much of the Late Cenozoic, the deformation of the rift was due to horizontal extension.

5 There is a structural east to west asymmetry to the rift. Within the eastern shoulder, major faulting was restricted to the immediate vicinity of the median graben. Downwarping of this shoulder was also confined to its western parts so that outer areas are essentially undeformed (except where cut by the early NW-trending faults). In contrast the whole of the western shoulder is deformed with several major structures (the Elgeyo–Kula Fault system and the Tiati Monocline) affecting the outermost rocks.

6 Due to the tectonic control on volcanism the structural asymmetry reflects a petrological asymmetry. An example of this is the essential confinement of Mio-Pliocene trachyte volcanoes to the inner part of the west shoulder—the area suffering almost continuous eastward downwarping. Williams and Chapman (1986) indicate that trachyte volcanism in the rift is confined to areas being warped rather than strongly faulted. Truckle (1977a) noted that the long history of eastward tilting or downwarping of the western shoulder suggests horizontal extension across the Rift. Vertical uplift is confined to specific periods of major normal fault movement: early Miocene movement on the Elgeyo Fault, late Miocene faulting of the inner parts of both shoulders, and Quaternary faulting of the median graben.

The E–W cross section in Figure 20 is based upon the aforementioned geological and geochemical data. It indi-

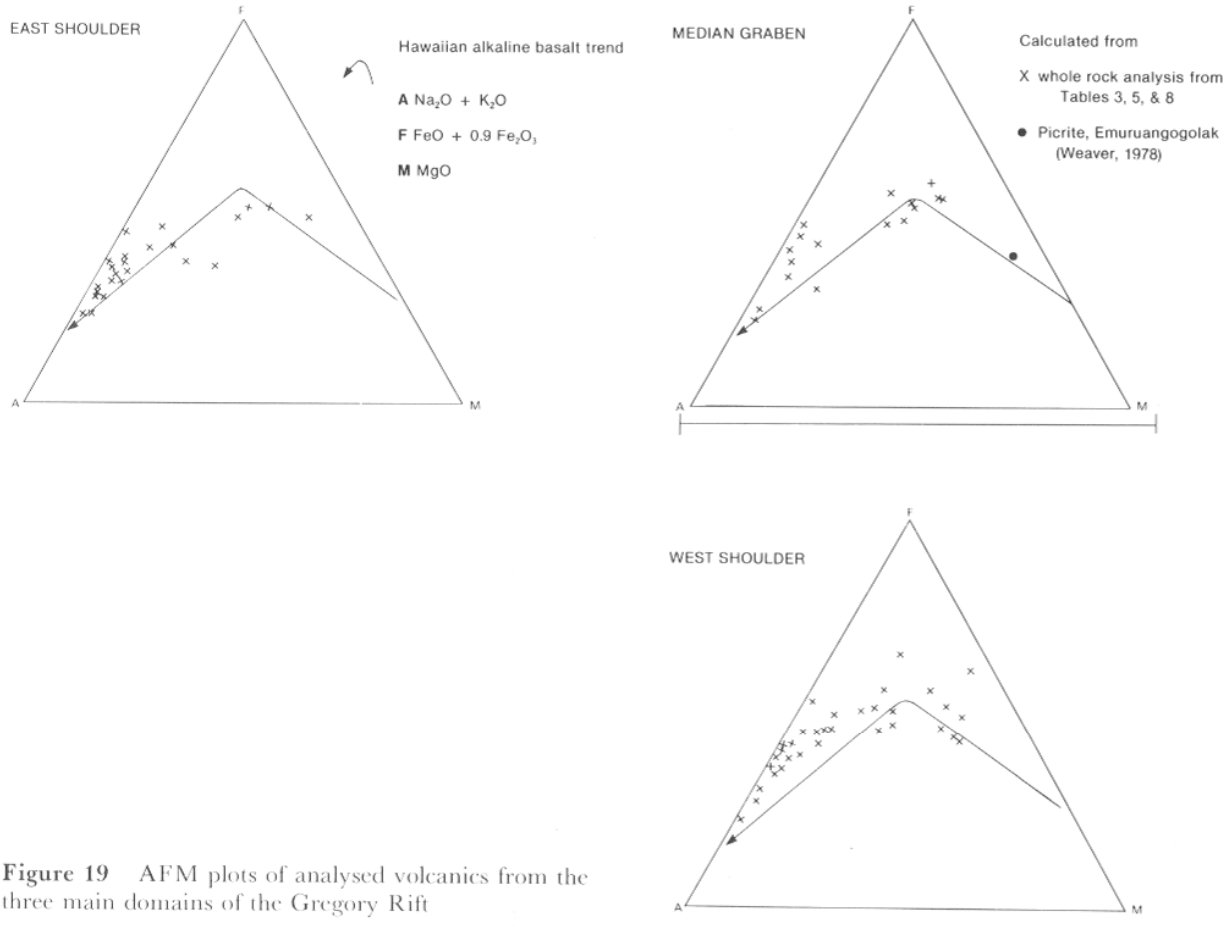


Figure 19 AFM plots of analysed volcanics from the three main domains of the Gregory Rift

cates a fundamental role (compare fig. 3.18 of King, 1978) for the Elgeyo Fault system and its major synthetic associates (e.g. the Kito Pass Fault eastwards to the Nakitoekirion Fault). Progressively smaller segments of the western shoulder slipped downwards along these faults towards, what is now, the median graben. The intriguing question is whether this process will lead to actual crustal separation in the median graben or if the rift eventually seals itself by cooling of the magma sources. The eastern shoulder appears to have played a less active role, vertical displacement into the median graben having been restricted to the innermost 'hinge' of the shoulder.

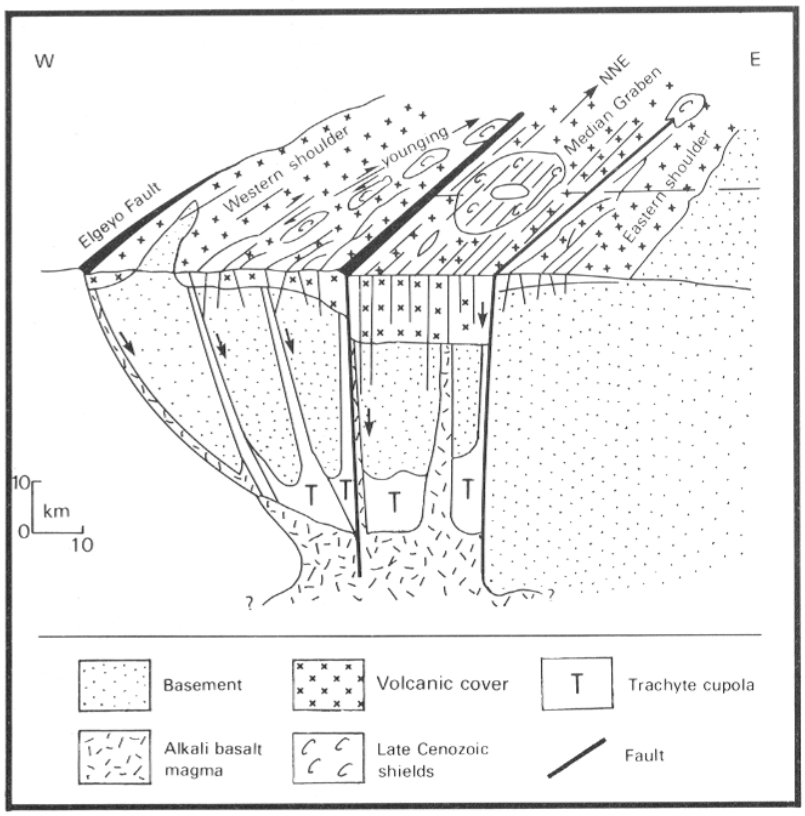


Figure 20 The deep structure of the Gregory Rift under Degree Square 27



Plate 15 Banded metagabbro, Kotikal



Plate 16 Granite offshoot from the Sartim Granite cutting folded mafic gneisses of the Siambu Complex



