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**GEOLOGY, GEOCHEMISTRY, AGE AND TECTONIC SETTING
OF THE GORE-GAMBELLA PLUTONIC ROCKS,
WESTERN ETHIOPIA**

by

Teklewold Ayalew Alemayehu, B.Sc., M.Sc.

**A thesis submitted to
the Faculty of Graduate Studies and Research
in partial fulfilment of
the requirements for the degree of
Doctor of Philosophy**

Department of Earth Sciences

Carleton University

Ottawa, Ontario

December, 1988

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Teklewold Ayalew Alemayehu, B.Sc., M.Sc.
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for the degree of Doctor of Philosophy

.....
Chairman, Department of
Earth Sciences

.....
Thesis Supervisor

.....
External Examiner

ABSTRACT

In a transect across the Birbir and Baro domains of western Ethiopia, distinct granitoid suites are recognized on the basis of their field relations, petrology, chemical features and age. The Baro Domain consists of migmatitic, upper amphibolite facies gneisses and meta-leucogranites. The Birbir Domain consists of lower amphibolite facies rocks with abundant intrusive and meta-intrusive rocks of mafic and intermediate composition. A ductile, transcurrent fault system, the Birbir Shear Zone, traverses the Birbir Domain. Kinematic indicators such as disrupted dykes and sills within the shear zone suggest major dextral movement which was succeeded by sinistral movement during its final stage. The pre- to syn-kinematic intrusives within the Birbir Domain are metamorphosed and mylonitized to variable degrees. Geochemical and isotopic data from early plutonic units in the Birbir Domain reflect arc-type igneous activity; late- to post-kinematic plutons are more alkalic and of intraplate character.

U-Pb zircon and Rb-Sr whole-rock isochron dates show plutonic activity between 830 and 540 Ma. A whole-rock Rb-Sr date of 760 Ma from a pre- to syn-kinematic pluton coincides with the age of low-grade metamorphism of arc-related rocks of the Red Sea Hills of NE Africa and the Jeddah terrane of Arabia. A U-Pb zircon age of 783 Ma was obtained from leucogranite sheets of the Baro Domain, considered to be the products of anatexis. A 635 Ma whole-rock, Rb-Sr date is considered to be a time of isotopic equilibration probably associated with major transcurrent movement. A U-Pb lower

intercept date of 582 Ma from the Baro leucogranite represents metamorphic overgrowth and coincides with the estimated time of amphibolite facies metamorphism in the Mozambiquian rocks of NW Kenya. Late- to post-kinematic intrusions at 570 and 550 Ma place a minimum age limit on deformation and metamorphism.

The Birbir Domain is a southward extension of the Pan-African crust of NE Africa and Arabia. The Birbir shear zone indicates a tectonically active continental margin along which magmatic arc rocks were accreted. The Baro Domain is interpreted as a reactivated pre- Pan-African continental margin linked to the Mozambique Belt of east Africa. A subduction model, involving closure of an ocean basin, is proposed for the evolution of rocks of the Birbir Domain. In a new synthesis, the Precambrian rocks of Ethiopia are divided into pre- Pan-African and Pan-African assemblages subdivided into nine major lithostructural units.

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CHAPTER 1

INTRODUCTION

1.1 GENERAL STATEMENT

The Precambrian Shield of western Ethiopia occupies a position of particular interest, lying at the interface between the predominantly gneissic terrane of the Mozambique orogenic belt to the south in East Africa and the volcanic-sedimentary-plutonic rocks of the north-east African Shield (Figs. 1.1 and 1.2). The north-east African Shield extends from eastern Egypt southward through eastern Sudan and Ethiopia and westward through Libya. The counterparts of the north-east African Shield found in the Arabian peninsula, Saudi Arabia and eastern Oman were once a segment of a formerly continuous orogenic belt, now separated by the Red Sea and the Gulf of Aden.

In western Ethiopia, the shield is made up of a mosaic of high-grade gneisses separated by lower-grade rocks. The contact and age relationships between adjacent rocks of contrasting metamorphic grade have been a source of controversy, and nowhere has the relationship between the two groups of rocks been conclusively resolved. This problem was first addressed by Shackleton (1967) who correlated lower-grade rocks from southern Ethiopia with similar rocks in Kenya. He considered these low-grade sequences to be younger, supracrustal rocks underlain by older metamorphosed and deformed basement. Mohr (1962), Kazmin (1971, 1972b) and Merla et al. (1973, 1979), amongst others, also proposed a basement - cover relationship. However, neither the physical relationships nor the geochronological work undertaken to date support this hypothesis.

Figure 1.1: Sketch map of the Precambrian rocks of the region showing geographical relationships between the western Ethiopian-NE African Shields (patterned), and rocks of the Mozambique Belt. WES: Western Ethiopian Shield. Small square in WES denotes study area.

Figure 1.2: Palinspastic sketch map of NE African, Arabian and western Ethiopian Shields showing the major tectonic terranes. Note the relationship between the Baro, Birbir and Geba domains of the study area and the distribution of the high-grade gneissic terranes and the lower-grade volcano-sedimentary terranes and ophiolitic rocks in the region. Note also the shift in geographic coordinates east and west of the Red Sea closure line. RSH, Red Sea Hills. Modified after Vail (1985).

In the Gore - Gambella area of western Ethiopia rocks belonging to both high- and low-grade assemblages are well-exposed and in mutual contact. An Ethiopian - Canadian Cooperative Geotraverse, carried out from 1984 to 1986, allowed systematic mapping of the area. This thesis was part of the supporting laboratory studies of the Geotraverse.

1.2 MOZAMBIQUE BELT AND THE PAN-AFRICAN EVENT

The Precambrian rocks of Ethiopia were originally regarded as the northward continuation of the Mozambique Belt (Fig. 1.3; Rogers et al., 1965; Cahen and Snelling, 1966; Cahen et al., 1984). The term "Mozambique Belt" was originally introduced by Holmes (1951), and subsequently re-defined by many authors, to describe a north-south trending complex, stretching sub-parallel to the present African east coast, superimposed on more ancient east-west structures. Cahen (1961), among others, described the Mozambique Belt as polycyclic, containing a number of stratigraphic units and exhibiting several periods of deformation. The Mozambique Belt contains early Proterozoic, and possibly Archean units, although most isotopic dates indicate late Proterozoic activity (Wendt et al., 1972; Martin and Porada, 1977; Black, 1984; Almond, 1984; Gabert, 1984).

Hepworth (1979) described the Mozambique Belt in Mozambique, Malawi, Tanzania, Kenya, Uganda and Ethiopia as containing numerous locally-named, lithostratigraphic groups which vary in lithology and metamorphic grade, from granulite and amphibolite facies gneisses and schists to low grade psammites and pelites. He restricted the use of the term "Mozambique Belt" to eastern Africa, with its possible extensions to the west and south: the Zambezi belt and the Damaran-Katangan. Clifford (1970, p. 13) suggested the

Figure 1.3: Sketch map of the northern part of the Mozambique Belt and rocks affected by the Pan-African event in NE Africa and Arabia, showing main structural trends. Solid arrows ABCDE represent known Mozambique Belt trends, and open arrows and dashed lines at GHIJK are inferred fold axis and axial surface trends. J is the Birbir Domain, N and O are the Baro and Geba gneisses of western Ethiopia respectively. L is the Adola zone in southern Ethiopia. M is the Surma mylonite zone. XY is the boundary between Pan-African rocks and pre-Pan-African crust (basement?). Modified after Hepworth (1979).

presence of "two types of zone within the Mozambique Belt, one "geosynclinal" and the other "vestigial" (rejuvenated basement). The NE African volcanic - sedimentary greenschist belts are considered examples of the geosynclinal type and the relatively high-grade gneissic terranes vestigial. Clifford's classification implies that the gneissic terranes are older than the lower-grade metamorphic rocks.

Rocks of the Mozambique Belt are supposedly affected by the "Pan-African" event (Kennedy, 1964) that involved the structural differentiation of Africa into cratons and mobile belts during the late Proterozoic and early Paleozoic (550 +/-100 Ma). Burke and Dewey (1973) proposed that the belt is the root of a Himalayan-type continental collision zone of the late Proterozoic age. Kroner (1977) pointed the transitional nature of the western margin and the continuity of lithology from the foreland into mobile belt.

Several discussions of the term "Pan-African", as related to the Precambrian shield of NE Africa and Arabia, have appeared in recent years but no consensus has yet been reached regarding its significance and its age. In contrast to the views held by Kennedy and others concerning the Pan-African event, the IGCP Project 164 (Al Shanti, 1979) used the term to define various cycles of magmatism, sedimentation and tectonism during the time span 1100-450 Ma. Jackson and Ramsay (1980) recommended the use of "Pan-African" as a general designation for a 550+/-100 Ma old "late Proterozoic pervasive crustal heating event" and suggested that local names be given to the various magmatic and metamorphic events or orogenic cycles. Gabert (1984) considered the Pan-African event of the Mozambique Belt as representing an aborted Hercyno-type orogeny, during which geothermal

gradients were raised over a large part of the Pangaeic supercontinent, resetting many isotopic systems. Dissenting views shared by Hepworth (1979) and Almond (1984) favour the use of the term Pan-African to define an age province or thermal event, without tectonic connotations.

Within recent years there has been considerable success in applying plate tectonic models to those areas affected by the Pan-African event. Numerous tectonic models have been proposed, most involving Wilson Cycles. Typical collision signatures such as sutures and ophiolites have been documented. Implicit in these models is a complex history of island arc formation and ocean closure that involved some combination of arc-arc, arc-continent and/or continent-continent collision. These models are supported by the thick calc-alkaline successions, and isotopic signatures that favour an ensimatic origin (Fleck et al., 1980) as well as the presence of ophiolitic rocks (Shackleton, 1977; Hussein et al., 1982).

These magmatic arc and ophiolitic assemblages constituted new additions to the continental crust and are considered to be of Pan-African age (Gass, 1981). Although individual models differ in detail, most investigators attribute cratonization of the Pan-African rocks to arc accretion (e.g. Shackleton et al. 1980; Gass, 1977, 1981; Vail, 1983, 1985). However, many workers are still skeptical about such broad generalizations. The absence of blue-schist assemblages, the longevity and episodic character of magmatic activity and lack of paleomagnetic data fail to add support to such a model.

It is also apparent that Pan-African mobilization has affected areas extending several hundred kilometres west of the margins of the known Pan-African rocks. Older rocks, unaffected by the Pan-African event, are recognized in the Central African Republic (the Congo craton); the western

limit of the late Proterozoic structures appears to extend through Sudan, west of longitude 34°E in a roughly N-S direction (Vail, 1985; Shackleton, 1986).

1.3 STATEMENT OF THE PROBLEM

There are still very many questions concerning the evolution of the western Ethiopian Shield, and its relationship to the NE African Shield.

Included among these are:

- 1 do the gneissic terranes in western Ethiopia represent an extension of the Mozambique Belt of East Africa?
- 2 do the small blocks of gneissic rocks represent fragments of pre- Pan-African continental crust?
- 3 do the ultramafic bodies represent remnants of ophiolite complexes? If so, do the sutures mark fossil subduction zones?
- 4 are the lower-grade, volcanic-plutonic-sedimentary rocks parts of magmatic arcs, and if so are the arcs continental or oceanic?
- 5 what was the time at which new crust was generated during the Pan-African event?

1.4 PREVIOUS WORK

Geological mapping of the Precambrian rocks of Ethiopia is fairly localized and consists mainly of either reconnaissance traverses or detailed surveys commonly associated with mineral prospecting sites, e.g. Libling and Nowack (1939), Dainelli (1943), Gherardi (1951), Usoni (1952), Molly (1960), Augustithis (1965), Jelenc (1966), Baines and Duesing (1966), Levitte and Kent (1968), Ball (1971), UNDP (1972), Nippon Mining Company (1974), and

Ciarke (1978).

Mohr (1962) summarized the results of early reconnaissance geologic mapping of the Precambrian rocks of Ethiopia, and Rogers et al. (1965) undertook the first K-Ar dating programme of Precambrian rocks of the region. Gilboy (1970) and Chater (1971) gave detailed accounts of the petrology, metamorphism, structure and geochronology of the Precambrian of the Gariboro and Megado region in southern Ethiopia. Both concluded that Precambrian rocks within southern Ethiopia are similar to those that crop out elsewhere within the northern portion of the Mozambique Belt and proposed a regional correlation between the two. Beyth (1971) and Morton (1975) also published detailed field relations of various parts of the Precambrian rocks of Ethiopia. Kazmin (1971, 1972b, 1975a and b) proposed stratigraphic subdivisions and, for the first time, attempted a tectonic synthesis of the Precambrian in Ethiopia. In 1973, the Geological Survey of Ethiopia published the first geological map of Ethiopia, at a scale of 1:2,000,000. The National Research Council of Italy also published a geological map of Ethiopia and Somalia (Merla et al., 1973, 1979). A reconnaissance geological and geochemical survey of southwestern Ethiopia was supported by CIDA (Canadian International Development Agency; Omo River Project Team, 1973, 1976; Davidson, 1983).

The relatively high-grade metamorphic rocks of western Ethiopia were considered by most as basement, equivalent to the Mozambique Belt rocks, reworked during the Pan-African event. Arguments were based mainly on field relationships, such as different styles of deformation, metamorphic changes and truncation of major rock units. In recent years, however, such views have been critically evaluated in terms of Precambrian plate tectonics. New

concepts of regional tectonics and the significance of ophiolite occurrences have been discussed by Kazmin (1976, 1978a and b), de Wit (1977), de Wit and Berg (1977), Kazmin et al. (1978, 1979), and de Wit and Chewaka (1981).

Interest in the Precambrian rocks of the Gore - Gambella area was stimulated by regional geological mapping of the study area by Teferra and Berhe (unpublished manuscript) of the Ethiopian Institute of Geological Surveys. An important account of the geology of the study area was also given by de Wit (1977).

1.5 PRESENT STUDY

The present study is part of a co-operative research project, the Gore-Gambella Geotraverse, undertaken by the the Department of Geology, Addis Abeba University and the Ethiopian Institute of Geological Surveys in collaboration with the faculty and staff of the Ottawa - Carleton Geoscience Centre. The geotraverse was financed by the International Development Research Centre of Canada.

Field operations began in February, 1984. They extended over two four month field seasons in 1984 and 1985, and 6 weeks in 1986. This thesis is based on work carried out during the traverse and deals mainly with the intrusive and meta-intrusive rocks in the area. Mapping was conducted using 1:50,000 aerial photographs, and strike-normal traverses were made at intervals of approximately 2 to 4 km. The author was involved in mapping during the entire first field season and part of the second field season. Most of the author's second field season involved detailed mapping of several plutonic units and sampling for geochronological and chemical studies. Laboratory work was undertaken at Ottawa - Carleton Geoscience

Centre and the Geological Survey of Canada.

Laboratory studies included:

1. Petrographic study of over 200 samples from all rock types of the study area.
2. Major and some trace element analysis of whole-rock samples using X-ray fluorescence.
3. Determination of trace elements, including rare earth elements (REE), by instrumental neutron activation analysis (INAA).
4. Electron microprobe analysis of some minerals from the intrusive and meta-intrusive rocks.
5. Rb-Sr and U-Pb geochronology.

1.6 LOCATION, PHYSIOGRAPHY, ACCESS AND CLIMATE

The Gore - Gambella Geotraverse area is situated between $34^{\circ}45'-35^{\circ}15'E$ and $8^{\circ}00'-8^{\circ}30'N$, in the western Ethiopian Shield. It lies in the central part of Illubabor administrative region, and covers an area of 2850 square kilometres (Fig. 1.4).

The area lies in the western fringes of the Ethiopian Highlands, which rise to altitudes greater than 2000 metres and form an extensive plateau massif, bisected by the main Ethiopian and Afar rift systems. The highlands generally slope to the west along the major course of the Baro river, that flows through the centre of the study area. The eastern limit of the area is an elevated, sub-mature plain, incised by streams draining to the Baro and Birbir rivers. Thin erosional remnants of Tertiary basalts are underlain by Precambrian rocks. The incised terrain merges westward into the Gambella

Figure 1.4: Location map of the Gore - Gambella Geotraverse area.

mature lowland plain, which is also underlain by Precambrian rocks and scattered Tertiary volcanic outliers, with an extensive cover of Quaternary alluvium.

An asphalt road from Addis Abeba to Metu, and an all-weather gravel road from Gore to Gambella run through the area (Fig. 1.4). The northwestern limits of the area are reached via Dembi Dollo by four-wheel drive vehicles. The remaining areas are virtually inaccessible apart from travel by foot. Most traverses were conducted on foot with the help of local labour. The highlands are inhabited by the Oromo people, who subsist on agriculture, and the lowlands by the Mesengo, a semi-nomadic people and the Anuwak, who farm along the river courses.

Insufficient data limit the precise determination of the volume and intensity of rainfall and the variations in temperature in the study area. The eastern and northeastern part of the area have seven rainy months from April to October, with a mean annual rainfall of 1400-1800 mm. Daily maximum and minimum temperatures range from 12 to 25°C, and a humid moisture level is reported (Gamechu, 1977). In contrast, the western lowland is dry and seasonally very hot. During the field work, between February and May, daytime shade temperatures in excess of 40°C were recorded, and rainfall has been estimated at 600 mm annually, falling mainly between May and September.

CHAPTER 2

GEOLOGY OF THE GORE - GAMBELLA AREA

2.1 INTRODUCTION

The area mapped during the Gore - Gambella Geotraverse consists of a 55 km section across strike that straddles the boundary between three litho-structural domains, identified on the basis of previous mapping (Teferra and Berhe, unpublished). The domains are, from west to east: 1) the Baro Domain, 2) the Birbir Domain and (3) the Geba Domain (Fig 2.1).

Lithologic assemblages in the Birbir Domain include metamorphosed pelite, wacke, volcanoclastic debris flow deposits, volcanic rocks, mylonite and voluminous plutonic rocks. The Baro and Geba domains consist of heterogeneous ortho- and paragneisses and migmatites. Along the eastern edge of the Baro Domain metasedimentary schists and gneisses occur that include pelites, calc-silicate rocks and amphibolites.

Field relationships and lithological descriptions of the map units are provided in this chapter. The map units are mainly composite, and names assigned to them are based on lithology. The order of many of these units in the legend is arbitrary, as their age relationships are unknown. The intrusive rocks, however, can be grouped according to degree of deformation and contact relations into pre- to synkinematic, and late- to post-kinematic plutons. The mapping units are divided into two broad groups of (1) intrusive and meta-intrusive rocks and (2) metamorphosed volcanic, sedimentary and hypabyssal rocks, and gneisses.

Intrusive and meta-intrusive unit names are given on Figure 2.2*. Modal mineralogy of selected samples from the intrusive and meta-intrusive rocks

Figure 2.1: Geological map of subsheets O, P, U, V, Sheet NC36-16. Heavy ornamented lines: domain boundaries. Order in legend of some mapping units is arbitrary (after Moore et al., 1987)

Figure 2.2: Geological map of the Gore-Gambella Geotraverse area showing main structural trend lines and sample locations for the intrusive and meta-intrusive units. (Appended in pocket).

Figure 2.3: Sketch map of the intrusive rocks and sample localities. (Appended in pocket).

Scale

5 0 5 10 km

Location map

are listed in Table 2.1 and sample localities shown on Figure 2.3^{*}. Detailed petrographic descriptions for representative samples from the plutons are given in Appendix 1. Typical mineral assemblages of the metasedimentary and volcanic rocks, and gneisses are shown in Table 2.2.

2.2 META-INTRUSIVE AND INTRUSIVE ROCKS

2.2.1 Pre- and/or syn-kinematic plutonic units

2.2.1.1 Bure pluton (mgb) ^{**}

The Bure pluton forms an elongate mass, well exposed south of Bure village. Here it forms ridges that extend SSW into the Baro River valley (Fig. 2.2). The pluton consists of hornblende gabbro and diorite with minor lenses of talc, chlorite and amphibole schist. The predominant rock is dark grey, fine to medium grained, and equigranular with a color index between 35 and 60. It shows faint compositional layering. Foliation parallels the contacts and is concordant with foliation in the country rocks. Plagioclase laths and hornblende prisms are progressively transformed into lenticles that define the foliation. Primary subhedral-granular and porphyritic textures are locally well-preserved.

In thin section, subhedral-euhedral plagioclase (An₆₀₋₇₀), 0.5 to 4 mm in length, makes up 30 to 60% by volume. Hornblende, the principal mafic mineral, can make up to 60% by volume of the modal minerals. Some hornblende has been partly replaced by fibrous actinolite. Hornblende

^{*} Figures 2.2 and 2.3 are in the back pocket.
^{**} map unit names and symbols as on Figure 2.2.

Table 2.1 (continued)

Locality Sample No.	Baro		4M	Mao		Bonga		
	4F	8F		7O	11O	6L	5K	8K
Quartz	38	36	39	26	29	15	32	33
Plagioclase	22	18	20	30	28	31	20	22
K-feldspar	32	38	33	37	32	42	40	38
Biotite	4	3	-	5	4	6	4	4
Hornblende	-	-	-	1	1	2	2	2
Muscovite	-	-	5	-	-	-	-	-
Epidote	+	+	-	+	+	-	-	-
Garnet	-	-	3	-	-	-	-	-
Apatite	+	+	+	+	+	+	+	+
Allanite	-	-	-	+	+	+	+	+
Titanite	+	+	-	+	+	+	+	+
Zircon	+	+	+	+	+	+	+	+
Calcite	+	+	+	+	+	-	-	-
Opaques	+	+	-	+	+	+	+	+

commonly occurs as recrystallized granular aggregates, although there are a few euhedral-subhedral phenocrysts. Sericite, epidote and calcite are found along cleavage traces and the margins of plagioclase crystals. Hornblende is commonly chloritized. Quartz, apatite and titanite are accessory minerals. Interstitial magnetite and ilmenite occur sparingly, and some samples contain trace amounts of clinopyroxene and biotite. Biotite, present in the strongly altered samples, has replaced hornblende.

2.2.1.2 Haya River pluton (mtn)

Along strike and spatially associated with the Bure gabbro (mgb) is a strongly deformed and metamorphosed pluton made up of tonalite and granodiorite (Fig. 2.2). The pluton, for the most part, is irregular in shape and shows elongation parallel to the regional NNE structural grain. The internal fabric is highly variable, from weakly to strongly foliated, over a few centimetres to hundreds of metres. Contact relationships with the adjacent rock unit (mvs) are not clear due to the strong degree of deformation.

The main rock type in the Haya River pluton is grey, fine to medium grained, and locally shows pervasive foliation, and compositional layering. The latter consist of lamellar aggregates of biotite and muscovite, about 1 cm thickness, alternating with granoblastic aggregates of plagioclase and quartz, minor microcline, and some garnet.

A granoblastic texture can be seen in thin section. Plagioclase forms 45 to 55% of the rock, and occurs as anhedral crystals between 2 to 5 mm in length, commonly altered to sericite, granular epidote and rare calcite. Microprobe analyses of the plagioclase feldspar show reverse zoning with sodic cores (An_{28}) and calcic rims (An_{32}). Quartz, 0.5 to 3 mm in size,

makes up 25 to 30% of the rock, and occurs mainly as anhedral grains with elongate irregular boundaries. Large grains show undulose extinction and, locally, show a preferred orientation. Anhedral microcline grains, 0.5 to 2 mm in diameter, constitute up to 10% of some samples. Biotite, the main mafic mineral, forms flakes 0.5 to 1 mm long commonly mantled by muscovite. Biotite and muscovite typically occur in association with fine, granular epidote. Euhedral epidote, <1 mm across, occurs in close association with biotite and plagioclase. Accessory minerals include apatite and titanite.

2.2.1.3 Birbir Intrusive Complex (mqd)

Rocks of the Birbir intrusive complex occupy the western half of the Birbir Domain, and are deformed to varying degrees in a major north-south, ductile, shear zone. Lenticular bodies of quartz diorite, with subordinate gabbro, diorite and leucogranitic rocks alternate with narrow, heterogeneous mylonitic units. Three strongly sheared intrusive sheets, 0.5 to 3 km wide, alternate with rocks of unit (mvs). The intensity of shearing varies over a few metres: strongly sheared rocks contain less-deformed zones, which retain igneous textures, on the scale of a metre to tens of metres. In the northern part of the map area, rocks of (mqd) grade into predominantly mafic schist and mylonite (mvs). East of the Baro-Birbir River confluence, the quartz diorite forms thin layers of variably sheared rocks, less than a metre to several hundred metres thick, that alternate with abundant, variably deformed, mafic and felsic dykes and sills. To the west, the unit is homogeneous and mylonitization is less severe than in other parts of the Intrusive Complex. The quartz diorite contains elongate xenoliths of microdiorite (Plate 2.1A).

The main rock type of the Birbir intrusive complex is a grey to dark grey, medium grained, with a well-developed foliation defined by biotite and hornblende streaks, and flattened quartz and plagioclase laths. Mineral lineations, common throughout the pluton, are defined by quartz rods and mafic streaks. In thin-section, the variable degree of deformation is expressed by a wide range of grain sizes, grain boundary types and grain shapes. The rock is composed of andesine - oligoclase, quartz, biotite, K-feldspar, and variable amounts of hornblende and epidote. Accessory minerals include ubiquitous, euhedral apatite, titanite and zircon. Fe-Ti oxides and sulfides occur as inclusions, mainly in biotite and hornblende.

Plagioclase occurs as subhedral or euhedral porphyroclasts, up to 6mm in size, and as much smaller anhedral grains in the matrix. The porphyroclasts are commonly fractured and bent, mantled by fine grained crystals, and slightly sericitized. Microprobe analyses of plagioclase show zoning. The cores are calcic (An_{36-49}) and the rims sodic (An_{29-32}); crystals contain small antiperthitic patches of K-feldspar.

Most quartz and feldspar form a fine-grained, polycrystalline matrix; some quartz grains are flattened parallel to foliation, or make up ribbons wrapped around feldspar porphyroclasts. Biotite commonly occurs as dark green to brownish green, subhedral plates with bent cleavage traces; it also forms lamellar aggregates molded around porphyroclasts of oriented hornblende and plagioclase. Hornblende forms euhedral to subhedral, poikiloblastic crystals with well developed pleochroism.

Epidote occurs both as intergranular discrete subhedral to euhedral crystals between biotite, hornblende and plagioclase grains, and as fine granular aggregates resulting from alteration of plagioclase and mafic

minerals. Euhedral epidote is characteristic of the Birbir (mqd) and to a lesser degree, the Haya pluton (mtn). The epidote shows no reaction relationship with other minerals, and is similar to magmatic epidote described by Zen (1985) and Zen and Hammarstrom (1984, 1986). Nancy (1981) presented the results of experimental data showing that epidote can crystallize from melts of intermediate composition only at pressures of about 8 kb, equivalent to depths of approximately 25 km. Assuming the epidote in the Birbir quartz diorite is magmatic the Birbir intrusive complex was seemingly formed at deep crustal levels. More detailed investigation would be however required in order to demonstrate conclusively whether epidote is a primary phase in the Birbir rocks.

2.2.1.4 Goma pluton (mgd)

The Goma pluton crops out as a prominent, broad north-trending ridge, rectangular in plan, with dimensions 10 by 4 km. It cuts metamorphosed volcanoclastic rocks, turbiditic wackes and pelites along its northern contact, and gabbroic rocks to the southeast. The pluton is generally massive and is cut by narrow shear zones along its western margin. The contact, where observed, is sharp and the pluton cuts the regional foliation along its northern contact. Foliation of the country rock is deflected around the southwestern corner of the pluton and is truncated by a major shear zone which forms the western boundary of the pluton (Fig. 2.2). The contact relationships between the pluton and country rock indicate that structures occur in the country rock that predate the emplacement of the pluton. The pluton, however, predates the mylonitic foliation.

Fine-grained, mafic xenoliths, 20 cm to 0.5 m in size, and mafic

mineral clots occur locally along the eastern margin. Most are spheroidal and a few are oblate. The xenoliths have sharp margins and no tectonic fabric.

The pluton consists of subhedral-granular, rarely blasto-porphyritic, granodiorite containing plagioclase, K-feldspar, quartz, and minor biotite. Large (4-6 mm), subhedral-euhedral oligoclase (An_{15-28}) laths are zoned and strongly altered to coarse epidote, sericite and chlorite. Locally, quartz occurs as polycrystalline aggregates that show strong undulose extinction. Most K-feldspar occurs as subhedral microcline grains up to 5 mm across; a few are perthitic, and some poikilitically enclose plagioclase laths. Plagioclase is altered, commonly along cleavage traces, to fine-grained muscovite and granular epidote. Biotite is extensively altered and intergrown with chlorite and K-feldspar. The chlorite is commonly associated with rutile. The latter mineral encloses exsolved opaques with titanite rims. Other accessory minerals include allanite, apatite, titanite, and zircon. Allanite occurs both as euhedral, and some metamict crystals; the latter are corroded and embayed, with discontinuous overgrowths of colourless to yellow anhedral epidote.

2.2.2 Syn-kinematic granite

2.2.2.1 Garnet-muscovite granite (mgt)

This intrusive unit is distinctive, although it is not volumetrically important. It forms concordant sheets and pods, from 2 to about 100 m across, within metagranodiorite that lies along strike from quartz-muscovite schist and conglomerate (qms) at the eastern edge of garnet-sillimanite gneiss (gsg) of the Baro Domain (Fig. 2.1). The main rock type is pale

reddish brown and medium-grained with a characteristic sub-horizontal stretching lineation parallel to the north-south trending structural grain. In places the rock is pegmatitic, and aplitic varieties have also been observed. There is a strong spatial association of these granites with late pegmatitic dykes and veins that clearly crosscut the regional foliation.

The rock contains quartz, microcline perthite, plagioclase (An_{15}), muscovite and garnet. Muscovite occurs as euhedral isolated plates, <3 mm in size; garnets occur both as discrete euhedral grains up to 1 mm and as small polycrystalline aggregates scattered throughout the rock.

2.2.2.2 Baro leucogranite (mgt)

Within the Baro Domain, sharp crested ridges are underlain by lenticular leucogranite bodies (Figs. 2.1 and 2.2). Observed contacts are parallel to foliation in the country rock, and have gentle dips. The rock is pink to light gray, medium-grained, generally equigranular, and locally contains enclaves of garnet-bearing, quartzofeldspathic gneiss. One of these intrusive sheets, located 3 km west of Bonga, is about 200 m thick and over 6 km long. It is strongly foliated along its margin, although relict subhedral-granular textures are preserved in its interior. Foliation is defined by biotite streaks as well as flattened quartz and feldspar aggregates.

Thin sections of rocks from this body contain sodic plagioclase (An_{3-7} by microprobe analysis), microcline perthite, quartz, minor biotite and hornblende. Accessory minerals include ubiquitous magnetite, ilmenite mantled by leucoxene, haematite, titanite, zircon and apatite. In other similar plutons in the area, magnetite forms the predominant mafic mineral.

2.2.3 Late- to post-kinematic plutons

2.2.3.1 Mao pluton (gt)

The Mao pluton, in the north - central part of the map area, forms an oval-shaped body, 9 km long and 5 km wide. It locally exhibits well-developed shear bands along its eastern margin and truncates mylonites in units (mss) and (mvs). It contains deformed dioritic xenoliths, and is cut by aplitic dykes and quartz veins. The rock is medium-grained, buff-pink to light grey and consists of plagioclase, microcline, quartz, biotite and minor hornblende. The eastern margin is generally more mafic than the interior parts and is typically coarser grained. The central parts contain large feldspar augen, up to 1 cm in diameter, surrounded by anastomosing biotite lamellae.

In thin sections, most undeformed samples have a porphyritic to subhedral-granular texture, in which large (5-6 mm long) subhedral to anhedral microcline and subhedral plagioclase (An_{15-20}) grains are surrounded by a matrix of fine-grained, plagioclase, microcline, quartz, biotite and hornblende. Microcline is commonly perthitic, containing fine albite stringers. Plagioclase is associated with marginal myrmekitic intergrowths and rare antiperthite. More deformed samples show quartz and feldspar with undulose extinction, kinked twin lamellae in plagioclase, microcline mantled by fine-grained recrystallized feldspars, sub-grain development in quartz and feldspar and a weak foliation defined by biotite. Biotite and hornblende occur as large discrete grains with ragged margins, and less commonly are poikilitically enclosed within the larger feldspars. Biotite is partially replaced by chlorite, stilpnomelane and opaque minerals, hornblende by chlorite, and feldspars by sericite, epidote and calcite.

2.2.3.2 Bonga pluton (gt)

This pluton is one of three undeformed, discordant granite stocks that lie along the boundary between the Birbir and Baro domains, north and south of the Baro River (Figs. 2.1, 2.2), ENE of Bonga. It comprises two nested intrusions: a hornblende-biotite granite encloses a quartz monzonite. Intrusive relationships suggest that the core-forming quartz monzonite is the earlier of the two types. The granite contains randomly distributed angular gneissic xenoliths. At its southeastern margin the granite is typically a heterogeneous mix of granodioritic gneiss, intrusive breccia with fragments of gneiss, and homogeneous granite. The southwestern part of the pluton exhibits sharp, discordant contacts. The core-forming quartz monzonite contains abundant rounded mafic xenoliths. Both the xenoliths and the host rock show a Rapakivi texture with K-feldspar megacrysts, up 5 cm in size, mantled by plagioclase (Plate 2.1B). In one locality a quartz-feldspar orbicule, about 20 cm in diameter, was observed in the quartz monzonite.

The quartz monzonite consists of coarse (up to 4 cm), subhedral microcline perthite phenocrysts, partly mantled by plagioclase, and quartz megacrysts (xenocrysts ?) set in a fine to medium grained matrix of interlocking anhedral quartz, subhedral plagioclase (An_{22}), K-feldspar, biotite and hornblende. Accessory minerals include allanite, apatite, magnetite, titanite and zircon. In some cases, feldspar megacrysts partly enclose tabular plagioclase and quartz. The presence of megacrysts in both the xenoliths and the matrix suggests that they are metasomatic in origin.

The pink granite is massive and porphyritic with coarse K-feldspar

phenocrysts set in a fine- to medium-grained subhedral-granular groundmass; the dominant mineral is microcline perthite followed by quartz, sodic oligoclase (An_{12} by microprobe analysis), biotite and hornblende. Accessory minerals include allanite, apatite, titanite and zircon. The plagioclase is altered to sericite, calcite and granular epidote along cleavage traces. Allanite grains are euhedral and zoned.

2.2.3.3 Gabbro, diorite and hornblendite (gb, hbt)

In the northern part of the study area, west of the Gebari River (Fig. 2.2), circular bodies of relatively undeformed, mafic rock occur. These consist of medium to coarse-grained, granular gabbro, containing plagioclase, clinopyroxene, orthopyroxene, olivine, hornblende, and Fe-Ti oxides. Zoned plagioclase (2-4 mm in size) makes up about 40% of the modal minerals and is slightly altered to sericite and epidote. The clinopyroxene (augite) is usually in equant granular aggregates with less abundant orthopyroxene and olivine and typically is surrounded by a narrow marginal zone of hornblende. Opaques are mainly skeletal Fe-Ti oxides.

Within the western part of the Birbir complex (mqd) there are lenses and dykes of hornblendite and hornblende gabbro (hbt; Fig.2.1) 10-100 m wide. Geologic contacts for unit (gb), particularly in the central and southern part of the area, are approximate as the rocks are difficult to separate from the least-sheared rocks of unit (mqd). The virtual absence of deformation in these rocks, their spatial association and mafic composition suggest a genetic relationship between the undeformed gabbro and the dykes. These mafic rocks mark a discrete igneous episode that post-dates any regional deformation.

2.3 METAMORPHOSED SEDIMENTARY, VOLCANIC AND HYPABYSSAL ROCKS; GNEISSES

2.3.1 Metasedimentary schist (mss)

These rocks underlie much of the area east of the north-south stretches of the Birbir River, and extend eastward, northwest of Bure where they are overlain by Tertiary basalts. They consist of buff-grey, weakly to strongly schistose, well-preserved metawacke and pelite that underlie the broad valley west of Bure; in areas immediately south of the Mao pluton they are mylonitic. The metasedimentary rocks are interlayered with sheets of mylonitic quartz diorite and biotite-hornblende schist, and are clearly cut by the Mao and Goma plutons. Strong deformation obscures the contact relationships with the mylonitized plutonic rocks.

