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GRAVITY FIELD STUDIES IN THE MAIN ETHIOPIAN RIFT AND SOUTHERN AFAR

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**By
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**ADDIS ABABA UNIVERSITY
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ABSTRACT

All available data, over 3000 stations, obtained from the EIGS and the Geophysical observatory (Dr. Abera's Ph.D work) were reduced to sea level with a uniform crustal density of 2.67 gm/cm^3 . Effects of Bouguer masses were calculated applying the simple Bouguer correction. Terrain correction was not applied in which case its effect was treated as systematic error in computing the over all mean square error of the simple Bouguer anomalies at each station. Theoretical gravity field was computed by means of the international gravity formula of 1967 (GRS67) and tied to the IGSN71. The accuracy of the Bouguer anomaly at each station is calculated to about $\pm 2.5 \text{ mgal}$. The regional and residual gravity field were calculated from the Bouguer anomaly map by means of a computer software.

The compiled Bouguer anomaly map shows a strong correlation between the regional pattern and topographic feature of the survey area. Owing to higher elevations, the adjacent plateaus are marked by negative Bouguer anomalies, the minimum of magnitude -270 mgals being located around Debreberhan for the study area. The maximum gravity values occur in the Afar depression corresponding to the relatively lower elevations. The main Ethiopian rift including Afar is characterized by a relatively positive anomaly as compared to the plateaus. The high on the rift accounts for the excess mass at depth on which the crust lies relative to the plateaus where a mass deficiency is assumed.

Gravity values on the main Ethiopian rift are generally less than those in the Afar. This accounts for the thinning of crust material towards the Afar along the rift, or it can equally be explained as by a thickening of a denser material beneath the crust (probably the anomalous mantle). The margins separating the rift from the plateaus are marked by steep gravity gradients with their magnitude representing the slope of the escarpment. Steep gradients characterize the transition of gravity field from the plateaus to the rift floor.

CHAPTER 1

1. INTRODUCTION

1.1 LOCATION

The area under survey is located (Fig. 1) in the northern part of the Main Ethiopian rift and the Southern Afar rift including the adjacent plateaus. It lies between 7.15° and 12° N latitude and 37° and 42° E longitude covering an area of 296,260 sq. kilometers. The compiled gravity data covers almost all the area with the exception of the Somalia plateau which lacked data. The northern limit of the main Ethiopian rift where the boundary with Afar lies is at about latitude 9.45° N.

1.2 GENERAL OUTLOOK/OBJECTIVES OF THE SURVEY

It is known that the East African rift system (EAR) has an active geological feature with recent tensional faulting and volcanism, and has long been recognized as a continental extension of the world rift system (Rothe, 1954; Ewing and Heezen, 1956). It comprises a series of rift zones (Rosendahl et al., 1986) and extends for some 3200 Km from the Afar triple junction at the Red sea - Gulf of Aden intersection to the Zambezi river in Mozambique. The depth and disruption to the Lithosphere and Asthenosphere is estimated to be at least 100 Km (Brown & Girdler, 1980).

Since several years gravimetric measurements have been carried out in the East African region for geological and geophysical applications to investigate the crustal and upper mantle structures. The Ethiopian rift system (the Afar depression, the Main Ethiopian rift, and Southern Rift) forming part of the EAR system provide a valuable platform on which to study this process within the crust and upper mantle.

The main Ethiopian rift (MER) is widely accepted to be due to the relative motion of the Nubian and Somalia plates with tensional and normal faulting with distribution of seismic, volcanic, and tectonic activity and zones of mineralization. The Afar depression, further northeast, is the triple junction of the Red sea, the Gulf of Aden and the Ethiopian part of the East African rift.

The geologic and tectonic features of the study region have drawn the attention of several researchers, and efforts have been made to give solutions to existing problems applying different geophysical methods. The application of gravimetry in conjunction with other geophysical methods has been effective in delineating areas of maximum crustal attenuation with recent volcanic and tectonic activities which correspond to volcanic complexes, and in outlining basement and concealed tectonic lines which are useful for exploration purposes. This becomes successful only when reliable gravity data of standard accuracy can be gathered.

Therefore, the themes of main scientific concern are:

- the study of the geophysical environment of the region for natural hazard reduction through monitoring, to help the development and settlement programs of the country.
- the use of geo-informatic data, that will be generated or have already been generated, to better understand the geology of the region in the efforts made to exploit mineral and energy resources.