Plate 17 Parasitic, recumbent Sabachian folds, Karisia Hill



Plate 18 Coarse anthophyllite asbestos, east of Sartim



Plate 19 The hot waterfall at Kapedo



Plate 20 Lorusio Hot Springs with Silali in the distance

9 Economic geology

For the most part prospecting within Degree Square 27 has been confined to private individuals examining the major pegmatites in the eastern metamorphic terrain. Initially the prospectors were attracted to the showing in these pegmatites of large muscovite plates, with traces of radioactive minerals (Baker, 1963). Of recent years, with the discovery of numerous gemstones within the Mozambique Orogenic Belt of southern Kenya and Tanzania their attention has turned toward beryl mineralisation in the same, muscovite-bearing pegmatites.

In the 1970's Anglo-American undertook a regional soil geochemical survey of northern Kenya to include the eastern part of Degree Square 27. This work was not properly documented. A multi-purpose airborne geophysical survey was flown by Geosurvey International (EEC funding) in 1982 over the eastern metamorphic terrain. Their results are filed in the Mines and Geological Department, Nairobi. One significant radioactive anomaly was located south-west of Siambu Hill within Degree Square 27.

The results of the project stream sediment sampling programme, essentially confined to the eastern metamorphic terrain, are discussed separately in Chapter 10 by the project geochemist.

The volcanic terrain is unprospected. Shackleton (1946) reported fluorite mineralisation along fault planes in the eastern shoulder of the Gregory Rift. Truckle (1977a) records numerous diatomite beds in the Suguta Trough and on adjacent parts of the western shoulder. Some stream sediment samples were collected during the present survey including four samples from drainage off the Kailimerlim Foyaite.

Surface showings of possible economic interest are described in this chapter. Significant finds of asbestos, chromite, diatomite, gemstones (beryl, sapphire, green garnet), graphite, radioactive minerals, and vermiculite warrant detailed follow up investigations. Large deposits of river sand and gravel are evenly distributed throughout Degree Square 27 and material suitable for roadstone and building stone occurs in both the volcanic and metamorphic terrains.

9.1 ASBESTOS

Veins, and to a lesser extent pods of anthophyllite and tremolite, cross and long-fibre asbestos occur in many of the larger ultramafic bodies within the Siambu Complex. Besides the localities mentioned by Baker (1963), other good occurrences were discovered at GR 2687 1188 (Plate 18), GR 2753 1128, GR 2721 1639 and GR 2707 1636. Brittle and soft-fibre lengths vary from about 2 cm up to 40 cm in veins up to 5 m wide and over 10 m in strike length. Exploration trenches cut the veins at GR 2707 1636.

Three veins from 10 cm to 1 m thick and up to 6 m long of coarse, brittle fibre anthophyllite asbestos are exposed in the ultramafic pod south-east of Maralal.

9.2 BUILDING STONE

Local demand is limited and the area is too far from the major population centres of Kenya to support a national requirement for building stone. Cliff faces of massive quartzofeldspathic gneiss at Loole Rock, Soit Pus Rock and the water tank hill could provide building stone for Maralal. A quarry at GR 2493 1803 into the basal air-fall tuffs of the Lopet Phonolite adjacent to the main Maralal track supplied stone for the Catholic Mission at Baragoi. Similar tuffs to the west could also provide building stone for Lokori, Lomelo and Kapedo.

9.3 CHROMITE

Baker (1963) describes the chromite pods in ultrabasics at Kangura. The ultrabasics should be drilled to ascertain the presence of concealed pods. A smaller chromite pod is present in an ultramafic body at GR 2612 1650 (north of Noljushin Hill). It is 1 m long and the chromite contains 27.7 per cent Cr_2O_3 and 9.5 per cent Fe_2O_3.

9.4 COPPER

The Siambu Complex would appear to be a good setting for copper mineralisation with its wealth of altered basic and ultrabasic rocks. However copper showings are limited to minor malachite impregnations; pyrite and chalcopyrite occur at GR 2649 1642 and azurite 2 km further SSW. Otherwise malachite is present infilling vesicles in basalts of the Kamolingaran basalts in the Baragoi River (Baker, 1963). Chrysocolla also occurs in a fault breccia cutting basalts at GR 2291 1557.

9.5 CORUNDUM

Well formed bipyramids of grey opaque corundum are a common constituent of float around many, if not all, of the ultrabasic hills in the Masiketa area. They weather out of plumasite veins in the ultrabasics with individual crystals usually several centimetres in length. They could be sold as mineral specimens.

9.6 GEMSTONES

Almandine garnet

Fragmented orange to red almandine garnets are a major component of stream sediments in the south-east where they have weathered out of underlying bedrock. They have no potential as gemstones but could be used for abrasives. The massive almandine garnets in banded calc-silicates at GR 2749 1607 and GR 2735 1606 and in quartz-limonite rocks at GR 2576 1565 and GR 2564 1628 are also only suitable as abrasives.

Amethyst and apatite

Well-formed amethyst crystals and translucent apatite crystals infill vughs in the banded amphibolite immediately south of Maralal. Both minerals are only present in minor quantities and the deposit has no commercial value.

Beryl

Transparent aquamarine and translucent green beryl form fragmented crystals embedded in milky quartz in many muscovite-bearing pegmatites in the north-east. Most localities have been pitted (e.g. the large pegmatites at GR 2723 1807 and GR 2679 1529), and all the easily accessible material removed, as Baker (1963) observed for the Nachola occurrences. Aquamarines seen during the present survey have a rather pale blue colour and were not of the best quality.

Petrified wood

Several petrified wood localities are recorded in the volcaniclastic sediments (Chapter 4). Large (up to several metres long) specimens litter the ground at GR 2413 2046. This occurrence is well known; it is recorded by Baker (1963) and the best specimens have been illegally taken over the last decade. It is reasonably accessible by a track from Baragoi and there is still sufficient material remaining to warrant a legitimate claim.

Sapphire

Grey opaque corundum crystals with deep blue transparent centres were noted in the float at GR 2646 1757. Detailed sampling is warranted in this area because of the fine colour of the central material. Diaspore locally replaces the corundum. Similar dark blue cores to grey corundum are also present at GR 2627 1581.

Tourmaline

A quartz vein at GR 2674 1789 contains nicely formed individual crystals and aggregates of schorl up to several centimetres in length. The deposit is of no economic value.

Tsavorite (green grossular garnet)

Subrounded transparent green tsavorite crystals less than 0.5 cm in diameter occur in the graphite gneisses at Lareodo (GR 2727 2060). These crystals are too small to be fashioned as gemstones but the whole of the graphite gneiss unit warrants detailed prospecting for large pods of tsavorite (the source of the southern Kenya gem material). An unidentified, attractive, orange, transparent mineral also occurs in the same gneisses, with green mica and green kyanite blades.

9.7 'GOSSAN'

A massive chert-rich 'gossan' crops out within the ultramafic pod south-east of Maralal. Similar iron oxide caps are common to ultramafic bodies in the Siambu Complex. Pitting, and if necessary drilling, into and through this material is required to ascertain whether they are due to sulphide mineralisation in underlying bedrock.