The typical mineral assemblage of rocks of unit (mss), shown in Table 2.2 is quartz-plagioclase-biotite-muscovite-epidote-calcite. Large porphyroblasts of andalusite (Plate 2.1C) and staurolite, pseudomorphs of cordierite (?) and traces of garnet and graphite occur locally. Lenses of polymictic, poorly-sorted and sub-rounded, matrix-supported coarse volcanoclastic rocks (Plate 2.1E) occur in contact with the Goma pluton along its NNE margin. The clasts are mainly rhyolite and andesite. Within unit (mss), marble occurs as thin interlayers, up to 6 m wide. At one locality, about 4 km NW of the Mao pluton, a large body approximately 1 km wide of fine- to medium-grained crystalline carbonate rock can be traced for at least 3 km.

The metawackes commonly show lenticular structures, mineral lineations, and feldspar augen wrapped by micaceous minerals. Primary sedimentary features such as bedding, cross-lamination, load casts, and graded bedding

Table 2.2: Typical mineral assemblages of the metasedimentary and volcanic rocks, and gneisses.

UNIT	Qz	Pc	comp.	Kf	Bi	Hb	Ep	Mu	Ch	Ti	Cc	Ox	Others
mss	X	X	ol	X	x	-	x	x	x	-	-	x	st,ad
mvs	x	X	ol-an	x	-	X	x	-	(x)	x	x	x	ac,ap
mvs	X	X	ol	-	x	X	x	(x)	-	-	-	x	
mvs	X	X	ol	X	x	-	x	x	-	-	x	x	
qms	X	X	ol	-	x	-	x	x	x	-	-	x	
mqs	X	X	ol	X	x	-	-	X	-	-	-	x	si
csg	X	X	ol-la	X	-	x	x	-	-	x	x	-	cp(>10%), sc
gsg	X	X	ol	x	X	-	-	(x)	-	-	-	x	ga,si,co,gd
hbg	X	X	ol	x	X	x	x	-	-	-	-	x	
big	X	X	ol	X	X	x	-	-	-	x	-	x	

NB: X: major mineral (>10%); x: minor mineral (<10%); (x): retrograde mineral

Mineral abbreviations: Qz, quartz; Pc, plagioclase; Kf, K-feldspar; Bi, biotite; Hb, hornblende; Ep, epidote; Mu, muscovite; Ch, chlorite; Ti, titanite; Cc, calcite; Ox, oxide; ol, oligoclase; an, andesine; la, labradorite; gd, anthophyllite (gedrite); ga, garnet; co, cordierite; si, sillimanite; sc, scapolite; ap, apatite; cp, clinopyroxene.

are common (Plate 2.1C). The sedimentary structures indicate that turbidity flows have been responsible for some of the sediments. Associated coarse volcaniclastic rocks suggest rapid deposition in a shallow water environment close to an area being uplifted and rapidly eroded. The slight rounding of the clasts may require subaerial transport and/or shoreline abrasion and imply a subaerial volcanic source area. The predominance of wackes and absence of sedimentary rocks of continental origin suggest that the succession is similar to that found in present-day, intra-oceanic island arc environments (Mitchell and Reading, 1978).

2.3.2 Metavolcanic schist and mylonite (mvs)

In the flat-lying area east of the Birbir River, metavolcanic rocks and mylonites occur between sheets of metaquartz diorite (mqd) and metasedimentary rocks (mss), on the scale of hundreds of metres thickness. North and south of Bure, metavolcanic rocks, mainly basic in composition, are interleaved with sheets of the Bure pluton (mgb). The main metavolcanic rock is a fine-, to medium-grained, hornblende/actinolite-plagioclase-epidote-calcite schist, commonly interlayered on a centimetre to metre scale with minor quartz-muscovite-sericite, and actinolite schists. Primary igneous textures include amygdules (Plate 2.1D) and plagioclase laths and hornblende phenocrysts in rocks of andesitic composition. Most mylonitic rocks have a lenticular structure, commonly with mineral lineation. Near the Baro-Birbir river confluence, mafic and felsic dykes cut the unit and these are disrupted by shearing. de Wit (1977) interpreted these rocks as part of a dismembered sheeted-dyke complex, while others have attributed them to multiple dyke injection synchronous with shearing (S.K. Hanmer and J.M. Moore, pers. comm. 1987).

2.3.3 Garnet-sillimanite gneiss (gsg), calc-silicate gneiss (csg) and muscovite bearing gneiss (msg)

The garnet - sillimanite gneiss unit underlies the area east of the Bonga River (Fig. 2.2). It consists of interlayered aluminous and calc-silicate gneisses and subordinate amphibolite. The main exposures occur near Bonga village and consist of sillimanite and/or garnet-bearing layers, up to tens of metres thick, associated with grey to pink biotite gneisses and schists. The sillimanite-bearing rocks commonly contain fibrolitic knots, a few centimetres in diameter, of intergrown sillimanite quartz and plagioclase (Plate 2.1F), and large garnet porphyroblasts. Locally, garnets up to 5 cm in diameter constitute as much as 60% of the rock and are intergrown with quartz, biotite, and sillimanite. Cordierite-orthoamphibole (gedrite)-bearing rocks are interlayered with the sillimanite-garnet-biotite gneisses. East of unit (gsg), a narrow belt of muscovite-bearing gneisses (msg) occurs that probably represents a downgraded equivalent of unit (big) to the west.

East of Bonga, along the eastern boundary of the sillimanite gneiss, is a well-exposed section of calc-silicate gneiss (csg). The rock is fine- to medium-grained, green-grey, weakly foliated, and commonly layered on a centimetre-scale. Mineral assemblages in these rocks include garnet-epidote-hornblende-diopside, garnet-epidote-microcline-biotite, and biotite-plagioclase-hornblende with epidote-rich bands (epidosite).

2.3.4 Quartz - muscovite schist and conglomerate (qms)

This unit, forming a NNE-trending wedge that lies to the east of the muscovite-bearing gneiss (msg), consists of quartz-muscovite schist with

rare garnet and staurolite(?), in addition to subordinate layers of quartzite. Some of these rocks are chlorite-rich. The predominant lithology is fine- to medium-grained, leucocratic schist, containing quartz and plagioclase, with minor microcline and variable amounts of muscovite (5-20%). Thin section examination shows a clastic texture. A lens of quartz-pebble conglomerate, containing pebbles up to 2 cm in diameter within the quartz-muscovite schist suggests a clastic origin for rocks of unit (qms).

2.3.5 Biotite gneiss (big) and hornblende-biotite gneiss (hbg)

Most of the area west of Bonga consists of layered and non-layered gneisses. Dominant rock types are strongly foliated, pink to light gray, medium-grained biotite and hornblende-biotite gneisses. Quartzofeldspathic gneisses with minor quartzite and amphibolite layers also occur within the (hbg) unit. The gneisses typically consist of plagioclase, microcline, quartz, biotite and hornblende, locally with traces of muscovite and garnet. Relationships between associated rock types, particularly the intercalations of amphibolite and thin quartzite layers, are obscured by strong deformation. Some of the biotite and hornblende-biotite gneisses are migmatitic and are cut by numerous sub-concordant lenses of granitic and pegmatitic material. Discordant layers and pods of pegmatite also occur. A consistent and appropriate mineralogical composition suggest that these rocks are predominantly tonalitic or granodioritic orthogneisses.

Plate 2.1: Structural and metamorphic features of some map-units from the Birbir (A,B,C) and Baro domains (D).

- A- Mafic xenoliths in Birbir quartz diorite (mqd). Main road, 10 km E of Bonga (Fig. 2.2). Small divisions on scale card are centimetres.**
- B- Rapakivi texture in post-kinematic quartz monzonite, core of Bonga pluton (gt), 1 km north of Baro River. p: plagioclase, f: K-feldspar.**
- C- Cyclic layering of mudstone and wacke in unit (mss) showing graded bedding. Truncated cross-laminae are top indicators shown by the direction of arrow; large andalusite crystals have grown in pelitic layers. There is a weak cleavage at -25° to the layering. One kilometre north of main road, 11 km west of Bure. Pen magnet is 12 cm long.**
- D- Andesitic flows (mvs) containing flattened quartz amygdules (am). Two kilometres north of main road, 10 km west of Bure.**
- E- Coarse volcanoclastic rock (mss) viewed on surface perpendicular to lineation. vc = flow banded rhyolitic clast. Main road, 9 km west of Bure.**
- F- Sillimanite-quartz nodules in unit (gsg) near the eastern boundary of the Baro Domain with the Birbir Domain. Hammer handle 50 cm long, lies on foliation plane, parallel to shallow lineation. Three kilometres north of Baro River, 2 km east of Bonga.**

Plate 2.2: Some aspects of metamorphic mineral assemblages in pelitic rocks (A, B); and strain features in Birbir Shear Zone (C, D, E, F).

- A- Assemblage almandine-cordierite-gedrite-sillimanite from unit (gsg) of the Baro Domain. cd: cordierite, s: sillimanite, gd: gedrite, g: garnet. PPL. Scale bar is 0.5 mm long. Sample # UP49a, locality: Bonga.
- B- Staurolite porphyroblasts in fine-grained matrix of quartz, biotite, muscovite and chlorite in unit (mss). st: staurolite, cl: Fe-rich chlorite, b: biotite. PPL. Scale bar is 1 mm long. Sample # OY032, locality: 11 km west of Bure.
- C- Flattened mafic xenolith in weakly to moderately foliated Birbir quartz diorite (mqd). Ductile shear zone at top (s) truncates xenolith and has sinistral sense. Main road 10 km east of Bonga. Pen magnet is 12 cm long.
- D- Mylonitic extension lineation in (mqd) defined by aggregates of biotite and quartz, parallel to pen. Birbir River, 5 km north of confluence with Baro River.
- E- Well-defined mylonitic foliation in metagranodiorite (mgd) defined by biotite and hornblende streaks, and flattened quartz and feldspar. sb: shear bands, indicating sinistral shear sense. Two kilometres north of main road, 11.5 km west of Bure.
- F- Folded mylonitic foliation in Birbir quartz diorite (mqd). One kilometre north of main road, 12 km west of Bure.

2.4 METAMORPHISM

2.4.1 Mineral parageneses and P-T conditions

Several diagnostic metamorphic mineral assemblages have been identified in the aluminous and calc-silicate rocks of units (gsg) and (csg) along the eastern margin of the Baro Domain. These include:

- 1) sillimanite-garnet(almandine)-cordierite-orthoamphibole-plagioclase-biotite-quartz
- 2) sillimanite-K feldspar-biotite-quartz
- 3) sillimanite-garnet (almandine)-K feldspar-biotite-quartz
- 4) hornblende-clinopyroxene (diopside)-epidote-titanite

Assemblages (1) and (2) define a very restricted PT range for the peak metamorphic conditions that affected the metasedimentary schist and gneiss unit (gsg). Assemblage (1) indicates a reaction isograd:

Part of a reaction grid developed by Carmichael (1985) for Fe-Mg rich pelites in a six component system: $\text{SiO}_2\text{-Al}_2\text{O}_3\text{-FeO-MgO-Na}_2\text{O-K}_2\text{O}$ (without sufficient K to form muscovite), has been used to interpret this reaction isograd. The reaction is univariant and occurs at pressures between 5.8 - 7.0 kb and temperatures between 630-730°C (Fig. 2.4a).

The absence of muscovite from assemblages (2) and (3) shows that peak metamorphic temperatures were above the reaction isograd:

Figure 2.4: Petrogenetic grid (after Carmichael, 1985) showing reactions among assemblages that contain biotite, plagioclase and quartz observed in metasedimentary schists and gneisses (gsg) of the Baro Domain (A), and metapelites (mss) of the Birbir Domain (B). Stippled areas indicate range of possible P-T conditions for observed assemblages and dots indicate invariant points.

Mineral abbreviations

A Al_2O_3
ky kyanite
and andalusite
sill sillimanite
C cordierite
Ch chlorite
G garnet
Gd gedrite
K K feldspar
L granitic liquid
M muscovite
P orthopyroxene
S staurolite
V vapour

The reaction, in combination with (1), restricts conditions of formation of the aluminous rocks to within 680-730°C and 6.3 - 7.0 kb (Fig. 2.4a).

The minerals shown in assemblage (3) possibly formed by a continuous reaction where K-feldspar appears over a range of temperatures (Ramberg, 1952). The reaction is as follows:

The presence of muscovite-bearing assemblages in unit (qms), from the eastern side of the Baro Domain, indicates an increase in metamorphic grade towards the west where muscovite-free, sillimanite-bearing assemblages prevail. The sillimanite-free, muscovite-bearing assemblages of unit (qms) have primary clastic textures suggesting they are not retrograde assemblages from unit (gsg).

In rocks of unit (gsg) containing assemblage (1), some garnet porphyroblasts contain quartz inclusions with an orientation that is continuous with foliation in the matrix. Such observations suggest a late- or syn-tectonic period of garnet growth with respect to foliation associated with the main deformation. Orientation of some of the orthoamphibole (gedrite) and sillimanite in the direction of the main foliation indicates that these minerals grew during the development of the gneissic fabric (Plate 2.2A); the growth of garnet porphyroblasts outlasted the deformation.

The origin of the bulk composition associated with assemblage (1) is interesting because of the high content of Mg- and Fe-rich minerals in these rocks. Three competing views are held. The first is that the rocks reflect the premetamorphic bulk composition. The second explanation involves

metasomatic addition of $\text{MgO}+\text{FeO}$ and simultaneous removal of CaO and alkalis, while the third involves enrichment of $\text{MgO}+\text{FeO}$ by metamorphic differentiation or anatexis. In the third alternative, MgO and FeO would have migrated from rocks undergoing granitization and concentrated near relicts of amphibolite and calc silicate rocks (Ramberg, 1952; Moorhouse, 1969). This may be brought about by strong penetrative deformation or by partial melting. Because of migmatization in the area and the presence of anatectic granites, enrichment in Fe and Mg may be a direct consequence of metamorphic differentiation and anatexis.

Metamorphic mineral assemblages from units (mss) and (mvs) of the Birbir Domain include:

- 1) quartz-muscovite-biotite-plagioclase-epidote-calcite
- 2) quartz-muscovite-biotite-staurolite-andalusite-chlorite
+/- garnet, and
- 3) quartz-oligoclase-microcline-hornblende-biotite-epidote
+/- actinolite,

assemblages that indicate lower amphibolite (epidote-amphibolite) facies metamorphism. Rocks with assemblage (2) contains large porphyroblasts of andalusite and staurolite co-existing with biotite, muscovite and Fe-rich chlorite.

Staurolite occurs as well-formed euhedral, sieve-textured, porphyroblasts, partly replaced by a fine-grained mixture of quartz, feldspar, biotite and muscovite (Plate 2.2B). Chlorite occurs as rims around

staurolite and also independently in the matrix. Both staurolite and andalusite overgrew the foliation and contain inclusions of quartz and biotite flakes that are continuous with the foliation in the matrix. A concentric overgrowth around andalusite consists of muscovite followed by chlorite with a few fine-grained staurolite crystals. The assemblage implies a reaction isograd involving biotite, plagioclase and quartz:

The appropriate part of a petrogenetic grid constructed by Carmichael (1985) is shown in Figure 2.4b. The reaction is univariant at constant H_2O activity; it has taken place within the andalusite stability field at temperature of about 520-530°C and 3.5-4.1 kb pressure.

The meta-igneous rock assemblages and sheared equivalents of the Birbir Domain can be represented by the association of quartz-microcline-oligoclase-biotite-hornblende-epidote-titanite, without chlorite, that broadly indicates epidote-amphibolite facies metamorphism. Metamorphic conditions associated with the mylonitic rocks and those virtually unaffected by the shearing thus both reflect lower amphibolite facies metamorphism and suggest that metamorphic conditions may not have changed throughout the duration of shearing. Nevertheless, the presence of actinolite after hornblende, and minor chlorite veinlets in the mylonitic rocks indicate limited retrograde effects during the shearing event.

2.5 STRUCTURE

2.5.1 Gneissic terranes

In the map area, the gneissic terranes consist of the Baro Domain to the west, and the Geba Domain to the east. Rocks of the Baro Domain contain planar structures with a general NNW strike and moderate dips of 30-60° to the east and west; locally they may be subvertical. Linear features plunge gently north or south. An exception to the latter occurs in an area immediately west of the Birbir Domain at the northern end of the map area, where lineations plunge moderately toward the east and down the dip of foliation. A fold pattern expressed by foliation directions (Fig. 2.2), is truncated with abrupt change in foliation patterns that coincides with changes in the attitudes of linear features mentioned above. The truncated fold pattern shows what appears to be a detached limb of a fold but could not be verified on the ground and no structural line can be drawn between the rocks of the Baro and Birbir Domains.

At least two stages of deformation can be documented in rocks of the gneissic terranes. The earliest is defined by a well-developed gneissic layering and penetrative mineral foliation. In rocks of unit (gsg), garnet porphyroblasts and sillimanite (fibrolite) overgrew the fabric suggesting that metamorphism outlasted regional deformation. Intra-folial folds related to this deformation are recognized in migmatites.

A second deformation resulted in kilometre-scale folds in the foliation of the earlier deformation. Foliation attitudes mapped west of Bonga and seen on aerial photographs suggest a major antiform with kilometre scale wavelength and a north-south trending axial-surface trace. Crenulation and

weak mineral lineations are developed with gentle northerly plunges, parallel to the axis of the inferred fold. The crenulation is probably related to the second deformation.

One of the Baro leucogranite sheets shows an upright, open fold pattern. The gneissic fabric and lenticular shape of the granite sheets suggest that they are either pre- or syn- tectonic relative to the first generation of structures. Similar granite sheets are found further to the south, and Davidson (1983) described them as having tectonic contacts related to thrusting. Granitoid pegmatites cut the main fabric and all are undeformed.

In the eastern part of the map area, gneisses of the Geba Domain also show pronounced layering and foliation. A major antiform at the western margin of the Geba Domain has been truncated against rocks of the Birbir Domain. The contact is marked by a mylonitic metagranite sheet (mgt).

2.5.2 Birbir Domain

Rocks of the Birbir Domain have predominantly north-south - striking foliations and typically dip steeply to the east or west. Sedimentary structures such as cross-lamination and graded bedding are well preserved in the metagreywacke (mss). In the area just west of Bure, bedding-cleavage intersections (S_0/S_1) at several localities give way-up directions consistent with those indicated by sedimentary structures, all of which indicate easterly younging. Most bedding and cleavage are subparallel.

Whitten (1969) discussed cleavage-bedding parallelism and suggested three main mechanisms of formation: 1) the development of foliation parallel

to axial planes of recumbent isoclinal folds 2) the development of foliation by minerals growing with preferred orientation along some pre-existing plane (eg. bedding planes or shear planes), the result of solutions percolating through the rock and 3) the rare development of foliation within incompetent units interlaminated between competent ones during flexural-slip folding. The second of these explanations is preferred by the author where foliation (S_1) developed subparallel to bedding (S_0) during shearing.

The development of the major fabric in rocks of the Birbir Domain was followed by a period of porphyroblast development. Rocks of unit (mss) contain staurolite and andalusite porphyroblasts that overgrew the foliation. The porphyroblasts contain aligned inclusions that suggest that their growth post-dated the development of the major planar fabric.

2.5.3 Birbir Shear Zone (BSZ)

The western half of the Birbir Domain is characterized by a NNE-striking transcurrent fault system consisting of numerous mylonite zones separated by less-strained rocks. These zones occupy an area at least 15 km wide, and are collectively termed the Birbir Shear Zone (BSZ; S.K. Hanmer and J.M. Moore, pers. comm. 1986). The rocks affected by the shearing commonly have extension lineations that plunge gently to the north or south suggesting transcurrent movement (Plate 2.2D). Dips of the mylonitic foliation are typically steep to vertical (Plate 2.2 C,E). Rocks of unit (mvs) show mylonitic foliation which is locally a crenulation of pre-existing cleavage; the mylonitic foliation is asymmetrically folded in places (Plate 2.2F). The crenulation may suggest more than one generation of shearing, although it could be produced as a continuum in one major shearing event.

Structures in the BSZ suggest emplacement of abundant mafic and felsic dykes synchronous with deformation (S.K. Hanmer, pers. comm., 1986). Most of the mafic dykes were emplaced in a plane oblique, either clockwise or anticlockwise, to the mylonitic foliation. The former are shortened, producing lozenge-shaped fragments, while those disposed anticlockwise are symmetrically boudined/extended in their own plane. Veins are also emplaced into either quadrant, suggesting dextral shear combined with shortening across the shear plane. A sinistral sense of movement can be identified from the displacement of late felsic dykes at several localities and rotated feldspar augen, on the eastern contact of the Mao pluton, that clearly cut the earlier mylonitic foliation. The BSZ, therefore, was initiated with dextral movement that subsequently converted to sinistral movement.

Successive stages of mylonitization are observed in the BSZ. Plutonic units within the Birbir Domain are variably deformed within the shear zone. Shearing is most extreme in the Birbir quartz diorite complex, where rocks range from protomylonite to mylonite; rocks of the Goma and Mao plutons are protomylonitic. Shear foliation and lineation are expressed by quartz rods and hornblende and biotite aggregates.

The Birbir quartz diorite sheets show a wide variety of grain sizes, grain boundary types and grain shapes on a very local scale. It is possible on the basis of petrographic studies to typify samples in order of decreasing shear strain:

(1) Rocks of high strain (Plate 2.3 F and G; samples # 2G, 8H, 2I; see Fig. 2.3 for localities). These are composed of a fine-grained lepidoblastic matrix of quartz, feldspar, biotite and hornblende (>80%) that contains

Plate 2.3: Photomicrographs showing various stages of mylonitization of unit (mqd) in the Birbir shear zone (ABCDEFGH); and some textures and mineralogy in other intrusive units. PPL, XPL: under plane and crossed polarized light respectively. Scale bar is 1 mm for 2.3 A and G, 0.5 mm for 2.3 BCDEFHI. Sample localities are given in Figure 2.3.

- A- Subhedral-granular texture, showing curved to slightly embayed grain boundaries of plagioclase laths, unoriented biotite and hornblende, and interstitial quartz. XPL. Sample # T4I.
- B- Bent and kinked plagioclase (p) showing twin lamellae; microfractures and zones of sericitization run at right angles to the twin lamellae. XPL. Sample # T1H.
- C- Plagioclase phenoclasts (p) showing deformation twins and subgrain development, mantled by recrystallized quartz (q) and feldspar. XPL. Sample # T5G.
- D- Granoblastic quartz (q) and feldspar (f) with plagioclase (p) phenoclasts showing straight grain boundaries and triple junctions. XPL. Sample # T3G.
- E- Biotite (b) and hornblende (h), euhedral epidote (e), titanite (t), and interstitial quartz (q) defining mylonitic foliation. PPL. Sample # T5G.
- F- Polygonized quartz ribbons (q) around K-feldspar porphyroclasts (f). Note that the feldspar shows drawn-out tails indicating dextral sense of shear. XPL. Sample # T8H.
- G- C-S fabric shown by microcrystalline aggregates of recrystallized feldspar and quartz in >80% fine matrix. Shear foliation C is defined by traces of matrix layers. XPL. Sample # OY107, locality: 5 km northeast of the Baro - Birbir Rivers confluence..
- H- Garnet-muscovite granite of Baro Domain (mgt). g: garnet, ms: muscovite, m: microcline, q: quartz. p: plagioclase. PPL. Sample # T4M, locality: 7.5 km east of Bonga.
- I- Large biotite crystals enclosing euhedral zircons (z) in Bonga granite (gt). m: microcline. XPL. Sample # T5K, locality: 7 km east of Bonga.

phenoclasts of plagioclase and hornblende (10-20%). The rocks are considered mylonites or ultramylonites (Sibson, 1977; Berthe et al., 1979). Plagioclase laths up to 5 mm are commonly bent and contain strain-induced albite and pericline twin lamellae. Some are mantled by very fine-grained feldspars. Some of the plagioclase phenoclasts have recognizable tails, formed by very fine-grained crystals, that can be used to determine the sense of shear. As biotite constitutes about 15% of the rock, a large part of the strain imposed through shearing was probably taken up by grain boundary sliding. As a result, some hornblende and plagioclase phenoclasts were relatively unaffected and survived shearing. Shear foliation is mainly defined by alignment of biotite flakes, and oriented hornblende and plagioclase phenoclasts. This orientation is probably a result of rigid body rotation. Fine-grained quartz and feldspar are polygonized and locally ribboned into strain-free grains that make up the bulk of the matrix. Deformation in these rocks seems to be essentially a product of ductile processes involving recrystallization and recovery.

(2) Rocks with intermediate strain (Plates 2.3 C, D and E; Samples # 3G, 5G, 9I). These rocks display similar strain features to those described above but have sub-equal proportions of matrix (composed of quartz, feldspar, biotite, hornblende and epidote) and phenoclasts (plagioclase and hornblende). Granoblastic quartz and feldspar show undulatory extinction.

(3) Least strained rocks (Plate 2.3 A and B; samples # 9G, 1H, 4I). These rocks are coarse grained, contain well-preserved igneous textures such as subhedral-euhedral plagioclase and hornblende phenocrysts, unoriented biotite plates that tend to occur in clots, and perthite, antiperthite and

myrmekite.

The protomylonite stage is represented by (3), mylonites by (2) and (1) and ultramylonite stage by (1). In the last, a small proportion of phenoclasts remain in the fine-grained matrix.

Evidence for non-penetrative internal deformation in the Goma and Mao plutons is shown by narrow shear zones, that are well-developed near the contacts of the plutons. Both ductile and brittle shear strain are seen in samples from the western contact of the Goma pluton. The strain features include undulose extinction in quartz and to a lesser extent in feldspars, feldspar phenoclasts mantled by fine-grained recrystallized grains along weak deformation bands, and microfractures in plagioclase phenoclasts, now filled by stringers of secondary minerals.

2.6 DISCUSSION

The present study reveals that the rocks of the Birbir Domain are bounded by tectonites. The western margin of the Birbir Domain is defined by a major shear zone; the eastern boundary is also marked by mylonite and by truncation of foliation patterns that define a map-scale antiform in the Geba Domain (Fig. 2.2). Deformation within the Birbir Domain has produced penetrative foliation and schistosity that run sub-parallel to those of the gneissic terranes. An apparently coeval regional deformation thus appears to have affected rocks of both the Baro and Geba high-grade gneissic terranes and the Birbir lower-grade terrane.

Several shear zones, striking generally north-south, occur within rocks of the Birbir Domain and also separate relatively low-grade rocks of the Birbir Domain on the east from the high-grade schists and gneisses of the Baro Domain to the west. Andalusite - bearing pelitic schists (mss) indicate low-pressure facies metamorphism (Miyashiro, 1973), whereas sillimanite - K - feldspar - orthoamphibole assemblages to the west indicate medium-pressure - high temperature facies metamorphism. The pelitic mineral assemblages in rocks of the Birbir Domain indicate low-amphibolite grade whereas the Baro and Geba domains have attained metamorphic grade of upper almandine-amphibolite facies. The change in metamorphic grade is observed in samples obtained within a strike-normal distance of about 4 kilometres suggesting an abrupt increase in metamorphic grade across the transcurrent fault system of the Birbir Shear Zone (BSZ). The metamorphic gradient is probably too steep to be accounted for by exposure of more deeply buried rocks by tilting, and there are no faults in the area that suggest vertical displacements. The two domains thus appear to have been tectonically juxtaposed.

Older east - west isoclinal recumbent folds documented in gneisses of the Geba valley outside the Geotraverse area (Kazmin et al., 1979; Teferra and Berhe, unpublished) have not been seen in the Baro Domain. Along the north-west frontier between Ethiopia and the Sudan, however, Warden and Horkel (1984) identified east-west trending folds from satellite imagery. There is little evidence to correlate the Baro and the Geba domains, but the presence in both of more than one generation of deformation, similar grade of metamorphism and their tectonic relationship with the lower-grade rocks all indicate that both are composed of reworked rocks, possibly pre- Pan-

African in age. Intense deformation along north-south Pan-African structures coupled with recrystallization may have obliterated any earlier metamorphic mineral assemblages and structures.

CHAPTER 3

PETROCHEMISTRY

3.1 GENERAL STATEMENT

Chemical analyses of seven plutons from the Gore-Gambella Geotraverse are used to (1) classify the intrusive rocks (2) infer their petrogenetic history and (3) evaluate the tectonic environments in which the magmas were generated. The averages and ranges of element abundances and CIPW norms of each pluton are listed in Table 3.1 to 3.4. Average abundances of some trace elements are compared to Le Maitre's (1976) average values for some igneous rocks and Taylor's (1979) and Taylor and McLennan's (1985) average values for continental crust (Table 3.5).

3.2 GEOLOGICAL SETTING

The Baro and Geba domains consist of high-grade heterogeneous ortho- and paragneisses and migmatites, whereas the Birbir Domain comprises lower-grade wackes, pelites, volcanic and plutonic rocks covering a wide compositional spectrum. Field relations, lithology and structural style of plutons in the Birbir and Baro domains can be used to divide these rocks into three groups: 1) pre- and/or syn-kinematic mafic and intermediate intrusive units 2) syn-kinematic leucogranite sheets and 3) late-to post-kinematic granitoid rocks. Rocks of Group (1) and (3) occupy the Birbir Domain and are metamorphosed and deformed to various degrees within the Birbir Shear Zone (BSZ). Deformation is most extreme in rocks of Group (1). The rocks range from protomylonite to mylonite and show steep mylonitic foliation and sub-horizontal extension lineation. Group (2) rocks are

Table 3.1: Average and range of chemical composition and CIPW norms of the low-K suite (21 samples) from Group (1) compared to analyses from the compilation by Le Maitre (1976). FeO(T) is total iron; An% is average anorthite content of plagioclase feldspar; n is number of analyses

	Bure (n=12)		Haya (n=9)		Le Maitre's average	
	Average	Range	Average	Range	Gabbro	Tonalite
SiO ₂	50.12	46.17-55.29	71.82	69.35-73.53	50.14	61.52
TiO ₂	0.65	0.07-1.23	0.14	0.07-0.19	1.12	0.73
Al ₂ O ₃	14.00	8.33-22.13	15.93	14.24-17.41	15.48	16.48
Fe ₂ O ₃	9.24	4.52-12.29	1.55	1.21-1.85	10.63	5.65
MnO	0.18	0.06-0.25	0.06	0.05-0.06	0.12	0.08
MgO	9.54	4.70-14.25	0.55	0.32-0.70	7.59	2.80
CaO	11.92	5.44-19.44	3.60	2.35-4.57	9.58	5.42
Na ₂ O	1.99	0.08-3.94	4.81	4.45-5.07	2.39	3.63
K ₂ O	0.43	0.03-1.61	1.26	0.96-1.63	0.93	2.07
P ₂ O ₅	0.07	0-0.21	0.04	0.02-0.06	0.24	0.25
Total	98.14		99.81	-	99.15	100.01
Q	0.86	0-4.31	30.12	26.39 -33.70	0.71	16.62
C	-	-	0.32	0-1.03	-	-
Or	3.37	0.18 -9.63	7.45	5.67 -9.63	5.49	12.24
Ab	21.06	6.14 -33.72	40.70	37.66 -42.90	20.26	30.67
An	36.02	23.97 -49.00	17.26	11.53 -22.35	28.60	22.58
Ne	0.09	0-0.44	-	-	-	-
Di	13.09	1.21 -21.42	0.29	0-2.35	13.70	1.49
Hy	4.73	0.15 -13.01	2.39	1.78 -2.78	22.13	9.68
Ol	7.52	0-20.21	-	-	-	-
Mt	3.14	2.35 -4.00	0.75	0.58 -0.89	4.36	2.66
Il	1.21	0.14 -2.36	0.27	0.13 -0.36	2.13	1.40
Cr	0.06	0.02 -0.13	-	-	-	-
Ap	-	-	0.09	0.05 -0.14	0.56	0.58
An%	68	62-78	29	23-35	-	-

Table 3.2: Average and range of chemical composition and CIPW norms of the calc-alkaline suite (43 samples) from Group (1) compared to analyses from the compilation by Le Maitre (1976).

	Birbir (n=29)		Goma (n=16)		Le Maitre's average		
	Average	Range	Average	Range	Diorite	Grano- diorite	Adamellite
SiO ₂	59.50	55.22-63.89	59.50	67.38-71.47	57.48	66.09	68.65
TiO ₂	0.72	0.53-0.91	0.46	0.29-0.59	0.95	0.54	0.54
Al ₂ O ₃	16.89	15.61-18.96	15.43	14.48-16.55	16.67	15.73	14.55
Fe ₂ O ₃	7.19	5.04-8.77	2.95	2.29-4.31	3.93	4.11	7.42
MnO	0.13	0.09-0.18	0.08	0.06-0.12	0.12	0.08	0.08
MgO	3.00	1.95-4.35	0.93	0.62-1.22	3.71	1.74	1.14
CaO	6.00	3.81-7.79	2.23	1.87-2.73	6.58	3.83	2.68
Na ₂ O	3.71	3.36-4.67	5.13	4.18-5.75	3.54	3.75	3.47
K ₂ O	2.28	1.38-3.91	3.25	2.64-4.04	1.76	2.73	4.00
P ₂ O ₅	0.17	0.09-0.31	0.13	0.06-0.19	0.29	0.18	0.09
Total	99.59	-	100.09	-	99.98	99.90	100.05
Q	11.24	2.04-15.46	19.48	14.85-26.24	10.28	22.36	25.17
Or	13.46	8.16-23.11	19.21	15.60-23.88	10.42	16.11	23.66
Ab	31.38	28.43-39.52	43.61	35.37-48.66	29.96	31.73	29.36
An	22.73	12.55-27.39	8.89	7.23-11.55	24.40	17.34	11.55
Di	4.43	1.46-9.47	0.72	0.03-1.41	4.67	-	-
Hy	10.74	6.01-14.44	4.11	3.05-5.36	12.56	7.40	5.66
Mt	3.19	2.43-3.49	1.62	1.11-2.08	3.63	2.00	1.79
Il	1.44	1.01-1.73	0.87	0.55-1.12	1.80	1.03	1.03
Ap	0.39	0.21-0.72	0.30	0.14-0.44	0.68	0.42	0.20
An%	42	26-47	18	14-20	-	-	-

Table 3.3: Average and range of chemical composition and CIPW norms of Group (2) plutonic units compared to analyses of S-type granites of White and Chappell (1983).

	Baro (n=10)		Gt-ms gt (n=10)		S-type granite
	Average	Range	Average	Range	
SiO₂	75.57	73.01-78.73	74.62	72.73-75.59	69.08
TiO₂	0.18	0.10-0.34	0.02	0.01-0.07	0.55
Al₂O₃	12.06	10.50-13.61	14.55	13.85-14.98	14.30
FeO(T)	2.75	1.59-3.53	0.87	0.39-1.03	3.96
MnO	0.04	0.01-0.09	0.18	0.01-0.32	0.06
MgO	0.12	0.02-0.17	0.13	0.02-0.20	1.82
CaO	0.42	0.05-0.96	0.64	0.41-0.74	2.49
Na₂O	3.32	2.27-4.11	4.02	3.59-4.48	2.20
K₂O	5.38	4.88-6.58	4.53	4.14-5.09	3.63
P₂O₅	0.01	0-0.01	0.02	0-0.04	0.13
Total	99.85		99.58		98.22
Q	34.68		33.15		
Or	31.86		25.82		
Ab	28.13		37.38		
An	1.51		3.39		
C	0.66		1.55		
Di	0.21		0		
Hy	0.10		0		
Mt	1.83		0.55		
Il	0.05		0.01		
Ap	0.02		0.02		
An%	5		8		

Table 3.4: Average and range of chemical composition and CIPW norms of Group (3) plutons compared to analyses from the compilation by Le Maitre (1976) and A-type granites of White and Chappell (1983).

	Mao (n=10)		Bonga (n=13)		Le Maitre's Average	
	Average	Range	Average	Range	Granite	A-type granite
SiO ₂	70.12	68.95-73.09	75.18	72.35-79.10	71.30	73.60
TiO ₂	0.43	0.13-0.52	0.14	0.10-0.24	0.31	0.33
Al ₂ O ₃	14.24	13.97-14.66	12.72	10.41-13.40	14.32	12.69
FeO _(T)	3.97	2.20-4.36	2.16	1.62-3.15	2.85	2.71
MnO	0.08	0.06-0.09	0.03	0.02-0.05	0.05	0.06
MgO	0.60	0.19-0.70	0.08	0.06-0.21	0.71	0.33
CaO	1.79	0.81-2.14	0.85	0.69-1.12	1.84	1.09
Na ₂ O	3.95	3.71-5.14	3.27	2.74-3.58	3.68	3.34
K ₂ O	4.74	4.18-4.97	5.35	4.14-5.93	4.07	4.51
P ₂ O ₅	0.12	0.01-0.16	-	-	0.12	0.09
Total	100.04		99.78		100.07	98.75
Q	24.69		30.58		29.06	
Or	27.88		34.08		24.50	
Ab	31.48		28.43		31.13	
An	8.18		4.21		8.04	
C	-		0.03		0.92	
Di	0.81		0.25		-	
Hy	2.98		0.16		3.37	
Ms	2.87		0.39		1.75	
Il	0.88		0.17		0.58	
Ap	0.37		0		0.28	
An%	19		12			

Table 3.5: Average abundance of some trace elements (ppm) in rocks of all plutons compared to Taylor and McLennan's (1985) averages for crust (Ct), basalt (Bs) syenite (Sy), granodiorite (Gdt) and granite(Gt); and White and Chappell's (1983) average analyses of I-, S- and A- types.