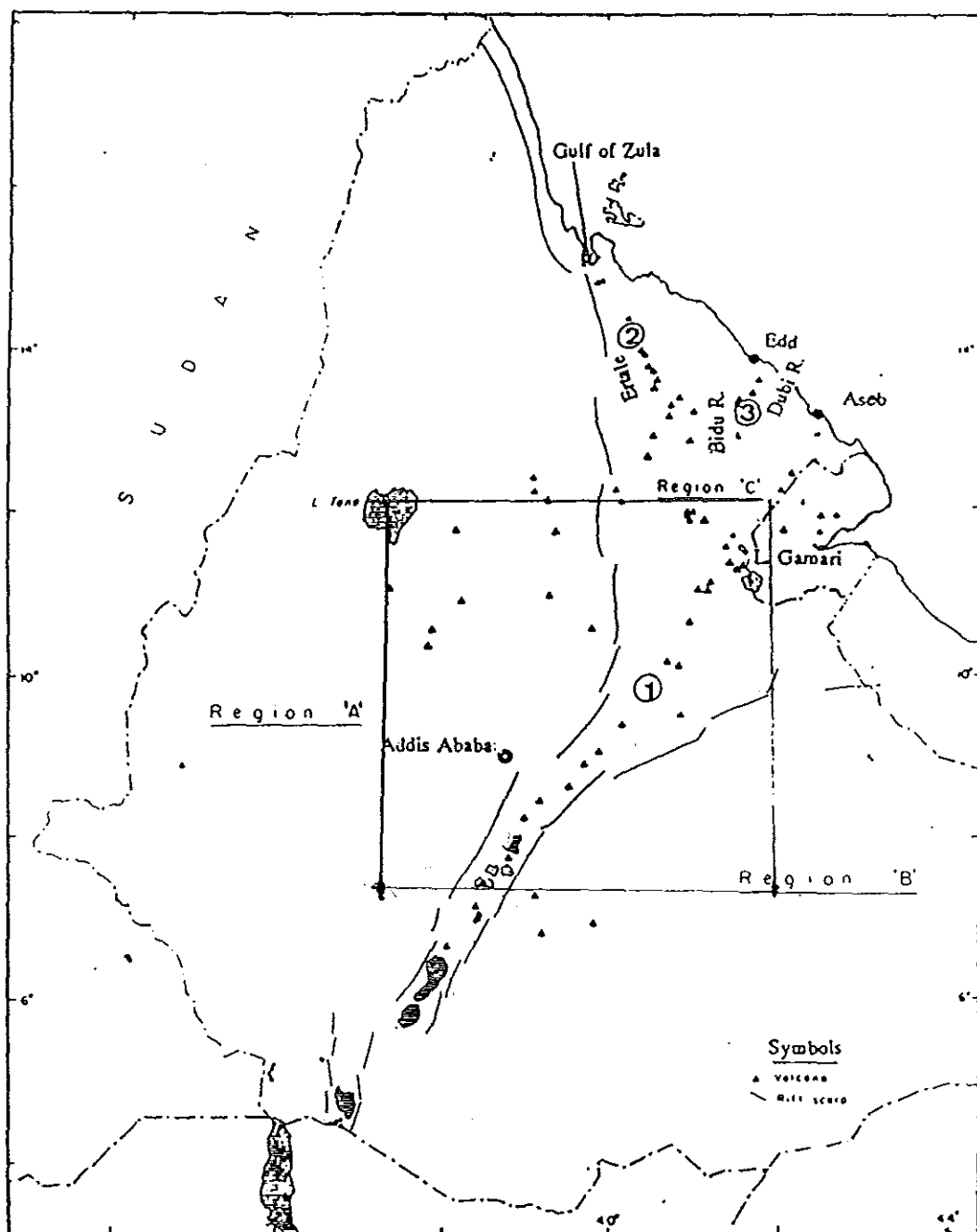


Fig.1. Location Map of the Study Region

- The aim of this thesis work is:

- to compile a new gravity map of the rift valley and associated adjacent plateaus by homogenizing all the available heterogeneous data obtained from the sited sources following standard reduction procedures.
- to delineate the major geologic and tectonic structures of the rift system and the neighboring regions by adopting the previously conducted geological and geophysical (gravity, seismic refraction and seismicity) survey results as a constraint.
- to present a preliminary qualitative interpretation (location, shape, ... etc.) of gravity anomaly signatures revealed by the main structural features of the region.

In general, the production of these maps of small (1:250,000) and large scale (1:50,000) is believed to give a first hand information in the distribution, location and orientation of the subsurface structures related to crustal deformation, mineral and energy resource deposits. Moreover, the compiled data is believed to contribute to the national challenge in the compilation of a gravity catalogue of the country.

CHAPTER 2

2. GENERAL GEOLOGY AND VOLCANO-TECTONICS

The African rift system has been the subject of intense geological study and debate since it was first described by Suess (1891) and Gregory (1896). Its role in the recently formulated theory of plate tectonics remains rather enigmatic, but the system is generally held to mark an extensive and complex zone of Cainozoic, crustal dilatation within broader regions of continental swell uplift (Baker et al., 1972).

The East African rift system meets the Red Sea and Gulf of Aden spreading zones at the Afar triple junction (Fig. 2), and is susceptible to plate tectonics analysis. Plate tectonic analysis yields an integrated rate, and further assumes that opening of all three rifts branching from Afar started together, about the beginning of the Miocene, according to McKenzie et al. (1970), but in the early Tertiary according to Mohr (1972a). In fact the influence of the African rift system at the Afar triple junction may have become significant only since the late Pliocene (Mohr, 1970b, 1972a).

The Ethiopian rift valley owing to its junction in the Afar depression, is an important part of this structure, where sea floor spreading is taking place. It is an area of graben fracturing showing an active geological feature with an intense tensional faulting and volcanism.

Rifting is believed to have started in the lower Miocene, and by the Pliocene it was a shallow trough with an in filling of silicic volcanic products erupted from centers close

to the rift margins (Baker et al., 1972). Rifting is followed by a regional uplift of the Afro - Arabian region. This can be understood from Almond (1986) in which he concluded that before upper Miocene times the Afro - Arabian dome was not recognizable.

Three main episodes of uplift were recognized in Ethiopia at different times, and the voluminous trap flood basalt flow was associated with the earliest episode of late Eocene. The present form of the rift is ascribed to the last uplift which occurred during late Pliocene to early Pleistocene.

The Ethiopian rift valley is a graben with an average width of about 70 - 80 Km in its central part. It is limited, to the west, by the Ethiopian plateau and, to the east, by the Somalia plateau. Its development is widely accepted to be due to tensional movements of the Ethiopian and Somalia plateau. A great number of step faults produces a total difference of altitude of more than 1000 m between the top of the plateau and the floor of the rift valley.

2.1 THE RIFT MARGINS AND ADJACENT PLATEAUS

The margins of the main Ethiopian rift continue southward from Afar without significant break. It is merely convenient to take the Yerer-Gugu cross - rift lineament as the separating line. The ESE-NNW Yerer volcano line (Scoriaceous basalt superimposed on Pliocene anorthoclase trachytes) marks a faulted and possibly warped boundary between plateau rocks to the north and rift ignimbrites and sediments to the south (Mohr, 1967c).

The eastern margin of the rift curving southward from Mt. Gugu terminates at mountain Bada, and resumes some 15 Km to the west of Assela, hence strong NNE-SSW faulting continues south to form the eastern margin of the rift above lake Zway and lake Langano. The northwards continuation of the rift margin faulting at Assela is directly along the wonji fault belt which also suffers a small dextral displacement at the Guraghe - Chilalo line.

On the western side, the rift margin is formed by the dominating NE-SW facing single in double escarpment. The Guraghe horst is faced by opposite faulting developed strongly enough to form a true marginal graben of Afar type. The marginal graben is about 6 Km wide and its western boundary faulting is associated with upper Pleistocene - Holocene Scoriaceous basalt, as well as explosion craters (Mohr, 1967).

The eastern escarpment of the MER is characterized for all its length by step faults with an important throw compared to the distance from one block to the other. The western escarpment shows in its NE sector an abrupt displacement, sometimes exceeding 1500 m (Mt. Guraghe), between the top of the plateau and the rift floor, whilst in its SW sector the main faults have a small down throw and progressively die out, so that, at its SW end the structural limit plateau rift floor becomes only physiographic feature.

2.2 THE ETHIOPIAN RIFT SYSTEM

2.2.1 PHYSIOGRAPHY

The general trend of the main Ethiopian rift is NNE but changes are observed in the study region compared to the southern part of the MER. These are shown by small

displacement of the eastern escarpment at latitude $8^{\circ} 20' N$ (Kazmin and Berhe, 1978) and westward displacement of the western escarpment at about $9^{\circ} N$ latitude.

The average altitude of the plateau on both sides of the rift is about 2500 m a.s.l while the floor of the rift valley gently decreases from an altitude of 1600 m a.s.l at Lake Awasa to an altitude of 1300 m a.s.l just north of the Bosseti - Gudda volcano. Proceeding northwards from Bosseti - Gudda volcano, the floor of the rift valley continues to decrease down to about 250m a.s.l at its end, that is, the southern shore of Lake Abbe. Proceeding southwestward from lake Awasa, the altitude of the rift floor suddenly reaches 1100m a.s.l at Lake Abbaya -about 50 Km south of Awasa - and it remains at that level up to the southern shore of Lake Chamo, where the MER terminates at Konso upland. As shown in the map (Fig. 2) four lakes of tectonic or volcano - tectonic origin (Zway, Langano, Abiata, Shalla) occur in the region described. Also other lakes of volcano - tectonic origin (Awasa, Chamo, Abbaya) occur outside the study area.