9.8 GRAPHITE

The graphite gneiss at Lareodo underlies a large hill several hundred metres in height and diameter in the nose of an antiform. It is the most important graphite locality within the SMP area and should be investigated in detail. Baker (1963) records up to 22.54 weight per cent graphite as small flakes up to 1 mm in length. The other graphitic gneiss bands along the eastern map margins have no economic potential for graphite. They contain less than 4 per cent of small flake graphite.

9.9 LIMESTONE

The various marbles shown on the geological map are essentially pure carbonate rocks. A single analysis of the marble east of Masiketa indicates it is a mixed calcite and dolomite rock with 26.9 wt per cent CaO and 12.0 wt per cent MgO. However the marbles east of Baragoi are mostly calcitic with 44.8 wt per cent CaO and 4.4 wt per cent MgO.

The marbles have potential as sources of lime, cement or facing stone.

9.10 MAGNESITE

Baker (1963) provides descriptions of the minor magnesite veins found in all serpentinites in the Siambu Complex. As he notes, there are no economic deposits of this mineral.

9.11 MOLYBDENITE

Shackleton (1946) reports trace amounts of molybdenite in a quartz vein near Kweribe waterhole (about 10 km NNW of Poro). The vein cuts biotite gneisses close to the interface with the eastern Rift shoulder.

The occurrence has no economic potential and no other molybdenite has been located subsequently in any other part of Degree Square 27.

9.12 MUSCOVITE

Books of silvery muscovite up to several centimetres in length are common in pegmatites in the Siambu Complex. However the economically viable deposits at Nachola (see Baker, 1963) and at GR 2580 1560 near the Barsaloi River have been exhausted.

9.13 PYRITE

Pyrite cubes are visible at GR 2699 1630 (Siambu Hill) and GR 2676 1544 as well as in the molybdenite-bearing quartz vein at Kweribe waterhole. They have no economic significance.

9.14 QUARTZ

Pure quartz reefs cut the Siambu Complex and the larger reefs are shown on the geological map. Smaller veins cutting the mafic gneisses near Maralal are exposed on hillsides just north of the town. They could meet a local demand for pure quartz.

9.15 RADIOACTIVE MINERALS

The airborne radioactive anomaly south-west of Siambu Hill is in an area of the Siambu Complex cut by G2 granites and later aplite veins. It is possible that the anomaly is related to one or other of these felsic intrusives. Initial ground follow-up failed to locate the source rock for the radioactivity. However anomalous scintillometer counts were recorded during the field mapping in the Barsaloi and Karisia Hills areas near Lorokomongo village (GR 2520 1406) and around Engineji and Lamalach Hills (GR 2712 1394). The anomalies are related to euxenite-bearing pegmatites. Baker (1963) lists samarskite, pitchblende, columbite, euxenite and monazite from pegmatites in the Nachola area.

9.16 RIVER SAND AND GRAVEL

All watercourses are infilled by alluvium, consisting of various mixtures of sand, gravel, silt and clay. Over the metamorphic terrains the alluvium is mostly composed of a quartz and feldspar sand with localised gravel lenses. Around Maralal, on the Karisia Hills and on the El Barta Plain there is a prominent silt fraction which is probably of volcanic origin. Over the volcanic terrain the watercourses are usually infilled by a coarser alluvium with rounded volcanic clasts up to boulder size. The Kerio and Suguta Rivers do have a sandy infill.

This river sand and gravel can supply all the construction requirements within Degree Square 27.

9.17 ROADSTONES

Any of the various (basaltic, phonolitic or trachytic) lavas could be crushed to provide roadstone for any future upgrading of the major tracks in the map area.

9.18 TALC

Numerous small pods of white to green talc are found in ultramafic rocks within the Siambu Complex. As Baker (1963) observes these pods are generally too small and too remote to be of economic potential. They also contain other calc-silicate minerals, notably tremolite. The area east of Sartim on the northern map-edge contains numerous such talc pods and is the most prospective area for this mineral, should there be a future demand.

9.19 VERMICULITE

Vermiculite flakes, up to 1 mm in length, are a noticeable component of the soils over large parts of the Karisia Hills north of Maralal. The thick forest cover precludes an accurate estimate of the area with vermiculite-bearing soils by limiting ground access. The vermiculite is derived from underlying mafic gneisses of the Loroki Gneisses: it is a secondary replacement of biotite in mafic gneisses cut by felsic pegmatites or remobilised quartzofeldspathic gneisses.

In one instance, vermiculite flakes were seen to be replacing clinopyroxene along cleavage planes and in another case individual vermiculite flakes up to 2 cm long were found adjacent to a thin pegmatite. Detailed follow-up exploration is warranted especially in the deforested areas around Maralal.

9.20 VOLCANIC TERRAIN

Shackleton (1946) recorded fluorite along fault planes in the Samburu basalts south of 1°30′N. The Rift as a whole is prospective for fluorite (note that groundwater in the volcanics is high in fluorine) and all fault planes are potential traps for this mineral.

Key (1987) located over 1 ppm gold values in Miocene and Quaternary volcanics related to the East African Rift in northern Kenya. Anomalous gold values were found in basic, intermediate and acid rocks and it is felt that all rocks in the Gregory Rift should be assayed for gold.

The Kailimerlim Foyaite is prospective for rare earth mineralisation. Four bulk samples from watercourses draining this intrusive gave high values for Ba (1149 ppm), Sr (457 ppm), W (37 ppm), Ni (135 ppm) and Nb (121 ppm). Table 12 provides the full results of this limited sampling programme.

Table 12 Geochemical results from stream sediments collected adjacent to the Kailimerlim Foyaite Plug

Element	641/01	641/02	641/03	641/04	Analyst
Ag	0.3	0.3	0.3	0.3	
Ba	140	130	80	100	
Co	19	22	17	30	
Cu	24	22	18	32	Geosurvey
Fe*	4.3%	5.2%	4.3%	5.7%	International
Li	6	5	6	6	Ltd, Nairobi,
Mn	940	1350	1200	1210	Kenya
Mo	2	2	2	2	(fine fraction)
Ni	26	14	15	37	
Pb	3	5	6	1	
Sr	162	130	118	127	
Zn	75	99	92	79	
TiO_2*	1.76%	2.44%	0.86%	1.76%	
Fe_2O_3*	9.61%	12.37%	7.85%	10.12%	
MnO*	0.67%	0.64%	0.37%	0.22%	
As	3	6	1	1	
Bi	0	0	0	0	
Ba	864	1149	659	362	
Co	34	23	11	49	Nottingham
Cr	706	81	69	811	University
Cu	31	30	14	47	UK
La	69	71	70	30	(panned
Mo	2	6	5	0	concentrate)
Ni	96	17	28	135	
Nb	57	85	121	36	
Pb	12	14	15	10	
Sn	0	0	0	0	
Sr	263	457	241	317	
Th	5	9	13	4	
U	4	3	2	0	
V	254	213	64	264	
W	6	37	23	2	
Zn	88	156	118	66	
Zr	209	338	400	184	

*weight %: other results in ppm

Diatomite beds within the sedimentary units of the 'Suguta Beds' are of variable purity and thickness (Truckle, 1977a). The most prospective areas are at Gwawsinyono Arrolio and Kamuge where beds of relatively pure diatomite are exposed, 6–10 m thick.