Bure gbdt	GROUP (1)				GROUP (3)				GROUP (2)				Taylor's average				White & Chappell's average		
	Haya tt	Birbir qzdt	E	Goma gdt	Mao mgt	Bonga gt	(1)	(2)	Baro gt	G-M gt	Ct	Bs	Sy	Gdt	Gt	I	S	A	
Rb	3	20	52	57	61	52	173	120	90	94	128	90	30	110	120	150	132	180	199
Nb	<2	2	2	5	4	10	29	13	23	5	-	20	20	40	20	20	9	11	22
Zr	38	75	158	161	185	275	351	219	430	648	45	165	150	500	140	180	143	170	342
Y	11	8	25	25	27	26	61	55	72	34	64	30	25	20	30	40	27	32	76
Ba	156	600	972	676	730	1136	870	303	1061	314	199	525	250	1500	500	600	520	480	605
Si	506	508	630	530	456	312	209	50	255	42	56	375	465	300	450	285	253	139	105
Ni	208	6	5	16	12	6	10	4	15	5	-	75	150	5	20	0.5	9	17	2

(1) Lithology: gbd: gabbro-diorite; tt: tonalite; mgt: monzogranite;

(1) alkali-feldspar granite; (2) monzodiorite;

G-M gt: garnet-muscovite granite.

(2) C: central; W: western; E: eastern sheet.

Table 3.5 (continued)

	Bure	Haya	W	Birbir C	E	Goma	Mao	Bonga (1)	Bonga (2)	Baro	G - M gt
K/Rb	380	620	360	350	323	510	245	370	400	615	326
Rb/Sr	0.01	0.04	0.08	0.11	0.13	0.17	0.83	2.4	0.35	2.2	3.4
Ba/Sr	0.31	1.18	1.5	1.3	1.6	3.6	4.2	6.1	4.2	7.5	4.1
Ba/Rb	52	30	19	12	12	22	5	2.6	11.8	3	2
Zr/Nb	38	38	79	32	46	28	12	17	19	130	45
Rb/Zr	0.08	0.27	0.33	0.35	0.32	0.19	0.5	0.6	0.21	0.15	2.8
Th/U	2.6	5.6	2	1.8	2.8	2.4	6.3	4.1	4.6	6.8	2.6
Ta/Nb	0.3	0.15	0.17	0.1	0.19	0.1	0.1	0.12	0.1	0.26	-
Ta/Hf	0.23	0.13	0.08	0.12	0.14	0.14	0.29	0.21	0.2	0.08	0.19
Ta/Zr	0.02	0.004	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01

strongly deformed, and form lenses and sheets within the high-grade rocks of the Baro Domain. Plutons of Group (3) occupy the Birbir Domain and its western boundary with rocks of the Baro Domain. These are circular, discordant and generally undeformed.

3.3 SAMPLING AND ANALYTICAL METHODS

A total of 120 rock samples were collected of which 106 were analyzed for major and some trace elements. Between 2 and 3 kg of fresh chips were collected. Weathered surfaces were removed, and the fresh material pulverized in a jaw crusher. A representative sample, made by coning and quartering, was then crushed to < 100 mesh using a Bleuler Mill (shatter-box).

Major elements as well as Ba, Cr, Zr, Sr, Rb, Y, Nb, Ni and V were determined by XRF spectrometry using a Philips PW 1410/20 AHP instrument at the University of Ottawa. Fused-glass pellets were prepared according to the methods of Norrish and Hutton (1969). One gram of powdered sample was mixed with 5 gm of $\text{Li}_2\text{B}_7\text{O}_4$ and 0.3 gm LiF, and fused to a glass pellet within a platinum holder. Accuracy was estimated by analysing reference standard samples DR-N and GS-N. Results are given in Appendix 2, Tables A2.1 and A2.2. CIPW normative calculations and various cartesian and triangular plots were made using the BMDP and IGPET programs.

Abundances of the rare earths La, Ce, Nd, Sm, Eu, Tb, Dy, Ho, Er, Yb and Lu as well as Sc, Co, Sb, Cs, Hf, Ta, W, Th and U from 10 samples were measured by instrumental neutron activation (INAA) in the Department of Nuclear Engineering, Ecole Polytechnique in Montreal and the data are given

in Appendix 2, Table A2.3. The compositions of plagioclase, K-feldspar, biotite, hornblende, apatite, epidote, muscovite and garnet were determined at Carleton University using a Cambridge Microscan 5 electron microprobe (Appendix 3).

3.4 ALTERATION AND METAMORPHISM

The extent to which regionally metamorphosed igneous rocks have behaved as closed systems is unknown. For mylonitic rocks "open" vs "closed" system behaviour is even more difficult to assess because fluid transport along microscopic slip planes causes localized selective enrichment and depletion of some elements. Despite the known mobility of some elements, however, many studies present data from meta-igneous rocks that show closed-system behaviour and reveal magmatic trends (Winchester and Floyd, 1976; Winchester and Max, 1984).

In the sheared samples of the pre-kinematic plutonic units in the study area, variable degrees of alteration are observed. The changes include partial replacement of plagioclase and hornblende by epidote and carbonate, biotite by chlorite and feldspar by sericite. However, chemical data from all of the plutonic units in this study show coherent behaviour. This coherence suggests that most of the samples have behaved as closed chemical systems.

3.5 CLASSIFICATION

The plutonic rocks vary considerably in modal mineralogy, particularly in the quartz content and proportion of plagioclase to K-feldspar. The rocks are classified according to Streckeisen (1976) using the modes of the

plutonic units (Fig. 3.1a). Group (1) rocks fall in the range of diorite through quartz diorite to granodiorite; rocks of groups (2) and (3) fall in the granite field. Because rocks from groups (1) and (2) are metamorphosed, chemical classification based on the proportions of normative minerals is preferable because it is somewhat difficult to establish primary modal mineralogy. Hietanen's (1963) and O'Connor's (1965) classifications in terms of the normative Ab-An-Or diagram show that samples of each pluton tend to cluster in distinct compositional fields (Figs. 3.1b and 3.2). In Group (1), the Bure pluton ranges from a mafic gabbro to diorite, data from the Haya pluton fall in the tonalite and monzotonalite fields, the Birbir data cluster predominantly in the granodiorite and quartz diorite fields, and the Goma data plot in the granite and trondhjemite fields. Data from groups (2) and (3) largely fall in the granite field, although a few fall in the quartz monzonite field.

Data from Group (1) rocks, plotted on an AFM ($\text{Na}_2\text{O} + \text{K}_2\text{O}$ vs. $\text{FeO}_{(\text{Total})}$ vs. MgO) diagram, lie close to the calc-alkaline differentiation trend (Fig. 3.3). Low $\text{FeO}_{(\text{T})}/\text{MgO}$ ratios (<1) from the Bure pluton suggest sub-alkaline characteristics (Mottana et al., 1968). The data from Group (1) rocks are plotted on the normative Qz-Ab-Or (Fig. 3.4a) and $\text{CaO}-\text{Na}_2\text{O}-\text{K}_2\text{O}$ plot (Fig. 3.4b), diagrams that have been widely used for chemical distinction between trondhjemites and calc-alkaline rocks (Barker and Arth, 1976; Barker, 1979). In the Haya pluton, the $\text{Na}_2\text{O}/\text{K}_2\text{O}$ ratios are mostly >2 , values suggesting trondhjemitic affinity (Condie and Hunter, 1976). The trondhjemitic affinity of the Haya pluton is distinct from that shown by the rest of Group

Figure 3.1: (a), Classification of the plutonic units in terms of typical modes of quartz (Q), plagioclase (P) and alkali-feldspar (A), after the IUGS system (Streckeisen, 1976). 1:gabbro/diorite/quartz diorite, 2:tonalite, 3:granodiorite, 4:granite. (b) Normative classification in terms of orthoclase (Or), albite (Ab) and anorthite (An), after Hietanen (1963). 1:mafic gabbro, 2:gabbro, 3:quartz diorite, 4: granodiorite, 5:tonalite, 6:monzo-tonalite, 7:quartz monzonite, 8:trondhjemitic, 9:granite trondhjemitic, 10:granite.

Figure 3.2: Normative proportions of Ab, An and Or with classification boundaries of O'Connor (1965); the dashed lines are feldspar solvi after Glikson and Sheraton (1972). Symbols same as in Fig. 3.1 (a).

Figure 3.3: AFM (A: $\text{Na}_2\text{O} + \text{K}_2\text{O}$; F: $\text{FeO}_{(\text{total})}$; M: MgO) diagram for the meta-intrusive rocks of the Birbir Domain. Solid lines separate fields of tholeiitic (TH), calc-alkaline (CA) and alkalic (AL) rocks. Also shown are fields for plutonic units from extensional (E) and compressional (C) tectonic settings (dashed lines) suggested by Petro et al. (1979). Symbols same as in Fig. 3.1 (a).

Figure 3.4: (a) Ternary plot of normative Q , Ab , and Or ; and (b) weight proportions of $CaO : Na_2O : K_2O$ showing trends for calc-alkaline and trondhjemitic suites. The trend lines are after Barker and Arth (1976). Symbols same as in Fig. 3.1 (a).

(1) plutonic units, which follow a strong calc-alkaline trend.

Both the Bure and Haya pluton, that lie close to one another spatially in the field, have generally low contents of large-ion-lithophile elements, particularly low K_2O content. The Birbir intrusive complex and the Goma pluton have medium to high K_2O content and follow a calc-alkaline trend (Figs. 3.3 and 3.4a and 3.4b). Group (1) may thus be divided on the basis of major element chemistry into two subgroups: a low-K suite and calc-alkaline suite each of mafic and intermediate chemical composition.

In Shand's (1951) and Pitcher's (1983) alumina saturation classification, in which they used molecular proportions of $Al_2O_3/(CaO+Na_2O+K_2O)$ (= A/CNK), intrusive rocks are divided into peralkaline ($A < NK$), metaluminous ($CNK > A > NK$), subaluminous ($A = CNK$), and peraluminous ($A > CNK$). All rocks in groups (1) and (3) are metaluminous and hypersthene-normative, other than the Haya tonalites which appear to be marginally peraluminous (A/CNK cationic ratio of about 1.04) and contain normative corundum. Rocks of Group (2), however, are typically peraluminous. Garnet - muscovite granite (mgt) has an average cationic ratio for A/CNK of 1.14 and normative corundum content of upto 2% whereas in the Baro leucogranites both A/CNK and normative corundum contents are < 1 . However, both plutonic units of group (2) contain garnet and/or muscovite, both aluminous minerals. These minerals coupled with the chemistry mean that rocks of Group (2) can be classified as peraluminous granites.

In terms of the alkalinity index, $(AI) = (ACNK/AC-NK)$, of Wright (1969), rocks of both groups (2) and (3) are alkaline ($AI > 2.5$), with moderate to high agpaitic ratios ((NK/A) ; Gerasimovski, 1956) ranging

between 0.82 and 0.87 for Group (3) compared to about 0.5 for Group (2). The high alkali contents of Group (3) granites and the composition of the feldspars (predominantly microcline and albite, see Chapter 2) allow them to be classified as alkali-feldspar granite.

Several different classification schemes for granitic rocks have been proposed in addition to the modal and normative mineral classifications. Chappell and White (1974) divided granitoid rocks into I- and S- types. The granites designated "I-type" were thought to represent melts derived from a metaluminous igneous rock source, and "S-type" from aluminous-metasedimentary source rocks. Ishihara (1977) classified granitic rocks in Japan into a magnetite and an ilmenite series that somewhat parallels Chappell and White's classification. Loiselle and Wones (1979) and Collins et al. (1981) suggested a group called A-type, characteristic of granites from anorogenic regions. The last are chemically distinct from the normal I- and S-type granites, and contain high Na_2O , Nb, Y, Zr, REE and low CaO. Pitcher (1981) introduced an M-type plutonic group, in addition to those above, which covered the most calc-alkaline plagiogranite.

For comparative purposes, trace element data of the various granite types of Chappell and White are shown in Table 3.5. Compositional data are compared in K_2O - Na_2O space showing the I-, S- and A-type granites (Fig. 3.5). The calc-alkaline suite essentially falls in the I-type field whereas the data from the low-K suite fall outside any of the defined fields. Group (2) and (3) data scatter within the A-type boundary. It is surprising that analyses of Group (2) do not overlap the compositional field defined for S-type granites as they satisfy most of the chemical criteria for these

Figure 3.5: Plot of Na_2O versus K_2O with fields for I-, S- and A-type granites of White and Chappell (1983) shown for comparison. Symbols same as in Fig. 3.1 (a).

granites. One potential explanation is that the Australian S-type granites owe their degree of alumina saturation to low Na_2O and not to high Al_2O_3 . Rocks of Group (2), however, owe their degree of alumina saturation, in part, to very low CaO whereas the abundance of Al_2O_3 seems to be as high as the Australian rocks. The low CaO is characteristic of many sedimentary rocks because it is easily removed in weathering cycle.

The aluminous nature, the presence of two micas and the overall low incompatible element contents of the Haya tonalite of the low-K suite are consistent with M-type granite, whereas the hornblende-bearing, calc-alkaline suite are similar to the I-type (Cordilleran) granitic rocks as described by Pitcher (1981). A summary of some lithologic and geochemical characteristics of the Birbir Domain plutons is given in Table 3.6.

3.6 MAJOR- AND TRACE ELEMENT CHEMISTRY

3.6.1 Pre- to syn-kinematic mafic and intermediate intrusive complex:

Group (1)

The spatially associated Bure gabbro and Haya tonalite of Group (1) have extremely low K_2O content. The tonalite has distinctly high Na_2O and low CaO in contrast to LeMaitre's (1976) average tonalite and other rocks of comparable SiO_2 content (Tables 3.1 and 3.2). Both the Bure and Haya plutons have low overall large-ion lithophile element abundance suggesting a common origin; they may be designated as a low-K suite, with bimodal SiO_2 values of 46-55 and 69-73 wt.%. MgO , $\text{FeO}_{(\text{T})}$, CaO , and TiO_2 decrease as SiO_2 increases. Al_2O_3 has mean values of 14.0 wt.% in the mafic and 15.93 wt.% in the felsic rocks.

Table 3.6: Summary of the distinctive petrologic and geochemical characteristics of the plutonic units within the Birbir Domain.

	Group (1) : Pre- and syn- kinematic		Group (3) : Late- to post- kinematic
	low-K suite	calc-alkaline suite	
Lithology	tonalite subordinate to composite gabbro and diorite; (gb,di,tt,tg)	quartz diorite dominant but broad compositional spectrum: diorite to granodiorite; major association with gabbro; gb,di,dq,tg,tt)	alkali-feldspar granite; (gm,tg,gt)
Modal mineralogy	hornblende in gb and two-mica in tt	hornblende and biotite	hornblende and biotite
SiO₂	Bimodal; 46-55 and 69-73 wt.%	Unimodal; 55-68 wt.%	Unimodal; 68-79 wt.%
Shand's A/CNK	1.04 (PA) for tt	0.96 (MA)	0.99 (MA)
C.I.P.W. norms	corundum-normative tonalites; and hypersthene in gabbro	hypersthene-normative	hypersthene-normative
AFM, Qz-Ab-Or trend	trondhjemitic for tt and subalkaline gb	calc-alkaline	alkaline
K₂O-SiO₂ trend	Oceanic island arc of Fijian type	Peruvian/Andinotype marginal continental arc type	anorogenic/intraplate
White & chappell's classification (1983)	I-type	I-type	A-type
Pitcher's classification (1983)	M-type	I-type (cordilleran)	A-type

Lithology: gb, gabbro; di, diorite; dq, quartz-diorite; tt, tonalite; tg, granodiorite; gm, monzogranite; qm, quartz monzonite; gt, granite.

ACNK: Al₂O₃, CaO, Na₂O, K₂O; PA: peraluminous; MA: metaluminous

Figure 3.6: Variation diagrams for the unimodal - calc alkaline suite of Group (1) plutonic units. (R = correlation coefficient)
Symbols same as in Fig.3.1 (a).

Figure 3.6 (continued):

2

The calc-alkaline suite, comprising the Birbir quartz diorite and Goma granodiorite, covers a broad range of SiO_2 values from 55-72 wt.%. There is a linear variation of SiO_2 with some major oxides (Fig. 3.6); Al_2O_3 , MgO , $\text{FeO}_{(\text{T})}$, TiO_2 , CaO , MnO and P_2O_5 decrease, and alkalis increase as SiO_2 increases. The Goma granodiorite shows a slight differences from the Birbir quartz diorite in its content of some oxides such as Na_2O . Variation diagrams (Fig. 3.6) with significant correlation coefficients (>0.9) suggest magmatic differentiation of the suite from a single parent magma. Overall, the major element chemistry follows a calc-alkaline differentiation trend (Fig. 3.4). Trace element distributions differ little between the Goma and Birbir plutons (Table 3.5). The overall mean values of the major elements for the Birbir intrusive complex and Goma pluton are similar to those given by LeMaitre for diorite and granodiorite respectively (Table 3.2).

There is a general decrease in the concentrations of Cr, Ni, Zr, Zn, Y and V, and an increase of Ba, Sr, Nb and Rb from mafic to felsic compositions of the low-K suite. The abundances of Rb, Nb and Y are not far removed from Wood's (1979) estimate of primordial mantle values for these elements of 15, 0.6 and 5 ppm respectively.

Barium, Sr, Zr, V and Y contents broadly decrease whereas Rb increases with increasing SiO_2 in the calc-alkaline suite. Strontium decreases with decreasing CaO and Ba, and Rb tends to closely follow K_2O . Chromium and Ni follow the MgO concentrations. Over the wide silica range shown by Group (1) rocks there is a general decrease of CaO and Sr and increase in K_2O , Ba and

Rb. The former suggests significant plagioclase fractionation, while the latter may be due mainly to biotite and to a lesser extent K-feldspar fractionation. The distinctive chemical difference between the calc-alkaline suite and the felsic rocks of the low-K suite is the general enrichment in the calc-alkaline rocks of large-ion-lithophile elements such as Rb, Ba, Th, U, the high Rb/Sr and Ba/Sr and low K/Rb, Ba/Rb, Th/U and Zr/Nb abundance ratios.

3.6.2 Peraluminous and alkali-feldspar granites: Groups (2) & (3)

Rocks of Group (2) differ markedly from those of Group (3). They show strong peraluminosity and moderate agpaitic ratio (~0.5). Group (3) plutons have distinctive chemical characteristics that include high K_2O , K_2O/Na_2O , Fe_2O_3 , low CaO, MgO, and Al_2O_3 ; alkalinity indices in excess of 1.5, high agpaitic ratios upto 0.87. They also occur as discordant intrusive bodies.

The trace element patterns of groups (2) and (3) shown in Table 3.5 are marked by high abundances of Rb, Sr, Ba and Th. Barium and Sr tend to decrease irregularly while Rb increases with increasing silica and tends to follow K_2O . On the basis of high differentiation indices (normative Qz + Or + Ab) that exceed 90 and Rb, Sr and Ba abundances, the Mao and Bonga plutons are relatively differentiated granites. In general terms, concentrations of many large-ion-lithophile elements, including Nb, Th, U, Zr, Y, Ta and Hf, are enhanced in Groups (2) and (3).

3.7 REE DISTRIBUTION

The REE data, given in Appendix 2, Table A2.2, and summarized in Table

3.7, characterize each of the plutonic units despite the limited number of analyses. The data are plotted on chondrite-normalized (cn) diagrams in Figures 3.7 a, b, and c. The patterns shown on these diagrams are different from one another and closely correspond to the groupings of the plutonic units based on geological and major element chemical differences. REE patterns of six samples from Group (1) plutons are shown in Figure 3.7a. The patterns depict slight enrichment in the light rare earth elements (LREE) $\text{La/Sm}_{(\text{cn})}$ ratios between 2.5 to 4.2 and flat to poorly fractionated heavy rare earth elements (HREE).

The low-K suite (Bure and Haya plutons, Fig. 3.7a) has a relatively low absolute REE abundances (40-75 ppm), flat to poorly fractionated HREE patterns and no Eu anomaly (Eu/Sm : 0.31-0.32). The Haya pluton shows a large LREE/HREE ratio ($\text{La/Lu}_{(\text{cn})}$ of about 10). Four analyses from different members of the calc-alkaline suite have extremely similar distributions and abundances, all have the same fractionated LREE patterns. The $\text{La/Lu}_{(\text{cn})}$ ratio lies between 5.1 and 7.5, and Eu/Sm of 0.23 to 0.27 (i.e. little or no Eu anomaly) and REE contents are low to moderate (99-121 ppm). The Haya pluton of the low-K suite is more LREE enriched than the calc-alkaline suite. Apart from this, there appears to be a gross similarity in patterns between the low-K and calc-alkaline suites that reflects little or no difference in the nature of the magma and source material.

The absence of any Eu anomaly in rocks of Group (1) can be explained in terms of the redox conditions of the melt. In oxidizing conditions Eu^{+3} tends to substitute for Ca in the large cation site of plagioclase and thus europium is removed during fractional crystallization. The oxygen fugacity

Table 3.7: Summary of REE distribution in terms of LREE enrichment, LREE/HREE ratio, Eu-anomaly and absolute abundances.

I represent the compositional ranges for calc-alkaline rocks of island-arc and continental subduction zones (after Cullers and Graf, 1986) cn: chondrite normalized.

Sample #	La/Sm _(cn)	La/Lu _(cn)	Eu/Sm	ΣREE(ppm)
Group 1				
4D(Bure)	4.12	5.3	0.31	75.01
1A(Haya)	4.18	10.1	0.32	40.96
11I (Birbir)	2.96	5.3	0.23	121.17
1H (Birbir)	2.68	7.5	0.27	115.45
3G (Birbir)	2.52	5.1	0.26	99.76
6J (Goma)	3.42	5.9	0.26	114.37
Group 2				
8F (Baro)	3.76	11.1	0.09	462.86
4M	2.50	2.3	0.09	74.43
Group 3				
9O(Mao)	3.79	7.8	0.14	308.28
8K(Bonga)	2.82	7.6	0.07	261.19
8L(Bonga)	3.51	8.8	0.14	484.62
* 1		0.31- 19.3	0.23- 0.38	12- 273

Figure 3.7: Chondrite-normalized REE abundance patterns of representative samples from: 3.7a) Group (1), 3.7b) Group (2), and 3.7c) Group (3). Shaded area shown in Fig 3.7a represents the range of REE distribution for magmatic products above present day subduction zones (Cullers and Graf, 1986). The chondrite factors used are those of Evensen et al. (1978). Sample localities are given in Fig. 2.3.

Figure 3.7 (continued):

during crystallization of Group (1) magmas appear to be low, a condition that is consistent with the little or no modal magnetite and ilmenite. The Group (1) REE patterns and abundances are similar to those shown by calc-alkaline rocks of island arc and continental subduction zones (Fig. 3.7a).

The REE data from rocks of Groups (2) and (3) differ from those shown by Group (1) rocks. Group (2) and (3) samples show strong negative Eu anomalies ($\text{Eu/Sm} = 0.07\text{--}0.14$) and most are more enriched in the LREE than Group (1) rocks. Rocks of Group (3) in comparison to those of Group (1) (Fig. 3.7c) show a significant enrichment of LREE over HREE ($\text{La/Yb}_{(\text{cn})} = 7.6\text{--}8.8$), and higher REE abundances (ΣREE : 261–485 ppm). The garnet - muscovite granite has lower REE abundances (74 ppm), and a flat to concave-upward pattern similar to patterns typically observed in peraluminous granites (Muecke and Clarke, 1981). Data from the Baro granite sheet, on the other hand, have high REE abundances (460 ppm) and are more LREE enriched ($\text{La/Lu}_{(\text{cn})} = 11.1$). The mineralogical differences between the two, particularly the more abundant biotite, zircon and apatite in the Baro leucogranite, could account for the difference in the REE abundances.

3.8 TECTONIC IMPLICATIONS

Trace element abundances and ratios in granitoid rocks (Pearce et al. 1984; Harris et al. 1986) from known tectonic settings have been used to discriminate empirically the geologic setting of granitic rocks. The data define four main groups that reflect (i) ocean ridge (ORG), (ii) volcanic arc (VAG), (iii) within plate (WPG) and (iv) collision granites (COLG). Inter-element plots were used for the granitoid rocks of the study area.

These include Rb vs (Y+Nb), Nb vs Y, Rb vs (Yb+Ta) and Ta vs Yb diagrams (Figs. 3.8 a-d, 3.10). In all of these diagrams the data from the Group (1) rocks suggest that these bodies are the product of magmatic arc activity (VAG), whereas Groups (2) and (3) reflect intra-plate activity (WPG).

Petro et al. (1979) used major elements to recognize granitoid rocks from compressional and extensional plate margins. The diagnostic characteristics of rocks from extensional environments are: alumina undersaturation, high FeO and low MgO and CaO, whereas rocks from compressional environments show alumina oversaturation, low FeO, and high MgO and CaO. Major element distribution fields defined on an AFM diagram (Fig. 3.3) suggest an extensional environment for the low-K suite and a compressional setting for the calc-alkaline suite. The low-K suite, however, has Nb, Zr, Rb, TiO_2 and K_2O more akin to the basaltic members of the island arc volcanic series of Gill (1981) than to the compositions of intrusives formed under conditions of spreading or back-arc rifting. The chemical characteristics of Group (1) rocks, especially the calc-alkaline suite, appear to be typical of granites from orogenic zones. The calc-alkaline suite is hornblende-bearing and exhibits most of the chemical characteristics normally associated with pre-collision calc-alkaline intrusions of a magmatic arc (VAG).

Magmatic arc granites can vary in setting from oceanic to continental. In Fig. 3.9 data from rocks of Group (1) are compared with Phanerozoic plutonic rocks in an attempt to evaluate their tectonic settings. The data from the calc-alkaline suite overlap the data from the Peruvian coastal batholith. Of the low-K suite, the Haya pluton shows similarities with the

Figure 3.8: Trace elements define three fields. These are ocean ridge granites (ORG), within plate granites (WPG) and collision granites (syn-COLG) plus volcanic arc granites (VAG) (after Pearce et al., 1984; Harris et al., 1986); (a) Nb - Y (b) Rb - (Y+Nb) (c) Rb - (Yb+Ta) (d) Ta - Yb. Symbols same as in Fig. 3.1 (a).

Figure 3.8 (continued):

Figure 3.9: Plot of K_2O - SiO_2 variation diagram. Shown for comparison with fields for plutonic rocks from various destructive plate settings. Data sources: Peru (Atherton et al. 1979), Sierra Nevada (Bateman and Dodge, 1970), Solomon Islands and New Britain (Hine and Mason, 1978), Fiji (Gill and Stork, 1979). Symbols same as in Fig. 3.1 (a).

Figure 3.10: Rb and Sr co-variation diagram (after Coleman and Peterman, 1975). The bimodal - low K suite has compositions more akin to those of oceanic plagiogranites than to those of continental granophyres and trondhjemites. Symbols same as in Fig. 3.1 (a).

Fiji tonalites described by Gill and Stork (1979) whereas the Bure rocks are very similar to gabbros from Solomon islands and New Britain both of which represent young, immature, ensimatic arcs. A Rb and Sr co-variation diagram (Fig. 3.10) for Group (1) rocks shows that data points from the low-K suite scatter between the fields for oceanic and continental trondhjemites. The calc-alkaline suite largely falls in the ranges shown by continental granitoid rocks.

Granites of Group (2) show chemical characteristics similar to those of syn-collisional peraluminous intrusions of (Harris et al., 1986). Of particular note are their high Rb, Ba and Ta, and low REE contents (Table 3.5 and 3.7). The presence in the Baro leucogranites of pelitic schlieren as well as substantial amounts of muscovite in the modal mineralogy suggest that they have originated by in-situ partial melting and thus they formed from a wet magma which could not have risen far through the crust without crystallizing. Their genesis is therefore attributed to anatexis of sialic crust, perhaps the metasedimentary schists and gneisses of the Baro Domain.

Group (3) rocks overlap those of Group (2) on plots of Rb vs (Y+Nb) and Nb vs Y, Rb vs (Yb+Ta) and Ta vs Yb diagrams that indicate intraplate affinity for both groups. The source melts may have been generated during thermal relaxation in the lower crust and adiabatic decompression in the upper mantle; they could contain components from both crustal melting, including partial melting of the Group (1) pre-collision suite, and mantle melting (England and Thompson, 1986).

3.9 MINERAL CHEMISTRY

Analyses of different minerals were undertaken in order to estimate magmatic or metamorphic temperature and pressure conditions of formation and, in the case of primary minerals, for testing whether differentiation has taken place by fractional crystallization. Compositions of plagioclase, K-feldspar, biotite, hornblende, apatite, epidote, garnet and muscovite were determined using an electron microprobe. Biotite and hornblende separates from the Birbir quartz diorite were also analysed by X-ray fluorescence and ferrous iron determinations were made by the modified Wilson cold acid method. The analyses are given in Appendix 3.

3.9.1 Pressure and temperature of crystallization

An attempt has been made to infer magmatic crystallization temperatures based on experimental determinations of alkali feldspar solubility relations for known pressures (Stormer, 1975; Whitney and Stormer, 1977). In order to estimate pressures of crystallization the normative rock compositions were plotted in part of the "granite system" (Qz-Or-Ab-H₂O system), projected onto the anhydrous base of the tetrahedron. Tuttle and Bowen (1958) showed that the minimum in the system is depressed and shifted away from quartz as H₂O pressure increases. The quartz - feldspar liquidus cotectic boundaries are given for water pressures of 2, 4 and 7 kb and for Ab/An ratios of 1.8 and 5.2 (Fig.3.11; after Tuttle and Bowen, 1958 as modified by Anderson and Cullers, 1978). Assuming a melt, undersaturated with respect to H₂O, and assuming that those samples with highest silica content represent the most evolved compositions, an estimate can be made of the pressure of

crystallization.

The post-metamorphic Bonga pluton contains large alkali feldspars, some up to 1 cm, and subhedral-euhedral plagioclase laths with straight grain boundaries. The alkali feldspar is perthitic microcline. The pluton has SiO_2 contents as high as 78 wt. % and Differentiation Indices exceeding 95, indicating that it is a highly evolved body. Apparent minimum melt compositions for the pluton suggest a minimum pressure of about 2 kb, and therefore minimum depth of emplacement of about 7 km (Fig. 3.11).

Orthoclase and albite become increasingly miscible with increasing temperature. In slowly cooled intrusions, albite exsolves from the alkali feldspar grains, and separates as unzoned plagioclase lamellae (Whitney and Stormer, 1977). The compositions of coexisting K-feldspar (microcline perthite) and plagioclase from sample T8K (Appendix 3) are $\text{Or}_{92} \text{Ab}_7 \text{An}_1$ and $\text{Ab}_{84} \text{An}_{14} \text{Or}_2$ (mole percent) respectively. These compositions plot outside all the solvi shown in Figure 3.12a, suggesting a temperature of less than 400°C for unmixing.

Fluorine (F) - hydroxyl (OH) exchange in coexisting biotite-apatite pairs has been used to infer temperature conditions of metamorphism (Ludington, 1978). Figure 3.12b shows the partitioning of F and (OH) between coexisting biotite and apatite from intrusives of known equilibration temperature (Ludington, 1978). The hydroxyl content was estimated from the structural formula for a given composition. Analyses of biotite and apatite from metamorphosed intrusive rocks of Group (1), the

Figure 3.11: Normative quartz, albite and orthoclase of Mao and Bonga plutons in comparison to the minimum melt experimental grid compiled by Anderson and Cullers (1978). Open stars represent data from Mao pluton and solid stars that of Bonga pluton.

Figure 3.12:

A) Alkali-feldspar solvi in relation to plagioclase composition (Haselton et al., 1983). Solvi are plotted for 1 Kb based on the phase equilibrium and given in order of decreasing critical temperatures. Feldspar compositions plot outside all solvi shown.

B) Fluorine - hydroxyl partitioning in coexisting biotite and apatite from the Birbir quartz diorite (diagram and reference from Ludington, 1978).

Birbir quartz diorite, have been superimposed; two of the three analyzed pairs indicate a temperature less than 600°C (Fig. 3.12b). These values are consistent with metamorphic equilibration at no higher than mid-amphibolite grade.

E-an Zen and Hammarstrom (1984) and other petrologists have suggested that epidote, commonly regarded as a metamorphic mineral, can be a magmatic product in granitoid rocks. Epidote crystals that seem to satisfy the criteria for magmatic epidote are described in Chapter 2. Diagnostic features include the common association of epidote with mafic minerals, euhedral contacts against biotite, and resorption of hornblende in contact with euhedral epidote and wormy intergrowth with plagioclase.

Average values of seven representative analyses of epidote from sample T9G (Appendix 3) are listed in Table 3.10 along with three analyses from magmatic epidote cited by Zen and Hammarstrom (1984, 1986). There is little difference between the two groups of analyses.

The occurrence of magmatic epidote has been used as an indicator of emplacement depths around 25-30 km (Naney, 1983). Mineral assemblages in the country rocks of epidote-bearing Group (1) plutons indicate a maximum regional metamorphic grade of lower amphibolite facies, suggesting that the depth of burial of these rocks did not exceed about 15 km. If the inferences of Zen and Nancy are applicable to Group (1) rocks, there was substantial uplift between the time of their emplacement and their regional metamorphism in the Birbir Shear Zone (BSZ).

Table 3.10: Average microprobe analysis of epidote from sample T9G of the Birbir complex (mqd) compared to Zen and Hammarstrom's (1986) representative analyses of magmatic epidote (Analyses # 1,2 and 3).

	mqd	(1)	(2)	(3)
SiO ₂	37.75	37.29	37.51	36.64
Al ₂ O ₃	23.99	23.62	25.99	23.62
FeO _(T)	12.02	13.95	12.40	13.69
MgO	0.03	-	0.14	-
CaO	22.74	24.26	23.01	23.34
Na ₂ O	0.05	0.01	-	-
TiO ₂	0.11	0.16	0.05	0.08
MnO	0.20	0.22	0.34	0.49
F	0.11	-	0.03	0.03
Total	97.07	99.51	99.46	97.88

3.10 PETROGENESIS

Fourcade and Allègre (1981) point out the difficulties involved in the petrogenetic modelling of plutonic rocks in comparison with volcanic complexes. During volcanic activity, the frequent sampling of the magmatic reservoir by eruptions enables a more or less continuous evolution in the chemistry of the liquid to be monitored. Granitoids, on the other hand, commonly represent the end products of the chamber itself which may consist of various mixtures of cumulate solids and crystallized liquid. In addition, postmagmatic textural and chemical changes or the metamorphosed nature of the intrusive units make it difficult to infer their history from the bulk composition. Thus, the testing of differentiation and/or fractional crystallization models in the section to follow should be viewed as preliminary.

3.10.1 Mafic and intermediate intrusive complexes: Group (1)

Harker diagrams of data from the calc-alkaline suite (Fig. 3.6) define approximate linear trends for most major elements, that suggest fractionation. Petrographic data for the calc-alkaline suite of Group (1) indicate that plagioclase, biotite, and hornblende are the significant primary and early-crystallizing phases. Most apatite and titanite are poikilitically enclosed in the biotite and hornblende, suggesting their crystallization from a melt. Fractionation of hornblende and plagioclase can explain the decrease in CaO and Sr with increased SiO₂. The decrease in MgO, FeO, P₂O₅, and TiO₂ as SiO₂ increases appears to be controlled through fractionation of biotite, hornblende, apatite and titanite.

The calc-alkaline suite also shows little or no negative Eu anomaly, and small LREE/HREE ($\sim \text{La/Lu}_{\text{(cn)}}$) ratios. These low ratios could be produced by partial melting of a mafic source involving minerals with low LREE/HREE ratios, or fractional crystallization of plagioclase from less differentiated melts with low LREE/HREE ratios (Cullers and Graf, 1986). According to Hanson (1980) europium anomalies could have been prevented in these plagioclase-rich rocks, in one of three ways: (1) large concentrations of Sr and Ba (2) fractionation of hornblende together with plagioclase or (3) a low oxygen fugacity of the melt. The oxygen fugacity during the crystallization of these rocks appears to be low, a condition inferred from the absence of magnetite and ilmenite from the modal mineralogy of these rocks.