The Afar depression is triangular in shape and is separated from the elevated areas of the Ethiopian plateau to the west, the south east plateau and Ogaden to the south, and the Danakil horst to the east by fault controlled escarpments. At the south western apex of the triangle the East African rift joins the depression. In the plateaus adjoining the Afar depression Mesozoic and Precambrian rocks outcrop, covered on the Ethiopian plateau by extensive Tertiary volcanic flows. By contrast with the MER Afar tends to be featureless and monotonously flat plane, relieved only by young volcanic cones and fresh graben.

2.3 THE MAIN ETHIOPIAN RIFT

2.3.1 GEOLOGY OF THE MER

The prominent geological feature (Fig. 3) in the region is the occurrence of a pre-trappean structure close the western rift margin. It exposed a 150 m thick mass of Precambrian crystalline basement rock overlain by Mesozoic marine sequences and a tertiary ignimbrite. The Precambrian rocks consist of biotite gneiss with quartzo feldspathic pegmatite veins and minor migmatites (Di Paula and Berhe, 1979). Sediments of Lacustrine origin are well represented; they cover large areas of the rift with sometimes great thickness. They are the only non - volcanic formation in this portion of the Ethiopian rift covering an area of 4000 Km² and whose thickness is sometimes considerable, ranging from about 40m in Bulbula river and 50 m in Boru and Maki rivers upto more than 100m between Mojo and Koka.

Two volcano - Tectonic units can be identified (Meyer, etal, 1975) in the Northern part of the MER: the Nazret series and the Wonji series. The Nazret series extends from the Turkana rift in Kenya through MER floor to South Afar and the South Eastern Escarpment (Baker, Mohr and Williams, 1972). It is composed of Silicic rocks such as Ignimbrites, Rhyolites and Pumice. The age of this unit is 5.2 m.y. and its known thickness is not more than 150 m. The Nazret series is overlain by Lacustrine and fluvial sediments.

The younger unit, the Wonji series, is of Pleistocene to Holocene age and lies with an unconformity of the Nazret series. The Wonji series mainly consisting of Basaltic flows

as well as some Silicic and intermediate rocks are generally located near the Wonji fault belt (Meyer, et al, 1975; Mohr, 1971). The rock units are called the Wonji group (Kazmin and Berhe, 1978) and include all the rift volcanics formed after the major episode of rift faulting following the accumulation of the Bofa basalt. Pantelleritic volcanic centers represents eruptions of Perkaline Rhyolites, Trachytes, Pumice and Obsidian. They are aligned "en echelon" along the segments of the Wonji fault belt. These centers in the study region include (from the south) Aluto, Tullu Moye, Kone, Gedmesa, Fantale and Dofan volcanoes.

The Wonji series is divided into two parts. The lower and upper parts (Meyer, et al, 1975). The lower part characterized by the Welenchiti basalt and its accompanying basaltic rocks adheres strongly to the NNE-SSW fractures and fissures. The upper part lies over the Welenchiti basalt with an unconformity, and a thick layer of loam is situated between the two.

2.3.2 VOLCANO - TECTONICS OF THE MAIN ETHIOPIAN RIFT (MER)

The MER floor consists of three caldera - related basins connected by the volcano-tectonically active Wonji fault belt (WFB) (Woldeghebriel et al., 1990): Zeway-Langano-Abiyata, Shala-Awasa and Bilate river drainage basin (Fig). It is also marked by a persistent belt of intense fresh faulting which has been termed as the Wonji fault belt (Mohr, 1967). The faults are short and of normal type, and are notably associated with tensional fissures first recognized by Gibson (1967b) from the Fantale region. The Wonji

fault belt is frequently, but not always, axial to the rift and is dextrally displaced along the same cross-rift lineament which displays the rift margin. The WFB is also a line along which recent lavas and Ignimbrites have erupted and the whole volcano-tectonic association is one suggestive of crustal tension acting across the rift.

The westward displacement is related to the latitudinal Yerer fault (Mohr, 1967a) although this transverse fault is not a surface exposure. At the latitude of the eastern displacement, the rift escarpment runs north east. This change in angle is related to the reactivation of the pre-rift tectonic lines and the course of rifting (Kazmin et al., 1978). These transverse tectonic lines are also inferred to follow the rectilinear pattern of the Meki and Awash river course (Di Paula and Berhe, 1979).

In the south, the group of Wonji faults follow a NNE-SSW direction. In the north, this trend of Wonji faults cut across the NE-SW trend of the MER indicating the development shearing effect (Gibson, 1969; Gibson and Tazief, 1970). Meyer, et al (1975) however stated the main tectonic stress hitherto has been tensional and shearing pattern can be observed in outcrops.

In the northern Ethiopian-South Afar rift sector, the orientation of the en echelon structure as a whole as well as the offsets of axial zones (progressively displayed eastward moving towards the north) and the orientation of the rift shoulders are probably related to the complex movement of the Somalia and Nubian plates.