10 The Geochemical Exploration Programme

J. Ridgway

10.1 GENERAL DESCRIPTION

Introduction

The geochemical exploration of Degree Sheet 27 formed part of a wider reconnaissance survey covering the area between 36° and 38°E and 0° and 2°N: for the most part the following summary is applicable to the whole of the survey (Degree Sheets 27, 28, 35 and 36). Sampling was conducted concurrently with geological mapping and was largely constrained by the requirements of that exercise. Thus samples were collected chiefly, although by no means exclusively, along traverse lines dictated by geological considerations. Little or no sampling was done in areas previously mapped by the Geological Survey of Kenya or the East African Geological Research Unit (EAGRU). More detailed descriptions of the methods used and the rationale behind the geochemical survey can be found in a series of open file reports of the Overseas Programme Directorate of the British Geological Survey (formerly the Institute of Geological Sciences) (Ridgway, 1981a; 1981b; 1982; 1984).

Multi-element anomaly maps and single element plots based on decile divisions of the stream sediment data set, all at 1:250 000 scale, can be obtained through the Mines and Geological Department, Nairobi. Data from both stream sediment and pan concentrate surveys are also held in computer readable form and can be made available as print out, decile maps covering specific areas or on magnetic tape and floppy diskette media. All requests for such data should be made to the Commissioner of Mines and Geology, Mines and Geological Department, Nairobi, from whom further details are obtainable. A service charge will be made for all maps and data sets.

The sample collection programme

Two sampling media were employed in the survey; normal stream sediments and panned heavy mineral concentrates. The intensity of sampling varied according to the geology and the degree of exposure. Well-exposed areas of complex geology were subject to higher density collection than regions of extensive superficial deposits, which received little attention. For Degree Sheet 27 sampling was restricted to the basement rocks in the eastern half of the area, the western portion being composed largely of volcanic rocks previously mapped by EAGRU. Figures detailing sample densities have little meaning in this context and the pattern of collection can be seen on the summary multi-element geochemical anomaly maps for each of the four degree squares comprising the survey area (see above). In general terms the initially recommended densities of 1 sample per 4–8 km² for normal sediments and 1 per 13–16 km² for pan concentrates were achieved, and exceeded in regions of special interest.

Stream sediments were collected by combining grab samples from 5 to 10 points within a 10 m radius of the selected site, the number of points varying according to the width of the channel. If necessary the sediments were dried before being sieved and the fine (−80 mesh) fraction retained for analysis.

Heavy mineral concentrates were obtained by taking 2 to 5 kg of material from the sample site and panning at the base camp, where water was available, or at the Mines and Geological Department headquarters in Nairobi.

The analytical programme

The fine fraction from the stream sediment samples was analysed by atomic absorption spectroscopy (AAS), after digestion in hot 1:1 hydrochloric acid, for Ag, Ba, Co, Cu, Fe, Li, Mn, Mo, Ni, Pb, Sr and Zn. Initially the analyses were carried out in the laboratories of the Mines and Geological Department but for a number of reasons it unfortunately became necessary to switch to Geosurvey International Limited after approximately one quarter of the total number of samples had been processed. The effects of this change on the interpretation of the results are discussed later. For Degree Sheet 27, the analyses of stream sediment samples were all carried out by Geosurvey, and only the eastern half of the map area was sampled, i.e. eight 1:50 000 map sheets. With both laboratories precision (defined as twice the standard deviation of a series of replicate analyses divided by the mean and expressed as a percentage) was monitored by the submission of control samples and replicates. For the Mines and Geological Department analyses precision generally varied between 20 and 40 per cent depending on the element and concentration, the results being worst at low concentrations. Low concentrations also yielded the worst precision figures for the Geosurvey data but monitoring charts of the type advocated by Thompson and Howarth (1973) show that for most elements the precision is better than 20 per cent, the chief exceptions being Pb and Li. A precision level of 20 per cent is adequate for geochemical exploration purposes and even where the figure is higher, examination of simple element concentration maps shows that the stream sediment chemistry reflects the geology with a fair degree of accuracy and there is therefore no reason to doubt the reliability of the data. Further details on the precision of the analytical results can be found in Ridgway (1981b; 1982 and 1984).

Pan concentrates were analysed by X-ray fluorescence methods in the Department of Geology at the University of Nottingham, UK. The elements determined were As, Ba, Bi, Co, Cr, Cu, Fe, La, Mn, Mo, Nb, Ni, Pb, Sn, Sr, Th, Ti, U, V, W, Zn and Zr. Because of the small amounts of concentrate usually available it was not possible to monitor the precision of the XRF programme by using replicates, but the laboratory claims a figure of better than 8 per cent at the 100 ppm level for all the elements determined.

Interpretation

Fine-fraction stream sediments

Interpretation relied heavily on the use of plots of log concentration against probability (Lepeltier, 1969; Sinclair,

1976) to determine threshold values, above which concentrations can be considered to be anomalous and possibly related to mineralisation. This approach was supplemented by the examination of maps showing the spatial variation in concentration for each element.

Because of the complexity of the geology in the survey area the data were grouped according to the dominance of one or more of a number of lithological units in the drainage basin upstream from the sample site before commencing the generation of log concentration/probability plots. The lithological units were chosen in consultation with the mapping geologists and are described briefly below.

The greater part of the area covered by the geochemical survey is underlain by rocks of the Mozambique Orogenic Belt. These are chiefly metasediments of probable middle to late Proterozoic age which have undergone polyphase metamorphism accompanied by widespread anatexis. The present metamorphic grade is almost uniformly of the upper amphibolite facies. Ten significant groupings have been recognised in the basement (letters in parenthesis are codes used in later tables and text); of these ten groupings the seven listed as follows are represented on Degree Sheet 27:

Ol Doinyo Ng'iro Gneisses and Il Busi Gneisses (OB) medium- to coarse-grained, banded, biotite gneisses with major metasedimentary intercalations of marble, graphite gneiss, garnet gneiss, sillimanite garnet gneiss and metaquartzite.

Kotim and Loroki Gneisses (KL) quartzofeldspathic gneisses, sometimes carrying muscovite, and minor, banded, biotite gneiss. Highly leucocratic with little biotite, but common zircon and magnetite/ilmenite. Variably migmatised.

Migmatites (M) coarse-grained quartzofeldspathic migmatites with some mafic hornblende-rich varieties and garnet-bearing types. Generally massive, heavily veined and pegmatised. Muscovite occurs in the pegmatites.

Siambu and Korr Complexes (SK) hornblende plagioclase gneisses and biotite gneisses. Mixed metasedimentary-meta-volcanic sequences with metabasic and meta-ultrabasic pods and marbles.

Metabasic and meta-ultrabasic pods (MBU) not obviously related to the Siambu or Korr Complexes.

Quartzofeldspathic gneiss and granite gneiss (QG).

Undifferentiated gneisses (X) used to designate drainage basins containing a mixture of gneiss groups, or for basement of doubtful affinity, especially inliers in the volcanic cover. Mostly banded biotite gneisses.

Tertiary and Quaternary volcanics overlie the basement and are particularly extensive on the eastern shoulder of the Rift Valley and in the southern part of the region, where they occur on the northern slopes of Mount Kenya and in the Nyambeni range. Four groupings have been recognised in the volcanics, only two of which are represented on Degree Sheet 27:

Phonolites and phonolitic trachytes (P) extensive flows of Miocene age found along the eastern shoulder of the Rift Valley and on the northern slopes of Mount Kenya.

Basalts and associated undersaturated rocks of Miocene to Pliocene age (SO) extensive thin flows on the eastern rift shoulder (Samburu Basalts) and at the foot of Mount Kenya (Osirua Basalts).

In addition to the above a small group of undifferentiated sediments has been defined.

Mafic and felsic intrusions of various ages are found throughout the basement but are generally of too limited extent to form useful divisions for the purposes of the stream sediment exploration programme.