The low-K suite is Fe-poor and Na-rich. Similar trends have been described from an arc-type setting where crystal-liquid fractionation of magmas near the base of the crust is one way of suppressing the tholeiitic Fe-enrichment trend, and resulting in the calc-alkaline trends (Ringwood, 1977; Jackson, 1986).

3.10.1.1 Quantitative modeling of major elements in the calc-alkaline suite

Fractional crystallization can be used to model the calc-alkaline suite, represented by the Birbir quartz diorite complex and Goma pluton. The model was evaluated using whole-rock and mineral compositions that involve mass balance calculations using the least-squares computer program XLFRAC, a program used for the interactive testing of magmatic differentiation models (Stormer, 1978). The calculation was made using only major

TABLE 3.8: Hypothetical liquid compositions in the calc-alkaline suite

	QL1	QL2	QL3	QL4
SiO ₂	55.0	60.0	65.0	70.0
TiO ₂	0.86	0.71	0.56	0.41
Al ₂ O ₃	17.8	16.8	15.98	15.1
Fe ₂ O ₃	9.13	7.02	4.91	2.80
MnO	0.16	0.13	0.10	0.08
MgO	3.99	2.85	1.71	0.57
CaO	7.69	5.68	3.67	1.65
Na ₂ O	3.22	3.87	4.52	5.16
K ₂ O	1.69	2.30	2.91	3.52
P ₂ O ₅	0.2	0.17	0.14	0.11

elements.

Hypothetical liquid compositions were compiled by reading the concentrations of the various component oxides from the Harker diagrams (Fig. 3.6) at four equally spaced silica values that span the observed compositional range. The oxides were then normalized to 100%. The liquid compositions are designated QL1 to QL4 (Table 3.8). Microprobe analyses of plagioclase, hornblende, biotite, and apatite are used to represent compositions of the early-formed phases that were subtracted from the derivative melt during fractional crystallization. Proportions of the crystallizing phases and compositions of residual liquid were then calculated using the XLFrac program.

The composition of QL1 was chosen as an approximation to the initial melt composition, differentiating successively to QL2, QL3 and QL4. Given two successive magma compositions, such as QL1 and QL2 and four mineral phases, the program determines the most likely amounts of phases that may have fractionated, and which must be subtracted to produce the observed difference between each composition.

Several criteria are used to evaluate the results of the least-squares mass-balance calculation. The sum-of-the-squares of the residuals, a statistical parameter, is the quantity that is minimized by the program. A low value is generally indicative of a good solution. It is impossible to give an absolute value for all cases which indicate an acceptable solution, but generally solutions with low values (<2) are better than ones with high values (>5) (Stormer, 1978). The proportions, as well as compositions, of precipitated minerals and the residual liquid for each one of the silica

Table 3.9: Testing the feasibility of fractional crystallization in mqd.

Recalculated analyses (Wt. % basis)			
Observed - calculated = Residuals			
	QL1 to QL2	QL2 to QL3	QL3 to QL4
SiO ₂	- 0.115	- 0.36	- 0.053
TiO ₂	- 0.381	- 0.336	- 0.147
Al ₂ O ₃	- 0.209	- 0.072	- 0.058
FeO	0.337	0.196	0.132
MnO	0.029	0.015	0.019
MgO	- 0.707	- 0.341	- 0.271
CaO	- 0.227	- 0.105	- 0.084
Na ₂ O	0.983	0.472	0.368
K ₂ O	0.145	0.099	0.044
P ₂ O ₅	0.144	0.108	0.050
Sum of the squares of the residuals			
	1.876	0.530	0.266
Amount as weight percent of subtracted phases			
plagioclase	45.23	44.04	45.28
hornblende	50.19	50.67	50.64
biotite	3.52	3.94	3.06
apatite	1.07	1.35	1.02

compositions were calculated. The results for major element fractionation are presented in Table 3.9. In all cases the sums of the squares of the residuals are less than 2 suggesting a good fit of the calculated compositions with those that were obtained for the variation diagram. The sum-of-the-squares of the difference between the observed and calculated values is 1.88 for QL1 to QL2, 0.53 for QL2 to QL3, and 0.27 for QL3 to QL4. The results of the mass-balance calculation are thus consistent with fractional crystallization producing the calc-alkaline suite, in which plagioclase, hornblende, biotite and apatite were the main fractionating minerals.

3.10.1.2 Origin of the Birbir intrusive complex

Chemical data indicate that Group (1) rocks represent the products of magmatic arc plutonism. Field relations suggest that these intrusive rocks are pre- to syn-kinematic. In terms of Rb-Sr, K_2O-SiO_2 and other chemical variables the calc-alkaline suite is considered to be the result of magmatism in an Andean-type continental margin whereas the low-K suite displays chemical characteristics of intrusive rocks identified with an oceanic environment.

The origin of the low-K suite is uncertain. It could be: 1) part of an oceanic island arc succession, 2) the product of an incipient rift representing transitional material along an attenuated continental margin or 3) the upper part of an ophiolite suite. Considering the K_2O content is relatively high for ophiolitic rocks, an island arc or back-arc basin mode of development is preferred. This choice is also supported by the nature of

the associated sedimentary material. The spatial coincidence of the calc-alkaline suite with the Birbir Shear Zone and the presence nearby of metasedimentary rocks (quartz - muscovite schist and conglomerate (qms)) suggestive of a continental shelf environment imply that the eastern boundary of the Baro Domain marks a continental margin. The presence of predominant wackes and pelites (mss), further east in the Birbir Domain, suggest an intraoceanic environment for the low-K suite.

Numerous models have been proposed for the origin of orogenic granites. These fall into one of three groups: (1) extraction directly from the mantle; (2) reworking of older continental crust; and (3) mixing of both mantle and crustal components. Wyllie et al. (1976) and Wyllie (1977) described phase relationships for granites and tonalites and showed that liquids less siliceous than granodiorite composition can not be generated by crustal anatexis under conditions of normal regional metamorphism. The chemical evolution of arc-type rocks is generally considered to be the result of the complex interplay of deep continental crust, subducted oceanic crust and the mantle wedge above a subduction zone, in response to the introduction of water and volatiles from the dehydrating slab of subducted oceanic lithosphere (Ringwood, 1974, 1977; Mysen, 1979).

According to Brown (1982), the space-time trends of granitoid rocks from modern compressional arcs reflect a stage of plutonism in ocean-continent arc settings, where dominant calc-alkaline magmas intrude mature continentalized margins with or without ancient sialic basement. Brown described overlapping tectonic and geochemical zones in South America and New Guinea; these are (1) an arc founded on oceanic crust which contains

diorites and low-K granodiorites; (2) an elongate thrust zone with quartz diorite and granodiorite that follow more closely the medium-K calc-alkaline trend of the Andean batholiths, and (3) the continental arc, where a high-K series of granitoid rocks intrudes the crystalline basement.

Zones (1) and (2) of Brown (1982) provide contemporary analogues that could represent the complete range of magma types for the low-K and calc-alkaline suites in the Birbir Domain. These two suites are strongly deformed, and their age relations cannot be inferred. Although the relative age of these bodies is unknown, if they are cogenetic then the low-K suite is best interpreted as predating the calc-alkaline suite.

Similar chemical paths to those deduced from the Birbir intrusive rocks have been documented by many workers (see Best, 1975; Gill, 1981; Mysen, 1979). The stages of development are: (1) progressive thickening of the continental crust above the subduction zone, upward migration of mantle-derived basaltic magma, and amphibole-pyroxene controlled, calc-alkaline fractionation; (2) change in the composition of the melt in the mantle wedge, with progressive enrichment in LIL elements and volatiles as more material is added from the subducted slab that is undergoing dehydration. The melt generated during subduction has a relatively high oxygen fugacity, thus Fe-enrichment in the crystallizing phases is suppressed. The early melts have tonalitic or trondhjemitic affinities (e.g. Fiji: Gill and Stork, 1979), similar to the low-K suite. The Haya tonalites could represent the felsic magmas generated by partial melts of tholeiitic crust, the

forerunners of arc magmatism.

3.10.2 Peraluminous granite suite: Group (2)

Chemical trends for rocks of Group (2) differ markedly from Chappell and White's (1983) peraluminous rocks as well as the metaluminous calc-alkaline suite of Group (1) and alkali-feldspar granite of Group (3). There is no sign that these rocks were derived from a more mafic parent, as discussed in previous sections and Chapter 2. Crustal sources with a significant metasedimentary component are often invoked for peraluminous melts (Hine et al., 1978; Clarke and Halliday, 1980). Low Na, Ca and Sr abundances, characteristic of these rocks, agree with what would be expected from sedimentary rock sources, as these elements are removed in solution during chemical weathering. Clemens and Wall (1981) described granites with chemical characteristics such as high K_2O/Na_2O , low CaO and Sr content, similar to analyses from Group (2). They considered these granites to be the product of anatexis of aluminous high-grade metamorphic rocks. The peak pressure and temperature in unit (gsg) (see Chapter 2 for details) reached minimum melting conditions for rocks of quartzofeldspathic composition. At temperatures above the muscovite - quartz stability field, partial melting in high-grade metamorphic rocks of peraluminous compositions involves sillimanite, garnet, cordierite, biotite, quartz and feldspars (Clemens and Wall, 1981) similar to the assemblages described from unit (gsg). One possible equilibrium reaction for the assemblage is:

The garnet-bearing schlieren within the Baro leucogranites could represent partially digested inclusions, or a residual crystalline assemblage.

3.10.3 Alkali-feldspar granite suite: Group (3)

Group (3) plutons post-date regional metamorphism. Low confining pressure during emplacement is inferred for the Bonga pluton from two lines of evidence. The most evolved compositions, with Differentiation Indices of 94, indicate crystallization at pressures less than 2 kb (~7 km). Granitic magmas increasingly approach the "ternary" minimum composition with increasing differentiation (Tuttle and Bowen, 1958). The absence of pegmatitic phases within the Bonga pluton suggests that the melt was vapour-undersaturated, with an H₂O content of less than 1.5 wt.% (Maaloe and Wyllie, 1975; Burnham, 1979).

The compositional field of the Mao pluton indicates pressures of about 2 to 4 kb (7.5 to 15 km) and could reflect fusion in the lower levels of a thickened crust. The lower amount of normative quartz, in the less differentiated Mao pluton, is expected because higher pressure conditions shift the pressure sensitive quartz-feldspar boundary plane (Presnall and Bateman, 1973). The felsic nature of this pluton indicates that it is probably still on the intersection of the two-feldspar and quartz-feldspar planes, not far removed from minimum melt composition.

The alkaline affinity of Group (3) rocks, indicated by major elements

and high Zr and Nb (i.e. Zr > 200 ppm; Nb > 15 ppm), suggests a possible derivation from an igneous source that had previously been partially melted (Pitcher, 1981). Altogether the field relations and chemistry suggest an epizonal plutonic emplacement as product of partial fusion of sialic crust.

CHAPTER 4

GEOCHRONOLOGY AND ISOTOPE GEOCHEMISTRY

4.1 GENERAL STATEMENT

Isotopic dating of intrusive and metamorphic events in the Precambrian rocks of Ethiopia is at a reconnaissance stage. The available data, from southern and northern Ethiopia only, are mostly either K-Ar or Rb-Sr mineral dates. The few published dates available are summarized in Table 4.1. Most results fall between 700 and 450 Ma. Prior to the present study, no geochronological data were available from the study area or from elsewhere in the Western Ethiopian Shield.

The high-grade rocks of the Western Ethiopian Shield have been regarded as the northward continuation of the Mozambique Belt, representing basement unconformably underlying the lower-grade metamorphic rocks of the Birbir Domain (Kazmin, 1971). The lower-grade calc-alkaline sequences and ophiolitic rocks, however, are without a recognizable older substrate.

Kroner (1977, 1979), Hepworth (1979, 1980) and many others consider the Mozambique Belt to be an ensialic mobile belt, polycyclic in origin, which contains reworked older crust. Shackleton (1986) regarded the Mozambique Belt to be a complex assemblage of Proterozoic belts of different ages, and argued that the strong Pan-African deformation and metamorphism has obliterated Archean and early Proterozoic dates within the belt.

The present geochronologic data neither support nor refute the view that the high-grade gneissic rocks thought to be Mozambiquian are older than the juxtaposed lower-grade (Pan-African?) rocks. The oldest event documented

Table 4.1: Isotopic age determinations of Precambrian rocks in Ethiopia.

LITHOTECTONIC UNIT ¹	LOCALITY	ROCK TYPE		METHOD	AGE/DATE ² (Ma)
Southern gneiss zone	El-Der, Sidamo	Hbd-Bi gneiss	Bi	K-Ar	739 +/-15 (1)
"	"	Amphibolite	WR	"	637 +/-20
"	Neghelli, Sidamo	Bi gneiss	Ei	"	507 +/-5
Adola magmatic zone	Moyale, Sidamo	Quartz diorite	Bi	"	517 +/-5
"	Mormora R., Sidamo	Granite	Bi	"	495 +/-20 (2)
Yavello para-gneiss zone	Gariboro, Sidamo	Granite	WR	Rb-Sr	680 +/-10 (3)
"	"	Granite	WR	"	515 +/-10
"	"	Pegmatite	Ms	"	530 +/-10
Northern volcanic-sedimentary terrane	Zula, Eritrea	Mica schist	Ms	K-Ar	976 (4)
"	"	Granite	Bi	"	675
"	Hauzien, Tigre	Granite	Bi	"	583 +/-20 (5)
"	Adigrat, Tigre	Quartz diorite	Bi	"	582 +/-22
"	Mareb V., Eritrea	Granodiorite	Bi	"	627 +/-10 (6)

¹ see Chapter 6

² Sources of data:

- (1) Rogers et al., 1965
- (2) Jelenc, 1966
- (3) Gilboy, 1970
- (4) Frazier, 1970
- (5) Ethiopian Institute of Geological Surveys
- (6) Miller et al., 1967

WR: whole-rock, Bi: biotite, Ms: muscovite, Hbd: hornblende

from Pan-African rocks in the region is 1165 Ma (Rb-Sr whole-rock isochron age from metavolcanic rocks in Saudi Arabia; Fleck et al., 1980). Rb-Sr whole-rock isochrons from volcanics in eastern Egypt, yielded youngest dates from 710 to 610 Ma (Stern, 1979). The oldest reported date of Pan-African granitoid rocks from the region is a Rb-Sr whole-rock isochron of 960 Ma (southern Saudi Arabia; Greenwood et al., 1976). The more fractionated granitoid rocks from southern Saudi Arabia have been dated at 785 - 600 Ma (Greenwood et al., 1976) and from southeastern Egypt at 670 - 620 Ma (Stern, 1979). Younger rocks include alkali-rich granites dated at 590 to 550 Ma (Fullagar and Greenberg, 1978; Abdel Monem and Hurley, 1979).

Gneisses at Ouweinat, west of Pan-African rocks at 25°E, 22°N, were dated by the U-Pb method at 2600 Ma by Klerkx and Deutsch (1977). Kroner et al. (1987) reported ion microprobe ages of detrital zircons that range from ca. 1000 to 2650 Ma. The zircons were separated from a high-grade gneiss terrane at Sabaloka on the Nile river north of Khartoum, Sudan. A U-Pb zircon date of 1700 Ma has been reported from psammitic gneisses of the Jabal Hafafit group of eastern Egypt (Dixon, 1981), a fine-grained granodiorite in the eastern edge of the Arabian shield yielded a zircon U-Pb date of 1630 Ma (Stacey and Hedges, 1984). These dates were reinforced by model Nd ages that yielded 1800 Ma from metapelites in eastern Egypt and 1550 Ma from quartzofeldspathic metasediments in the Bayuda desert of Sudan (Harris et al., 1984).

None of the documented accounts about the Mozambique Belt has been able to assess the time of amphibolite and granulite metamorphism with any degree of certainty. The dates of the latest metamorphism of the Mozambique Belt

rocks in NW Kenya indicate much younger events (ca. 580 Ma; Ries et al., unpublished) than the metamorphic ages of the Pan-African rocks of NE Africa (ca. 760 Ma; Ries et al., 1985). Suwa (1979) obtained a Rb-Sr whole-rock date on gneissic granite from central Kenya of 766 ± 29 Ma, while Leggo (1974) interpreted a lower concordia intercept of 650 Ma as the time when the Pan-African tectono-thermal event affected the Ugandan basement. A similar pattern has been recorded from Tanzania, where a U-Pb zircon lower intercept of 652 ± 10 Ma is thought to be the age of the granulite facies metamorphism (Coolen et al., 1982). These late Proterozoic dates from the Mozambique Belt and the Pan-African rocks in NE Africa contrast with those from the craton to the west in Uganda and Tanzania, where Rb-Sr whole-rock ages of ca. 2.7 Ga and U-Pb zircon ages up to ca. 2.9 Ga have been determined (Spooner et al., 1970; Leggo, 1971, 1974; Bell and Dodson, 1981).

4.2 INTRODUCTION

U-Pb zircon and Rb-Sr whole-rock and mineral dating have been carried out in an attempt to learn the time of intrusion of the plutons in the Gore-Gambella area, and their subsequent history. In addition, the isotopic data from these granitoid rocks contributes to an understanding of the regional age framework.

The present isotopic work identifies distinct periods of plutonism and tectonic activity throughout the Birbir Domain and the Baro Domain, where detailed mapping has provided structural control on a complete cross section across the western Ethiopian Shield (see Chapter 2). Three generations of granitoid rocks are recognized on the basis of field relations and

geochemical data. The earliest consists of pre-kinematic, arc-type mafic and intermediate intrusive complexes in the Birbir Domain. The second group is made up of anatectic granites from the high-grade metamorphic rocks of the Baro Domain. The third group consists of late- to post-kinematic plutons that occur in both domains. Petrographic study, and major and trace element analyses were performed on all of the samples that were used for the isotopic work (Appendices 1 and 2).

4.3 U-Pb ANALYSES OF ZIRCON AND THE CONCORDIA DIAGRAM

U-Pb zircon ages were determined of five plutonic units. The data are listed in Table 4.2 and plotted on concordia diagrams (Figs. 4.1 - 4.5).

Zircon is used in U-Pb dating because it is widely distributed in many rocks and exhibits atomic substitution of U^{+4} for Zr^{+4} whereas Pb^{+2} is essentially excluded from the mineral during crystallization. The greatest geological interest in zircon stems from its common ability to retain U, Pb and intermediate decay products over geologic time, i.e. preserve its isotopic signature, despite partial melting or high grade metamorphism to temperatures as high as 800°C (Speer, 1982).

Pb has four naturally occurring isotopes of which only ^{204}Pb is considered non-radiogenic. The decay of the two radioactive uranium isotopes ^{238}U and ^{235}U to their respective daughter products ^{206}Pb and ^{207}Pb can be expressed in the form of two equations (Faure, 1986):

$$^{206}Pb/^{204}Pb = (^{206}Pb/^{204}Pb)_0 + ^{238}U/^{204}Pb (e^{\lambda_1 t} - 1)$$

$$^{207}Pb/^{204}Pb = (^{207}Pb/^{204}Pb)_0 + ^{235}U/^{204}Pb (e^{\lambda_2 t} - 1)$$

where:

$^{206}\text{Pb}/^{204}\text{Pb}$, $^{207}\text{Pb}/^{204}\text{Pb}$ = isotope ratios of lead in the mineral at the time of analysis; $(^{206}\text{Pb}/^{204}\text{Pb})_0$, $(^{207}\text{Pb}/^{204}\text{Pb})_0$ = initial isotope ratios of lead incorporated into the mineral at the time of its formation; $^{238}\text{U}/^{204}\text{Pb}$, $^{235}\text{U}/^{204}\text{Pb}$ = isotope ratios in the mineral at the time of analysis; λ_1 , λ_2 = decay constants of ^{238}U and ^{235}U , respectively; t = time elapsed since closure of the mineral to uranium, lead and all intermediate daughters.

The ratios $^{206}\text{Pb}^*/^{238}\text{U}$ and $^{207}\text{Pb}^*/^{235}\text{U}$ (*: radiogenic lead) can be calculated for any given time, t . Plotting these two values for all t yields a curved locus of points called **concordia** (Wetherill, 1956). Any system which has not suffered uranium or lead disturbance should plot on this curve. A system that yields similar ^{238}U and ^{235}U dates is said to be concordant; if they are not, the dates are discordant. Concordant dates represent the age of the mineral with the following assumptions:

- 1- the mineral has behaved as a closed system throughout its history,
- 2- correct values are used for the initial lead isotope ratio,
- 3- the decay constants are known accurately,
- 4- the isotopic composition of U is normal and has not been modified by isotope fractionation,
- 5- analytical results are accurate and free of systematic errors.

For every time point on concordia, there is a unique $^{207}\text{Pb}/^{206}\text{Pb}$ ratio, and this varies from about 0.45 at 4 Ga to 0.04607 for the present. Ages calculated using only this ratio are referred to $^{207}\text{Pb}/^{206}\text{Pb}$ ages.

Discordancy may result from loss or gain of Pb or U. Several lead-loss models have been proposed to interpret discordant zircons. However, Krogh

(1982) noted that the exact cause of discordia lines is largely unknown and introduced two important techniques in an attempt to eliminate rather than explain discordances. These are the air-abrasion technique and separation of the least paramagnetic grains. Both techniques are designed to eliminate from the sample the most damaged zircon fraction, that is most subject to lead loss. According to Krogh, discordancy in abraded zircon populations is of two general types:

- 1) **Collinear discordia:** The abraded samples yield data that move toward concordia along a line defined by the more discordant points. This results in a linear array on the concordia diagram. In the simplest interpretation, the upper intersection of this line with concordia represents the age of zircon formation, whereas the lower intercept represents the time of Pb loss or U gain. Lead can be lost from the system by either continuous diffusion (Tilton, 1960), or by loss of water together with accumulated Pb contained in metamict zircons (dilatancy), the result of uplift and decrease in pressure (Goldich and Mudrey, 1972).
- 2) **Multiple age population:** Inheritance can result either (1) from the mixture of magmatic zircon with xenocrystic zircon from the source region of the magma or (2) solid state metamorphic growth of zircon under amphibolite or granulite facies conditions, around pre-existing crystals (Krogh and Davies, 1973; Parrish and Roddick, 1985). Mixing by any process of concordant grains with two different ages results in a linear array of data points that when extrapolated meets concordia at lower and upper intersections representing the two periods of zircon growth. As a measure of

the goodness of fit of regression lines fitted to these arrays, the mean square of weighted deviates (MSWD) is employed (Ludwig, 1980). Abrasion allows the true ages of the two-component mixture to be determined only to the extent that discordance for Pb loss can be eliminated. Krogh also indicated that the degree of discordance can vary between grains in a single cogenetic fraction. Grains with their surfaces removed and the least paramagnetic grains provide the most concordant data.

4.3.2 Analytical methods

U-Pb analyses of zircons from the study area were made at the geochronology laboratory of the Geological Survey of Canada. Details of the analytical procedure are given in Appendix 4.1. Heavy mineral concentrates were obtained using a heavy liquid (methylene iodide with 3.325 density at room temperature) and run through the Frantz magnetic separator at 1.0 amp. with a forward slope of 25° and side tilt of 15° . The non-magnetic fraction was run through the separator repeatedly until most magnetic grains were removed by progressively increasing the amperage. The zircon concentrate was then split into different size fractions (-149, +149 to -105, +105 to -74, +74 to -62, and +62) with nylon sieves. The final zircon concentrate for analysis was obtained by hand picking. An air abrasion technique was used to remove the outer parts of the zircon in order to reduce the degree of discordancy (Krogh, 1982). For extraction of U and Pb, the chemical procedure developed at the geochronology laboratory of the Geological Survey of Canada (Parrish et al., 1987) was used.

Table 4.2: U-Pb analytical data.

Sample size ¹ (mm)	Weight (mg)	U. (ppm)	Pb ² (ppm)	²⁰⁶ Pb/ ²⁰⁷ Pb	Pb _c ⁵ (pg)	²⁰⁸ Pb ² (%)	(²⁰⁶ Pb/ ²³⁸ U) ⁶ P	²⁰⁷ Pb/ ²³⁵ U ⁶	²⁰⁷ Pb/ ²⁰⁶ Pb ⁶	²⁰⁷ Pb/ ²⁰⁶ Pb age(Ma)
1. Baro granite (sample # T8F)										
a)-149	0.126	616.65	72.74	30970	18	9.4	0.1168+/-0.06	1.0238+/-0.09	0.06359+/-0.04	777.9+/-1.6
b)-149+105	0.062	638.02	74.57	11240	26	8.6	0.1168+/-0.06	1.0231+/-0.09	0.06356+/-0.04	726.9+/-1.7
c)-105	0.028	651.5	78.61	16220	9	9.4	0.1194+/-0.09	1.054+/-0.11	0.06398+/-0.04	741.1+/-1.7
d)-105+74	0.067	706.66	81.36	19630	17	8.5	0.1152+/-0.06	1.0056+/-0.09	0.06334+/-0.04	719.6+/-1.6
e)-105	0.010	492.9	60.8	2441	15	11.3	0.1196+/-0.12	1.055+/-0.14	0.06401+/-0.06	741.9+/-2.6
f)-74+62	0.032	650.8	74.44	14260	11	8.7	0.1142+/-0.10	0.9957+/-0.12	0.06322+/-0.04	715.8+/-1.7
2. Birbir (west., sample # T3G)										
a)-149	0.081	141.47	19.73	6578	15	10.3	0.1363+/-0.08	1.253+/-0.10	0.06671+/-0.05	828.6+/-2.0
b)-149+105	0.109	141.61	19.51	8085	16	9.8	0.1355+/-0.08	1.245+/-0.10	0.06665+/-0.04	826.9+/-1.8
c)-105+74	0.105	131.48	18.38	8670	14	10.7	0.1361+/-0.07	1.250+/-0.10	0.06663+/-0.04	826.1+/-1.8
d)-74+62	0.0413	140.79	19.35	1791	28	9.9	0.1350+/-0.14	1.241+/-0.15	0.06665+/-0.09	826.7+/-3.8
3. Goma (sample # T6J)										
a)+105,NA	0.062	149.97	17.84	2298	29	11.7	0.1145+/-0.10	1.043+/-0.13	0.06602+/-0.08	807.0+/-3.1
b)-105+74	0.053	97.41	13.57	3350	13	12.1	0.1335+/-0.11	1.219+/-0.14	0.06620+/-0.06	812.5+/-2.6
c)-74+62	0.037	111.85	15.69	2829	12	12.5	0.1338+/-0.10	1.222+/-0.13	0.06625+/-0.06	814.3+/-2.6
4. Mao (sample # T9O)										
a)-149+105,NA	0.117	215.2	18.86	4219	33	9.3	0.0873+/-0.11	0.7109+/-0.13	0.05907+/-0.05	569.6+/-2.1
b)-105+74	0.0275	309.8	28.75	1141	43	10.5	0.0912+/-0.10	0.7417+/-0.18	0.05901+/-0.05	567.4+/-5.2
c)-105+74	0.0181	444.5	41.73	5194	9	11.8	0.0909+/-0.10	0.7404+/-0.12	0.05908+/-0.05	570.0+/-2.2
d)-105+74	0.025	157.9	15.27	2424	10	11.9	0.0934+/-0.13	0.7698+/-0.16	0.05975+/-0.08	594.4+/-3.5
e)-74+62	0.031	207.7	21.46	2426	17	11.9	0.0997+/-0.11	0.8368+/-0.14	0.06086+/-0.08	634.3+/-3.4
5. Bonga (sample # T8L)										
a)-149	0.051	175.33	15.99	4449	12	8.3	0.0919+/-0.10	0.7484+/-0.13	0.05910+/-0.06	570.6+/-2.5
b)-149+105(1)	0.035	196.37	17.87	2295	17	9.3	0.0906+/-0.10	0.7384+/-0.14	0.05910+/-0.08	570.7+/-3.3
c)-149+105(2)	0.029	351.12	31.39	3272	18	7.6	0.0907+/-0.10	0.7383+/-0.13	0.05906+/-0.06	569.1+/-2.6

¹ Sieve size in microns.
² radiogenic Pb.
³ Corrected for fractionation and spike.
⁴ Total common Pb in the analysis.
⁵ Corrected for fractionation, spike, blank and common Pb, blank U.
⁶ Errors are 2 sigma in Na.
⁷ NA not abraded

4.3.3 Pre- and/or syn-kinematic intrusive rocks

Zircons from the Birbir quartz diorite complex (mqd) and Goma granodiorite pluton (mgd) broadly display similar morphological characteristics. The zircons are euhedral to subhedral, prismatic, mostly clear to partly turbid, and show multiple growth zoning under transmitted and reflected light. The zoning suggests that these zircons have a primary igneous origin. Some fractions contain distinctly turbid, flat, subequant grains, that may represent zircons strained by shearing; these zircons were not analyzed.

A nearly-concordant upper intercept of $828 \pm 9/-2$ Ma was obtained from four abraded size fractions extracted from a protomylonitic quartz diorite of the Birbir complex (sample # T3G, Fig. 4.1; see Figs. 2.2 and 2.3 for sample localities). The small degree of uncertainty attached to the upper intercept unambiguously indicates the emplacement age of the Birbir quartz diorite complex. Three data points (2 abraded and 1 not abraded) from the Goma granodiorite pluton yield an age of $814 \pm 2/-2$ Ma (sample # T6J; Fig. 4.2) which is also interpreted as the intrusion age. The data point for the unabraded +105 mm size fraction plots below concordia and clearly shows the importance of abrasion in producing more concordant results from zircons.

4.3.4 Syn-kinematic Baro leucogranite

Highly-deformed leucogranites characterize the Baro Domain. In both reflected and transmitted light, the zircons from one of these granite sheets, the Baro leucogranite (sample # T8F; sample locality shown in Fig. 2.2), are mainly subhedral and tabular to ellipsoidal in shape. The least

Figure 4.1: Concordia diagram for the Birbir pluton.

Figure 4.2: Concordia diagram for the Goma pluton.

damaged crystals in six fractions were selected, and all fractions were abraded.

Data are shown in Fig. 4.3. They define a linear array below concordia, with an upper intercept of $783 \pm 19/-14$, a lower intercept of $582 \pm 29/-33$ and an MSWD of 1.59. The large uncertainty attached to both intercept ages reflect the restricted spread of data and the long extrapolations to concordia. The dates so obtained could be a result of different degrees of Pb loss within a cogenetic zircon population or alternatively reflect a mix of isotopic ratios of zircons of different generations. An interpretation of the discordant dates thus required further investigation, particularly as to the internal morphology of the zircons.

Photomicrographs with a scanning electron microscope (SEM) of HF-etched zircons from T8F show a complex internal morphology. Plates 3.1 A, B and C show that these zircons have readily recognizable euhedral zoned cores rimmed by a clear, unzoned overgrowth. Such a pattern is typical of metamorphic overgrowth on igneous zircon. Some contain small rounded inclusions of an earlier generation of zircon within the zoned crystals. In addition the igneous zircons contain hematite, feldspar(?), and fluid inclusions.

The data are typical of the multiple age population of Krogh (1982) that represent mixtures of magmatic zircon with metamorphic overgrowth. The upper intersection age is interpreted as a first approximation to the age of the magmatic zircons. This age is thus attributed to igneous emplacement near the time of anatexis under peak metamorphic conditions in the Baro Domain (see Chapters 2 and 3). It coincides with metamorphic age of similar

Figure 4.3: Concordia diagram for the Baro pluton.

Plate 4.1: Scanning electron photomicrographs of HF-etched zircon grain mounts from the Baro leucogranite (Plates 4.1: A,B,C) and the Mao pluton (Plates 4.1 D, E, F). Plates are described in text.

rocks from W Kenya (Suwa, 1979).

The lower intersection is inferred as a time of high-temperature metamorphic overgrowth of zircon at 580 ± 30 Ma. There is also independent evidence from high-grade rocks in NW Kenya, that are similar to rocks found in the Baro Domain that is suggestive of metamorphic activity at 584 Ma (Ries et al., unpublished) Thus, the zircon overgrowths are evidence of upper amphibolite facies metamorphism at ca. 580 Ma ago in the Baro Domain.

4.3.5 Late- to post-kinematic plutons

U-Pb analyses of four zircon fractions from sample # T90 of the Mao pluton yielded discordant dates when plotted on a concordia diagram (Fig. 4.4). The pluton is discordant with respect to the foliation and layering in the country rocks. The sample comes from the eastern margin of the pluton and shows foliation and shear bands, superimposed on an igneous texture. The zircons extracted from the sample are predominantly clear and prismatic but some are tabular and of turbid appearance. Four size fractions of the former were analyzed, and fit a chord with an MSWD = 4.22, with an upper intercept of $772 \pm 70/-63$ Ma and a lower intercept at $541 \pm 10/-16$ Ma.

Plates 4.1 D, E and F show SEM photomicrographs of zircons from sample # T90. The lowermost zircon grains shown in Plate 4.1D consists of a core with dark, possibly hematite inclusions, mantled by well-developed igneous zonation while those shown in 4.1E and F are free of cores and display well-developed igneous zoning. A few contain re-entrants filled with homogeneous zircon that may have resulted from microfracturing and filling of indentations, metamorphic regrowth or resorption possibly at the time of shearing.

Figure 4.4: Concordia diagram for the Mao pluton.

No simple interpretation of discordia is possible here for this complex internal morphology. The apparently inherited cores shown in Plate 4.1 D and the data points shown in Fig. 4.4 that trend away from concordia appear to result primarily from a mixture of two generations of zircon. Data points of fractions that plot close to the lower intersection are considered relatively free of inherited zircon. Thus, the convergence of data from abraded zircons on a lower intercept of about 550 ± 10 Ma is considered as the best estimate for the age of igneous zircon crystallization, and therefore the age of intrusion. The upper intersection has a large uncertainty but may be tentatively interpreted as the age of the inherited cores. The infilled re-entrants that characterize some of the grains may have contributed to the large uncertainty.

It is interesting to note that the upper intersection overlaps within analytical uncertainty the age of the pre- and/or syn-kinematic intrusive rocks in which the Mao pluton was intruded. It is interpreted as the age of the zircon generated at the same time as the Baro leucogranites.

Analyses of three zircon fractions from quartz monzonite of the Bonga pluton define a nearly concordant data set at $571 \pm 11/-3$ Ma (Fig. 4.5). The pluton is discordant with respect to the regional fabric; it straddles the boundary between the Birbir and Baro domains. The sample, massive and containing K-feldspar megacrysts, was obtained from the quartz monzonite core of the pluton which is cut by granite that forms the margin of the pluton. The analyzed zircons are all clear, prismatic and show typical

Figure 4.5: Concordia diagram for the Bonga pluton.

igneous growth zonation under transmitted light. The $571 \pm 11/-3$ Ma date is thus interpreted as representing the time of emplacement of the quartz monzonite of the Bonga pluton.

4.4 Rb-Sr SYSTEMATICS AND THE ISOCHRON DIAGRAM

The number of ^{87}Sr daughter atoms produced by decay of ^{87}Rb in a rock or mineral since time of formation, t years ago, is given by an equation similar to those already used to treat the decay of U:

$$^{87}\text{Sr}_T = ^{87}\text{Sr}_i + ^{87}\text{Rb} (e^{\lambda t} - 1)$$

where $(^{87}\text{Sr})_T$ is the total number of atoms at the present time; $(^{87}\text{Sr})_i$ is the number of atoms of this isotope that were initially incorporated into the system at the time of its formation; ^{87}Rb is the number of atoms of this isotope at the present time; λ is the decay constant of ^{87}Rb ; t is the time elapsed since the time of formation of the system. Rather than use the absolute number of atoms of a given nuclide, ratios are used by dividing through by ^{86}Sr , and the equation is rewritten as:

$$(^{87}\text{Sr}/^{86}\text{Sr})_T = (^{87}\text{Sr}/^{86}\text{Sr})_i + ^{87}\text{Rb}/^{86}\text{Sr} (e^{\lambda t} - 1)$$

This equation has two unknowns: the age t and the initial ratio, $(^{87}\text{Sr}/^{86}\text{Sr})_i$. The $^{87}\text{Sr}/^{86}\text{Sr}$ of a rock sample increases with time, the rate of increase depending on the Rb/Sr ratio of the sample. Similarly the initial ratio of a suite of rocks depends on the Rb/Sr ratio of the source of those rocks. In order to determine the age of an igneous body, the present day $(^{87}\text{Sr}/^{86}\text{Sr})_T$ and $^{87}\text{Rb}/^{86}\text{Sr}$ ratios of a suite of samples are measured. All rock specimens belonging to a comagmatic suite will plot as points on a straight line in coordinates of $^{87}\text{Rb}/^{86}\text{Sr}$ and $^{87}\text{Sr}/^{86}\text{Sr}$. This line is called an isochron. Its slope is proportional to t , and the intercept defines $(^{87}\text{Sr}/^{86}\text{Sr})_i$. A suite of co-magmatic rocks of age t will define an isochron only when each member of that suite had the same

$(^{87}\text{Sr}/^{86}\text{Sr})_i$ and the rocks have behaved as a closed system since crystallization.