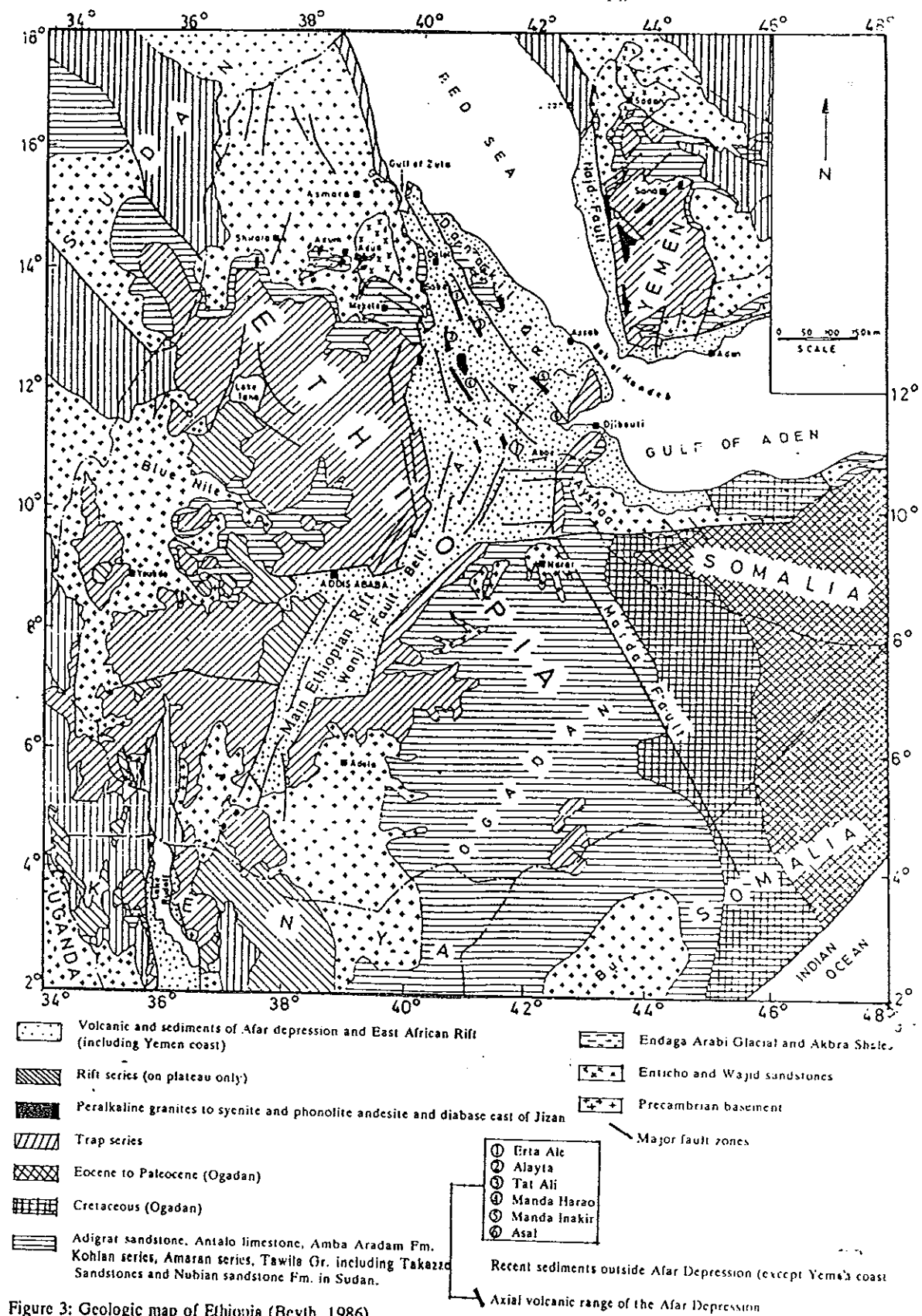


Figure 3: Geologic map of Ethiopia (Beyth, 1986).

2.4 THE AFAR DEPRESSION

2.4.1 GEOLOGY OF THE AFAR DEPRESSION

The rift structures are frequently marked by central volcanoes (Ayelu, Amoissa, Yagundi, Gabilema) which erupted significant amounts of Silicic products. The floor of the depression is covered by several series of quaternary volcanics with some Pleistocene Lacustrine deposits. Beneath the cover of Pliocene - Quaternary volcanics and sediments, most of Afar is flooded by a great thickness of flood Alkali basalts, extruded during the Paleocene down warping of the rift system. These basalts, together with the older cover rocks, are cut by intense fault belts of probable late Pleistocene age, formed during a major tectonic paroxysm in the evolution of the Ethiopian swell - rift system.

In central Afar, Aden series basalts cover an extensive area south of lake Julietti; they are also found at the northern end of the Danakil horst and salt plain, and in the Djibouti republic. In southern Afar these basalts are restricted in their occurrence as in the main Ethiopian rift, being found as small patches principally associated with the Wonji fault belt (Mohr, 1967).

2.4.2 VOLCANO - TECTONICS OF AFAR

The junction of the Kenyan rift and the Afar depression and its effect on the fault trends is clearly evident, showing the characteristics of triple junction with normal and sheared pattern of deformation (Courtillet, 1980).

In northern Afar the early phase of the rift began between 25 and 23 m.y ago. This phase features a climax in volcanic activity as well as geochemical change in the volcanic products. In the southern Afar, the initial phase of rifting can not be clearly determined, probably beginning at approximately the same time as that of northern Afar.

The culmination of tectonic activity occurred between 9 to 11 m.y ago. This tectonic phase known as chorora, is regarded as responsible for the present geological configuration of the southern Afar. The evolution of the graben was accompanied by significant sinking, which becomes more intense moving northward.

A progressive process of crustal attenuation with transition from continental to an oceanic type of rift, characterizes the transition from the northern Ethiopian to the Afar rift. Being the site of intense tectonic activity, with crustal thinning, Asthenosphere uprise and associated volcanism, both rift branches are certainly the site of a thermal regional anomaly. The association of the observed geothermal manifestations (hot springs, geysers, fumaroles) on the surface of both rift branches is clear illustration of this phenomena.

In the southern part of lake Abbe a very intense E-W fault belt extends eastward from the Wonji fault belt. Upthrows are dominantly to the south, but the variability permits the existence of a large graben in the northern part of the E-W fault belt. A huge margin of this belt against the sediments of a once more extensive Lake Abbe. The E-W belt curves eastward to an ENE-WSW trend (Gulf of Aden trend), but becomes largely obliterated by the ESE-WNW faults of the Djibouti republic, and indeed further south merges with this faulting. There is therefore no surface faulting extending inland WSW-

wards from the Gulf of Tajura directly towards Lake Abbe. Only south of latitude 11.14° N (at the longitude of the Gulf of Tajura) does weak Gulf of Aden faulting appear.

In south-western Afar the northwards - flowing Awash river follows a narrow but widening belt between the Ethiopian Plateau - Afar margin and the Wonji fault belt (and its branch north from Amo-Issa). Indeed, between Dofan volcano and Lake Hertale the Wonji fault belt abuts against the plateau - Afar margin. North of Lake Hertale and the clearly expressed Ajelu - Amoissa cross lineament, the faulting of the Awash valley is of two trends: NNW-SSE parallel to the Ethiopian - Afar margin, and ENE-WSW as some linear and very persistent cross - rift lineament: the geysers and hot-springs south of Tendaho are situated on one of these cross lineaments. The strongest of these cross rift lineaments extend from Karakore east to the E-W fault belt immediately south of Lake Abbe.

The silicic centers of the Wonji fault belt are situated where there is an intersection from a cross rift lineament, and their calderas or craters are elongated in the direction of the cross rift lineament. The cross-rift lineaments trend WNW-ESE in the main Ethiopian rift and the Lake Tana rift, but whilst this trend also occurs in the southern Afar, the dominant trend in Afar is WSW-ENE, perpendicular to the Red Sea rift.

followed into south Afar where it changes and becomes parallel to the Red sea trend. It is offset en echelon from the spreading area of the Red sea, which is marked by an alignment of gravity maxima.

Mohr and Gouin (1967/68) concluded that the main Ethiopian rift is in gravity 'high' as compared to the adjacent plateaus, and the Free-air values for the rift floor range between 0 and -50 mgals. They tentatively identified the cause of the rift 'high' is the underlying silicic magma chambers and their extensive volcanic effluvia (welded tuffs, lavas, pumice), and with thick Lacustrine sediments, all deposited in the subsiding proto-rift, pre-middle Pleistocene trough.

Again Mohr and Rogers (1966) related the gradient over the northern Ethiopian plateau, to the change in crustal thickness and/or to the thickness of underlying anomalous material striking towards the center of the Ethiopian swell. "The southern limit of this regional gradient must occur near the latitude of Addis Ababa, coinciding with the southward introduction of westward downwarped Trap series basalts in southern Shoa and northern Kaffa".