After separation of the data into results produced by the Geosurvey laboratory and those emanating from the Mines and Geological Department laboratories, the above groupings were used to generate threshold values applicable to the whole of the surveyed area; (Table 13). Thus, depending on the relationship between the areal distribution of the lithological groups and the time the samples were processed each division may have one or two threshold values per element. This does not affect the appearance of the summary multi-element anomaly maps for the four degree sheets. On these the anomalies are plotted in a standardised form (Govett and Galanos, 1974) using the formula:

$$\frac{(\text{Anomaly} - \text{Threshold}) \times 100}{\text{Threshold}}$$

Results equal to the threshold level were regarded as anomalous and given the standardised value of 0.

The practice, frequently adopted in the absence of well-defined inflection points in the log concentration/probability curves, of regarding the top 2.5 per cent of high values as being anomalous has not been followed here. Threshold values have been defined only where significant changes in slope at high concentration levels could be identified.

For the maps showing the areal variation in element concentration the data from the two analytical sources were separately divided into decile groupings for the individual element concentration ranges and the decile intervals plotted as variable sized symbols.

In assessing the significance of the anomalies it should be borne in mind that they are not all necessarily related to occurrences of economically important minerals. Analytical interferences, secondary environmental effects and concentration in 'non-economic' minerals may singly or in combination give rise to anomalous results for some elements. Detailed consideration of these factors is

Table 13 Threshold (upper) and background (lower) levels in relation to drainage basin lithology groups (see text for explanation of letter codes)

Drainage Basin Lithology	Zn	Pb	Ni	Co	Mn	Cu	Sr	Ba	Li
OB	100	16	70	50	1200	60	—	—	—
	45	1	17	15	438	23	33	132	5
KL	150	20	70	35	1600	45	250	350	22
	44	2	15	12	476	15	39	107	6
M	125	20	40	70	—	55	135	350	30
	42	1	15	13	422	19	30	103	5
SK	170	20	600	—	3000	200	250	500	—
	37	1	20	14	384	25	46	132	5
MBU	100	—	1000	—	2000	170	250	—	—
	38	1	21	17	453	23	36	130	5
QG	150	20	50	40	3000	80	155	500	19
	41	1	14	12	468	18	32	118	5
X	80	10	35	27	1500	50	150	240	—
	35	2	13	12	510	17	30	114	5
P	200	40	40	50	5000	—	160	500	—
	69	4	10	14	1184	13	31	161	7
SO	200	15	70	70	—	60	190	500	—
	62	2	14	16	752	17	63	181	7

For fine fraction stream sediments where no threshold value is given, the log concentration versus cumulative frequency graph indicted the presence of only one population. The background level is taken as the 50th percentile of the distribution and is thus equivalent to the median.

beyond the scope of this report but some of the more obvious possible effects are worthy of comment.

Background absorption can lead to large errors when AAS methods are used to measure low concentrations of an element in geochemical samples (Fletcher, 1970; Foster, 1971). In the present survey Pb, Co and Ni could have been affected but the low threshold levels for Pb in some lithological groups make anomalies in this element the most likely to be spurious. Ca and Fe are two of the most important interfering elements and care is needed in the evaluation of Pb anomalies where Fe concentrations, marbles or calcrete are prominent in the catchment area.

'Scavenging' by Fe/Mn oxides can preferentially concentrate a number of elements and give rise to anomalies unrelated to mineralisation. Mn oxides are probably more efficient scavenging agents than those of Fe (Chao and Theobald, 1976). Co, Zn, Ni and Cu have all been reported to be affected, whereas Pb commonly is not (Levinson, 1974; Carpenter and others, 1975; Whitney 1981). Nowlan (1976), however, considers Pb to be weakly scavenged. Low-level anomalies in the above elements in association with high Mn values are the most suspect, but it should be remembered that some mineral deposits are accompanied by enhanced levels of Mn.

Erosion of ultramafic rocks can give rise to anomalies in Ni, Co and Cu because of their characteristically high content of these metals. Calcrete deposits might be enriched in strontium and produce low-level anomalies, particularly in flat-lying plains and areas of poorly defined drainage.

Pan concentrates

A number of factors can contribute to variability in the analytical results from a pan concentrate exploration programme. Among these factors are the selection of the collecting site in the stream, the depth of collection, the consistency of the degree of panning and the 'nugget effect' in the fraction selected for analysis. Given the size of the present survey, the time allotted for completion and the difficulties of panning in a semi-arid environment it was not possible to reduce significantly the variability inherent in the methodology.

Interpretation at this stage has relied on the identification of anomalous samples from scatter plots of individual element concentrations against the combined concentration of Ba and Sr; this latter factor being introduced as a rough measure of the degree of panning. Thus relatively low values of an element might be considered anomalous if high values of Ba + Sr indicate a poor degree of concentration of heavy minerals. Threshold levels for the pan concentrates are shown in Table 14.

As with the fine-fraction stream sediments, anomalies in pan concentrates are not necessarily directly attributable to the presence of ore minerals. Many of the elements determined in this survey are found in small quantities in common rock-forming and accessory minerals. If any of these minerals occur in quantity in the concentrates, anomalies in a variety of trace elements could result.

Predicted mineralisation

Neither the Mozambique Belt nor the East African Rift volcanics contain many significant recorded mineral deposits.

However, a wide variety of mineral deposits could occur in a complex metamorphic terrain and metasediments, metavolcanics, acid and basic intrusives are all present within the Project area. Thus ores related to igneous, metamorphic, sedimentary and surficial environments may be present in some form. Base and precious metals, tin, tungsten, chromium, niobium/tantalum, rare earths, uranium and pegmatitic deposits are all possible targets for the exploration programme.

In the volcanic areas carbonatite and kimberlite intrusions and a variety of vein deposits associated with faults and volcanic centres could be present.

Minor amounts of economically interesting minerals such as beryl, mica, chromite and corundum, along with indications of Ag, Au, Bi, Cu, Mn, Nb and rare earth mineralisation, have been noted in the Project area.

Table 14 Threshold levels for pan concentrates established from examination of scatter plots of individual elements against combined Ba and Sr values

Element	Threshold	Element	Threshold
As	12	Nb	250
Ba	2000	Ni	250
Bi	5	Pb	50
Co	110		Ba + Sr > 2500: 10
	Ba + Sr > 600: 70	Sn	50
	Ba + Sr > 1300: 40		Ba + Sr > 1000: 25
Cr	3500	Sr	900
Cu	70	Th	800
La	2000	*Ti	33
*Mn	1.9	U	60
	Ba + Sr > 2000: 1.0	V	2300
Mo	45	W	50
	Ba + Sr > 1500: 35	Zn	900
	Ba + Sr > 2400: 10		Ba + Sr > 700: 500
		Zr	1750

*Values in % MnO and TiO_2

10.2 RESULTS FOR DEGREE SHEET 27

Fine-fraction stream sediments

A summary map of the anomalies is given in Figure 21. In the following discussion the anomaly numbers refer to those on Figure 21. Some details of the numbered anomalies are given in Table 15.

Pan concentrates

Pan concentrate anomalies are shown on Figure 22.

Comments on the anomalies

The majority of anomalies occur in the southern half of the area. In the north there are no coincident pan concentrate and stream sediment anomalies and no individual anomalies which appear particularly interesting. Three of the pan concentrate anomalies are in strontium and have relatively low levels of iron: these may represent poorly panned samples with a high feldspar content. The low lead anomaly associated with one of the strontium-rich samples is probably also related to the feldspar content and the accompanying Mo value is only anomalous at a high level of combined Ba and Sr (see Table 14). The two Co anomalies are similarly related to relatively high levels of Sr and Ba while the U and W are associated with, and possibly genetically connected (through adsorption?) with the formation of, the highest MnO content recorded in the pan concentrate programme.