Dating of metamorphosed igneous rocks is based on the assumption that a whole-rock domain of a certain size remains as an effectively closed system during a metamorphic event. This is achieved by collecting samples exhibiting a range of Rb/Sr ratios, so that the resultant isochron may be well defined. If a data point does not fit the isochron within random analytical errors, then the coordinates of the data point either have systematic analytical error or the sample does not satisfy the assumptions inherent in the isochron method (Faure, 1986).

Whole-rock Rb-Sr dating of granitoid rocks commonly yields younger ages than the U-Pb ages from the same rock (Kennan and Murphy, 1987). Bell and Blenkinsop (1978) and Field and Raheim (1979) have shown that the whole-rock systems can be completely re-set by metamorphic activity. It is also possible that an apparent isochron age produced by incomplete re-setting may yield an age that may have little meaning and be unrelated to any geological event. In other cases, more than one isochron date from the same pluton could result from heterogeneous deformation and recrystallization.

The Rb-Sr isotopic data for five plutonic units are listed in Table 4.3. In addition, results from various regression analyses using different sample groups are summarized in Table 4.4.

4.4.1 Sampling and analytical methods

Samples were collected from a total of nine plutons. Between 8 and 14

samples per pluton were selected representing the compositional and textural variability of the plutons. Samples that showed visible alteration, fractures, veins and xenoliths were avoided. The geographical spread of the samples typically extends over several kilometres (Fig 2.3). Two to 3 kg samples were considered adequate for Rb-Sr geochronological studies. Rb and Sr concentrations were determined by X-ray fluorescence spectrometry on 10-15 gm pellets, using G-2 as the reference standard. Mass absorption coefficients were determined using the method of Norrish and Chappell (1967). The estimated uncertainties are +/-2% for Rb and Sr abundance, and 1% for the Rb/Sr ratios.

Standard ion exchange procedures were employed in Rb-Sr separation and collection (see Appendix 5). Isotopic analyses were made on a Finnigan-MAT 261 multicollector, solid source mass spectrometer operated in the static mode (see Bell et al., 1987).

Errors are quoted throughout at the 1 sigma level and the decay constant used in the age calculations is the value recommended by the IUGS subcommission for geochronology (Steiger and Jager, 1977). Regression lines were calculated using a least squares method based on that of York (1969). Where the mean square of weighted deviates (MSWD) exceeds 2 as the limiting or critical value, the scatter of data about the line can not be accounted for entirely by experimental error. If a line is fitted through such data it is termed a "scatterchron", and the values quoted for age and $(^{87}\text{Sr}/^{86}\text{Sr})_i$ should be viewed with caution.

Table 4.3: Rb-Sr whole-rock and mineral analytical data.

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Sample	Rb ¹ (ppm)	Sr ¹ (ppm)	Rb/Sr (atomic)	⁸⁷ Rb/ ⁸⁶ Sr (atomic)	⁸⁷ Sr/ ⁸⁶ Sr ²
Birbir (west) - whole-rock					
T5G	66	609	0.108	0.311	0.70648
T9G	57	564	0.101	0.290	0.70632
T12G	73	517	0.142	0.409	0.70728
T6G	49	584	0.084	0.242	0.70608
T7G	49	656	0.076	0.219	0.70588
T11G	30	891	0.034	0.097	0.70454
T4G	69	569	0.124	0.357	0.70694
Birbir - Biotite					
T5G	320.8	4.4	72.262	238.6	2.15168
T3H	289.5	4.7	62.128	216.9	2.82419
T8I	233.4	4.9	47.356	152.79	1.88806
Birbir (central) - whole-rock					
T3H	63	429	0.148	0.428	0.70742
T5H	49	495	0.099	0.287	0.70671
T8H	42	443	0.095	0.275	0.70646
T10H	59	462	0.127	0.368	0.70725
T2H	62	511	0.122	0.353	0.70728
T7H	56	499	0.112	0.324	0.70723
T4H	55	498	0.109	0.316	0.70706
Goma - whole-rock					
T4J	64	281	0.229	0.662	0.70989
T5J	71	296	0.241	0.699	0.71027
T6J	52	342	0.151	0.436	0.70753
T21J	68	304	0.222	0.643	0.70946
T23J	71	283	0.251	0.725	0.71070
T24J	59	325	0.181	0.524	0.70838
T1J	67	311	0.214	0.618	0.70955
T2J	53	321	0.164	0.475	0.70857
T3J	56	341	0.164	0.474	0.70834
T8J	54	358	0.151	0.436	0.70790

¹ Rb and Sr abundances for the whole-rock samples were determined by XRF, and for biotite by isotope dilution analysis.

² Measured ⁸⁷Sr/⁸⁶Sr ratio is normalized to a value of ⁸⁶Sr/⁸⁸Sr = 0.1194

Table 4.3 (continued)

Sample	Rb ^I (ppm)	Sr ^I (ppm)	Rb/Sr (atomic)	⁸⁷ Rb/ ⁸⁶ Sr (atomic)	⁸⁷ Sr/ ⁸⁶ Sr
Mao - whole-rock					
T7O	178	229	0.780	2.258	0.72110
T9O	176	238	0.739	2.139	0.72057
T11O	189	215	0.886	2.567	0.72386
T5O	176	221	0.798	2.312	0.72181
T6O	185	224	0.826	2.393	0.72220
Bonga - whole-rock					
T6K	139	38	3.672	10.699	0.78540
T8K	111	36	3.066	8.912	0.77646
T10K	117	45	2.598	7.554	0.76409
T11K	115	44	2.632	7.657	0.76574
T1K	111	61	1.827	5.303	0.74667
T4K	92	40	2.322	6.748	0.75652

Table 4.4: Results of various regression of whole-rock Rb-Sr data.

Pluton/ complex	Sample Nos.	Date ¹	Intercept	MSWD
Goma	TJ:4,5,6,21 23,24,1,2,3,8 (10 pts.)	677 +/-12	0.703576 +/- 0.0001	20.6
	TJ:6,24,1,21, 4,5,23 (7pts.)	750 +/-17	0.702813 +/- 0.0001	3.6
	TJ:6,24,1,4,5, 23 (6 pts.)	759 +/-18	0.702773 +/- 0.0001	1.3
	TJ:8,3,1,4,5, 23 (6 pts.)	640 +/-14	0.703948 +/- 0.0001	2.3
Birbir (west.)	TG:5,9,12,6,7, 11,4 (7 pts.)	698 +/-67	0.703745 +/- 0.00003	17.3
	TG:5,9,12,11, 4 (5 pts.)	632 +/-8	0.703675 +/- 0.00003	2.2
Birbir (central)	TH:3,5,8,10,2, 7,4 (7 pts.)	510 +/-18	0.703259 +/- 0.0001	37.7
	TH:5,8,10,2, 4 (5 pts.)	633 +/-31	0.704062 +/- 0.0001	12.1
Mao	TO:7,9,11,5, 6 (5 pts.)	552 +/-41	0.703553 +/- 0.0014	1.3
Bonga	TK:8,10,11,1, 4,6 (6 pts.)	540 +/-10	0.705784 +/- 0.001	7.2

¹ Isochron dates calculated using a ⁸⁷Rb of $1.42 \times 10^{-11} \text{ y}^{-1}$.

4.4.2 Birbir quartz diorite complex

Seven samples from the western quartz diorite sheet (Fig. 2.2) yielded a scatterchron corresponding to a date of 698 ± 67 Ma, an intercept of 0.70375 ± 0.00003 , and $\text{MSWD} = 17.3$. The MSWD is significantly reduced when two data points are omitted from the calculation. A five-point isochron yields a date of 632 ± 8 Ma, an intercept of 0.70368 ± 0.00003 and an $\text{MSWD} = 2.2$ (Fig. 4.6b). The two samples excluded from the calculation have a high $^{87}\text{Sr}/^{86}\text{Sr}$ present day ratio, exhibit relatively high mylonitic strain and contain a higher proportion of granular aggregates of epidote and apatite than the other five samples. Epidote, rich in Ca, preferentially incorporates radiogenic Sr (Steiger and Wasserburg, 1969).

The central quartz diorite sheet yields a seven point scatterchron with a slope of 510 ± 18 Ma, an intercept of 0.70326 ± 0.0001 , and an $\text{MSWD}=37$. When two samples were excluded from the regression, a five point scatterchron yields a slope of 633 ± 31 Ma, an intercept at 0.70406 ± 0.0001 , and a substantially reduced MSWD of 12. The latter date from the central quartz diorite sheet overlaps, within analytical error, that obtained from the western quartz diorite sheet. The dates are much lower than the U-Pb zircon upper intersection age of 828 Ma from the Birbir quartz diorite, considered to be the age of intrusion. The significantly different Rb-Sr date is attributed to subsequent resetting of the Rb-Sr system. The reset isochron dates of ca. 633 Ma are interpreted here as the time when the isotopic system re-equilibrated, perhaps reflecting a major deformation and recrystallization event that affected the Birbir quartz diorite complex

Figure 4.6: Isochron plot for Birbir quartz diorite.
Open circles represent data points omitted from calculation.

during major transcurrent movement along the Birbir Shear Zone.

4.4.3 Goma granodiorite pluton

Data from ten samples yield a scatterchron corresponding to a date of 677 ± 12 Ma, an initial ratio of 0.70358 ± 0.0001 , and an $\text{MSWD}=20.6$. The points do not fit the isochron within analytical errors, thus no simple interpretation is possible. The data points, however, can be used to define two different regression lines. The two groups of samples are not geographically related; they have four common data points: samples T1J, T4J, T5J and T23J, all of which display subhedral-granular texture and broadly can be classified as protomylonites.

The group that yields the older isochron date (Table 4.4), consists of the four common points together with samples T6J, T21J and T24J, that are protomylonitic. The seven data points yield a date of 750 ± 17 Ma with an intercept at 0.70281 ± 0.0001 , and an $\text{MSWD}=3.6$. If sample T21J, which shows significant chloritization of biotite and sericitization of feldspar, is excluded then the remaining six points yield a figure of 759 ± 18 , an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.70277 ± 0.0001 and an $\text{MSWD}=1.3$ (Fig. 4.7a).

Grouping samples T3J and T8J, both of which are more highly deformed than the others, with the four common points however yields a younger isochron of 640 ± 14 Ma and an intercept of 0.70395 ± 0.0001 with an $\text{MSWD}=2.3$ (Fig. 4.7b).

The Rb-Sr isochron dates point to events significantly younger than the $814 \pm 2/-2$ Ma event indicated by the concordant upper intercept U-Pb zircon

Figure 4.7 a,b: Two re-set isochron plots from the Goma pluton. Open circles represent data points omitted from calculation.

age, which almost certainly represents an emplacement age. The two different Rb-Sr dates from the Goma pluton clearly reflect resetting of the isotope system. The two Rb-Sr isochrons may be related to variable degrees of deformation and recrystallization.

The 759 ± 18 Ma isochron date has a low MSWD of 1.3 and is based on data obtained from the protomylonite, those least affected by mylonitization associated with the Birbir Shear Zone. Isotopic re-equilibration is thought to represent an early thermal event within the Birbir Domain. Correlation of the Rb-Sr whole-rock date with a significant thermal event is supported by the zircon age of $783 \pm 19/-14$ Ma from a presumed, syn-metamorphic leucogranite within the Baro Domain.

The second set of data points yields a significantly different slope and intercept, but the slope of the isochron is controlled mainly by the two data points from the mylonitic samples. Although the 640 Ma date may be spurious, it does coincide with the dates recorded by the more highly mylonitized samples from the Birbir quartz diorite (Fig. 4.10).

4.4.4 Late- to post-kinematic plutons

Dates from the discordant Mao and Bonga plutons are in marked contrast to those of the Birbir quartz diorite sheets and the Goma granodiorite pluton discussed above. Data from 5 samples from the Mao pluton (Fig. 4.8) yield a Rb-Sr whole-rock isochron date of 552 ± 41 Ma, an intercept of 0.70355 ± 0.001 , and an MSWD=1.3. The large uncertainty attached to the calculated age is mainly due to a very small spread in $^{87}\text{Rb}/^{86}\text{Sr}$ ratios.

The near-concordant lower intercept U-Pb age of 550 Ma and the Rb-Sr isochron date of 552 Ma from the same pluton are interpreted as the best estimate of the age of intrusion of the Mao pluton. The pluton is later than the major shearing in the BSZ, because it cuts the main mylonitic fabric in the surrounding country rock. Because the pluton is, however, sheared along its eastern margin, the latest movements on the BSZ post-date ca. 550 Ma.

The granite rim of the Bonga pluton yielded a six point scatterchron of 540 ± 10 Ma, an intercept of 0.70578 ± 0.001 and an MSWD=7.2 (Fig. 4.9). A U-Pb age of $571 \pm 11/-3$ Ma was obtained from the quartz monzonite core of the same pluton. An intrusive contact of the granite into the quartz monzonite indicate that the rim is later than the core. Both varieties are undeformed, and straddle the boundary between the Birbir and Baro domains. Thus, the 571 Ma emplacement date clearly puts a minimum age limit on strain history along the border and perhaps on the timing of most transcurrent movement in the Birbir Domain.

Figure 4.8: Isochron plot for the Mao pluton.

Figure 4.9: Isochron plot for the Bonga pluton.

Figure 4.10:

WHOLE ROCK Rb-Sr RESULTS OF REGRESSION ANALYSES

4.4.5 Rb-Sr mineral dates

Of all of the plutons, the Birbir quartz diorite complex is the most affected by the BSZ. The complex is made up of three major sheets, separated by schists and mylonites. One sample was selected from each of the sheets, and biotite separated for Rb-Sr analyses. Biotite and hornblende define a mylonitic foliation in these rocks. Biotite constitutes 8 to 12 % of the mode and occurs as elongate flakes, 0.4 to 3 mm long and 0.1 to 1 mm wide.

Using the whole-rock and the biotite data from the same sample it was possible to calculate three two-point isochrons. Calculated dates are 426 Ma (sample # T5G; sample locality shown in Fig. 2.3) from the western quartz diorite sheet, 543 Ma (T3H) from the central sheet and 680 Ma (T8I) from the eastern sheet. These dates represent the time at which the biotites became closed systems, i.e. marking the time since they dropped below the temperature at which radiogenic Sr stops diffusing and begins to accumulate in the system. Estimated blocking temperatures have been determined at 250-311°C for biotites (Hoffmann and Gilletti, 1970). These dates are consistent with K-Ar and Rb-Sr mineral dates from rocks affected by the Pan-African thermal event of Kennedy and others (see Cahen et al., 1984).

The dates from the three biotite - whole-rock pairs decrease from east to west. This may result either from westward migration of shear strain in the BSZ with time, or progressive cooling of a heat source located to the west, provided by thickened continental crust of the Baro Domain.

4.4.6 Rb and Sr concentrations and Sr isotopic compositions

Much use has been made of the petrogenetic application of Rb-Sr

systematics (De Paolo and Johnson, 1979; Allegre and Othman, 1980). The Rb/Sr ratios for the low-K suite (Bure and Haya plutons) range between 0.01 and 0.04, in contrast to the ratios for the calc-alkaline suite that range between 0.11 and 0.17. The trend in Rb/Sr ratio of Group (1) pre-kinematic rocks is consistent with other field and chemical data suggesting that both low-K and calc-alkaline suites could be members of an evolving consanguineous series. If so, these data indicate that the low-K rocks are older than the calc-alkaline suite. Rb/Sr ratios of the late to post-kinematic, alkali-feldspar granites are generally higher than 0.8 (mean values for Mao = 0.83, Bonga = 2.4). The increase in Rb/Sr ratio also agrees with other chemical distinctions of the late- to post-kinematic plutons of Group (3) that reflect their derivation from much more evolved magmas than those of Group (1) plutons. The low Rb/Sr ratios (<0.2) which characterize the pre- to syn-kinematic plutonic units from the Birbir Domain resulted in high errors attached to whole rock isochron ages (see York, 1969).

$^{87}\text{Sr}/^{86}\text{Sr}$ intercepts from the re-set isochrons are uniformly low: < 0.704 for Group (1) plutons, 0.7036 for the Mao pluton, and 0.7058 for the Bonga pluton. Because the systems are re-set, their intercepts represent maximum values for the initial Sr isotope ratio of the magma. The re-set isochron intercept values therefore are considered to be maximum values for the initial ratios of the original magmas.

All of the calculated initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are low, close to the late Proterozoic mantle values (Powell and Faure, 1972), suggesting that the plutons are derived from a source with low Rb/Sr ratio similar to that of

the mantle or lower crust, or generation from a fairly young crustal segment. The Rb/Sr and initial Sr ratios are sufficiently low to preclude any significant contribution from a long-lived, upper crustal source.

4.5 SUMMARY

The results of this geochronological study help to constrain the ages of the plutons and their subsequent history. No evidence of older basement to the oldest Pan-African rocks was obtained. A number of significant conclusions can be drawn from the isotopic data coupled with the geological and geochemical data provided in previous chapters. These are:

1- U-Pb zircon ages of 828 ± 5 and 814 ± 2 Ma were obtained from Group (1) pre-kinematic plutonic units, indicating two distinct ages of emplacement in what may be part of a continuum of magmatic events in the Birbir Domain. The low initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.7028 and 0.7040) from the same plutonic units denote values very close to that of the late Proterozoic mantle and thus indicate that there was no contribution to the magmas by crustal rocks substantially older than the age of isotopic redistribution. The ages date the earliest known igneous activity in the Birbir Domain.

The low-K suite could not be dated during this study because of unfavourable Rb/Sr ratios and low zircon content.

2- The Baro leucogranite zircons yielded a mixing line on the concordia plot with an upper intersection of $783 \pm 19/-14$ Ma and a lower intersection at $582 \pm 29/-33$ Ma. The semi-concordant relationship with country rocks, presence of pelitic schlieren and overall aluminous chemistry suggest that the Baro

granite resulted from large scale anatexis in the Baro Domain. The upper intersection date is inferred to represent the age of syntectonic intrusion of an anatectic melt. The re-set Rb-Sr isochron corresponding to a date of 759 ± 18 Ma, indicated by the less-deformed samples from the Goma pluton, points to a thermal event that postdates its emplacement but agrees within uncertainty with the time of intrusion of the Baro leucogranite. Emplacement of the Baro leucogranite denotes a high-grade regional metamorphic event that is also reflected in resetting of isotopic systems at ca. 760 Ma in the Birbir Domain. If an anatectic origin for the Baro leucogranite is accepted, then the 783 Ma zircon age represents an approximate time of high-grade metamorphism in the Baro Domain, and the ca. 783-760 Ma interval defines a period of regional metamorphism across the Birbir and Baro domains. The metamorphism is represented by high-grade metamorphic mineral assemblages such as sillimanite, gedrite and K-feldspar, which formed contemporaneously with the development of gneissic foliation in the Baro Domain.

The timing of this metamorphism also correlates with the age of deformation and epizonal metamorphism at 761 ± 22 Ma of arc-related rocks from Abu Hamed, in the Southern Red Sea Hills of the Sudan (Ries et al., 1985), at 750 ± 29 Ma from the Bayuda desert in central Sudan (McInhold, 1979), and in the Western Arabian Shield (Jackson and Ramsay, 1980). It also correlates with a metamorphic event loosely defined for Mozambiquian rocks of Kenya in the range of 750-850 Ma (Shibata, 1975; Suwa, 1979).

The $582 \pm 29/-33$ Ma lower intersection is interpreted as the age of the latest high temperature metamorphic overgrowth of zircon. This event probably corresponds to the growth of large garnet porphyroblasts and

randomly oriented sillimanite (fibrolite) in unit (gsg). The ca. 580 Ma date correlates with the age of amphibolite facies metamorphism in the nearby Mozambique Belt of NW Kenya (Ries et al., unpublished).

3- The Rb-Sr isotope systems of the Group (1) pre-kinematic intrusive units show variable degrees of isotopic re-equilibration which can be attributed to heterogeneous deformation. The regressed values of different data sets from the Birbir quartz diorite and Goma pluton of the Birbir Domain cluster of about 635 ± 8 Ma (Fig. 4.10), providing evidence for significant isotopic homogenization during a late Proterozoic event. This event is probably associated with major transcurrent movement on the Birbir Shear Zone, age of mylonitization.

The ca. 635 Ma time of isotopic homogenization in rocks of the Birbir Domain and the metamorphic age at ca. 580 Ma inferred from rocks of the Baro Domain may represent parts of a protracted metamorphic event that affected rocks of both domains.

4- U-Pb data from the late-tectonic Mao granite, that cuts the main mylonitic fabric of the Birbir Shear Zone but is itself marginally deformed, also define a mixing line, in this case between pre-intrusion zircon cores and predominant igneous overgrowths. The lower intersection dates the time of intrusion at $541 \pm 10/-16$ Ma, and the poorly-constrained upper intersection indicates zircon of ca. 770 Ma, that may have been generated at the same time as the Baro leucogranites. The Bonga granite, intruded across the boundary zone of the Birbir and Baro domains, contains a concordant zircon population whose age is $571 \pm 11/-3$ Ma. Together the dates mark the

termination of tectonic and intrusive activity at about 550 Ma.

5- Initial Sr isotope ratios for groups (1) and (3) plutons of the Birbir Domain are estimated at <0.704 and <0.706 respectively, implying no significant contribution of older crust to the generation of these rocks. The intercepts, obtained from re-set systems, are maximum values, thus they point to a mantle-like source.

CHAPTER 5

DISCUSSION AND INTERPRETATION

5.1 GENERAL STATEMENT

Mapping at the scale of 1:50,000, supported by chemical studies, has indicated distinct periods of plutonism and tectonothermal activity within the Birbir and Baro Domains. Three generations of plutonic rocks are recognized. Group (1), the earliest generation, consists of pre- to syn-kinematic, mafic and intermediate intrusive complexes. A second group is made up of marginally peraluminous, probably anatectic granites and a third, younger, group comprises alkali-feldspar granites.

Because the area has been affected by regional metamorphism and shearing, it is not clear to what extent the rocks have remained as closed chemical systems. Systematic variation, however, in major and trace element distribution, including the REE, suggests restricted elemental mobility. The absence of significant retrograde metamorphic features, typical of many rocks affected by shearing, implies either a limited fluid circulation, an anhydrous environment, or perhaps cessation of shearing before drop in temperature.

The trace element variations and REE patterns of samples from Group (1) reflect arc-type magmatism, while Groups (2) and (3) reflect intraplate activity. Chemical data from Group (1) define two distinct suites: a low-K suite of oceanic affinity and a calc-alkaline suite similar to that of an Andean-type arc. A steeply-dipping ductile transcurrent fault system, the Birbir shear zone (BSZ), transects the two suites of Group (1) rocks. The

calc-alkaline suite possesses low $^{87}\text{Sr}/^{86}\text{Sr}$ initial ratios, less than 0.7037, implying little or no contribution from older continental crust. The overall petrologic and chemical features suggest that the Group (1) rocks formed in a subduction-related tectonic setting.

5.2 REGIONAL LITHOTECTONIC ASSEMBLAGES

Although geological investigations of the Precambrian rocks in the Afro-Arabian region began in the first half of this century, an understanding of the stratigraphic, structural and tectonic features has remained elusive. It is possible however to recognize three types of regional lithostructural domain (Figs. 1.2 and 5.1) that are similar to those in western Ethiopia and the Geotraverse area. Recent geological investigations of the Precambrian rocks of NE Africa and Arabia are represented by the contributions to the International Geological Correlation Programme (IGCP, Project 164). Of particular importance are the contributions of Al Shanti and Mitchell (1976), Greenwood et al. (1976), Gass (1977, 1979), Kroner (1977), Shackleton (1977), Kazmin et al. (1979), Engel et al. (1980), Bokhari and Kramers (1981), de Wit and Chewaka (1981), Duyverman et al. (1982), Al Shanti and Gass (1983), Ries et al. (1983), Vail (1983, 1985), Harris et al. (1984), Kroner (1985), Jackson (1986) and Shackleton (1986). Their ideas are summarized in the following sections in an attempt to relate the domains in the study area to those found elsewhere in the region.

5.2.1 Volcano-sedimentary terranes

The volcano-sedimentary terranes contain calc-alkaline basalt, andesite

Figure 5.1: Palinspastic sketch map of the region showing main tectonic terranes. (Modified after Vail, 1985).

and rhyolite flows, tuffs and pyroclastics. Voluminous sedimentary rocks are common and include wackes, thin layers of marble and quartzite, pebbly mudstones and conglomerates and rare ironstones. These units are characteristically at greenschist to lower amphibolite facies grade of metamorphism and are extensively intruded by plutons of mafic and intermediate to granitic composition.

At least five such belts can be identified in NE Africa and Arabia (Fig. 5.1); each seems to be geologically discrete and each appears to have evolved independently. It is now generally accepted that most of these terranes developed as island arc-ocean basin complexes < 1 Ga ago (Engel et al., 1980; Fleck et al., 1980; Bokhari and Kramers, 1981; Calvez and Kemp, 1982; Duyverman et al., 1982; Roobol et al., 1983; Stacey and Stoesser, 1984). To the north, these arc-ocean complexes appear to terminate in the gneissic terrane of northeastern Egypt and southern Sinai (Stern, 1979). These can be traced into Arabia where their origin is ascribed to a series of events called the "Hijaz cycle" (Ramsay, 1979). To the south they branch into Ethiopia and the Sudan, where they narrow and become truncated by a major fault system, the Surma mylonite zone (Fig. 1.3).

5.2.2 Ophiolitic rocks

Rocks that may be members of an ophiolite suite ("suture zones" in Fig. 5.1) are widespread throughout NE Africa, and have been described in Ethiopia (Kazmin, 1976, 1978a, 1979; de Wit and Chewaka, 1981), Egypt (Shackleton, 1977; Shackleton et al., 1980), Sudan (Fitches et al., 1983) and Kenya (Vearncombe, 1981). Narrow discontinuous belts of ultramafic

rocks, gabbros, serpentinites, rare sheeted dykes, pillowed tholeiitic basalts and associated sedimentary rocks such as turbidites, chert, marble and volcanoclastic rocks have been described. Because of their linear nature, the ophiolitic belts are considered to represent boundaries of convergence or suture zones developed between continental and island arc terranes (Gass, 1979; de Wit and Chewaka, 1981; Vail, 1983; Stoesser and Camp, 1985; Kroner, 1985).

The absence of well-defined accretionary prisms, ophiolitic melanges and high-pressure metamorphic assemblages led Kroner (op.cit.) to suggest that either accretion did not occur along margins with deep ocean trenches, or extensive overthrusting took place so that the forearc segments were overridden and concealed. Fyfe (1979) commented on the lack, in the late Proterozoic rocks of NE Africa and Arabia, of low-temperature - high pressure metamorphic facies commonly associated with subduction environments, and concluded that this lack of blueschists reflects plate movements of only a few hundred kilometres. An ophiolite emplacement mechanism related to subduction processes seems nonetheless to be widely accepted by most investigators (see Vail, 1985; Jackson, 1986a).

5.2.3 Gneissic terranes

These consist of heterogeneous ortho- and paragneisses and migmatites of generally upper amphibolite to granulite grade of metamorphism. These terranes are thought to be older than the volcano-sedimentary terranes, but few rocks older than late Proterozoic have been conclusively identified. Although the data base is limited, there are a few dates from Sudan and

Egypt that suggest Archean to a mid-Proterozoic foreland/basement underlying the western side of the Pan-African rocks in NE Africa (see section 4.1).

On a regional scale, some of the gneisses occur as isolated inliers within wide belts of the volcano-sedimentary terranes. These are now mostly interpreted as remnants of pre- Pan-African microcontinents, representing exotic terranes accreted by oblique collision and strike-slip translation (Kroner, 1985; Stacey and Agar, 1985).

5.3 RELATIONSHIP OF THE BIRBIR DOMAIN TO THE PAN-AFRICAN ARC-TYPE ROCKS

Rocks of the Birbir Domain are traceable northwards into the Pan-African Asir, southern Red Sea Hills and Butana terranes of north-east Africa and Arabia (Figs. 1.2 and 5.1). The stratified rocks of these terranes are like those of the Birbir Domain are now generally in the lower amphibolite facies of metamorphism, and they have been extensively intruded by pre- and syn-kinematic, low-K and widespread calc-alkaline rocks of batholithic dimensions and by late alkali-feldspar granites. The voluminous calc-alkaline magmatism is ascribed to subduction processes giving rise to a number of magmatic arcs, that have been accreted to one another and to oceanic and continental fragments. A magmatic arc affinity is also suggested by trace element discriminant plots for the pre- and syn-kinematic rocks of the Birbir Domain. $^{87}\text{Sr}/^{86}\text{Sr}$ intercepts for the Birbir calc-alkaline suite indicate that they are unlikely to have been derived from much older crustal sources. On the eastern border of the Birbir Domain, to the north of the study area, the linear Tullu Dimtu mafic/ultramafic belt, that is thought to

be composed of dismembered ophiolitic rocks, can be traced into the ophiolitic rocks of the Baraka lineament of northeast Africa. If the basement were merely thinned underneath the Birbir Domain, there would not be widespread calc-alkaline rocks nor would oceanic rocks be associated. The evidence thus follows a cycle of ocean opening and closure, rather than an ensialic origin, for the Birbir Domain, similar to the model for the Pan-African terranes to the north.

South of the study area, rocks of the Akobo Domain are continuous with those of the Birbir Domain and have a wedge-shaped configuration; they terminate at a major tectonic front, the Surma mylonite zone (Davidson, 1983). The narrowness of the Birbir Domain in the study area compared to its northerly extension, where it reaches a strike - normal width of over 200 km, was probably due to a reduction in width of a proto-ocean basin at the latitude of the study area, and also possibly to increased shortening at this latitude. The boundary between the Birbir and the gneissic terrane to the west probably represents a mylonitic cryptic suture. To the north, where the wedge-shaped Pan-African rocks widen, ocean opening may have been considerable and plate convergence with subduction of oceanic crust could have continued for an extended period.

In conclusion, the Birbir Domain of western Ethiopia exhibits similarities in field relations and lithological, elemental and isotopic characteristics to the magmatic arc systems of Pan-African rocks to the north in northeast Africa and Arabia.

5.4 RELATIONSHIP OF THE BARO AND GEBA DOMAINS TO THE MOZAMBIQUE BELT OF EAST AFRICA

Rocks of the Mozambique Belt can be traced from NW Kenya through the Hamar Domain of the Omo Project area (Davidson, 1983) into the Geba Domain of the Geotraverse area. Although the Baro Domain can not be traced directly into the Mozambique Belt, its predominantly sialic lithology, metamorphic grade indicating mid-crustal level (medium-pressure, high-temperature metamorphism), and apparently polyphase deformation suggest that it may be the northward continuation of the Mozambique Belt in NW Kenya.

Field relations and chemical data from Group (2) rocks in the Baro Domain suggest anatexis of a sedimentary source. These rocks are interpreted as having been emplaced during folding and migmatization at ca. 780 Ma, initiated by thickening of a continental margin following the emplacement of the calc-alkaline suite. The Baro leucogranite yielded a U-Pb zircon lower intersection date of about 580 Ma, that clearly marks a later major thermal event contemporaneous with amphibolite facies metamorphism in NW Kenya. According to Gabert (1984), the Mozambiquian "event" should be restricted to a thermal event at about 550 ± 100 Ma; this period includes the time of the ca. 580 Ma high-grade metamorphic event in the Baro Domain. Thus the latest high-grade amphibolite facies metamorphism, without major contemporaneous deformation, in the Baro Domain of western Ethiopia can be correlated with the thermal event that affected rocks of the known Mozambique belt. This coincidence adds strength to the correlation of Baro Domain rocks with the Mozambique belt of NW Kenya.

5.5 RELATIONSHIP OF THE MOZAMBIQUE BELT ROCKS TO THE PAN-AFRICAN ROCKS IN WESTERN ETHIOPIA

There is as yet no geochronological evidence from the study area that supports the view that rocks of the Baro and Geba domains are older than those of the Birbir Domain. Geochronological evidence from areas adjacent to the Mozambiquian rocks, not influenced by the Pan-African event, yield ages as old as 2900 and 2600 Ma for the Dodoman System in the Tanzanian Craton (Bell and Dodson, 1981) and about 1900 Ma for the Usagaran - Ubendian fold belt (Gabert, 1984). Detrital zircons from granulites at Sabaloka on the Nile north of Khartoum yielded ages that range from ca. 2650 to 1000 Ma (Kroner et al., 1987) that mark the involvement of Archean to late Proterozoic continental crust in the evolution of the high-grade rocks to the west.

Nd isotope studies have identified important differences in the character of the various crustal provinces across northeast and east Africa. Peraluminous granites (Marich granite) from an area of high metamorphic grade in NW Kenya, similar to Group (2) plutonic units, have negative $\epsilon_{Nd}(T)$ values and suggest crustal precursors older than 1650 Ma (Harris et al., 1984). Nearby lower-grade Pan-African rocks have positive $\epsilon_{Nd}(T)$ values, suggesting a mantle-type source and little or no contamination by crustal rocks, and younger model ages than the rocks from the high grade metamorphic terrane. This interpretation supports the view that the rocks of the Birbir Domain, considered to be Pan-African, are younger than those of the Baro Domain in the western Ethiopian Shield.

The ages obtained in this study however demonstrate that part of the plutonic and metamorphic evolution of the Baro Domain is synchronous with that of the Birbir Domain. At least two distinct metamorphic events with ages between 780-760 and 635-580 Ma have affected both domains. Similar age relations were found in Pan-African rocks of the Bayuda desert in the northeast African shield (Fig. 5.1; Pohl, 1981).

Although field relationships are not conclusive, high-grade rocks of the Baro Domain appear to be overlain by wedges of metaclastic sedimentary rocks (unit qms) of continental derivation. These wedges are thought to have been formed at a plate margin, on the basis of the chemical characteristics of the associated plutonic rocks and the presence of a ductile transcurrent fault system, the Birbir Shear Zone (BSZ), that marks a tectonic contact between the contrasting Baro and Birbir domains. The Baro Domain is thus suggested to be a zone of tectonic reworking in the foreland to the Mozambique Belt, during the Pan-African events. Kroner (1977) similarly regarded the gneissic terranes to the north, in Egypt and Sudan as indicating transformation of a passive continental margin into a tectonically-active belt along which ophiolites and magmatic arc rocks were thrust over each other and/or onto the gneisses.

The western boundary of the Birbir Domain contains plutonic units chemically similar to those of an active continental margin (Andean-type). The Baro Domain contains leucogranites whose petrology and chemistry implies a sedimentary source. The available evidence for the origin of these plutonic rocks suggests anatexis in a tectonically active continental margin on the eastern side of the Baro Domain. The eastern boundary of the Birbir

Domain is also a tectonic one, marked by truncation of a major antiform in rocks of the Geba Domain.

In the western marginal zone of the Mozambique Belt of NW Kenya, Vearncombe (1983) reported ophiolitic rocks in the Sekerr area that may correlate with those in western Ethiopia. These features led Shackleton (1986) to conclude that the Mozambique Belt, at least as far south as NW Kenya, evolved as a result of continent-continent collision.

Penetrative deformation accompanied by migmatization and upper almandine amphibolite facies metamorphism may have obscured earlier structures and re-set isotopic systems within these high-grade metamorphic rocks. In order to better understand their history, precise U-Pb dating of minerals such as zircon and monazite will be necessary. Although it can not yet be proven, the Baro and Geba domains appear to be older than the crust generated during the Pan-African event.

5.6 AGE AND CORRELATION OF PLUTONIC AND METAMORPHIC EVENTS

The age spectrum of plutonic units from the Gore - Gambella area is broadly similar to that established for Pan-African rocks to the north along strike in Sudan, Egypt and Saudi Arabia, and in the Mozambique Belt of East Africa (Table 5.1).

Calc-alkaline plutonic rocks of the Birbir Domain are dated at ca. 830-815 Ma and are probably related to subduction. The earlier presence of oceanic crust is supported by mafic/ultramafic rocks of probable ophiolitic affinity to the north (Fig. 5.1), in the Tullu Dimtu - Dallati - Yubdo Belt. The period ca. 760 Ma is considered to be the age of earliest regional

Table 5.1: Tentative age correlation of plutonic and metamorphic rocks from western Ethiopia with Pan-African rocks in the Sudan and Saudi Arabia, and known rocks of the Mozambique Belt from NW Kenya.