Gravity surveys in Afar (Mohr, 1967) indicate that Free air values are about zero, without either the positive Bouguer values of the central Red sea or the negative values of the east African rift system and is thought to be attributed to flood basalts which have filled the Afar downwarp, By contrast with the thick silicics under the rift yielding a gravity 'low'.

Crustal density models (Alemu, 1992) in the study region demonstrate under compensation in the plateau areas and overcompensation in the depression at shallow depths. The study suggested that the compensation of the Ethiopian and Kenyan topographic Domes are due, at least in part, to crustal thickening. It also suggested that there are significant variations in the crustal structure along the rift axis.

3.2 SEISMICITY OF THE MER AND AFAR

Many parts of the MER system are seismically active with seismic energy release often taking place in form of intermediate and small earthquakes. Regarding seismicity, Gouin (1979) has thoroughly discussed the seismic activity of the whole Ethiopian rift system and neighboring regions prior 1977. Laike and Fekadu (1981) summarized the seismicity of the Ethiopian rift system based on instrumental and historical data. They suggest that the western escarpment of the MER system is more active than the Eastern Escarpment.

The westward shift of epicenters in the southern most rift tends to bridge the seismicity gap between the Ethiopian side of the East African rift and the western rift (Fig. 4) the seismicity gap between the Ethiopian rift and the Eastern rift is evident from the seismic data available (Shah, 1986).

An earthquake swarm took place near Dofane and Fantale volcanic centers in 1981. Prior to the swarm, a micro earthquake survey had been conducted by Molnar et al (1970). During this survey the region had very small seismic activity.

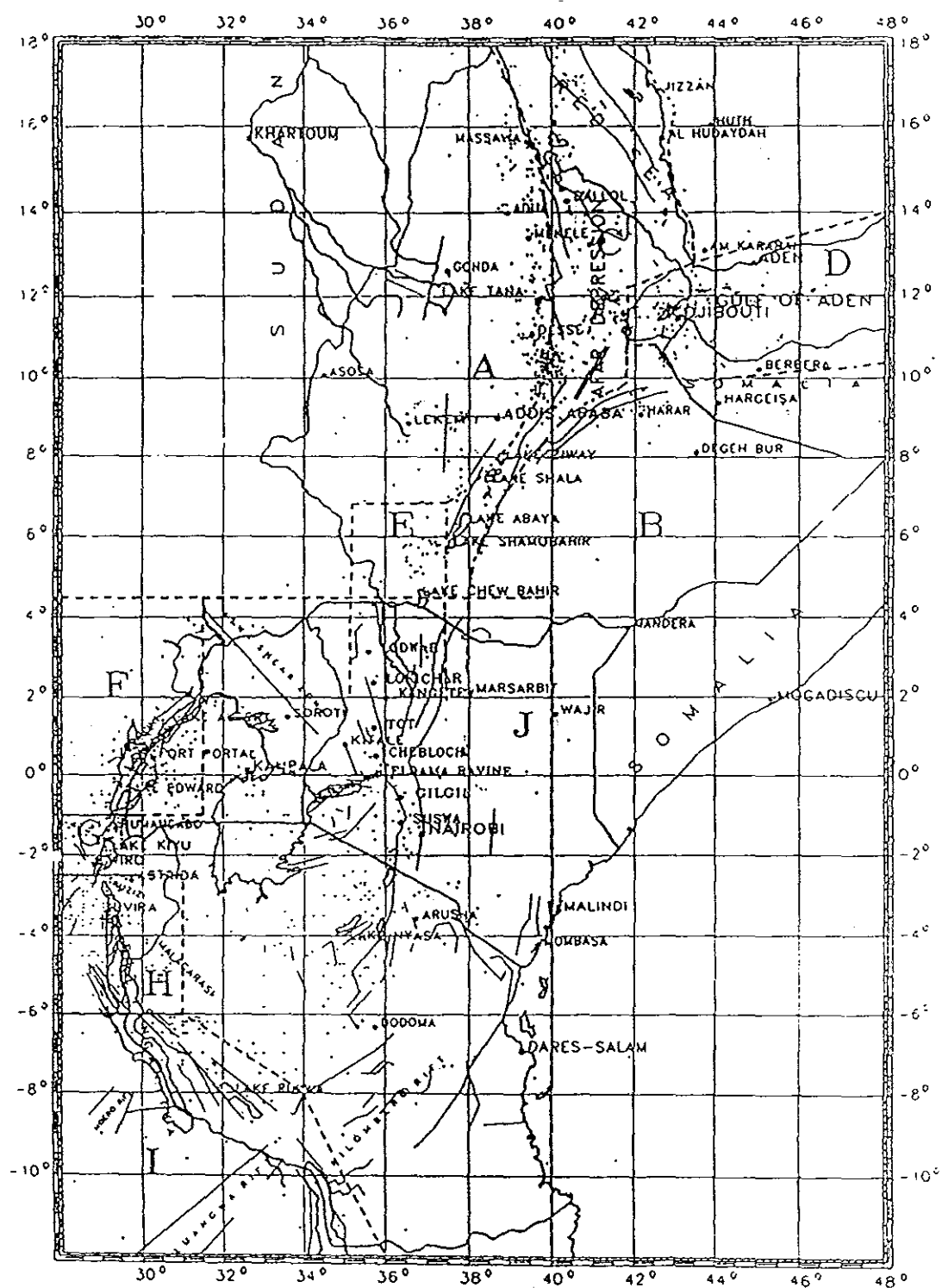


Fig.4. Seismicity in the East African Rift System

Seismicity and focal mechanism studies of the East African rift system (Sykes, 1967, Fairhead and Girdler, 1971; Fairhead and Stuart, 1982; Shudofsky, 1985) show that the rift system is a constructive plate boundary which is rather complex and forms a series of interconnecting rifts through eastern and southern Africa. These studies show that the eastern and western rifts of the East African rift system are characterized by continuous belts of normal faulting and graben structures. However, they differ from each other in that the eastern rift is thought to have a connection with the Gulf of Aden and the Red sea whereas much of the western rift appears to be older and is often assumed to be unrelated to the ocean ridge system.

Results of Kebede and Kulhanek, (1991) show that greater focal depths for the MER are found for earthquakes in the southern part of the East African rift rather than in the northern part.

3.3 SEISMIC REFRACTION RESULTS

Refraction profiles (Fig. 5) were measured (Berckhemer et al., 1975) along the main trends of crustal thinning and one west of Addis Ababa in order to compare results within the depression with a normal continental crust. The crustal structure of the Afar depression as obtained from the refraction profiles selected by the author departs from the normal continental crust in both thickness and velocity distribution in the crust and upper mantle. The results show an upper crust with a velocity of 6.1 to 6.2 km/s being observed on all profiles. It is however thinner than normal compared with the thickness of the upper crust on the plateau (18 to 20 km). The total thickness less than that of a continental crust being over 20 km over most of the area. However one should note that even in the north,

near Dallol, where the thickness of the crust is reduced to 14 km, 6.1 km/s upper crust is still present.