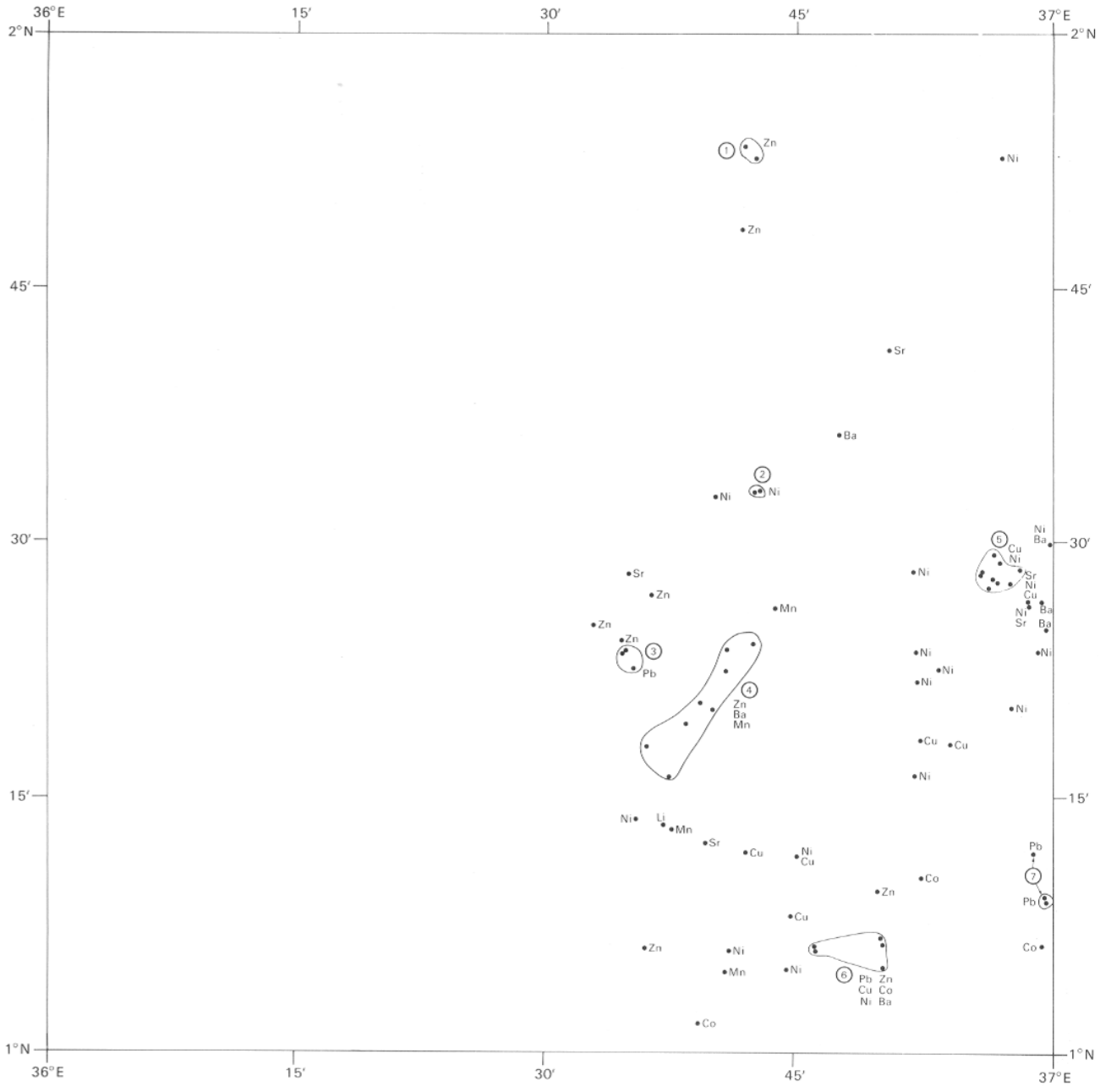


Figure 21 Anomalies in −80 mesh stream sediments. See text and Tables 13 and 15 for further details

Of the northern stream sediment anomalies, number 1 (Figure 21, Table 15) comprises single samples from two streams draining the same area. The drainage basin lithology has been classed as MBU but the surrounding area is of Siambu Complex rocks (SK) with a Zn threshold of 170 ppm: neither sample would be anomalous at this level. Anomaly 2 is at the confluence of streams draining the same area of the Lopet Plateau: the dominant lithology is considered to be phonolites (P) and the anomalies are unlikely to be of any significance.

Further south the Pb anomalies of 3 arise from a small inlier of gneisses surrounded by volcanics. All are still anomalous if the higher of the two gneiss thresholds is used but only that nearest the head of the drainage basin remains if the threshold level of phonolites (P) is applied. Two Zn anomalies nearby are related to the same gneiss inlier but would not be anomalous at either QG or P thresholds. Pan concentrate highs in Sn occur downstream of the Pb anomalies and also in a neighbouring drainage basin. Slightly elevated Cr values accompany the Sn anomalies in the concentrates.

The stream sediment anomalies of 4 are related to both gneisses and phonolitic volcanics in an area dominated by the volcanics. Most samples would not remain anomalous if phonolite (P) thresholds were used throughout. However, the area of anomalies is elongated along the line of a NE–SW fault zone which is cut by a second NW–SE set of faults. Zn, Pb, Mn, Ba and Li are all generally high in the region but may merely reflect the higher background values of the volcanics. Minor pan concentrate anomalies in Sn and W are found in the same area along with a Ba anomaly which is accompanied by high Fe, Mn and Nb values and may thus indicate the presence of small amounts of primary Ba minerals rather than feldspars in a poorly panned sample. At the north-eastern end of the fault zone, and probably related to an