TECTONO-THERMAL EVENT	WESTERN ETHIOPIA ¹ Birbir	SUDAN ² Baro	SAUDI ARABIA ³	NW KENYA ⁴
Uplift, post-tectonic plutonism	550+/-10 571 +/-5	549 +/-12	ca. 570-585	None
Metamorphism		582+/-30		ca. 584 +/-25
Development of major transcurrent fault system	635+/-8			
Closure of ocean basin, regional metamorphism	760 +/-18	761 +/-22	ca. 760	
Emplacement of syn-tectonic gneiss domes and granitoids		783+/-15	ca. 800	ca. 593 +/-50
Subduction -related magmatism and sedimentation	814 +/-2 to 828 +/-5	800 +/-83 (Abu Hamed, (sequence "B" Sudan)	816 +/-4 (sequence "B" of Hijaz cycle)	ca. 663 +/-49
Formation of oceanic crust	(?) Tullu Dimtu		ca. 890 (Jabal Al Wask)	

- Sources:
- (1) Present study
 - (2) Ries et al. (1985)
 - (3) Greenwood et al. (1976); Fullagar and Greenberg (1978); Jackson and Ramsay (1980a)
 - (4) Ries et al. (Unpublished)

metamorphism in Pan-African rocks of western Ethiopia. This interval seems to represent the time of juxtaposition of an intra-oceanic island arc, represented by the low-K suite, with the calc-alkaline suite of continental margin-type along the BSZ. High-grade metamorphism in the Baro Domain at ca. 780 Ma is also attributed to this event.

The timing of the regional metamorphism across the Birbir and Baro domains is similar in age to low-grade metamorphic activity associated with arc-related rocks at 761 ± 22 Ma from Abu Hamed, in the central part of northern Sudan (Ries et al., 1985) and 750 ± 29 Ma from the Bayuda desert (Meinhold, 1979). It also correlates with a thermal event loosely defined for Mozambiquian rocks of Kenya in the range of ca. 750-850 Ma (Shibata, 1975; Suwa, 1979).

A later metamorphic event in the Birbir Domain may also be defined, at ca. 635 Ma, as the time of an isotopic readjustment probably related to major transcurrent movement along the Birbir Shear Zone. Metamorphic overgrowth of zircons in the Baro leucogranite at 582 Ma is considered to represent the last metamorphic event recorded in Mozambiquian rocks in western Ethiopia. Thus a second distinct period of metamorphism, including one or more events in adjacent, contrasting metamorphic domains, can be constrained to ca. 635-580 Ma. In NW Kenya, the 584 Ma event from rocks of the Mozambique Belt also marks a second regional metamorphism younger than that of the Pan-African rocks in NE Africa (~ 760 Ma).

5.7 TECTONIC MODEL

The abrupt change in metamorphic grade from the Birbir into both the

Baro and Geba domains, the tectonic contacts separating them and the chemical distinction of the plutonic units and all suggest tectonic juxtaposition of newly generated crustal material with pre- Pan-African continental crust. The model of shield microplate accretion that has been applied to the Afro-Arabian region (Gass, 1981; Vail, 1985) thus seems to fit the observations from the western Ethiopian Shield. A mylonitic cryptic suture is suggested to occupy the study area, containing arc material represented by the low-K suite, welded between older continental masses: the Baro Domain to the west and Geba Domain to the east. Both the Baro and Geba domains are interpreted to represent continental crust contiguous with the Mozambique Belt to the south. The Baro Domain is here interpreted as part of the basement system of the African foreland that has been affected by Pan-African remobilization.

The proposed tectonic evolution of the western Ethiopian Shield involved the following steps: (a) breakup of a continental mass and subsequent development of an ocean basin, (b) initiation of a short-lived marginal basin and/or intra-oceanic subduction zone producing magmatic arcs both within the continental margins and on the oceanic crust, and (c) subsequent closure. There is no evidence from this study concerning the nature and timing of continental rift and ocean development. The remainder of the evolutionary scheme can be broadly divided into three periods: pre-accretionary, accretionary and cratonization as shown in Figure 5.2. The pre-accretionary period started before ca. 830 Ma (Fig. 5.2A; stage 1) and involved formation of an island arc and the subsequent destruction, through subduction, of oceanic lithosphere. As subduction continued this arc was accreted to the

(A) Pre-accretionary period

1. Subduction: pre (?)830 Ma

(B) Accretionary period (830–635 Ma)

2. Subduction (830–760 Ma)

Continental crust

Volcanic and sedimentary rocks

Oceanic crust

Granitoid rocks

Figure 5.2: Cartoon to illustrate the tectonic evolution of western Ethiopia during late Proterozoic.

3. Development of transcurrent fault system
(760-635 Ma)

(C) Cratonization period (635-540 Ma)

- 4. Back-arc spreading (?)
- 5. Uplift and erosion

Figure 5.2 (continued);

continental margin (Fig. 5.2B; stage 2). A westerly dipping subduction zone is shown in Figure 5.2, based on a generalized westward tectonic transport direction including fold vergence suggested by de Wit and Chewaka (1981) and Shackleton (1986) for rocks along strike immediately north and south of the study area respectively. This arc-accretion model would also be consistent with an easterly-dipping subduction zone, where the low-K suite was developed in a marginal basin that lay directly to the east of the subducting slab.

The accretionary period started about ca. 830 Ma with emplacement of the calc-alkaline suite and continued to about ca. 635 Ma. The steep mylonitic foliation and subhorizontal extension lineation of the BSZ together with syntectonic igneous emplacement suggest that final accretion was achieved through transcurrent movement (Fig. 5.2B; stage 3). The BSZ in this model represents a cryptic suture along which rocks of the low-K suite were tectonically accreted to the calc-alkaline suite and the pre- Pan-African gneissic complex represented by the Baro Domain.

Ultimately there was a collision between a continental fragment represented by the Geba Domain with the proto-African continental margin (Baro Domain) and arc-type rocks of the Birbir Domain. Upon collision, the subducting lithosphere was stripped from the oceanic crust and obducted on to the margin of the Geba Domain (Fig. 5.2B stage 3). Kazmin et al. (1979) and de Wit and Chewaka's (1981) observations of extensive serpentinization along west-dipping fault planes and sheared margins on the Tullu Dimtu rocks support the tectonic emplacement of the ophiolite and are consistent with this obduction model. The absence of ophiolitic rocks in the study area, as well as their scarcity to the south, is probably related to the depth of plate

underthrusting (Dewey and Bird, 1972).

The cratonization period (Fig. 5.2C) began after 635 Ma with igneous emplacement and regional metamorphism across the sutured boundary. The mafic intrusions of unit (gb) in the Birbir Domain, directly west of rocks of the calc-alkaline suite, have potentially important implications in this respect. These mildly deformed rocks suggest late-kinematic mafic plutonism. Unfortunately these rocks have not yet been dated nor are their chemical characteristics known. Similar occurrences have been reported north of the study area along strike for over 200 kilometres (Amenti Abraham, 1984, pers.comm.) associated with rocks similar to those of Group (1). In the absence of more rigorous evidence, they may be tentatively viewed as having formed in a back-arc environment (stage 4). According to Bruhne et al. (1978), back-arc rifting can be explained in terms of the physical separation of an Andean-type arc from a continent. This is followed by diapiric rise of partly melted mantle material, initiated in response to subduction, together with release of horizontal compressive stresses across the destructive plate boundary causing back-arc extension. Cratonization (Fig. 5.2C, stage 5) included emplacement of late- and post-tectonic granites together with minor reorganization of the accreted terrane by horizontal and vertical movements, uplift and erosion following collision.

The study has served to outline part of the tectonic history of the Ethiopian shield and to show that it is consistent with the Pan-African evolution determined elsewhere in the Afro-Arabian shield. The model is not one of ensialic orogeny, as earlier proposed, but rather of plate convergence with suturing of oceanic and continental crust.

CHAPTER 6

PRECAMBRIAN STRATIGRAPHY AND TECTONICS OF ETHIOPIA

6.1 GENERAL STATEMENT

Interpretation of the stratigraphy and tectonic setting of the Precambrian rocks of Ethiopia has been uncertain, mainly because of the lack of systematic dating and scattered nature of the exposures. Kazmin (1971, 1972, 1975) assigned the Precambrian rocks to Lower, Middle and Upper complexes. These divisions were based on the supposition of basement - cover relationships and stratigraphic succession from one complex to the next. Kazmin's scheme, which is in general use in Ethiopia, has come under increasing criticism by many workers in the region who have recognized temporal and spatial overlap and transition among his divisions (Warden and Horkel, 1984; Kozyrev et al., 1985).

More critically, the present mapping has revealed that the relationship between some rocks formerly described as basement and cover is rather that of tectonic assembly. Geochemical and isotopic data for the so-called "cover" rocks also imply collision-related calc-alkaline magmatic suites, the development of which may have involved subduction of oceanic lithosphere. The discussion in Chapter 5 calls into question the view that rocks of the Birbir Domain were ensialic mobile belts that formed within older continents.

The recognition of these relationships among major lithostructural domains, regarded in earlier nomenclature as stratigraphic units, requires a critical review of the whole of the Precambrian stratigraphy of Ethiopia.

In this chapter a new division of the Precambrian rocks of Ethiopia is attempted.

6.2 INTRODUCTION

The greater part of the Ethiopian plateau is underlain by Precambrian rocks succeeded by Mesozoic sedimentary rocks and Tertiary volcanics. The Precambrian rocks are exposed in four separate areas. The northern area includes parts of Eritrea, Gondar and Tigray administrative regions; the western area, parts of Wollega, Illubabor, Keffa, and Gojjam administrative regions; the southern area, parts of Sidamo and Gemu-Gofa administrative regions; and the eastern area is in Hararghe administrative region (Fig. 1.4). The stratigraphic and tectonic framework has been reviewed by Gilboy (1970), Chater (1971), Kazmin (1971, 1972 b and c, 1975b, 1976), Beyth (1971), Omo River Project Team (1973, 1976), Kazmin et al. (1978, 1979), de Wit and Chewaka (1981), Davidson (1983), Warden and Horkel (1984), Kozyrev et al. (1985) and in numerous unpublished reports by the geologists of the Ethiopian Institute of Geological Surveys.

Gilboy and Chater divided the Precambrian rocks of southern Ethiopia into three units, referred to as the Lower, Middle and Upper Groups. The transition between the Lower and Middle Groups is gradual, and no major stratigraphic break is present (Chater, 1971). Gilboy described four phases of deformation, and two major regional metamorphic events at ca. 630-680 Ma and ca. 500-550 Ma. Chater recognized four tectonic events and three periods of metamorphic mineral growth in the Lower and Middle Groups. Sedimentary structures, preserved in rocks of the Upper Group indicate that these

rocks are the right way up stratigraphically and structurally overlie the Middle Group. Chater suggested that tectonism in the region ceased at least ca. 630 Ma ago, and both Gilboy and Chater concluded that the Precambrian rocks of southern Ethiopia are typical of rocks of the Mozambique Belt elsewhere.

In view of the fact that traditionally the lower-grade rocks have been regarded as deposited unconformably on older gneissic basement, a regional stratigraphy erected on that basis is in common use. Kazmin (1971, 1972) adopted Gilboy and Chater's three-fold division of rocks in southern Ethiopia and extended it to all of the Precambrian rocks of Ethiopia. These divisions are reviewed and assessed below.

6.2.1 Lower Complex

The Lower Complex comprises gneisses and migmatites, regarded as reactivated Archean basement and believed to have a widespread distribution in Sudan, Egypt and possibly Saudi Arabia (Almond, 1969; Kazmin, 1975). They were also considered similar to those found in Kenya, Uganda and Tanzania (Hepworth, 1972, 1979)). The basis for this view was the occurrence of east-west trending relict structures in mafic granulites from southern Ethiopia (Kazmin, 1972), and in the gneisses of upper amphibolite grade in the Geba Domain of western Ethiopia (Kazmin et al. 1979; Mengesha and Seife, unpublished manuscript). The Complex is represented by the Awata, Yavello, Alge and Konso gneisses in southern Ethiopia and the high-grade metamorphic terranes in western and southwestern Ethiopia.

6.2.2 Middle Complex

Rocks of this complex have been recognized in southern Ethiopia as the Wadera Group and in eastern Ethiopia as the Boye and Soka Groups (Kazmin, 1975b; Kazmin et al., 1978). Kazmin considered the Wadera Group as unconformably overlying the Lower Complex. Chater (1971) and Kozyrev et al. (1985) indicated, however, the absence of such an unconformity between the Yavello gneiss (Kazmin's Lower Complex) and the Wadera Group. According to Kozyrev et al. (1985), the Awata and Yavello gneisses both belong to the Middle Complex.

6.2.3 Upper Complex

The rocks assigned to this complex are the Birbir Group, Akobo Domain and ophiolitic rocks of the Tullu Dimtu - Dallati - Yubdo belt in western Ethiopia (Kazmin et al., 1978); the Awata and Mormora groups of southern Ethiopia (Kozyrev et al., 1985); and the Tsaliect and Tambien Groups in northern Ethiopia (Beyth, 1971). The Birbir Domain of the Geotraverse area forms part of this complex. In southern Ethiopia the interface between the Middle and Upper Complex is described as a tectonic contact (Kozyrev et al., 1985).

6.2.4 Other Divisions: Omo River Project

A large area in southwestern Ethiopia, as far north as the southern edge of the Gore - Gambella area, was described in geological maps and reports prepared by the Omo River Project (Omo River Project Team, 1973, 1976; Davidson, 1983). Davidson (1983) recognized three domains on the basis of marked lithological, metamorphic and structural contrasts; these are

referred to as the Hamar, Akobo and Surma domains. Two major groups of rocks were identified, one probably much older than the other. Rocks of the Hamar Domain, part of the older group, have been metamorphosed twice and were intruded by plutonic rocks both before and after the first metamorphic event. Younger sedimentary and volcanic rocks of the Akobo Domain were deposited west of the Hamar domain and subsequently deformed and thrust toward the west, prior to the youngest period of metamorphic activity. During the second metamorphic event, gneissic rocks of the Surma Domain were juxtaposed against rocks of both the Hamar Domain and the relatively low-grade Akobo Domain along a major shear belt, the Surma mylonite zone.

6.3 LITHOSTRUCTURAL UNITS

Lithological, metamorphic and structural discontinuities in the Precambrian rocks of Ethiopia indicate that these rocks represent different lithostructural units. A new lithostructural subdivision and nomenclature are proposed here, based on the present mapping and a review of previously published information (Kazmin et al., 1979; de Wit and Chewaka, 1981; Warden, 1982; Davidson, 1983; Warden and Horkel, 1984; Kozyrev et al. 1985). A geological sketch map of the distribution of Precambrian rocks of Ethiopia, with the proposed names, is given in Figure 6.1. A generalized stratigraphic division, applicable only to western Ethiopia, and based mainly on the Gore - Gambella Geotraverse, is given in Table 6.1. For comparative purposes the earlier divisions are indicated in the text where appropriate, together with the newly proposed nomenclature.

Figure 6.1: Sketch map of the Precambrian rocks of Ethiopia showing the distribution of the proposed lithostructural units.