The crustal thickness under the Southern extension of the Danakil Horst is somewhat greater, but not significantly so. The crust throughout the depression is underlain by an upper mantle with an abnormally low velocity 7.4 m/s. The unusual structure of the crust could not be explained in terms of intrusions which form dykes and sills and extrusion in the Afar depressions. They follow fault lines and must have reached the surface during the breakup of the crust due to extension (Morton and Black, 1975)

The crustal structure of Djibouti area (Ruegg, 1975) is similar to the rest of the Afar area, having a Granite layer overlying a 6.8 km/s lower crust (referred to as an upper mantle by Ruegg. This layer overlies a Low velocity (7.1 to 7.4 km/s) upper mantle. He has noted a higher than usual poisson's ratio of 0.28 to 0.33 compared with 0.25 to 0.26 for old oceanic and continental crusts. Sereale (1975) has found a high V_p/V_s ratio across the Afar depression. These results indicate a partially melted upper mantle material.

The isostatic Moho-depth map (Alemu, 1992) was computed to check the regional isostatic trend of the Main Ethiopian Rift and the adjacent plateaus, according to the Airy-Heiskanen model of compensation. The seismic results (Berckhemer et al, 1975) and the 2-D gravity models reveal that the normal mantle beneath the Western Plateau lies at a depth of 38-42 km. Elevations 2000m were taken to correspond to the plateaus and those of -2000m to the rift.

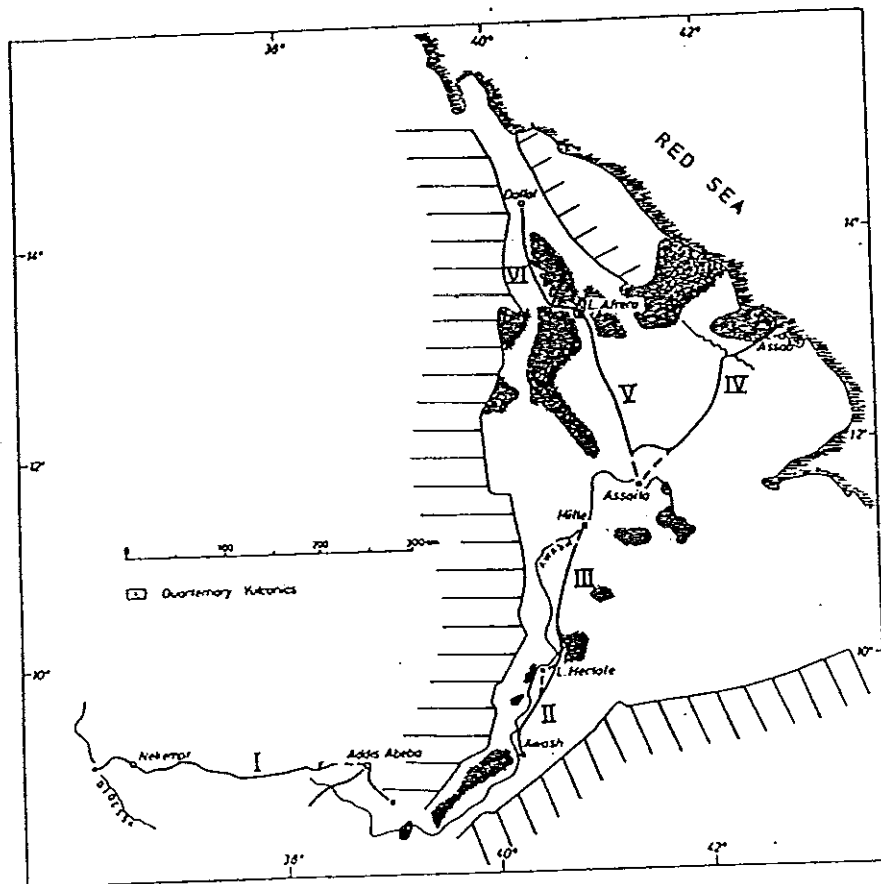


Figure 5: Location of the seismic refraction profiles on the Western Ethiopian plateau, the northern part of the Main Ethiopian Rift and the Afar depression (from Berckhemer, et al. 1975)

CHAPTER 4

4. THE GRAVITY FIELD AND GRAVIMETRY

The method involves the measurement at the surface of small variations in the gravitational field. Small differences or distortions in that field from point to point over the surface of the earth are caused by any lateral variation in the distribution of mass in the earth's crust. Therefore, if geologic movements involve rocks of different densities, the resulting irregularity in mass distribution will make corresponding variation in the intensity of gravity. The measured variations are interpreted in terms of probable subsurface mass distributions, which in turn are the basis for inferences about probable geologic conditions.

4.1 BASIC PRINCIPLES

The basis of the method is Newton's law of gravitation. This states that every particle of matter exerts a force of attraction on every other particle, this being directly proportional to the product of the masses and inversely proportional to the square of the distance between them. i.e.

$$F = GM_1M_2/R^2 \dots\dots\dots(1)$$

Where F is the force between the two particles of masses M_1 and M_2

R - their separation and

G - the universal gravitational constant ($6.672 \times 10^{-11} \text{m}^3 \text{s}^{-2} \text{kg}^{-1}$).

Let M_2 is replaced by M which is equal to the mass of the earth ($6 \times 10^{24} \text{kg}$), and R equal to the radius of the earth (6371 Km), and the M_1 mass of a body on the earth's

surface. Then, the attraction force of the earth per unit mass acting on the body is given by

$$g = GM/R \dots\dots\dots (2)$$

Where g -is the gravitational acceleration and its unit is the 'gal' ;

$$1 \text{ gal} = 1 \text{ cm/ s}^2$$

The possible forces acting on a body of mass m (Fig. 6) on the surface of a rotating spherical Earth (Heiskanen and Moritz, 1967) are: the attraction force of the earth, the centrifugal force (due to the rotation of the earth about its axis) and the coriolis force (for a body in motion).

The attraction force is given by:

$$F = - GMm R/R^3 \dots\dots\dots (3)$$

The centrifugal force, as an inertial force in a rotating earth is given by

$$f_{cg} = - m\omega \times (\omega \times R) \dots\dots\dots (4)$$

and the z-component is

$$(f_{cg})_z = -m\omega^2 R \cos\phi$$

where $\omega = 7.292115 \times 10^{-5}$ rad/sec is the angular velocity of the earth. It is known to a high degree of accuracy from astronomy.