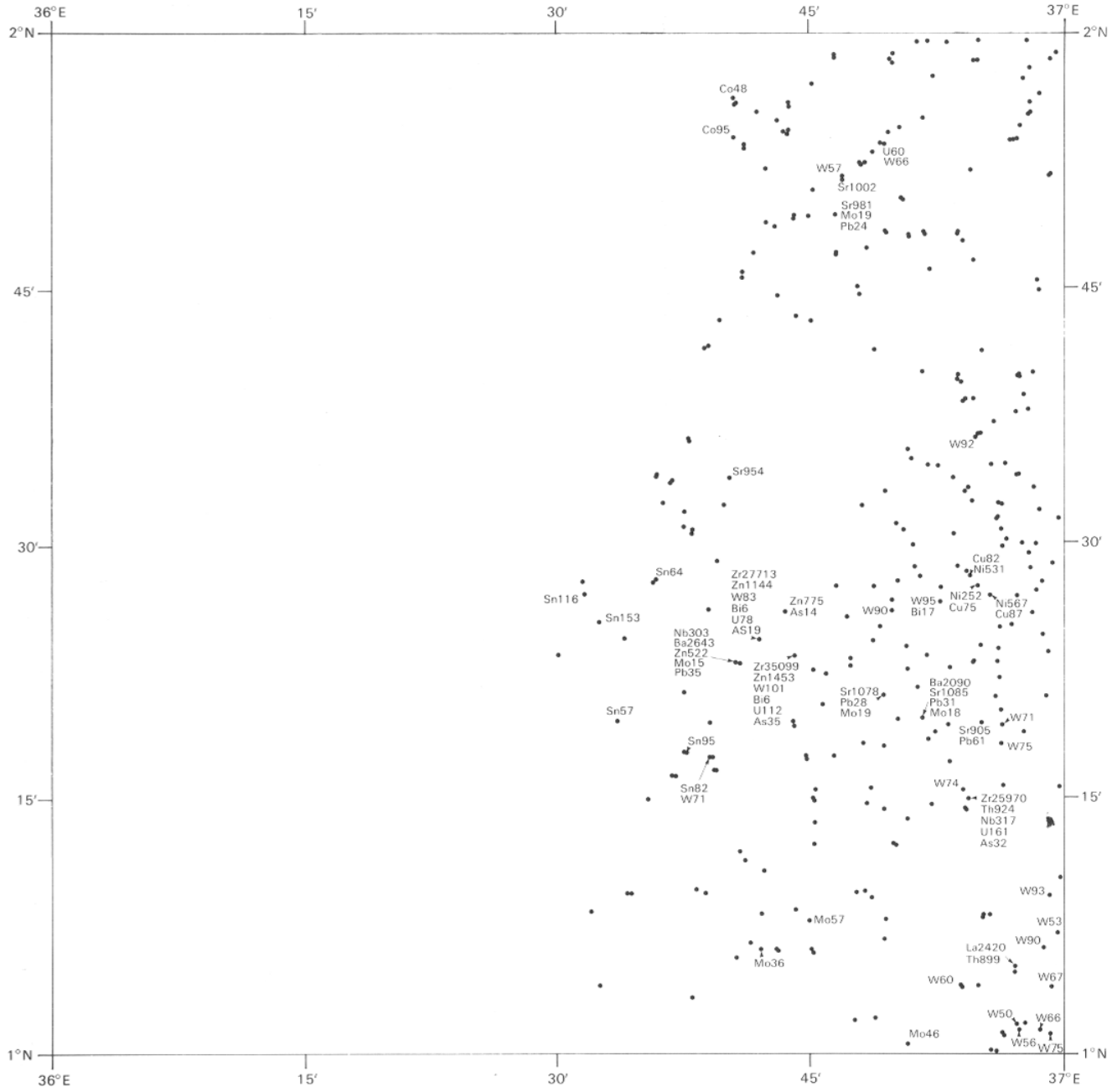


Figure 22 Anomalies in pan concentrates. See text and Table 14 for further details

area of Loroki Gneisses, are two pan concentrate anomalies in Zr, Zn, W, Bi, U and As. High Fe levels are associated with these anomalies and it is possible that they may be derived from accumulations of accessory minerals such as magnetite, zircon and apatite.

A series of fine-fraction sediment anomalies in Ni and Cu (5) occurs in streams draining radially from the meta-ultramafic rocks of Siambu Hill. The Cu anomalies would be lower and the Ni ones higher if threshold levels for the surrounding Siambu Complex (SK) were used. Pan concentrate anomalies in Ni, Co and Cu are also present and the area clearly merits more detailed investigation.

To the south and south-west of Siambu Hill three pan concentrate anomalies characterised by high Sr in association with low Fe are thought to represent poorly panned samples and a multi-element anomaly in Th, Zr, U, Nb, W and As may merely reflect accumulations in common accessory minerals.

The anomalies collectively numbered 6 on Figure 21 and Table 15 occur in streams draining the Loroki Forest area. Migmatites (M) occur in the drainage basin of three of the five anomalous samples and the use of M thresholds would have the effect of increasing the standardised values of the Ni, Cu and Ba anomalies.

Near the south-eastern boundary of the sheet Pb anomalies (7) are found in two separate drainage basins containing both Ol Doinyo Ng'iro and Undifferentiated Gneisses. The latter could possibly be classed as migmatites (M) and if the higher M thresholds were used one sample from each stream would still remain anomalous. Although not accompanied by any pan concentrate anomalies, apart from a single W high, the most northerly of the anomalous Pb values was the highest encountered in the geochemical survey.

Throughout the area, but particularly in the extreme south-east, there are a number of single element, low-

Table 15 Summary of the main stream sediment geochemical anomalies

Anom. No.	Elem.	D.B. Lith.	Thr. Value	Values ppm	Standard	No. of Samps.
1	Zn	MBU	100	110–155	10–55	2
2	Ni	P	40	63–80	58–100	2
3	Pb	X	10	33–36	230–260	2
		QG	20	62	210	1
	Zn		80	120–169	50–111	3
	Ba	X	240	327–427	36–78	2
	Mn		1500	3280	119	1
	Zn		150	153–185	2–23	3
4	Ba	QG	500	625	25	1
	Mn		3000	4460	49	1
	Zn		200	210	5	1
	Ba	P	500	665	33	1
	Mn		5000	6000	20	1
5	Cu	MBU	170	170–316	1–86	9
	Ni		1000	1200	20	1
	Pb		20	27–78	35–290	2
	Zn		150	162–167	8–11	2
6	Cu	QG	80	80–85	1–6	2
	Ni		50	67	34	1
	Co		40	40	1	1
	Ba		500	770	54	1
7	Pb	X	10	19–130	90–1200	3

Notes
Anom. No. refers to the numbered anomalies in Figure 21, Elem. = element, D.B. Lith. = drainage basin lithology, Thr. Value = threshold value, Standard = standardised anomaly value, No. of Samps. = number of anomalous samples.

All samples were analysed by Geosurvey International Limited.

level W anomalies. These are invariably associated with relatively high levels of Fe and Mn in the pan concentrates and the W is probably being carried in Fe/Mn minerals.

The Mo pan concentrate anomalies in the south-east are apparently unrelated to unusual levels of any other element and are therefore of uncertain significance.

Summary

The geochemical exploration programme in Degree Sheet 27 has indicated a number of areas where further investigations are merited. The five areas which are considered to have the greatest potential for the discovery of mineralisation are listed below.

1 The anomalies in the region of Siambu Hill (5, Figure 21), although all at a relatively low level, can be directly related to lithologies which have a known potential for Ni-Cu sulphide mineralisation and are the most obvious target for more detailed follow-up geochemical prospecting.

2 The Loroki Forest area, which has not been sampled in any detail and also gives rise to anomalies in Degree Sheet 35 immediately to the south.

3 The area around Anomaly 4, where there is a conjunction of stream sediment and pan concentrate anomalies with a well-defined intersecting fault pattern. The combination of faulting and anomalous trace element levels makes the area seem worthy of further exploration for a variety of vein or stockwork deposits.

4 The region of Anomaly 3 which has stream sediment anomalies in Pb and Zn and pan concentrate highs in Sn.

5 The area of Pb stream sediment anomaly 7, where the highest Pb value of the survey occurs.

These conclusions are based on a preliminary examination of the data. Further study may reveal more information on the possible origin of the anomalies and indicate additional areas of interest.

Addendum Uranium

Selected stream sediment samples collected from the basement outcrop area were analysed by Geosurvey International Ltd in Nairobi for uranium. Thirty-two samples gave values of 1.0 to 3.0 ppm uranium: half of these samples were collected from the area of the upper Baragoi Valley (Kolowaton River), on the outcrop of Siambu Complex hornblende-biotite gneisses. A value of 18.5 ppm uranium for one sample (from GR 2575 2097) was confirmed by re-analysis. The area has not been further investigated.

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11 Water resources and geothermal energy

11.1 EAST OF THE SUGUTA VALLEY

Surface water

Perennial supplies of surface water are confined to the upper reaches of watercourses in the Karisia Hills and western parts of the eastern shoulder of the Gregory Rift. These include the Logolin, Moridjo, Nashoda, Parasoro and Tinka Rivers on the volcanic terrain and the Loikas, Lulu, Ndadapo and Yamo Rivers in the Karisia Hills. The rivers are fed from cold springs. The small dam at Poro also holds permanent water because of the high rainfall in this highland area. The dam on the Nontoto River which supplies most of Maralal's water dries up in times of drought.