3

Table 6.1: IDEALIZED LITHOSTRATIGRAPHIC SEQUENCE IN WESTERN ETHIOPIA

Alluvium		QUATERNARY

1- Central-type volcanoes (mainly basalt, trachyte & phonolite)		TERTIARY
2- Fissural basalts		
----- UNCONFORMITY -----		
Birbir magmatic zone		
alkali-feldspar granite plutons, aplites & pegmatites (late- to post-kinematic)		550-570 Ma
mafic plutons, dykes and sills (late-kinematic gabbro & hornblendite)		
-----intrusive contact-----		
calc-alkaline intrusive rocks (pre-kinematic; predominantly quartz diorite & granodiorite)		810-830 Ma
calc-alkaline volcanics & volcanogenic sediments (mainly andesite and minor silicic flow, turbidite & volcaniclastic rocks)		LATE PROTEROZOIC
-----intrusive contact--?-----		
low-K intrusive rocks (predominantly gabbro & tonalite)		
Tullu Dimtu ophiolitic rocks		
ultramafic rock, gabbro, pillowed basalt turbiditic sediments, chert & marble		
===== TECTONIC CONTACT =====		
leucogranite sheets		LATE PROTEROZOIC
clastic sediments (quartz-muscovite schist, quartzite and conglomerate layers)		OR OLDER (?)
paragneisses, schists & amphibolite (sillimanite-garnet schist and gneiss, and subordinate calc-silicate rocks)		
~~~~~ probable unconformity ~ ~ ? ~ ~ ~ ~ ~		
Western gneiss zone	Central gneiss zone	
quartzofeldspathic gneiss with amphibolite bands and minor magnetite bearing quartzite layers	heterogeneous ortho- & para- gneiss, migmatite	
layered and flaggy hornblende- biotite gneiss, migmatite		

6.3.1 Western gneiss zone (WGZ)

In parts of western and southwestern Ethiopia a high-grade gneissic complex, that contains the Baro Domain of the Geotraverse area and is correlated with rocks to the south in the Surma Domain of Davidson (1983), is defined as the Western gneiss zone. The Baro Domain consists of biotite and hornblende gneisses, generally well-layered, with subordinate amphibolite and quartzite. It is similar to the Grey Gneiss Group of the Bayuda desert in the north-central Sudan (El-Nadi, 1987). A strong lithological similarity exists between the Baro and the Surma Domain to the south. The latter is separated by a major NW- trending shear zone from lower-grade rocks of the Akobo Domain to the east, in the same way that the Baro Domain is separated from the Birbir Domain rocks by the N-S trending Birbir Shear Zone. Mylonitic foliation is developed along the eastern contacts of both the Surma and Baro domains. Across both transcurrent fault systems, there is an abrupt change in metamorphic grade from upper amphibolite facies to epidote amphibolite facies grade of metamorphism (Davidson, 1983; Chapter 2).

The Baro Domain is characterized by widespread leucogranite sheets, such as the Baro leucogranite, that are strongly deformed into upright open folds with north - south axial traces and sub-horizontal axes. The sheets are interpreted as anatectic granites (see Chapter 2), although some may represent sills intruded parallel to gneissic layering. Davidson (1983) interpreted the granite sheets as highly-flattened slices, each associated with a folded, east-dipping thrust plane.

On the eastern edge of the Baro Domain, a wedge of high-grade

metasedimentary schists and gneisses occurs. These rocks include pelites, calc-silicates and minor psammites, best exposed near Bonga. To the north, near Dembi Dollo, these rocks are layered, garnet-bearing, biotite gneisses with minor interlayers of quartzite. These rocks may represent shelf sediments, deposited along a continental margin.

6.3.2 Central gneiss zone (CGZ)

Rocks of the Geba Domain and the western half of the Hamar Domain, as well as the Didessa gneisses in west - central Ethiopia, make up the Central gneiss zone (Fig. 6.1). The CGZ occupies the area northeast of Lake Turkana and east of the Omo River, where it comprises the Hamar Domain of Davidson (1983) and its northward continuation along the Geba and Didessa valleys. The Hamar Domain consists of gneisses and granulites of both igneous and sedimentary origin, partly migmatized and cut by a suite of younger plutonic rocks (Davidson, 1983). Kazmin et al. (1979) described the gneisses of the Geba and Didessa valleys as biotite and amphibole - biotite banded gneisses, locally garnetiferous, that grade into homogeneous grano- dioritic and granitic rocks.

In the Geba valley, transposition of east-west, map-scale, recumbent isoclinal folds into north-south trending isoclinal folds has been reported (Kazmin et al., 1979; Mengesha and Seife, unpublished manuscript). The east-west structures are similar to those found in the Archean East African cratons (Spooner et al. 1970; Leggo, 1971) that form the basement in Kenya, Tanzania, NE Uganda and SE Sudan (Kazmin, 1976; Vail, 1976; Almond, 1980). Kazmin also noted the abundance of large, late-tectonic granite bodies in

the northern part of CGZ and the extensive migmatization in the southern part of the CGZ. He attributed the distribution of these rocks to uplift and erosion of progressively more deeply buried and metamorphosed rocks southwards.

6.3.3 Southern gneiss zone (SGZ)

This zone contains the Alghe and Konso gneisses that mainly occur in Sidamo and Gemu Gofa administrative regions of southern Ethiopia, and the inliers of Dengego gneiss in Hararghe administrative region. The Alghe gneiss consists of layered biotite - hornblende gneiss and migmatite; it is structurally overlain by the Awata mafic gneiss (Warden, 1982; Warden and Horkel, 1984). Chater (1971) described the Alghe gneisses in Megado region, referred to as the Lower Group, as forming a broad domal core that represents an east-west trending fold system, reworked by the Pan-African event that produced north-south structures. Warden and Horkel (1984) supported Chater's view, and considered that the Alghe gneiss of the SGZ is the lowest structural unit in southern Ethiopia.

The Konso gneisses are made up mainly of granulites with subordinate amphibolite facies rocks of both sedimentary and igneous origin. The granulites and the amphibolite grade rocks within the SGZ are supposedly separated by Pan-African isograds, that cut across primary layering (Omo River Project Team, 1973; Davidson, 1983). The reported structural continuity in these rocks indicates that the isograds represent different structural levels in reactivated crust. Kazmin (1975b) arbitrarily assigned the granulites to his lowest stratigraphic unit, the Lower Complex.

The Dengego gneiss mainly consists of migmatites; the relationship between these rocks and those of the Alghe and Konso gneisses is not clear.

6.3.4 Yavello paragneiss zone (YPZ)

The Yavello paragneiss zone represents the Awata, Yavello and Burji gneisses in addition to the Wadera Group of the Middle Complex of Kazmin (1975b, 1978b). Kozyrev et al. (1985) subdivided rocks of the YPZ into the Awata and Mormora groups. According to Kozyrev et al., rocks of the Awata Group lie unconformably on the Alghe gneiss of the SGZ in one locality (Awata River), but is faulted against it elsewhere. The Awata Group is represented by the Bora and Buluka Formations, that contain mainly amphibole and biotite-amphibole gneisses. The Mormora Group is represented by the Zembaba, Aflata and Kenticha Formations in ascending stratigraphic order. The Zembaba Formation (equivalent to Kazmin's Wadera Group) consists of quartzofeldspathic rocks with minor intercalations of biotite-hornblende gneiss and marble. The Aflata Formation consists of biotite gneisses alternating with amphibolites and thin beds of marble in the upper part. The Kenticha Formation is characterized by graphitic and micaceous schists and metapelites, containing kyanite, staurolite and garnet. The prevailing psammites and conglomerates have well-preserved sedimentary structures such as cross bedding and ripple marks clearly indicating a continentally-derived, metaclastic succession (Gilboy, 1970; Chater, 1971).

Although adequate structural constraints are lacking, Warden and Horkel (1984) suggested that the YPZ rocks were subaerially deposited in a block-faulted graben system underlain by rocks of the SGZ. Contact relations

between the YPZ and the structurally underlying rocks of the SGZ are masked by migmatization and intense shearing in the basal part of the Awata and Yavello gneisses. In the absence of any marked stratigraphic break, distinction between the YPZ and the SGZ is made on contrasting lithologic types and moderate changes in metamorphic grade, from middle amphibolite to upper amphibolite and granulite grade respectively.

A Rb-Sr whole-rock isochron from a phyllite within the YPZ has yielded the oldest date in Ethiopia: 1030 $\pm$ 40 Ma (Chater, 1971), which places a minimum age on the time of sedimentation.

In eastern Ethiopia, lithologically similar occurrences to those of the Mormora Group, known as the Boye and Soka Groups (Kazmin, 1975), appear to be a northward continuation of rocks of the YPZ. The Boye Group consists of quartzite, arkose, graphitic and stromatolitic marble that are overlain by the Soka Group of graphite, talc, and chlorite schists as well as phyllite.

6.3.5 Tullu Dimtu ophiolitic rocks

The Tullu Dimtu-Dallati-Yubdo mafic/ultramafic rocks of western Ethiopia are thought to be ophiolitic in nature (Kazmin, 1976; de Wit and Aguma, 1977; Shackleton, 1986). These are probably equivalent to those found in northern Ethiopia along the Baraka lineament, and to the south in subdomain 4 of the Akobo Domain in the Omo River project area (de Wit and Chewaka, 1981; Davidson, 1983).

The Tullu Dimtu ophiolitic rocks form a belt 10 - 15 km wide and 120 km long belt (Kazmin, 1976; de Wit and Aguma, 1977) containing dunite, pyroxenite and gabbro with associated pillow lavas and breccia overlain by

chert, limestone and black shale. The metasedimentary rocks are interlayered with thick units of mafic metavolcanics. Sheeted-dyke complexes reported from the area look like vertical partings or fractures in metamorphosed mafic flows. Kazmin (1976) pointed out that the belt consists of numerous tectonic slabs separated by steep, westerly-dipping thrusts (overthrust to the east) and narrow shear zones. Davidson (1983) described elongate masses, lenses and slivers of mafic/ultramafic rocks in the Akobo and Alanga basins in Illubabor administrative region.

6.3.6 Birbir magmatic zone (BMZ)

Rocks of this zone occupy western and southwestern Ethiopia and are probably traceable northwards into the Butana and southern Red Sea Hills volcano-sedimentary terranes of Sudan. These are in turn traceable into the Asir volcano-sedimentary terrane in Arabia (the equivalent of the Jeddah terrane). The BMZ consists of calc - alkaline volcanic and plutonic rocks, and associated sedimentary rocks equivalent to the Birbir Group of Kazmin (1976, 1978). In southwestern Ethiopia, rocks of this zone can be traced as far south as latitude 6° N, where they form subdomains 1 and 3 of the Akobo Domain (Davidson, 1983). The zone extends to the north across the Gore-Gambella Geotraverse area, where plutonic rocks make up three-quarters of the rocks in the area. The present study provides a basis for sub-dividing the BMZ into four types using geochemistry and geochronology of the plutonic rocks.

The first consists of strongly-deformed gabbro, diorite, tonalite and granodiorite, referred to as the low-K suite; the second comprises quartz

diorite to granodiorite and minor granite assigned to the calc-alkaline suite. According to the present mapping and also further north (Kazmin et al., 1979), rocks associated with the low-K suite include metamorphosed andesitic to rhyolitic volcanics, wacke, pelite, and minor intercalations of coarse volcanoclastic rock, Fe-rich quartzite, intraformational conglomerate, marble and chert. In the preceding chapters, it is proposed that the low-K suite formed on oceanic crust in an island arc environment.

The strongly-deformed, elongate bodies of the calc-alkaline suite have been attributed on chemical and isotopic grounds to continental margin arc-type magmatism. The metasedimentary rocks associated with the group include wackes, pelites, volcanoclastic rocks and subordinate marble and chert.

The third group of plutonic bodies, the alkali-feldspar granite suite, is late- to post-kinematic, while the fourth, and least known group of plutonic rocks, on the west side of the BMZ, is characterized by massive to slightly deformed gabbro, diorite and hornblende in the Geotraverse area. Undeformed, circular gabbros intruding abundant mafic volcanics, interlayered graphitic schist and quartzite were described to the north along strike (Kazmin et al., 1979, Amenti Abraham, pers.comm., 1984). These seemingly late-kinematic gabbros imply that mafic magmatism was repeated late in the tectonic evolution of the BMZ, perhaps in an episode of back-arc spreading.

6.3.7 Northern volcanic-sedimentary terrane (NVST)

The northern volcanic-sedimentary terrane (NVST) occupies a large part of north and northeastern Ethiopia, in Tigray and Eritrea administrative

regions. The NVST is traceable into the full width of the Asir south (Abha) and Asir east volcanic-sedimentary terranes (Stoesser and Camp, 1985) along the Red Sea in Arabia (Fig. 1.2). The NVST consists of the Tsaliet Group that grades into rocks of the Tambien Group, which is in turn unconformably overlain by the Didikama, Shiraro and Matheos Formations (Beyth, 1971, Morton, 1975).

The Tsaliet Group consists of meta-andesite and tuff, tuffaceous slate, greywacke and rhyolitic breccia. Morton (1975) described the Tsaliet Group of the Enticho area as a succession of interbedded mudstone and intraformational conglomerate, greywacke, limestone and volcanic and pyroclastic rocks. Pillow structures in volcanics and sedimentary structures such as ripple marks have been reported. Morton also interpreted the Tsaliet Group as indicating deposition within both marine and subaerial environments. Syn- and post-tectonic plutons, that range from ultrabasic to granitic in composition, cut the Group. K/Ar dates of 690 ± 4 Ma (Miller et al., 1967), and 800 Ma (Metal Mining Agency of Japan, 1974) were obtained from post-tectonic granites and granodiorites.

The Tambien Group consists of slate and dolomitic and algal limestone. The Didikama, Shiraro and Matheos Formations comprise a succession of saccharoidal, stromatolitic limestones that alternate with thin beds of clastic sedimentary rocks.

6.3.8 Adola magmatic zone (AMZ)

The Adola magmatic zone occurs in southern Ethiopia and corresponds to the "Upper Group" of Gilboy and Chater (1970), later described as the Adola

Group that consists of the Chakata and Finkilcha Formations and Kajimiti beds (Kozyrev et al., 1985). The Chakata Formation makes up the lower part of the Adola Group and consists of strongly sheared, discontinuous ultramafic lenses associated with amphibolite, schist, mafic plutons and volcanics showing relict pillow structures, considered parts of an ophiolite complex. The upper part consists of graphitic phyllite and argillaceous rocks forming the Finkilcha Formation. Rocks of the Chakata Formation can be divided into narrow, parallel tectonic zones, considered to be easterly-thrusted nappes (Kazmin, 1976). The Kajimiti beds lie unconformably on rocks of the Chakata and Finkilcha Formations and consist of psammite, biotite schist and calcareous sedimentary rocks. The sedimentary rocks of the Finkilcha Formation and Kajimiti beds have been interpreted as products of sedimentation within the Megado graben-syncline (Kozyrev et al., 1985).

Rb/Sr whole-rock isochron dates of 680 ± 30 , 450 ± 10 and 505 ± 10 Ma (Gilboy, 1970; Chater, 1971) were reported from post-tectonic granites intruded into the AMZ.

6.4 SUMMARY

The lithostructural units described above appear to be representative of both the pre- Pan-African crust and Pan-African rocks. The portion of the pre- Pan-African crust reactivated during the late Proterozoic comprise the Mozambique Belt in east Africa. The Mozambique Belt has a complex history of superposed deformation and metamorphism, and reworked rocks as old as Archean have been identified within it (Kroner, 1977; Almond, 1982;

Shackleton, 1986). The Pan-African rocks, on the other hand, exhibit a more restricted set of events and have evolved mainly in the interval ca. 1000-550 Ma (Greenwood et al., 1976; Ries et al., 1985). This study as well as more recent publications (e.g. de Wit and Chewaka, 1981; Kozyrev et al., 1985; Shackleton, 1986) have supported an accretionary history more consistent with the emerging views concerning the surrounding region. Thus the eight lithostuctural units described above may be grouped into two broad categories:

6.4.1 Reactivated Pre- Pan-African rocks

The reactivated pre- Pan-African rocks of Ethiopia include the Baro and Geba Domains of the Geotraverse and comprise the Western, Central, and Southern gneiss zones. These are high-grade polycyclic rocks with more than one generation of metamorphism and structure (Gilboy, 1970; Chater, 1971; Davidson, 1983; Kozyrev et al. 1985). Relict east-west - trending recumbent isoclinal folds are preserved in the high-grade gneissic terranes of western (Kazmin et al. 1978) and southern Ethiopia (Chater, 1971; Warden, 1984; Warden and Horkel, 1984). In western Ethiopia, along the frontier zone between Ethiopia and the Sudan, the presence of large east-west - trending folds is reported from satellite imagery (Warden and Horkel, 1984). These folds represent probable remnants of pre- Pan-African basement structure, similar to those described from the basement rocks of Kenya and Uganda (Hepworth, 1972, 1979). The Pan-African event, however, reworked and largely obliterated the east-west structures. This event superimposed foliation and gneissic layering on earlier structures and these strike NNE to NNW.

6.4.2 Pan-African rocks

Rocks of Pan-African age within Ethiopia include ophiolitic rocks, calc-alkaline volcanic and plutonic units and associated sediments. In Ethiopia, the Pan-African rocks include the Yavello paragneiss zone, the Tullu Dimtu ophiolitic rocks, the Birbir magmatic zone, the northern volcanic-sedimentary terrane and the Adola magmatic zone. In western Ethiopia, these rocks have been considered as wedges of oceanic crust associated with rifted zones, early igneous activity and later crustal thickening during ocean closure and collision (Kazmin et al., 1978; Ayalew et al., 1987). The lithologic association, grade of metamorphism and style of deformation are similar to those described in other NE African and Arabian Pan-African rocks. These can be traced directly in a palinspastic reconstruction of the NE African-Arabian Shield, into two branching systems extending from within Ethiopia (Vail, 1985; Fig. 5.1). The eastern belt, comprising the northern volcanic-sedimentary terrane (NVST) and possibly the Adola magmatic zone (AMZ) to the south, joins the Asir south (Abha) and Asir east terranes within the Arabian peninsula. The western belt of the Birbir magmatic zone (BMZ) is traceable into the southern Red Sea Hills and the Jeddah terrane or Asir volcano-sedimentary terranes in Sudan and Saudi Arabia respectively. This extrapolation suggests that the two discrete belts each result from more than one phase of island arc formation.

6.5 RELATIONSHIP OF PRE- PAN-AFRICAN AND PAN-AFRICAN ROCKS IN WESTERN ETHIOPIA

There seems little doubt that early east-west trending structures in the Geba gneisses and Konso granulites of the CGZ (Teferra and Berhe,

unpublished; Kazmin, 1975b) are overprinted by north-south - trending structures of Pan-African age. South of the Western Ethiopian Shield similar structures have been documented in rocks of the known Mozambique Belt, that are known to be older than late Proterozoic (for example: Sanders, 1965; Hepworth, 1972; Shackleton, 1986). In the foreland zone to the west of the Mozambique Belt, the north-south structures are absent and rocks that have been dated yield Archean and early Proterozoic ages. Rocks of WGZ and CGZ have yet to be dated; however, they contain mineral assemblages typical of high temperature and medium pressure metamorphism. Most isotope systematics are easily disturbed under these physical conditions, so it is difficult to establish the true age of these rocks by conventional methods. Older ages may have been obliterated or distorted by the Pan-African event.

The chemical character of rocks of the Birbir magmatic zone and its association with ophiolitic rocks imply that it originated through both ensimatic and ensialic processes. These rocks appear to be disposed in a major steeply-dipping ductile transcurrent shear zone interpreted as a plate boundary. This boundary, marked by the presence of ophiolitic rocks, is continuous as far north as Saudi Arabia and defines a suture between pre-Pan-African rocks and Pan-African rocks. The subduction model that has been proposed in Chapter 5 clearly explains the tectonic juxtaposition of lower-grade and high-grade rocks in western Ethiopia. The model of tectonic assembly virtually negates Kazmin's stratigraphic model based on basement - cover relationships. Much more geochronological work will be needed, however, in order to rigorously test this new tectonic model.

CHAPTER 7

CONCLUSIONS

7.1 GEOLOGY

The Gore-Gambella area straddles the boundary between three geologic domains. The Baro and Geba domains consist of heterogeneous ortho- and paragneisses and migmatites while the Birbir domain consists of metasedimentary and calc-alkaline volcanic and plutonic rocks. There is an abrupt increase in metamorphic grade across the domain boundaries. Maximum P-T conditions of metamorphism inferred from rocks of the Baro and Birbir domains are 6.5kb, 700°C and 3.5-4kb, 525-545°C respectively.

Field relations of the seven plutonic units studied suggest three groups: 1) pre- to syn-kinematic mafic and intermediate intrusive complex, 2) syn-kinematic leucogranite sheets, and 3) late to post-kinematic alkali-feldspar granite. Rocks of Group (1) occupy the Birbir Domain, and are metamorphosed and deformed to variable degrees within a major north-south, ductile, transcurrent fault system, the Birbir Shear Zone (BSZ). Deformation is most extreme in rocks of Group (1). Protomylonites and mylonites alternate with narrow, heterogeneous mylonitic units and show steep to vertical mylonitic foliation and sub-horizontal stretching lineation. Group (2) rocks are strongly-deformed, elongate plutons with long axes and internal foliation parallel to the foliation found in the enclosing gneissic rocks of the Baro Domain. Plutons of Group (3) are more equidimensional, discordant and generally undeformed; they have intruded the western Birbir and Birbir/Baro boundary zone.

7.2 CHEMISTRY

Group (1) rocks are divided into two genetically distinct suites, a low-K gabbro-tonalite suite and calc-alkaline suite that ranges from diorite to granodiorite. REE in the low-K suite show low absolute abundance (40-75ppm), no Eu anomaly (Eu/Sm :0.31-0.32) and $\text{La/Lu}_{(\text{cn})}$ =5.3. The calc-alkaline suite has high absolute REE abundance (99-121ppm), a slight Eu anomaly (Eu/Sm :0.23-0.27) and $\text{La/Lu}_{(\text{cn})}$ =5.1-7.5. Rb/Sr ratios for the low-K suite range from 0.01 to 0.04 whereas for the calc alkaline suite they are significantly higher and range from 0.11 to 0.17. On the basis of trace element data (Hf, Nb, Rb, Ta, Y and Yb) Group (1) rocks are considered to be the product of magmatic arc activity, whereas Groups (2) and (3) reflect intraplate activity. Major and trace element data also suggest that the low-K suite is oceanic in affinity, whereas the calc-alkaline suite resembles the continental margin type.

Major and some trace element data for samples from groups (2) and (3) are very similar but the two groups have distinct REE patterns. Group (2) plutonic units are products of large scale anatexis, resulting from crustal thickening along the eastern part of the Baro domain. Group (3) rocks are thought to have been formed by partial melting of the evolved arc, related to post-collisional uplift.

7.3 GEOCHRONOLOGY

Five U-Pb zircon and four Rb-Sr isochron dates constrain the ages of the plutons and their subsequent history. The data show plutonic activity between 830 and 540 Ma, similar to the time span established for the Pan-

African rocks of the Southern Red Sea Hills of the Sudan and the Jeddah terrane of Saudi Arabia. U-Pb zircon ages of 828 ± 5 and 814 ± 2 Ma, obtained from Group (1) rocks, date the time of their emplacement. One of these plutons, the Goma pluton, also yielded a whole-rock Rb-Sr isochron date of 759 ± 18 Ma. Granite sheets of Group (2) within the high-grade Baro domain, that are believed to be anatectic in origin, were dated by U-Pb at 782 ± 15 Ma. The 780-760 Ma event is interpreted as the period of Pan-African regional metamorphism in western Ethiopia. This event also correlates with the age of low-grade metamorphism of arc-related rocks elsewhere in NE Africa. The Birbir domain, therefore, represents a southward extension of the Pan-African crust of NE Africa and Arabia. Other reset Rb-Sr whole rock isochrons that cluster around 635 ± 8 Ma provide evidence of significant isotopic homogenization, probably associated with major transcurrent movement parallel to a plate margin undergoing oblique collision. A U-Pb lower intercept age of 582 Ma from rocks of Group (2) denotes a second high-grade metamorphic event and correlates with the age of the Mozambiquian amphibolite facies metamorphism in NW Kenya. The Baro domain is here interpreted as a continental margin to the Mozambiquian basement system to which juvenile crust accreted.

U-Pb ages of 571 ± 5 and 550 ± 10 Ma were obtained from the Group (3) late- to post-kinematic plutons. Rb-Sr biotite dates from Group (1) pre-kinematic plutons indicate resetting of mineral systems until about 540 Ma ago.

7.4 CONCLUDING NOTES

The principal geologic features of western Ethiopia are consistent with

a plate tectonic mode of development. The geology, chemistry and isotopic data presented above show distinct and separate episodes of plutonism.

Chemical characteristics of plutons from the Birbir domain suggest a subduction model for magma generation. The presence of ophiolitic rocks in the Tullu Dimtu belt immediately north of the study area supports such a model. The inferred tectonic evolution of the area is summarized below:

- 1- The chemical and isotopic characteristics of rocks of the Birbir Domain imply both ensimatic and ensialic processes of origin. The Birbir Domain was developed and aggregated through processes of magmatism, sedimentation and tectonism that accompanied closure of an ocean basin.
- 2- The peraluminous plutonic units are restricted to the Baro Domain that may be related to crustal thickening and large scale anatexis along a tectonically active continental margin.
- 3- The Birbir Shear Zone (BSZ) is evidence for major translation during tectonic assembly of the Birbir and Baro domains. The transcurrent movement on the BSZ was initiated as dextral displacement, possibly parallel to an active plate margin undergoing oblique subduction, and subsequently changed to sinistral movement during the terminal stages of collision. The BSZ represents a mylonitic cryptic suture. The suture between the Geba and Birbir domains is marked to the north of the study area by the Tullu Dimtu ophiolitic rocks.

4- A culminating event at about 580 Ma, probably a continental collision, generated high-grade metamorphism and granitic plutonism during 580-550 Ma.

In conclusion, the recognition in this study of contrasting genetic types of plutonic rocks, represented by Group (1) on one hand and Group (2) on the other, and the identification of a wide zone of intervening mylonites lends support to an hypothesis for the existence of two distinct rock suites. The association of ophiolitic rocks (occurring north and south of the Geotraverse area but not mapped during this study) and the primitive Sr isotope ratios indicate the close association of oceanic crust in the region at the time. Rocks of the Birbir magmatic zone were apparently accreted to the eastern margin of the pre- Pan-African crust during a relatively long span of time, ca. 830-630 Ma. The zone presumably represents one or more suspect terranes, but the extent of the individual terranes and the processes through which they were accreted are essentially unknown. Considerable progress has been made in understanding the nature of the contact between the lower grade and high-grade rocks in western Ethiopia. There are no data yet that define the original age of the high-grade metamorphic rocks, seen in the Baro and Geba domains and elsewhere in Ethiopia, that appear to be continental crust to which the arc rocks have been accreted. Further chemical and isotopic studies of the plutonic and volcanic rocks, detailed geochronologic studies of selected critical samples, and detailed analysis of structural and metamorphic history are required to test the tectonic model in its entirety.

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APPENDIX 1

PETROGRAPHY

Typical modes of intrusive units were determined in part using a point counting procedure and these are listed in Table 2.1. Those obtained from metamorphosed sedimentary and volcanic rocks and gneisses are given in Table 2.2. Brief mineralogical and textural descriptions are provided here for representative samples from the Birbir, Goma, Haya, Bure, Baro, Bonga and Mao plutonic units. Sample localities are given in Figure 2.3.

Birbir: Sample # T3G. Quartz diorite

Mesosopic features:

- fine-grained, strongly foliated; foliation defined by compositional layering of biotite and hornblende between layers approximately 0.5cm thick.

Microscopic features:

- Quartz: commonly as anhedral microcrystalline aggregates (recrystallized with mosaic microstructure), some flattened; few ribbons up to 2mm by 0.2mm and as matrix less than 0.5mm across, straight grain boundaries, uniform to undulose extinction. Minor K-feldspar occurs as anhedral interlocking grains with quartz.
- Plagioclase: subhedral-euhedral laths up to 5mm by 1.5mm occur as porphyroclasts, and 0.5mm in the matrix; some show concentric zoning, deformation twins, mantled by very fine-grained recrystallized plagioclase; slightly altered along cleavage traces and edges to epidote and calcite.
- Biotite: subhedral-euhedral plates, dark brown to light greenish-brown pleochroism, wrapped around hornblende porphyroclasts, define mylonitic foliation by their orientation.
- Hornblende: euhedral-subhedral crystals, strong pleochroism from light green to dark green, show preferred orientation

- Accessory minerals: euhedral titanite, euhedral zircon commonly included in biotite; abundant apatite generally included in quartz and feldspar.
- Matrix minerals approx. 60-70%, composed of fine-grained continuous layers quartz- plagioclase and biotite-hornblende-epidote

Goma: Sample # T6J. Granodiorite

Mesosopic features:

- uniform, medium-grained, subhedral-granular, weakly developed planar fabric

Microscopic features:

- Quartz: commonly as fine-grained anhedral matrix and less commonly as stubby phenocrysts, 1 to 3 mm across, with strain features, including undulose extinction and microfractures penetrated by stringers or irregular patches of secondary minerals, mainly epidote.
- K-feldspar: mainly microcline with subordinate orthoclase, microcline occurs as subhedral perthite phenocrysts, 3 to 5 mm in diameter, often clouded by sericite alteration.
- Plagioclase: subhedral laths, 2 to 5 mm in diameter, straight to somewhat embayed grain boundaries, oligoclase (An_{12-20}), some filled with granular aggregates of epidote and minor calcite, particularly along cleavage traces.
- Biotite: anhedral to subhedral grains, ragged to curved edges, pleochroic : greenish brown to brownish green, strongly altered and intergrown with saenitic rutile and chlorite.
- Accessory minerals include allanite, titanite, zircon and apatite.

Haya: Sample # T4A. Tonalite

Mesosopic features:

- Fine-grained, pale grey, strong foliation defined by about 5 mm to 1 cm scale compositional layering of biotite and muscovite

Microscopic features:

- **Plagioclase:** anhedral, slightly rounded to elongate laths, up to 4 mm in length, curved to slightly embayed grain boundaries, often sericitized, and containing abundant epidote and calcite.
- **Quartz:** anhedral, equant to elongate habit, some form long rods, often embayed boundaries, some with undulose extinction, associated with rather rare K-feldspars.
- **Biotite and muscovite flakes,** subhedral laths with straight grain boundaries, straight extinction; planar orientation defines the foliation plagioclase and quartz crystals.
- **Accessory minerals** are very rare: apatite, titanite.

Bare: Sample # T5D. Hornblende gabbro**Mesosopic features:**

- **Fine-grained, strongly deformed, lenticular hornblende-rich layers** alternate with plagioclase-rich layers.

Microscopic features:

- **Hornblende :** abundant, often rimmed by fibrous actinolitic amphibole.
- **Plagioclase:** lath-shaped, filled with very fine-grained epidote and calcite especially along the edges.
- **Accessory minerals** include titanite, apatite and abundant opaques.

Baro: Sample # T8F. Biotite granite**Mesosopic features:**

- **medium-grained, weakly deformed; leucocratic: colour index <10.**

Microscopic features:

- **Alkali-feldspar:** mainly microcline and some perthitic microcline, pervasive sericitization
- **Plagioclase:** subhedral, predominantly albite, some are myrmekitic

- Quartz: mainly recrystallized matrix; forms interlocking grains with feldspars, some with undulose extinction; and some occur as fracture fillings.
- Biotite: pale yellow to rusty brown pleochroism, some show chloritization with extensive hematite staining, numerous pleochroic haloes - radiation damage.
- Accessory minerals: zircon, apatite, titanite, and opaque oxides.

Bonga pluton - core: Sample # T8L. Quartz monzonite

Mesosopic features:

- massive grey to pinkish grey, colour index of about 35; coarse K-feldspar and minor quartz megacrysts; some feldspars are as large as 5 cm with plagioclase rims about 1 mm thick (rapakivi texture), fine-grained melanocratic matrix; shows a significant variation in quartz content and the relative proportion of feldspars as well as textural differences (hybrid feature).

Microscopic features:

- Microcline: large subhedral phenocrysts (1 to 4 cm) often mantled by plagioclase, slightly altered to sericite; also less commonly enclose plagioclase
- Plagioclase: subhedral-euhedral laths, both as large phenocrysts and matrix, some zoned (An₄₀₋₂₈).
- Quartz with anhedral, sub-rounded or globular habit.
- Biotite: subhedral - euhedral plates and laths; poikilitically enclose allanite and zircon. Allanite occurs as non-metamict, euhedral reddish brown zoned crystals. Subhedral, prismatic hornblende pleochroic: dark green to light green is generally associated with biotite

Plagioclase, quartz, biotite, hornblende together with ubiquitous accessory minerals (allanite, apatite, zircon and titanite) constitute the matrix grains, about 0.5 to 1 mm in size.

Bonga pluton - rim: sample # T8K. Alkali-feldspar granite

Alkali-feldspar granite that cuts the megacrystic feldspar quartzmonzonite consists of compositionally homogeneous rock compared to T8L. A slight mineralogical variation with a general decrease in the biotite and hornblende content. The major part of the light minerals are represented by interlocking grains of microcline, quartz and plagioclase. Compared to sample # T8L, contains less mafic minerals and more abundant accessory minerals, biotite in large poikilitic form or in small crystals included in other minerals, hornblende as dark green to bluish-green pleochroic, as isolated subhedral crystal.

Mao: Sample # T90

- Sample, from the eastern margin of the pluton, shows shear band foliation; colour index of about 20; composed of quartz, microcline, plagioclase (sodic oligoclase), biotite and hornblende, and ubiquitous zircon, allanite, apatite and titanite. Most samples contain stilpnomelane and chloritized biotite, and slightly sericitized feldspar.

APPENDIX 2

MAJOR- AND TRACE ELEMENT ROCK ANALYSES,
ANALYTICAL PROCEDURES AND ACCURACY

Major elements, Ba, Cr, Nb, Ni, Rb, Sr, V, Y, Zn and Zr were determined by XRF spectrometry using a Philips PW 1410/20 AHP instrument at University of Ottawa. Fused-glass pellets were prepared according to the methods of Norrish and Hutton (1969). One gram of powdered sample was mixed with 5 gm of $\text{Li}_2\text{B}_7\text{O}_4$ and 0.3 gm LiF and fused to a glass pellet within a platinum holder. Analyses were undertaken using the wavelength-dispersive method.

The precision of XRF data produced in the laboratory, expressed as percent relative standard deviation is approximately as follows (standard deviation/concentration) * 100 :- SiO_2 (0.9), TiO_2 , Al_2O_3 , FeO, CaO (0.8), K_2O (1.2), MgO (2.0), Na_2O (2.5), MnO (4), trace elements (-10). Hence 95 percent confidence limits (+/- 25) to all data may be calculated. For example, 63.52 % SiO_2 should read 63.52 +/- 1.14, or 63.5 +/- 1.2. Similarly, 243 ppm Ba should read 243 +/- 48, or 240 +/- 50. There are, therefore only two significant figures in the tabulated analyses; the first is fixed and the second is doubtful. The third figure may be retained for calculations, and the fourth, where given, is meaningless.