The coriolis force, if the body is moving on the surface of the Earth with velocity V is given by

$$f_{cor} = 2m\omega \times V \dots\dots\dots (5)$$

and the z-component is

$$(f_{cor})_z = 2mv\sin\phi$$

It is evident from Eq. 4 that the centrifugal force is maximum at the equator and zero at the poles and it opposes the attractive force of the Earth and from Eq. 5, the coriolis force is zero for a body at rest ($V=0$) on the surface of the Earth. By definition, acceleration due to gravity (γ) is the resultant of the gravitational (g) and centrifugal (z) accelerations.

i.e.,

$$\gamma = g + z$$

$$\gamma = -GM R/R^3 + \omega \times (\omega \times R) \dots\dots\dots (6)$$

The representation of the gravity field and related computations are simplified if we consider the scalar quantity 'Potential' instead of the vector quantity 'acceleration'. Since

$$\text{Curl } g = 0, \text{ Curl } z = 0 \dots\dots\dots (7)$$

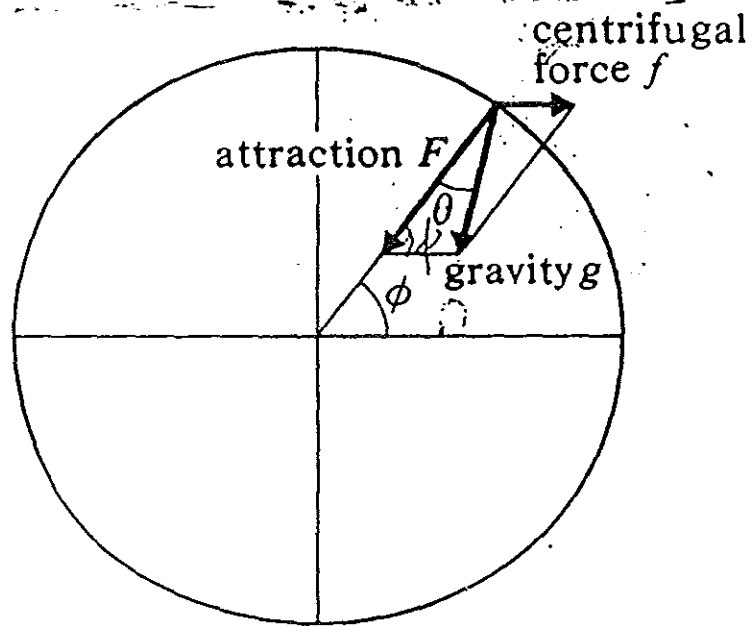


Fig.6. Forces Acting on a Body, on the surface of a Spherical Earth

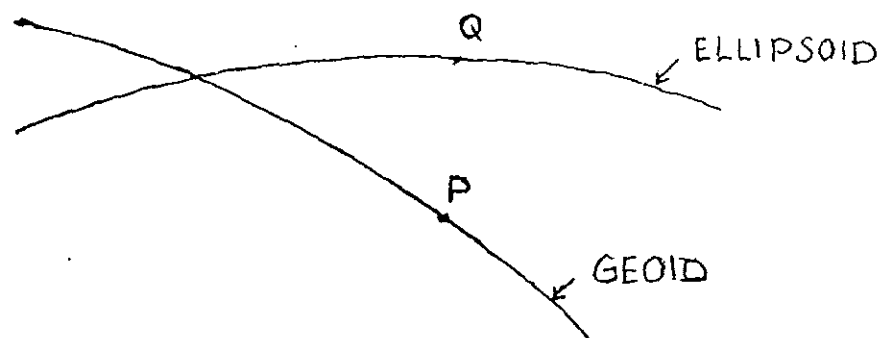


Fig.7. The Geoid and Ellipsoid

Corresponding potentials V and Φ exist for the gravitational and centrifugal fields with the relationships,

$$g = \text{grad } V, z = \text{grad } \Phi \dots \dots \dots (8)$$

The gravitational potential of the earth (v = volume of the earth), considering a mass element dm and density ρ can be expressed as a function of position 'r' as

$$V(r) = G \int \frac{dm}{r}$$

and the centrifugal potential is given by

$$\Phi = \frac{1}{2} \omega^2 d^2$$

where d is the distance perpendicular to the axis of rotation.

Sometimes it is convenient to use cylindrical coordinates. since the volume element in the Cartesian coordinates, $dv = dx dy dz$ is transformed in cylindrical coordinates as $r dr d\phi dz$, the expression for the gravitational potential becomes,

$$V = G \rho \iiint r dr d\phi dz$$

In spherical coordinates, $dx dy dz = r^2 \sin \theta dr d\phi d\theta$; hence we've

$$V = G \rho \iiint r \sin \theta dr d\phi d\theta$$

The acceleration in the direction of the z -axis (that is, the only direction in which g can be measured directly) is given by

$$g_z = dV/dz = -G \rho \iiint (z/r^3) dr d\phi dz, \text{ (cylindrical)}$$

and,

$$g_z = G \rho \iiint z \sin \theta dr d\phi d\theta = -G \rho \iiint \sin \theta \cos \theta dr d\phi d\theta, \text{ (spherical)}$$

4.1.1 POTENTIAL FIELD EQUATIONS

The divergence theorem (Gauss' theorem) when applied to the gravity field over a region of space is equivalent to the integral of the outward normal component of the field over the surface enclosing the region. Mathematically,

$$\int_v \nabla \cdot \mathbf{g} dv = \int_s \mathbf{g}_n ds \dots\dots\dots (9)$$

If there is no attracting matter contained within the volume, the integral are zero, and

$$\nabla \cdot \mathbf{g} = 0 \dots\dots\dots (10)$$

i.e the potential satisfies Laplace's equation in free space. But since the gravitational force is the gradient of scalar potential, then

$$\nabla \cdot \mathbf{g} = \nabla \cdot \nabla V = \nabla^2 V = 0 \dots\dots\dots (11)$$

If on the other hand, there is a particle of mass 'm' within the volume and, in particular, if we consider it at the center of a spherical surface of radius 'r', then the integral

$$\int_s \mathbf{g}_n ds = -(Gm/r^2)(4\pi r^2) = -4\pi Gm. \dots\dots\dots (12)$$

This result holds regardless of the shape of the surface and the position of the particle m within the surface. If the surface encloses several particles of total mass M, we can write

$$\int_v \nabla \cdot \mathbf{g} dv = \int_s \mathbf{g}_n ds = -4\pi GM.$$

If this volume v is made very small, enclosing a point on the region, we may remove the integral sign to get

The splitting of the gravity field into a "normal" and a remaining small "disturbing" field considerably simplifies the problem of its determination (Heiskanen and Moritz, 1967). Therefore we assume that the ideal reference ellipsoid, which is related to the mean sea level surface with excess land masses removed and ocean deeps filled, is an equipotential surface of a normal gravity vector field. Denoting the potential of the normal gravity field by

$$U = U(x, y, z) \dots \dots \dots (16)$$

we see that the reference ellipsoid is a surface with

$$U(x, y, z) = U_0 = \text{constant} \dots \dots \dots (17)$$

The potential (U) of the normal gravity vector field is the sum of the attractive potential (V) and the centrifugal potential (Φ):

$$U = V + \Phi = G \int \frac{dm}{r} + \frac{1}{2} \omega^2 (x^2 + y^2) \dots \dots \dots (18)$$

The normal gravity vector γ_r at a given latitude f on the reference ellipsoid is the gradient of U :

$$\gamma_r = - \text{grad} U \dots \dots \dots (19)$$

whose magnitude is the normal gravity and the direction is that of the plumb line, i.e., the plumb line is vertical at all points on the equipotential surface.