Seasonal pools remain in shaded parts of the larger rivers for short periods following heavy rains. Seasonal lakes or pans are widespread, but only hold small amounts of rainwater for short periods. These include (see Baker, 1963) pans at Lokongori, Nderentei, 5 km NW of Gelai, Ildebessare, 5 km SW and 2 km NW of Baragoi, Noyarot (on the Tirr Tirr plateau), and at GR 2430 1730 and GR 2460 2742.

Small earth dams have generally been constructed near most larger settlements but these have either become silted up or breached, e.g. at Rarait, Opiroi, Bauwa, Baragoi, in the Marti area at GR 2426 1480. Artificial pans have also been scooped on the El Barta Plains and alongside the main track from Baragoi to Maralal.

Groundwater

Incomplete data on 15 water boreholes were obtained from various sources, including the Ministry of Water (Table 16). Eleven of the boreholes were sited on metamorphic rocks and all eight which penetrated below about 60 m (200 ft) struck water, the remaining three were stopped at shallow depths for unknown reasons. According to Baker (1963) the water (in the NE) is of good quality with high yields of up to 22 800 gallons per day. Two boreholes into the eastern volcanics, and a third which penetrated into subvolcanic sediments, all struck water. A shallow (30 m) borehole into the sandy bed of the Barsaloi River initially yielded 896 gal/hour and is now fitted with a manual pump. All other operating boreholes are equipped with mechanical pumps which are constantly breaking down.

The nomadic populations rely on spring water or hand-dug wells into the sandy beds of the larger rivers, e.g. the Baragoi River (downstream of Baragoi), Barsaloi River, and Masiketa River (around Masiketa). Springwater is relatively common, either on the high basement terrains (the Karisia Hills or foothills of the Ndotos) or in the western volcanic terrain. Springs in the Karisia Hills have already been mentioned. Good quality water trickles out of the massive quartzofeldspathic rocks of the Ndotos at Lesirikan whilst 'oily' water is present at Lareodo in the Kileshwa River (from graphitic gneisses). To the west there are spring-fed pools in the Akangbelai River (in the extreme north); at GR 2836 1922, GR 2262 1837, GR 2239 1730 and GR 2249 1709 in the Baragoi River, Echua Etom, Kongia, at GR 2325 1735 and GR 2325 1849 in the Kangaurak area, Lukuto, in the Lopurr River, Losuk, Logobuk River; Lakwarsignen Valley north of Naramarsiro; Loruman at GR 2386 1571; Leparmara at GR 2334 1512; Nkabai at GR 2428 1482 and GR 2409 1473.

Future needs

The rural areas have an adequate supply of drinking water: their main problem is a lack of grazing. However, existing supplies could be improved by cleaning out and repairing the small earth of concrete dams; re-equipping, preferably with manual pumps, the boreholes which have been allowed to deteriorate; connecting springs to pipelines to disseminate the water supply over relatively larger areas.

Maralal has a water shortage problem for several reasons. Firstly the pumps attached to the water boreholes are continually breaking down, secondly the reservoir on the Nontoto dries up following long periods of low rainfall: and finally, the gravity-fed exploitation of the subsurface flow from the Loikas River is constantly being sabotaged by Samburu 'moran' spears. To solve the water problem the boreholes need to be cleaned and fitted with reliable pumps. The plastic pipeline from the Loikas River to Maralal should either be protected, or replaced by concrete piping.

Future water boreholes into basement rocks should always penetrate more than 60 m and avoid the massive migmatites and granites. There appears to be a good supply of groundwater (of uncertain quality) throughout the gneisses. Obvious targets for high yields are the major fracture zones, especially where they intersect or are followed by large watercourses. Boreholes on the volcanic terrain should either penetrate to basal sediments or gneiss basement, or intersect weathered or porous horizons in the volcanic succession: such include basaltic lavas, various tuffs or sediments.

11.2 THE SUGUTA VALLEY AND ITS WESTERN SHOULDER

Surface water

Perennial supplies of surface water are confined to the upper reaches of the Suguta River and the upper parts of its west bank tributaries: the Kasamanang, Katatarope, Lokwaleibit and Nasaken Rivers. The rivers are fed from cold and hot (saline) springs. No dams have been constructed to trap seasonal run-off. Seasonal pools remain in the shadier parts of large rivers. Seasonal pans and lakes are rare, although the northern Suguta Valley floods following heavy prolonged rains. The Kerio River may dry up between August and November. During the rains

Table 16 Water borehole data

Borehole number	Grid reference	Location	Major geological setting	Depth m	Water struck m	Rest level m	Yield (gallons/per day)
C1613	253 218	Kawab	Basement	123	69	66	16 800 (Baker, 1963)
C4449	251 217	Kawab	Basement	26	DRY	—	—
C1639	245 176	Marti	Sub-volcanic Sediment	103	70	61	22 800 (Baker, 1963)
C1553	247 178	Marti	Volcanic	137	71	60	3600 (Baker, 1963)
C1505	246 118	Maralal	Basement	106	26	23	3840
C1723	246 118	Maralal	Basement	123	35	30	2688
C3609	246 119	Maralal	Basement	72	38, 42	36	32 400
C3692	244 117	Maralal	Basement	50	DRY	—	—
C3833	241 115	Maralal	Volcanic	123	38	12	11 760
C479	244 118	Maralal	Basement	91	56, 70	11	2160
C3114	256 142	Lusien	Basement	123	30, 46, 65	25	21 504
C3651	275 198	Lesirikan	Basement	?	55	45	29 280
C3855	256 198	Baragoi	Basement	183	22	11	9600
C4105	121 197	—	Basement	25	DRY	—	—
C3461	184 113	Kapedo	Volcanic	134	108	101	23 040

this river may flood in the Lokori area, where water levels may suddenly rise by up to 10 m. Water trapped in the sandy infill to the larger rivers is generally palatable: it is exploited in shallow, hand-dug wells. Unfortunately this near-surface supply in the Suguta River is alkaline.

Groundwater

A shallow water borehole at Lokori augments supplies from wells into the sandy bed of the Kerio River. A water borehole has also been sunk at Losikiriomo in the Amaya embayment (GR 1908 1147). No details concerning this borehole are available; it is not known whether it has been equipped. The borehole at Kapedo has a high yield of water of uncertain quality (Table 16).

Seepage of coldwater from springs have been noted at Kapedo, the Nasaken area, north side of Silali, north side of Emuruangogolak, and Kamuge. Hot saline water seepages occur at Kapedo (Plate 19), Lorusio (Plate 20), 3 km south of the confluence of the Kamuge and Suguta Rivers, Nasaken, western slopes of Silali and Elboitong (north-east escarpment). Exact localities of all springs are shown on the geological map.

Future needs

Very little grazing takes place in this area because of the hot arid climate, the lack of vegetation and the ruggedness of the topography. Therefore the few cold fresh-water springs suffice to meet the needs of nomads. Shallow, manually operated boreholes into the sandy beds of the Kerio and Suguta Rivers could augment existing supplies in the few settlements.

11.3 GEOTHERMAL ENERGY

In addition to the hot springs in the Suguta Valley mentioned in Section 11.2 there are also steam jets on the north side of Paka (at Cheptomas) and fumaroles on Silali. Weaver (1977a) makes no mention of hot springs or fumaroles on Emuruangogolak but in view of its very recent volcanicity they must almost certainly be present. Kapedo is locally famous for its waterfall of hot water fed from several springs up to several kilometres upstream (Plates 19 and 20). Therefore there is considerable potential for geothermal energy within the median graben and adjacent parts of the Gregory Rift in Degree Square 27.

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