Geochemical analyses invariably exhibit uncertainty due to systematic error (inaccuracy), random error (imprecision or precision error), and occasionally to other causes such as impurity (Kretz, 1985). The uncertainty due to systematic error or inaccuracy of the

XRF analyses was estimated by co-analysing two reference standards, DR-N (Diorite) and GS-N (Granite); their preferred composition (R.V.) values on a dry basis were given by Abbey (1983). These values together with means (X) and standard deviations (S.D.) for these reference standards determined in this study, are tabulated below (Table A2.1). The standard deviation is also given as a percentage of the mean, known as the relative standard deviation (coefficient of variation) or relative error (R.E.). Standard error for most major elements is within $\pm 1\%$ ($\pm 10\%$ for P_2O_5 , MnO and trace elements).

XRF analytical data are given in Table A2.2, together with CIPW norms, calculated by computer at University of Ottawa are expressed in weight percent. $FeO(T)$ is total iron, and DI = Differentiation Index. Sample localities are given in Figure 2.2.

Table A3.1: Accuracy determined for XRF major- and trace element analyses of reference standards (N: number of analyses; Fe₂O₃: total iron as Fe₂O₃).

	ANRT1 - DR-N(Diorite) N=11				ANRT 2 - GS-N(Granite) N=8			
	R.V.	X	S.D.	R.E. (%)	R.V.	X	S.D.	R.E. (%)
SiO ₂	52.98	52.56	0.15	0.28	65.98	65.66	0.22	0.34
Al ₂ O ₃	17.56	17.42	0.06	0.35	14.71	14.65	0.05	0.35
Fe ₂ O ₃	9.72	9.64	0.10	1.07	3.58	3.78	0.01	0.19
MgO	4.41	4.30	0.03	0.78	2.31	2.33	0.03	1.32
CaO	7.07	6.97	0.03	0.50	2.51	2.46	-	-
Na ₂ O	3.00	3.01	0.07	2.35	3.78	3.81	0.05	1.19
K ₂ O	1.70	1.70	0.01	0.42	4.64	4.65	0.01	0.15
TiO ₂	1.09	1.03	0.02	2.06	0.68	0.63	0.01	1.12
P ₂ O ₅	0.25	0.19	0.02	9.85	0.28	0.29	0.02	5.97
MnO	0.22	0.22	-	-	0.06	0.05	0.01	14.14
Ba	390	338	112.4	33.27	1400	1436	8.40	0.63
Cr	42	46	5.13	11.16	55	64	2.12	3.31
Zr	125	130	6.27	4.82	240	219	1.58	0.72
Sr	400	399	2.19	0.55	570	608	3.81	0.63
Rb	70	67	2.80	4.18	190	181	1.00	0.55
Y	30	29	1.73	5.97	-	19	1.23	6.45
Nb	6	4	1.47	7.36	-	19	1.73	9.12
Zn	145	144	10.09	7.01	48	54	7.21	13.35
Ni	16	10	4.45	34.53	34	22	1.58	7.19
V	230	220	10.72	4.87	62	62	6.25	10.07

TABLE A 2.2
WHOLE-ROCK CHEMICAL DATA AND CIPW NORMS

Sample #	4J	5J	6J	7J	11J	1J	2J
(Wt. %)							
SiO ₂	69.35	69.74	68.34	71.47	66.30	67.38	67.57
Al ₂ O ₃	14.80	15.11	15.51	14.49	16.55	15.30	15.55
FeO _(T)	3.49	3.24	3.30	2.29	3.90	4.31	3.55
MgO	0.87	0.83	0.92	0.62	1.06	1.22	0.99
CaO	2.05	2.05	2.29	1.87	2.73	2.38	2.30
Na ₂ O	5.09	4.85	5.36	4.18	5.75	5.13	5.18
K ₂ O	3.49	3.53	3.09	4.04	2.64	3.13	3.23
TiO ₂	0.52	0.44	0.47	0.29	0.54	0.59	0.49
P ₂ O ₅	0.19	0.12	0.15	0.06	0.19	0.18	0.13
MnO	0.09	0.08	0.09	0.06	0.12	0.11	0.09
(ppm)							
Ba	1125	1035	1271	1279	1175	1093	1218
Cr	87	87	84	126	87	170	185
Zr	326	283	290	195	300	348	291
Sr	274	281	331	281	405	304	319
Rb	55	68	42	102	40	56	47
Y	27	27	28	20	29	33	25
Nb	11	8	8	7	6	14	11
Zn	90	66	41	36	118	79	54
Ni	1	8	12	41	39	-	-
V	34	24	20	15	28	25	37
TOTAL	100.20	100.21	99.76	99.63	100.05	99.94	99.35
CIPW							
Q	20.60	21.84	18.78	26.24	14.85	18.13	19.32
Or	20.63	20.86	18.26	23.88	15.60	18.50	19.31
Ab	43.07	41.04	45.36	35.37	48.66	43.41	44.30
An	7.23	9.03	9.14	8.84	11.55	9.48	9.74
C	-	-	-	-	-	-	-
Di	1.41	0.29	1.03	0.03	0.62	0.95	0.68
Hy	3.67	4.03	3.90	3.05	4.89	5.36	2.44
Mt	1.69	1.56	1.59	1.11	1.88	2.08	2.92
Il	0.99	0.84	0.89	0.55	1.03	1.12	1.28
Ap	0.44	0.28	0.35	0.14	0.44	0.42	0.31
Rb/Sr	0.20	0.24	0.13	0.36	0.10	0.18	0.14
DI	85.46	84.83	83.88	86.77	80.29	81.20	82.94

FeO_(T) is total iron.

DI is Differentiation Index

Sample #	3J	8J	9J	21J	23J	25J	26J
(Wt. %)							
SiO ₂	67.42	67.21	68.03	68.52	69.71	67.69	68.16
Al ₂ O ₃	16.05	15.91	15.95	15.15	14.48	15.62	15.76
FeO(T)	3.47	3.55	3.41	3.30	2.69	3.41	3.43
MgO	0.94	0.96	0.95	0.92	0.76	0.91	0.97
CaO	2.36	2.47	2.36	2.19	1.98	2.32	2.36
Na ₂ O	5.53	5.50	5.35	4.83	4.52	5.34	5.49
K ₂ O	3.05	3.02	3.06	3.42	3.47	3.07	2.97
TiO ₂	0.47	0.49	0.46	0.45	0.35	0.45	0.48
P ₂ O ₅	0.13	0.13	0.13	0.11	0.08	0.12	0.12
MnO	0.09	0.09	0.09	0.09	0.07	0.08	0.08
(ppm)							
Ba	1148	1108	1085	1176	1087	1149	979
Cr	173	156	191	132	127	106	111
Zr	266	275	275	278	213	281	268
Sr	326	344	321	297	279	324	337
Rb	47	45	47	60	67	49	40
Y	32	26	26	26	19	27	27
Nb	10	9	11	8	6	13	12
Zn	57	32	58	61	48	30	47
Ni	-	-	-	-	-	-	-
V	28	39	17	29	24	28	31
TOTAL	99.76	99.78	100.06	99.24	98.34	99.26	100.03
CIPW							
Q	16.65	16.80	18.16	20.73	24.25	18.12	17.98
Or	18.02	17.85	18.08	20.21	20.51	18.14	17.55
Ab	46.79	46.54	45.27	40.87	38.25	45.19	46.46
An	9.97	9.81	10.47	9.56	8.97	9.59	9.19
C	-	-	-	-	-	-	-
Di	0.74	1.32	0.32	0.49	0.27	0.94	1.10
Hy	4.24	4.03	4.44	4.21	3.54	4.03	4.07
Mt	1.68	1.71	1.65	1.59	1.30	1.65	1.66
Il	0.89	0.93	0.87	0.85	0.66	0.85	0.91
Ap	0.30	0.30	0.30	0.25	0.19	0.28	0.28
Rb/Sr	0.14	0.13	0.15	0.20	0.24	0.15	0.12
DI	82.94	82.65	82.76	83.76	85.49	83.34	83.23

Sample #	28J	29J	2G	3G	5G	9G	12G
(Wt. %)							
SiO ₂	68.64	69.14	56.67	61.75	62.08	60.27	63.89
Al ₂ O ₃	15.17	15.52	18.11	17.09	16.94	16.63	16.46
FeO(T)	3.34	3.27	7.85	7.36	6.28	7.35	5.61
MgO	0.89	0.89	3.58	2.55	2.06	2.72	1.98
CaO	2.12	1.95	7.22	5.34	4.97	5.62	4.67
Na ₂ O	5.02	5.36	3.92	3.37	3.80	3.58	3.61
K ₂ O	3.39	3.40	1.50	2.49	2.62	2.36	2.79
TiO ₂	0.45	0.43	0.71	0.70	0.61	0.67	0.53
P ₂ O ₅	0.12	0.12	0.22	0.16	0.14	0.17	0.14
MnO	0.09	0.08	0.14	0.13	0.11	0.13	0.10
(ppm)							
Ba	1132	1116	767	1328	1666	1091	1068
Cr	110	132	65	74	87	81	79
Zr	268	251	102	212	186	124	241
Sr	283	293	905	546	591	518	497
Rb	56	58	29	59	64	51	69
Y	28	29	21	28	24	32	29
Nb	10	11	-	2	-	-	1
Zn	35	60	74	76	68	86	65
Ni	-	-	22	3	-	-	12
V	27	25	133	120	88	120	104
TOTAL	99.47	100.40	100.18	101.24	99.96	99.76	100.05
CIPW							
Q	20.13	18.94	6.45	15.45	15.14	13.06	18.42
Or	20.03	20.09	8.86	14.67	15.48	13.95	16.49
Ab	42.48	45.36	33.17	28.40	32.15	30.29	30.55
An	8.85	8.25	27.39	24.05	21.43	22.34	20.47
C	-	-	-	-	-	-	-
Di	0.73	0.53	5.73	0.60	1.90	3.65	1.46
Hy	4.04	4.10	12.69	11.93	8.54	10.88	7.77
Mt	1.61	1.58	3.20	3.18	3.06	3.15	2.94
Il	0.85	0.82	1.35	1.34	1.16	1.27	1.01
Ap	0.28	0.28	0.51	0.37	0.32	0.39	0.32
Rb/Sr	0.19	0.19	0.03	0.08	0.11	0.10	0.14
DI	84.37	85.34	48.89	58.52	63.37	57.97	65.89

Sample #	4G	6G	7G	10G	11G	1H	3H
(Wt. %)							
SiO ₂	60.88	60.12	59.40	62.48	56.64	55.22	58.89
Al ₂ O ₃	17.01	16.89	17.30	16.42	16.78	18.96	17.23
FeO(T)	7.41	7.28	7.20	5.98	8.77	7.58	7.45
MgO	2.53	2.74	2.63	2.37	4.35	3.61	3.32
CaO	5.26	5.94	6.11	4.56	7.79	7.10	5.61
Na ₂ O	3.56	3.36	3.80	4.06	3.54	4.67	3.63
K ₂ O	2.72	2.05	1.79	2.86	1.48	1.38	2.45
TiO ₂	0.65	0.56	0.59	0.70	0.69	0.91	0.80
P ₂ O ₅	0.13	0.13	0.13	0.19	0.24	0.31	0.21
MnO	0.13	0.13	0.12	0.10	0.18	0.14	0.13
(ppm)							
Ba	1306	826	790	896	676	633	775
Cr	149	150	143	155	145	77	71
Zr	173	106	119	236	87	203	132
Sr	532	574	639	609	898	959	466
Rb	63	42	44	80	27	28	70
Y	29	23	21	25	17	23	24
Nb	1	2	3	8	4	4	8
Zn	73	71	74	77	64	103	92
Ni	-	-	7	1	8	35	34
V	145	125	124	99	181	137	143
TOTAL	100.58	99.44	99.32	99.96	100.72	100.17	99.96
CIPW							
Q	13.03	14.23	12.07	14.27	6.62	2.04	10.14
Or	16.13	12.12	10.58	16.90	8.75	8.16	14.48
Ab	30.20	28.43	32.15	34.35	29.95	39.52	30.72
An	22.46	24.95	24.86	18.13	25.53	26.70	23.49
C	-	-	-	-	-	-	-
Di	1.30	3.03	3.79	2.64	9.47	5.29	2.43
Hy	12.13	11.57	10.64	8.08	14.44	11.73	12.61
Mt	3.13	2.99	3.03	3.19	3.18	3.49	3.33
Il	1.26	1.06	1.12	1.33	1.31	1.73	1.52
Ap	0.30	0.30	0.30	0.44	0.56	0.72	0.49
Rb/Sr	0.14	0.07	0.07	0.13	0.03	0.03	0.15
DI	59.36	55.60	55.70	66.02	45.49	50.11	55.85

Sample #	5H	8H	10H	2H	4H	6H	7H
(Wt. %)							
SiO ₂	58.00	57.02	59.08	58.53	57.40	63.01	56.95
Al ₂ O ₃	17.55	17.33	16.96	16.57	17.25	15.61	17.33
FeO(T)	7.38	8.25	7.43	7.56	7.99	5.04	8.94
MgO	3.56	3.77	2.81	3.42	3.37	1.95	3.09
CaO	6.54	7.10	5.91	5.98	6.54	3.81	6.76
Na ₂ O	3.78	3.53	3.54	3.64	3.57	4.12	3.57
K ₂ O	1.91	1.83	2.50	2.21	2.14	3.91	2.02
TiO ₂	0.78	0.75	0.82	0.68	0.82	0.63	0.89
P ₂ O ₅	0.18	0.11	0.19	0.09	0.15	0.15	0.22
MnO	0.13	0.15	0.13	0.14	0.14	0.09	0.15
(ppm)							
Ba	506	591	677	593	680	1073	574
Cr	93	108	84	142	128	165	144
Zr	152	150	163	117	140	237	154
Sr	523	454	477	500	480	557	489
Rb	50	44	61	54	52	86	49
Y	23	30	29	25	27	17	28
Nb	5	5	5	5	6	7	5
Zn	89	49	71	83	86	59	86
Ni	40	10	20	14	-	-	6
V	155	178	113	131	145	72	185
TOTAL	100.03	100.04	99.58	99.04	99.60	98.60	100.14
CIPW							
Q	8.68	7.56	11.23	9.86	8.39	13.0	8.09
Or	11.29	10.81	14.77	13.06	12.65	23.11	12.04
Ab	31.99	29.87	29.95	30.80	30.21	34.86	30.44
An	25.28	26.04	23.01	22.35	24.73	12.55	25.49
C	-	-	-	-	-	-	-
Di	4.88	6.94	4.16	5.51	5.53	4.39	5.06
Hy	11.99	13.08	10.37	12.04	12.04	6.01	12.52
Mt	3.31	3.26	3.36	3.16	3.36	2.43	3.49
Il	1.48	1.42	1.56	1.29	1.56	1.20	1.70
Ap	0.42	0.25	0.44	0.21	0.35	0.35	0.51
Rb/Sr	0.09	0.09	0.13	0.11	0.11	0.15	0.09
DI	52.38	48.69	56.68	54.75	51.94	73.05	50.57

Sample #	9H	2I	4I	7I	9I	11I	12I
(Wt. %)							
SiO ₂	60.66	61.46	57.42	58.64	60.27	61.77	61.19
Al ₂ O ₃	16.25	15.93	17.36	17.06	15.64	16.47	16.89
FeO(T)	7.43	6.35	8.28	7.92	6.56	6.59	7.42
MgO	2.90	2.77	3.70	3.06	2.82	2.51	2.57
CaO	5.32	5.11	6.91	6.35	5.62	5.23	5.33
Na ₂ O	3.40	3.69	3.44	3.89	3.37	3.65	3.28
K ₂ O	2.74	2.33	2.37	2.00	2.19	2.70	2.46
TiO ₂	0.73	0.70	0.90	0.90	0.61	0.76	0.68
P ₂ O ₅	0.16	0.10	0.20	0.23	0.11	0.16	0.17
MnO	0.13	0.11	0.16	0.14	0.12	0.12	0.13
(ppm)							
Ba	662	630	661	687	925	746	1274
Cr	161	130	105	91	79	112	79
Zr	158	208	167	182	144	222	210
Sr	397	397	524	480	460	420	543
Rb	73	61	65	49	55	74	54
Y	28	27	24	28	22	32	25
Nb	7	7	5	-	3	4	-
Zn	76	71	92	95	63	81	91
Ni	-	15	24	-	9	9	-
V	146	117	148	145	134	99	128
TOTAL	99.94	98.82	100.97	100.42	97.55	100.18	100.42
CIPW							
Q	13.43	15.43	7.34	9.62	15.46	14.70	15.60
Or	16.97	13.77	14.01	11.82	12.94	15.96	14.62
Ab	28.77	31.22	29.11	32.92	28.52	30.89	27.87
An	20.99	20.02	24.93	23.18	21.08	20.58	24.20
C	-	-	-	-	-	-	-
Di	3.58	3.80	6.58	5.59	4.97	3.53	1.02
Hy	11.25	9.14	12.54	10.79	9.42	8.78	11.77
Mt	3.23	3.19	3.48	3.48	3.06	3.28	3.18
Il	1.39	1.33	1.71	1.71	1.16	1.44	1.30
Ap	0.37	0.23	0.46	0.53	0.25	0.37	0.40
Rb/Sr	0.18	0.15	0.12	0.10	0.12	0.17	0.13
DI	58.94	61.71	50.47	54.37	58.85	61.90	58.09

Sample #	5I	1D	2D	3D	4D	5D	6D
(Wt. %)							
SiO ₂	58.20	46.17	48.09	50.55	55.29	47.27	49.58
Al ₂ O ₃	17.65	17.75	12.21	18.88	16.91	22.13	10.34
FeO _(T)	8.35	11.87	8.47	7.32	8.29	4.52	11.77
MgO	3.88	6.60	9.68	8.66	5.27	10.11	11.86
CaO	7.31	12.31	19.44	9.71	6.54	9.90	12.39
Na ₂ O	3.53	2.23	0.81	2.61	3.94	2.64	1.48
K ₂ O	1.87	0.55	0.03	0.24	1.61	0.37	0.43
TiO ₂	0.76	1.06	0.30	0.46	1.23	0.07	0.63
P ₂ O ₅	0.08	0.11	-	-	0.21	-	0.03
MnO	0.15	0.18	0.15	0.12	0.15	0.06	0.23
(ppm)							
Ba	654	246	92	141	453	147	180
Cr	114	69	608	298	140	202	231
Zr	117	102	12	28	124	8	41
Sr	463	999	311	479	633	573	412
Rb	43	3	-	-	21	8	3
Y	28	17	12	11	19	3	22
Nb	4	3	1	-	5	-	-
Zn	87	84	8	60	83	45	105
Ni	16	25	93	159	69	372	109
V	191	364	322	110	129	28	141
TOTAL	101.99	99.18	99.52	98.79	99.66	97.25	98.98
CIPW							
Q	7.91	-	-	-	4.32	-	4.03
Or	10.94	3.32	2.26	1.45	9.63	2.26	2.54
Ab	29.53	19.28	23.07	23.12	33.72	23.07	12.52
An	26.49	37.61	49.00	39.53	23.97	49.00	20.30
C	-	-	-	-	-	-	-
Di	7.24	19.57	1.36	7.65	5.98	1.35	30.89
Hy	12.97	13.41	18.99	22.60	15.39	21.56	15.22
Mt	3.24	3.79	2.35	2.90	4.00	2.35	3.77
Il	1.43	1.52	0.14	0.89	1.72	0.14	0.49
Ap	0.18	0.26	-	-	0.49	-	0.07
Rb/Sr	0.09	0.01	-	-	0.03	0.01	0.01
DI	48.38	22.60	6.76	24.56	47.66	25.33	19.09

Sample #	7D	8D	9D	10D	11D	12D	1A
(Wt. %)							
SiO ₂	54.15	51.70	52.51	49.05	48.75	48.39	71.40
Al ₂ O ₃	17.01	17.22	8.48	8.33	9.50	9.35	16.35
FeO(T)	9.79	8.56	9.04	11.01	10.93	12.29	1.85
MgO	4.70	5.61	12.46	14.25	12.14	13.18	0.49
CaO	8.44	11.41	13.45	14.30	15.87	12.30	3.76
Na ₂ O	3.57	3.32	1.63	0.34	0.08	1.15	5.00
K ₂ O	1.03	0.18	0.18	0.22	0.08	0.26	1.06
TiO ₂	0.87	0.97	0.34	0.40	0.44	0.99	0.17
P ₂ O ₅	0.17	0.16	-	-	-	0.12	0.05
MnO	0.17	0.17	0.21	0.22	0.25	0.24	0.06
(ppm)							
Ba	432	74	140	89	48	94	409
Cr	770	153	1078	2037	2317	731	93
Zr	80	95	26	20	26	68	78
Sr	260	876	214	217	455	157	497
Rb	13	-	-	4	4	3	16
Y	25	23	17	12	20	24	8
Nb	-	5	-	3	-	7	2
Zn	114	90	121	120	156	134	36
Ni	9	120	117	478	571	162	-
V	254	158	223	202	181	266	9
TOTAL	100.11	99.51	98.59	98.58	98.57	98.49	100.34
CIPW							
Q	7.73	4.52	6.03	5.23			28.99
Or	6.09	1.06	1.06	1.30			6.26
Ab	30.21	28.09	13.79	2.88			42.31
An	27.35	31.55	15.29	20.55			18.33
C	-	-	-	-			0.26
Di	8.60	16.58	29.76	28.81			-
Hy	12.72	10.29	22.65	27.49			2.58
Mt	4.79	4.56	9.04	11.01			0.89
Il	0.36	0.36	0.45	0.47			0.32
Ap	0.39	0.37	0.25	-			0.12
Rb/Sr	0.03	-	-	0.02			0.03
DI	44.03	33.67	20.88	9.41			77.56

Sample #	2A	4A	5A	6A	7A	8A	26A
(Wt. %)							
SiO ₂	72.60	71.97	70.96	73.53	71.18	69.65	73.05
Al ₂ O ₃	16.57	16.72	17.41	14.92	14.24	16.04	15.16
FeO _(T)	1.69	1.55	1.66	1.35	1.21	1.76	1.34
MgO	0.59	0.56	0.64	0.56	0.32	0.70	0.57
CaO	3.95	4.13	4.57	2.83	2.35	4.32	2.91
Na ₂ O	4.77	4.89	4.88	4.63	4.45	5.07	4.80
K ₂ O	1.21	1.08	0.99	1.62	1.54	0.96	1.63
TiO ₂	0.19	0.16	0.18	0.11	0.07	0.17	0.11
P ₂ O ₅	0.05	0.02	0.05	0.03	0.02	0.06	0.03
MnO	0.06	0.06	0.06	0.06	0.05	0.06	0.06
(ppm)							
Ba	554	511	548	566	1019	443	608
Cr	109	106	93	195	191	157	192
Zr	86	93	99	56	59	83	60
Sr	587	594	639	373	429	634	387
Rb	16	15	13	22	31	13	24
Y	5	6	7	9	7	6	13
Nb	-	-	-	-	3	-	4
Zn	125	48	56	30	54	35	30
Ni	19	5	3	-	27	-	-
V	11	-	12	7	-	4	-
TOTAL	101.86	101.30	101.57	97.80	96.85	99.95	99.83
CIPW							
Q	30.47	29.25	27.64	33.11	33.70	26.39	31.42
Or	7.15	6.38	5.85	9.57	9.10	5.67	9.63
Ab	40.36	41.38	41.29	39.18	37.66	42.90	40.62
An	19.27	20.36	22.35	13.84	11.53	18.18	14.24
C	0.35	0.05	0.12	0.48	1.03	-	0.28
Di	-	-	-	-	-	2.35	-
Hy	2.67	2.52	2.78	2.44	1.78	1.88	2.46
Mt	0.82	0.75	0.80	0.65	0.58	0.85	0.65
Il	0.36	0.30	0.34	0.21	0.13	0.32	0.21
Ap	0.12	0.05	0.12	0.07	0.05	0.14	0.07
Rb/Sr	0.03	0.03	0.02	0.06	0.07	0.02	0.06
DI	77.98	77.04	74.78	81.86	80.46	74.96	81.67

Sample # 27A	10	70	80	90	110	50	
(Wt. %)							
SiO ₂	74.11	71.59	70.33	73.09	69.16	71.08	69.18
Al ₂ O ₃	14.74	14.66	14.38	14.16	14.31	14.24	14.12
FeO(T)	1.17	4.06	4.18	2.20	4.24	4.06	4.20
MgO	0.32	0.60	0.62	0.19	0.67	0.52	0.67
CaO	2.41	1.85	1.88	0.81	2.14	1.78	1.90
Na ₂ O	4.71	3.79	3.73	5.14	3.91	3.89	3.94
K ₂ O	1.58	4.80	4.84	4.18	4.64	4.80	4.11
TiO ₂	0.07	0.49	0.48	0.13	0.52	0.45	0.46
P ₂ O ₅	0.02	0.13	0.15	0.01	0.12	0.10	0.16
MnO	0.06	0.09	0.08	0.06	0.08	0.08	0.08
(ppm)							
Ba	1038	921	1000	333	960	1032	904
Cr	180	117	92	81	98	110	165
Zr	59	379	361	244	386	361	354
Sr	436	207	215	170	222	200	209
Rb	30	170	172	211	172	184	163
Y	7	67	59	59	62	61	60
Nb	3	25	22	55	25	28	25
Zn	29	91	65	62	58	72	91
Ni	-	22	18	10	24	22	-
V	5	7	27	-	10	25	3
TOTAL	99.41	102.31	100.91	100.12	100.05	101.27	99.76
CIPW							
Q	31.99	25.69	24.89	25.37	23.39	25.23	23.14
Or	9.67	27.88	28.50	24.74	27.57	28.17	28.65
Ab	40.74	31.48	31.42	43.52	33.23	32.65	33.57
An	14.28	8.18	8.15	3.21	7.81	7.15	6.67
C	0.28	0.17	-	-	-	-	-
Di	-	-	0.13	0.58	1.68	0.79	1.45
Hy	1.42	2.52	2.77	0.21	2.05	2.07	2.24
Mt	1.34	2.83	2.86	2.04	2.94	2.81	2.86
Il	0.09	0.91	0.92	0.25	0.99	0.85	0.88
Ap	0.07	0.29	0.35	0.02	0.28	0.23	0.37
Rb/Sr	0.06	0.82	0.80	1.24	0.77	0.92	0.78
DI	81.39	85.05	84.88	93.64	84.18	86.05	85.36

Sample # 60	100	30	40	5K	6K	3K
(Wt. %)						
SiO ₂	68.95	68.98	69.77	69.10	74.04	72.35
Al ₂ O ₃	14.20	14.11	13.97	14.21	13.27	13.40
FeO _(T)	4.20	4.16	4.06	4.36	2.21	3.15
MgO	0.66	0.70	0.66	0.66	0.06	0.16
CaO	1.95	1.91	1.82	1.90	0.88	1.12
Na ₂ O	3.76	3.71	3.76	3.85	3.28	3.26
K ₂ O	4.97	4.77	4.75	4.82	5.64	5.93
TiO ₂	0.46	0.44	0.45	0.46	0.14	0.24
P ₂ O ₅	0.12	0.17	0.13	0.14	-	-
MnO	0.08	0.07	0.08	0.09	0.04	0.05
(ppm)						
Ba	928	884	821	920	333	546
Cr	145	158	149	174	95	108
Zr	359	347	355	367	202	205
Sr	213	207	195	210	57	37
Rb	165	162	170	162	119	146
Y	58	55	66	61	59	70
Nb	28	28	30	29	11	17
Zn	76	88	92	97	50	56
Ni	-	-	1	-	-	16
V	16	18	1	24	-	2
TOTAL	99.59	99.26	99.70	99.83	99.67	100.46
CIPW						
Q	23.15	24.25	25.03	23.24	29.81	30.59
Or	29.66	28.56	28.31	28.70	33.54	34.09
Ab	32.09	31.77	32.06	32.79	30.81	28.43
An	7.24	7.84	7.26	7.29	2.62	4.21
C	-	-	-	-	-	0.03
Di	1.41	0.51	0.79	1.04	0.53	-
Hy	2.38	2.93	2.48	2.82	0.42	0.22
Mt	2.87	2.85	2.85	2.86	1.06	1.58
Il	0.88	0.85	0.86	0.88	0.26	0.27
Ap	0.28	0.40	0.30	0.33	0.02	-
Rb/Sr	0.77	0.79	0.87	0.77	2.12	1.32
DI	84.90	84.58	85.40	86.73	94.15	93.10

91.35

Sample #	4K	7K	8K	10K	11K	2L	6L
(Wt. %)							
SiO ₂	79.10	76.70	76.05	75.63	76.53	68.92	67.02
Al ₂ O ₃	10.41	13.56	12.51	13.15	12.61	13.53	13.98
FeO _(T)	1.72	2.27	1.98	1.93	1.62	5.56	5.68
MgO	0.09	0.08	0.07	0.11	0.09	0.66	0.98
CaO	0.69	0.81	0.77	0.96	0.73	2.20	2.49
Na ₂ O	2.74	3.40	3.38	3.60	3.29	3.30	3.58
K ₂ O	4.14	5.94	5.07	5.17	5.35	4.80	4.53
TiO ₂	0.10	0.15	0.11	0.12	0.10	0.68	0.74
P ₂ O ₅	-	-	-	-	-	0.29	0.25
MnO	0.03	0.03	0.03	0.03	0.02	0.09	0.09
(ppm)							
Ba	296	294	175	240	275	954	1120
Cr	140	98	107	108	123	65	57
Zr	226	239	193	207	196	338	442
Sr	38	52	32	45	42	227	245
Rb	92	129	119	118	120	101	89
Y	52	56	45	53	37	68	70
Nb	12	10	12	12	9	26	23
Zn	49	55	56	58	27	96	98
Ni	-	-	-	-	13	29	30
V	11	2	-	6	-	9	20
TOTAL	99.12	99.44	100.07	100.12	100.46	100.05	99.61
CIPW							
Q	46.11	31.40	36.28	32.51	35.07	25.72	22.18
Or	24.68	34.15	29.62	30.37	31.53	28.54	27.07
Ab	23.78	27.96	27.99	30.25	27.74	28.06	30.60
An	3.25	3.91	3.72	4.41	3.61	7.96	8.78
C	0.11	0.06	0.06	-	0.07	-	-
Di	-	-	-	0.25	-	0.92	1.63
Hy	0.24	0.19	0.34	0.16	0.22	3.63	4.35
Mt	0.07	1.31	0.50	0.54	-	3.18	3.28
Il	0.19	0.28	0.21	0.23	0.02	1.30	1.42
Ap	0.05	0.03	-	-	-	0.68	0.59
Rb/Sr	2.29	2.53	3.85	2.68	2.86	0.45	0.36
DI	94.56	99.44	93.89	93.13	94.34	82.31	79.86

Sample #	8L	11L	12L	4M	5M	7M	10M
(Wt. %)							
SiO ₂	68.21	64.92	64.11	75.59	76.01	75.53	74.35
Al ₂ O ₃	13.59	13.93	14.19	14.98	14.28	14.75	14.48
FeO _(T)	6.05	7.26	7.86	0.98	0.73	0.88	0.92
MgO	0.78	1.09	1.23	0.08	0.09	0.11	0.06
CaO	2.29	2.91	3.11	0.69	0.71	0.68	0.56
Na ₂ O	3.48	3.55	3.77	4.04	4.17	3.61	4.87
K ₂ O	4.76	4.37	4.13	4.43	4.39	5.09	4.44
TiO ₂	0.76	1.00	1.13	0.01	0.01	0.01	0.01
P ₂ O ₅	0.35	0.42	0.52	-	0.02	0.01	0.02
MnO	0.10	0.12	0.12	0.14	0.12	0.27	0.27
(ppm)							
Ba	914	1211	1107	125	124	474	109
Cr	75	52	45	113	108	109	127
Zr	395	508	461	48	37	26	27
Sr	230	286	284	63	62	108	31
Rb	109	86	65	94	90	176	122
Y	77	77	66	35	30	60	72
Nb	26	21	18	-	-	-	-
Zn	112	164	141	-	1	20	20
Ni	15	1	-	26	22	28	25
V	11	27	28	6	-	-	-
TOTAL	100.62	99.86	98.93	100.99	100.59	101.07	100.81
CIPW							
Q	23.54	19.64	17.68	33.20	33.14	33.12	27.79
Or	28.17	26.09	24.43	25.95	25.82	29.82	26.28
Ab	29.45	30.32	31.87	33.85	35.09	30.25	41.18
An	7.39	9.24	9.62	3.39	3.37	3.28	2.65
C	-	-	-	2.26	1.41	2.07	0.75
Di	1.40	2.14	2.00	-	-	-	-
Hy	3.47	5.97	6.86	-	-	-	-
Mt	3.28	3.66	4.11	0.34	0.27	0.76	0.55
Il	1.44	1.92	2.16	0.02	0.02	0.02	0.01
Ap	0.81	0.98	1.21	-	0.05	0.02	0.04
Rb/Sr	0.48	0.30	0.23	1.50	1.48	1.64	3.95
DI	81.16	76.06	73.99	93.01	94.06	93.19	95.25

Sample #	11M	12M	13M	16M	18M	19M	1F
(Wt. %)							
SiO ₂	75.41	74.61	73.76	73.43	74.70	74.49	75.56
Al ₂ O ₃	14.54	14.68	14.68	14.97	13.87	13.85	12.25
FeO _(T)	0.72	0.99	1.01	0.92	0.39	1.03	1.59
MgO	0.04	0.18	0.17	0.17	0.19	0.20	0.17
CaO	0.41	0.67	0.74	0.64	0.72	0.63	0.45
Na ₂ O	4.48	4.19	4.11	3.94	3.87	3.59	3.88
K ₂ O	4.39	4.14	4.14	4.51	4.65	4.96	5.00
TiO ₂	0.01	0.01	0.01	0.01	0.03	0.07	0.10
P ₂ O ₅	0.04	0.04	-	0.01	0.02	0.01	-
MnO	0.11	0.32	0.32	0.12	0.01	0.02	0.04
(ppm)							
Ba	112	91	94	134	450	245	215
Cr	128	142	153	148	158	136	197
Zr	30	35	25	40	83	101	162
Sr	10	20	24	61	145	38	15
Rb	194	100	101	94	114	197	90
Y	16	86	80	27	153	77	30
Nb	6	-	-	-	-	32	9
Zn	13	25	7	21	10	30	46
Ni	-	-	-	-	7	-	-
V	-	6	-	-	-	-	-
TOTAL	100.23	99.88	98.99	98.79	98.59	98.95	99.15
CIPW							
Q	34.38	32.84	32.49	32.04	33.11	33.36	42.56
Or	25.65	24.52	24.74	27.02	27.93	29.67	31.17
Ab	34.80	35.50	35.14	33.76	33.25	30.72	20.91
An	1.74	3.07	3.71	3.15	3.49	3.09	0.87
C	2.27	2.18	2.11	2.50	-	1.47	0.32
Di	-	-	-	-	-	-	-
Hy	-	-	-	0.43	0.48	0.50	0.15
Mt	0.23	0.98	0.99	0.33	-	-	2.43
Il	0.02	0.02	0.01	0.02	-	0.02	0.36
Ap	0.07	0.09	-	0.02	0.05	0.02	-
Rb/Sr	7.18	5.00	4.40	1.57	0.79	5.19	6.52
DI	94.83	92.86	92.37	92.82	94.29	93.76	94.63

Sample #	5F	8F	9F	10F	2F	3F	4F
(Wt. %)							
SiO ₂	77.55	73.01	74.70	75.85	75.56	75.49	76.36
Al ₂ O ₃	11.41	13.61	13.32	12.96	12.25	10.81	10.54
FeO _(T)	2.71	3.53	2.63	2.71	1.59	2.97	2.84
MgO	0.02	0.15	0.09	0.12	0.17	0.15	0.13
CaO	0.12	0.96	0.63	0.76	0.45	0.19	0.05
Na ₂ O	2.27	4.09	4.11	4.10	3.88	2.74	2.96
K ₂ O	6.58	5.43	5.46	5.00	5.00	5.17	4.88
TiO ₂	0.12	0.34	0.24	0.23	0.10	0.11	0.12
P ₂ O ₅	-	0.01	-	-	-	-	-
MnO	0.03	0.09	0.06	0.07	0.04	0.03	0.03
(ppm)							
Ba	297	515	469	632	215	117	23
Cr	124	87	94	113	197	156	162
Zr	841	726	499	453	162	989	1626
Sr	80	62	43	68	15	12	8
Rb	137	69	66	60	90	129	107
Y	12	53	41	36	30	26	27
Nb	-	20	13	10	9	-	-
Zn	101	155	80	86	46	122	94
Ni	19	28	6	5	-	-	-
V	-	-	-	-	-	-	-
TOTAL	101.03	101.45	101.41	101.97	99.15	97.86	98.18
CIPW							
Q	38.46	25.95	28.36	30.85	33.14	39.76	40.82
Or	38.64	31.78	31.93	29.08	29.85	31.35	29.52
Ab	19.07	34.24	34.38	34.11	33.14	23.77	25.61
An	0.59	2.70	1.74	2.14	1.24	0.97	0.25
C	0.33	-	-	-	-	0.37	0.30
Di	-	0.58	0.48	0.63	0.79	-	-
Hy	0.28	0.32	-	-	-	0.43	0.83
Mt	2.33	2.64	2.01	2.31	-	2.39	2.40
Il	0.23	0.64	0.45	0.43	0.06	0.21	0.23
Ap	-	0.02	-	-	-	-	-
Rb/Sr	1.72	1.13	1.58	0.89	6.42	10.76	15.33
DI	96.18	91.98	94.66	94.04	96.13	94.88	95.94

Sample #	7F	11F
(Wt. %)		
SiO ₂	73.47	75.27
Al ₂ O ₃	12.83	12.25
FeO _(T)	2.91	2.19
MgO	0.16	0.15
CaO	0.49	0.42
Na ₂ O	3.15	3.41
K ₂ O	5.63	5.30
TiO ₂	0.20	0.12
P ₂ O ₅	-	-
MnO	0.06	0.03
(ppm)		
Ba	443	247
Cr	145	186
Zr	513	315
Sr	54	28
Rb	77	86
Y	35	45
Nb	3	3
Zn	59	76
Ni	-	-
V	2	-
TOTAL	99.06	99.26
CIPW		
Q	32.53	34.31
Or	33.71	31.63
Ab	26.98	29.11
An	2.46	2.10
C	0.67	0.14
Di	-	-
Hy	0.74	0.38
Mt	2.50	1.36
Il	0.39	0.23
Ap	-	-
Rb/Sr	1.42	3.06
DI	93.22	95.06

TABLE A 2.3
TRACE ELEMENT DATA

Sample No.	Sc	Co	Sb	Cs	La	Ce	Nd	Sm	Eu	Tb
T4D	19.9	29	0.08	0.83	14.7	31	15.9	3.7	1.15	0.55
T1A	2.8	2.4	0.1	0.27	9.6	17.2	7.5	1.44	0.46	0.18
T3G	16.9	15.4	0.11	1.59	19.2	41	22	4.8	1.23	0.75
T1H	18	17.8	0.20	0.85	23	50	25	5.4	1.47	0.80
T11I	16.3	13.2	0.42	2.5	24	53	25	5.1	1.17	0.85
T6J	6.8	4	0.15	0.84	25	50	23	4.6	1.20	0.70
T4M	7.8	0.32	0.1	1.80	15.7	29	12.5	2.4	0.22	0.56
T8F	4.1	1.11	0.1	0.2	104	215	96	17.4	1.68	2.4
T9O	7.2	4.3	0.08	1.57	68	142	59	11.3	1.65	1.86
T8K	1.21	0.69	0.09	0.83	51	120	57	11.4	0.75	1.60
T8L	10.3	7.4	0.17	1.32	103	224	100	18.5	2.6	2.8
Sample No.	Dy	Ho	Er	Yb	Lu	Hf	Ta	W	Th	U
T4D	3.7	0.72	1.6	1.70	0.29	2.6	0.60	2	2.8	1.06
T1A	1.4	0.18	0.72	0.60	0.10	2.3	0.28	2	1.5	0.27
T3G	4.5	0.79	2.8	2.3	0.39	4.3	0.34	2	1.16	0.58
T1H	4.4	0.82	2.3	1.94	0.32	4.1	0.50	2	1.25	0.68
T11I	5.5	0.98	2.4	2.7	0.47	5.4	0.76	2	5.9	2.1
T6J	4.0	0.83	1.9	2.7	0.44	6.7	0.92	2	3.8	1.59
T4M	4.6	1.13	3.3	4.3	0.72	2.3	0.43	2	4.1	1.60
T8F	12.3	2.1	5.3	5.7	0.98	15.6	1.28	2	12.4	1.82
T9O	10.7	2.16	5.1	5.6	0.91	10.2	3	2	18.2	2.9
T8K	8.6	1.74	4	4.4	0.70	7.1	1.52	2	10.6	2.6
T8L	14.7	2.9	7.3	7.6	1.22	11.1	2.2	2.5	13.2	2.9

APPENDIX 3

MICROPROBE ANALYSES OF MINERALS

BIOTITE

	T9G 1	2	3	4	5	6	7	*T12G 1	*T7I 1
SiO ₂	35.34	35.99	35.98	35.86	35.58	35.84	35.77	37.80	37.3
TiO ₂	2.15	2.15	2.13	2.11	2.21	2.20	2.13	1.92	1.98
Al ₂ O ₃	15.69	15.71	15.66	15.48	15.46	15.67	15.60	16.10	16.4
FeO _(T)	21.76	22.54	22.11	22.21	22.12	22.36	22.11	23.46	22.66
MnO	0.41	0.38	0.36	0.43	0.40	0.42	0.43	0.39	0.32
MgO	9.70	10.06	10.09	10.14	9.95	9.91	9.98	9.94	10.51
CaO								0.54	0.69
Na ₂ O	0.09	0.10	0.10	0.10	0.10	0.12	0.16	0.18	0.23
K ₂ O	9.44	9.34	9.38	9.36	9.06	9.46	9.32	9.42	8.89
Cl	0.13	0.09	0.10	0.10	0.16	0.16	0.15		
F	0.25	0.37	0.27	0.27	0.25	0.18	0.29		
Total	94.96	96.73	96.17	96.04	95.29	96.31	95.94	99.75	98.98

* minerals analyzed by XRF and for which ferrous determinations are done with the modified Wilson cold acid method.

*T12G: FeO = 17.48
Fe₂O₃ = 5.98

*T7I: FeO = 17.72
Fe₂O₃ = 4.94

HORNBLLENDE

	*T7I 1	T7I 2	T9I 1	2	3	4
SiO ₂	44.1	43.06	40.66	40.42	40.90	41.43
TiO ₂	1.19	1.76	0.69	0.80	0.81	0.80
Al ₂ O ₃	10.6	10.44	12.39	12.76	12.28	11.17
FeO(T)	20.51	19.34	22.11	22.0	22.4	21.88
MgO	8.57	9.08	7.51	7.45	7.61	8.12
MnO	0.49	0.54	0.44	0.47	0.50	0.42
CaO	11.4	11.52	11.26	11.47	11.57	11.3
Na ₂ O	1.09	1.07	1.30	1.31	1.34	1.29
K ₂ O	1.09	1.08	1.53	1.53	1.53	1.32
Cl		0.12	0.19	0.13	0.11	0.12
F		0.03	0.16	0.05	0.13	0.18
Total	99.04	98.05	98.23	98.38	99.16	98.04

* XRF analysis.

FeO = 12.23
 Fe₂O₃ = 8.28
 FeO(T) = 20.51

EPIDOTE

	T9G 1	2	3	4	T7I 1	2	3
SiO ₂	37.53	37.37	37.95	38.03	37.91	37.65	37.8
TiO ₂	0.14	0.14	0.07	0.13	0.14	0.10	0.06
Al ₂ O ₃	23.72	23.92	23.97	23.70	24.01	24.05	24.53
FeO(T)	12.17	12.31	12.15	12.46	11.98	11.63	11.46
MgO	0.06	0.03	0.02	0.03	0.04	0.03	0.03
MnO	0.19	0.22	0.15	0.22	0.19	0.23	0.21
CaO	22.82	22.94	23.11	21.82	22.63	22.97	22.86
Na ₂ O	0.04	0.03	0.02	0.04	0.06	0.08	0.06
K ₂ O	0.09	0.02	0.03	0.04	0.05	0.29	0.04
Cl	0.03	-	0.01	0.01	0.01	0.06	-
F	-	0.12	-	0.20	0.05	0.03	0.17
Total	96.79	97.10	97.49	96.69	97.06	97.12	97.22

PLAGIOCLASE

	T9G 1	2	3	4	5	6	7	8
SiO ₂	59.45	59.28	59.18	59.35	59.34	59.73	59.95	59.09
Al ₂ O ₃	25.71	24.98	25.75	25.20	25.58	25.48	25.15	25.34
CaO	7.34	6.78	7.44	6.94	7.26	7.08	6.76	7.12
Na ₂ O	7.41	7.60	7.31	7.52	7.46	7.3	7.75	7.49
K ₂ O	0.16	0.15	0.19	0.19	0.10	0.19	0.10	0.08
Total	100.08	98.78	99.87	99.20	99.75	100.02	99.72	99.11

PLAGIOCLASE

	T9G 9	10	11	12	T7I 1	2	3	4
SiO ₂	59.47	59.31	59.96	59.07	60.83	58.07	59.52	59.42
Al ₂ O ₃	25.22	25.47	25.77	25.36	24.61	26.14	25.86	25.10
CaO	6.93	7.18	7.27	7.14	6.11	8.02	7.45	6.85
Na ₂ O	7.56	7.46	7.58	7.44	8.21	7.02	7.49	7.70
K ₂ O	0.17	0.15	0.08	0.14				
Total	99.36	99.57	100.06	99.14	99.75	99.24	100.32	99.06

PLAGIOCLASE

MICROCLINE

	T7I 5	6	T7M 1	2	3	4	1	2
SiO ₂	59.34	54.09	64.53	64.70	65.97	65.84	64.19	64.48
Al ₂ O ₃	25.15	27.50	21.67	21.53	22.13	21.75	18.22	18.23
CaO	6.90	10.07	2.82	2.67	2.86	2.58	0.05	-
Na ₂ O	7.66	5.58	10.07	10.15	10.07	10.21	1.06	0.80
K ₂ O					0.34	0.26	15.12	15.64
Total	99.05	97.24	99.09	99.05	101.38	100.65	98.64	99.15

MICROCLINE

PLAGIOCLASE

	T7M 3	4	1	T8F 2	3	4	5	6
SiO ₂	64.28	64.35	68.81	67.83	66.21	67.05	67.25	66.96
Al ₂ O ₃	18.18	18.24	19.74	20.08	20.5	20.75	20.61	21.09
CaO	-	0.03	0.23	0.74	1.47	1.47	1.31	1.78
Na ₂ O	0.76	0.81	11.63	11.29	10.75	10.90	10.99	10.79
K ₂ O	15.64	15.56	0.16	0.14	0.17	0.17	0.12	0.12
Total	98.86	98.99	100.57	100.08	99.1	100.34	100.29	100.75

MICROCLINE

	T8F							
	1	2	3	4	5	6	7	8
SiO ₂	67.03	67.46	65.58	64.03	67.05	67.25	67.33	64.61
Al ₂ O ₃	20.58	20.56	18.60	18.11	20.75	20.61	20.58	18.36
CaO	1.34	1.22	0.04	-	1.47	1.31	1.27	0.07
Na ₂ O	10.90	10.33	0.44	0.7	10.9	10.99	10.78	0.53
K ₂ O	0.23	1.27	16.44	15.67	0.17	0.38	0.52	16.04
Total	100.09	100.84	101.11	98.51	100.34	100.29	100.49	99.6

PLAGIOCLASE

MICROCLINE

	T8k							
	1	2	3	4	5	1	2	3
SiO ₂	68.54	68.27	65.14	67.90	67.36	66.21	66.4	65.94
Al ₂ O ₃	19.78	19.80	22.10	19.79	19.93	18.73	18.81	18.70
CaO	0.33	0.41	3.03	0.48	0.73	0.01	0.03	0.05
Na ₂ O	11.55	11.44	9.92	11.39	11.23	0.90	0.74	0.81
K ₂ O	0.16	0.20	0.27	0.14	0.11	15.92	16.2	15.97
Total	100.36	100.13	100.46	99.71	99.37	101.78	102.19	101.46

MICROCLINE

	T8K	
	4	5
SiO ₂	65.77	65.36
Al ₂ O ₃	18.72	18.48
CaO	0.10	-
Na ₂ O	0.68	0.59
K ₂ O	16.10	16.19
Total	101.36	100.61

MUSCOVITE

GARNET

	T7M						
	1	2	3	4	5	1	2
SiO ₂	45.13	45.58	45.23	45.76	46.08	36.29	36.32
Al ₂ O ₃	33.37	33.55	33.06	32.13	31.65	20.83	20.83
FeO	4.24	3.99	4.36	4.28	4.70	25.61	25.83
MgO	0.51	0.49	0.54	0.52	0.50	1.21	1.24
CaO	-	-	-	-	-	0.84	0.89
MnO	0.02	0.01	0.03	-	-	14.86	14.75
Na ₂ O	0.41	0.48	0.41	0.27	0.17	-	-
K ₂ O	10.67	10.42	10.36	11.0	11.11	-	-
F	0.25	0.23	0.24	-	-	-	-
Total	94.65	94.81	94.27	93.96	94.21	99.64	99.87

APATITE

	T9G		
	1	2	3
CaO	55.34	55.37	53.99
P ₂ O ₅	42.58	42.71	41.75
Cl	0.10	0.13	0.08
F	3.49	3.48	3.27
Total	101.51	101.69	99.09

APPENDIX

U-Pb ANALYTICAL METHODS

U-Pb zircon geochronology laboratory procedures employed at the geochronology laboratory of the Geological Survey of Canada are given in Parrish et al. (1987). These include mineral preparation and selection, addition of tracer (spiking), chemical dissolution, chemical separation and purification, mass spectrometry, error analysis, and linear regressions.

Zircons were separated from 1-2 kg specimens by standard crushing, grinding, heavy liquid, and magnetic separator, followed by sieving in ethyl alcohol into a number of size fractions. All minerals were selected for analysis by handpicking in ethyl alcohol. The criteria used in mineral selection, particularly zircon, optimize the clarity and diagnostic morphology of the crystals, the lack of cracks and inclusions, and the general absence of cored grains unless these are specifically of interest. A portion of some zircon mineral separates was mounted in FEP Teflon or epoxy grain mounts, polished to an optical finish, and etched in HF vapour to reveal details of the internal structure. Optical microscopy, photomicrography, and scanning electron photomicrography were utilized to observe the morphological aspects which assist in the interpretation of U-Pb zircon results. The zircon fractions were then air abraded following the technique of Krogh (1982) using pyrite as a medium in the abrasion cells. Abraded zircons were leached 15 minutes in luke warm 3N HNO_3 to remove abrasion residues prior to the final mineral selection stage.

The clean chemistry laboratory for U-Pb has a source of high efficiency particulate filtered air. Critical chemical work and weighing was done in specially designed, mainly polypropylene, hoods with their own separate source of the high efficiency particulate air filtered room. Laboratory air is not recirculated, and the laboratory has a minimum of exposed metal. Minerals are weighed using aluminum boats in a micro-balance which has a quoted precision of ± 0.2 micrograms. In practice when weighing samples, experience indicates the uncertainty to be ± 0.5 -1.0 micrograms. Tracer solution is dispensed into Teflon dissolution crucibles and weighed by difference using a balance which has a quoted precision of ± 0.02 mg. Following chemical separation and purification, Pb and U solutions are collected and evaporated in 30ml polymethylpentene beakers. The collection beakers are pre-cleaned by boiling in reagent grade 6N HCl followed by refluxing for at least 1-2 hours on a hot plate using 6-8N HCl which was double distilled in a quartz sub-boiling still.

Zircon is dissolved in 0.1 to 0.5 ml of high purity 48% HF and 0.01 ml of high purity 70% HNO_3 are added; capsules are then sealed and heated in an oven at 230 - 240°C for 30-60 hours. Following dissolution is evaporation to fluorides and the salts are dissolved by heating in an oven at 210°C overnight. Subsequent element separation employs 3.1N HCl anion exchange resin column chemistry outlined by Roddick et al. 1987. Isotopic composition of Pb and U are measured on a Finnigan-MAT 261 variable multicollector mass spectrometer in the static mode as described by Roddick et al. 1987. Roddick (1987) has

detailed a numerical procedure for propagating the uncertainties in all the relevant components in the data reduction of the isotopic ratios and ages. Blanks for typical sample amounts of 0.01 to 1mg are less than 10 pg U and 6-40 pg Pb. Discordant U-Pb results are assessed on concordia plots using regression techniques. Criteria for assigning a U-Pb age in the case of zircon inheritance with scattered data are based on the convergence of data of nearly concordant analyses near the lower intercept of the concordia curve; otherwise linear regression is used on data with least inheritance or where very nearly concordant relative to other data and defines the lower intercept.

APPENDIX 5

Rb-Sr ANALYTICAL METHODS

A5.1 DISSOLUTION AND PREPARATION OF THE WHOLE-ROCK POWDERS FOR Rb-Sr GEOCHRONOLOGY

Rb and Sr concentrations were determined by X-ray fluorescence for whole-rock samples, and by isotope dilution using ^{87}Rb and ^{84}Sr spikes for minerals. Separations were carried out by conventional ion exchange techniques as follows:

- 1) 200 mg of rock powder is taken into solution in a Teflon evaporating dish using 20 ml of HF and 3-4 drops of HClO_4 .
- 2) The sample is taken to dryness on a hot plate at low temperature setting (100°C).
- 3) The sample is heated strongly (about 250°C) to drive off any remaining acid.
- 4) After the dish has cooled, the sample is taken into solution with about 5 ml of 2.5N HCl.
- 5) The sample is taken into solution with 5 ml of 2.5N HCl, transferred into a centrifuge tube and is centrifuged for approximately 10 minutes.
- 6) The solution is then transferred to a disposable beaker and taken to dryness at 100°C .
- 7) Then 2 ml of 2.5N HCl is added to the dried sample and the solution is centrifuged again for 10 minutes.

The sample is now ready to be loaded on to the columns.

A5.2 SEPARATION OF Rb AND Sr FROM WHOLE-ROCK POWDERS FOR MASS SPECTROMETRY

Apparatus: Teflon tube 22 cm in length with an internal diameter of 1.1 cm, Bio-Rad AG50W-X8 cation exchange resin (200-400 mesh, hydrogen form). The resin is equilibrated with 2.5N HCl. The columns are packed with 13 ml of the resin, cleaned with 100 ml of 6N followed by 2.5N HCl.

- 8) The prepared sample (after step 7 above) is loaded directly on top of the resin.
- 9) After the sample is completely absorbed into the resin, it is washed with 5 ml of 2.5N HCl in 1 ml portions to ensure that the sample begins to pass through the resin.
- 10) When all of the acid is absorbed, the resin is washed with 30 ml of 2.5N HCl in clean, labelled disposable beakers.
- 11) Rb is then collected with 10 ml of 2.5N HCl in clean, labelled disposable beakers.
- 12) The columns are washed with 25 ml of 2.5N HCl.
- 13) The Sr is eluted with 15 ml of 6.2N HCl.
- 14) The resin is then washed with 20 ml of 2.5N HCl.
- 15) The resin is also washed with 15 ml of 6.2N HCl.
- 16) The collected solutions are put on the hot plate.
- 17) When the Rb and Sr aliquots are evaporated to 1-2 ml, they are transferred into clean, small Teflon dishes and taken to dryness. These samples are now ready for analyses in the mass spectrometry laboratory.

Summary

Load sample	2 ml	2.5N HCl
wash	5 ml	2.5N HCl
wash	30 ml	2.5N HCl
collect	10 ml	2.5N HCl Rb
wash	25 ml	2.5N HCl
collect	15 ml	2.5N HCl Sr

A5.3 MASS SPECTROMETRY

Isotopic analyses were made on a Finnigan-MAT 261 multicollector, solid source mass spectrometer operated in the static mode. The sample is taken into solution with 0.1 ml of high purity doubly distilled water (DDW). This solution is loaded on the outgassed filament for analysis. Sr was loaded with phosphoric acid on a Ta single filament, and Rb was run as chloride on Re double filaments. Masses ^{86}Sr , ^{87}Sr and ^{88}Sr were measured simultaneously on three different channels, and the observed intensities were corrected for the gain of the amplifier and the baseline readings.

A typical analysis, which requires about 60 to 90 minutes, has a 2 sigma of about 0.002% (Bell et al., 1987). The reproducibility of the isotopic ratios, estimated on the basis of replicate analyses of samples and chemical standards, is about $\pm 0.004\%$ of the quoted values. The Eimer and Amend Sr standard gives a value for $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.70802 ± 0.00002 ; and the NBS 987 Sr standard, 0.71023 ± 0.00003 . Sr ratios are normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$. Blanks for typical analyses averaged 2 ng for Sr, and 0.01 ng for Rb.

4 of / de 4

MICROD

GEOLOGICAL MAP OF THE GORE - GAMBELLA GEOTRANSVERSE AREA

WESTERN ETHIOPIA Sheet NC36 - 16 (P,V,O & U)

Scale

LEGEND

Q Alluvium	hbg Hornblende-biotite gneiss and amphibolite
Tv Tertiary volcanics	big Biotite and quartz-feldspathic gneisses
PRECAMBRIAN TO LOWER PALEOZOIC ROCKS	
SEDIMENTARY AND METAMORPHIC ROCKS	
mgs Metasedimentary schists, predominantly metawacke and pelite containing andalusite and staurolite, minor volcanoclastic debris flow deposit	gt Late- to post-kinematic; biotite-hornblende granodioritic rocks, mainly granitic monzogranite and quartz monzonite
mvs Metavolcanic schist mainly hornblende-biotite-plagioclase-epidote and quartz-sericite schists, and mylonite (my)	gb Late-kinematic gabbro and hornblende
mb Marble	mgt Syn-kinematic metagranite
qms Quartz-muscovite schist and conglomerate	Pre-kinematic and/or syn-kinematic metamorphosed (gabbro/mg) quartz diorite (mgd), granodiorite (mgd) and tonalite (mtn)
msq Muscovite-bearing schist and gneiss, and minor quartzite	
csq Calc-silicate gneiss	
gsg Garnet-synkinematic and garnet-cordierite-gedrite gneisses	

- Kinematic indicators**
- Shear zone (a) brittle (b) ductile
  - Shear sense from (a) extensional (b) contractional
  - Shear sense from (a) extensional (b) contractional
  - Shear sense from (a) extensional (b) contractional

- Minor folds related to shear zone (with trend, plunge of axis, strike and dip of axial plane)**
- Minor folds related to shear zone (with trend, plunge of axis, strike and dip of axial plane)

- o Town/village**
- Road
  - Major river
  - Perennial/intermittent river

- Sample location**
- Birba west central east
  - Goma
  - Bure
  - Haya
  - Baro
  - Garnet-muscovite granite sheets
  - Mao
  - Bonga

SYMBOLS

- Geological contact (a) definite (b) inferred
- Foliation trend
- Strike and dip of foliation
- Fault
- Trace of antiform, synform

Sample locations for U-Pb geochronology
T8F, T3G, T8I, T9Q, T6J

35°E

Figure 2.3: Sketch map of the intrusive units and sample localities

SCALE

0

5 kilometres

-8°N

BONGA

BIRBIR

(TH)

(TI)

GOMA
(TJ)

BURBI
(TD)

7D

8D

BONGA

(TH)

(TL)

(TK)

(TM)

(TG)

BARO
(TF)

35°E

1H
2H

3H

(TH)

4H

5H

6H

8H

7D

8D

9D

11D

10D

12D

35'E

END

1|7·0|2·90

FIN