Using the definition of moment of inertia in mechanics one can arrive at,

$$U = MG/r [1 + (K/2r^2)(1 - 3\sin^2\phi) + w^2 (r^2/2MG) \cos^2\phi] \dots\dots (20)$$

where M is the total mass of the earth, and K is a constant determined by the moments of inertia about x, y, and z axes and the mass M (Garland, 1979, Tsuboi, 1983; Telford et al., 1990, Heiskanen and Moritz, 1967). Hence, using Equations 19 & 20, one can easily show that the normal gravity (theoretical gravity) value, γ_ϕ , as a function of the latitude angle, ϕ , at any point on this ellipsoid is given by

$$\gamma_\phi = \gamma_0 (1 + B_1 \sin^2\phi - B_2 \sin^2 2\phi) \dots\dots\dots (21)$$

where γ_0 is the value of γ_ϕ at the equator ($\phi = 0$), B_1 and B_2 are constants. Making use of all the observed values of g at a number of points over the earth, numerical values of the constants B_1 and B_2 have been determined in 1930 by the International Association of Geodesy (IAG) and adopted the formula known as the 1930 International Gravity Formula and is given by

$$\gamma_{1930} = 978049(1 - 0.0052884\sin^2\phi - 0.0000059\sin^2 2\phi) \text{ mGal} \dots\dots (22)$$

with $\gamma_0 = 978049 \text{ mGal}$, equatorial radius $a = 6378.388 \text{ km}$. polar radius $b = 6356.909 \text{ km}$, the ellipticity (polar flattening) given by $f = (a-b)/b = 1/297$.

Recent studies on the orbits of satellites have provided more precise values for constants B_1 and B_2 and the following is the revised theoretical gravity formula established by GRS in 1967.

$$\gamma_{\phi:1967} = 978031.85(1 + 0.0053024\sin^2\phi - 0.0000059\sin^2 2\phi) \text{ mGal} \dots (23)$$

With $\gamma_0 = 978031.85 \text{ mGal}$, $a = 6378.160 \text{ km}$. $b = 6356.909 \text{ km}$, and $f = 1/298.25$. Even in its most refined state, the standard theoretical gravity formula (Eq.23) is a very crude

approximation. It assumes that there are no undulations on the Earth's surface, whereas, in fact, we have elevated lands (hills) and oceanic depressions. Hence, for a practical work, i.e., measurement of gravity on the physical surface of the Earth we must define a physical equipotential surface on the Earth. This physical surface is known as the Geoid, a surface such that gravity g is perpendicular to it.

Geoid is a zero reference (elevation datum $h = 0$) for elevations and ocean depths, as given on topographic maps. It is the undisturbed mean sea level surface continued into continents so as to encircle the Earth, water seeking its level in imaginary shallow canals until it is at rest.

The value of gravity at a point calculated by the standard theoretical formula (Eq. 23) and that observed and reduced to the Geoid do not agree with each other. This is because the effect of attraction of an invisible anomalous mass under the point is involved in the observed value. The small difference between the actual gravity potential W and the normal (theoretical) gravity potential U is denoted by T , (Heiskanen and Moritz, 1967) so that

$$W(x, y, z) = U(x, y, z) + T(x, y, z) \dots\dots\dots (24)$$

T is called the anomalous potential, or disturbing potential.

We compare the potential on the Geoid

$$W(x, y, z) = W_0$$

with the potential on the reference ellipsoid

$$U(x, y, z) = U_0$$

of the same potential $U_0 = W_0$. Consider now the gravity vector g_P at a point P of the Geoid (Fig. 7) and the normal gravity vector γ_Q at a point Q of the ellipsoid. The gravity anomaly vector Δg is defined as their difference:

$$\Delta g = g_P - \gamma_Q \dots\dots\dots (25)$$

The difference in direction is the deflection of the vertical (θ). Here, g_P is the gravity value observed (g_{obs}) at a point on the surface of the Earth and reduced to the Geoid point (P). γ_Q is the theoretical gravity value on the ellipsoid point (Q).

Gravity anomalies develop in consequence of differences in the density distribution of the earth, particularly in the upper layers known as the crust. Therefore, they reflect the internal constitution of the crust and indicate the presence of various geological structures connected with the dislocation of rocks of different densities. This enables us to study the internal structure of the earth and for gravity prospecting.

4.2 DATA COLLECTION

To produce the necessary information for the preparation of the final Bouguer gravity map, the following essential data must be determined at each observation point.

1. Relative gravity difference from one or more references or base stations.
2. Relative elevation for making the elevation correction.
3. Relative position for making latitude corrections and for mapping the final results.

Field observations produce a series of meter readings. The usual final map is of Bouguer gravity, which is in gravity units (mgal) with meter readings multiplied by the

4.3.2 FREE AIR CORRECTION

Gravity varies with elevation. A point at higher elevation is farther away from the center of the earth and therefore has a lower gravitational acceleration than the one at lower elevation. The free air correction accounts for only changes of gravity with height. So it is necessary to correct for changes in elevation between stations to reduce observations to datum surface. The correction is obtained by differentiating the scalar equation for a spherical earth,

$$g = GM/R^2$$

where, M is the total mass of the earth

The gravity at a point located at a height h above the Geoid is

$$g_h = GM/(R+h)^2.$$

By the Binomial expansion

$$\begin{aligned} g_h &= GM/R^2 (1 - 2h/R + 3h^2/R^2 - \dots) \\ &= g - 2gh/R + \dots \end{aligned}$$

From this expression, the free air correction (δg) is written as

$$\delta g_{FA} = g - g_h = 2g_m h/R_m = 0.3086h \text{ mGal/m} \dots\dots (27)$$

where R_m and g_m are the average radius and gravity of the Earth, respectively. Thus, with the average values $R_m = 6371 \times 10^3 \text{ m}$ and $g_m = 980 \text{ Gal}$ inserted, the free air correction for h measured in meters is $\delta g_{FA} = 0.3086h \text{ mGal}$. The free air correction is added to the gravity reading when the station is above the Geoid and subtracted when below it.

4.3.3 BOUGUER CORRECTION

The Bouguer correction accounts for the attraction of the material between the station and datum plane that was ignored in the free air correction. It is the correction for the attraction as approximated by considering the material as an infinite horizontal slab. The gravity attraction for a point on the surface of a slab obtained by calculating the effect of an infinite disc is given by

$$\delta g_B = 2\pi G\rho h$$

where h is the height of the gravity station above the Geoid and G is the universal gravitational constant. For mean crustal density (2.67 g cm^3) and h in meters, the Bouguer correction reduces to

$$\delta g_B = 0.1119h. \dots\dots\dots (28)$$

This is referred to as the Bouguer reduction, which moves the mass between the Earth's surface and the Geoid to infinity and then reduces the point to the Geoid. The Bouguer correction is applied in opposite sense to the free air correction, i.e., it is subtracted when the station is above the Geoid and vice versa.

4.3.4 TERRAIN CORRECTION

The terrain correction allows for surface irregularities in the vicinity of the station. There are several methods for calculating terrain corrections, all of which require detailed knowledge of relief near the station and a good topographic map extending considerably beyond the survey area.

The correction is done by computing graphically the gravity effect at the observation point of all hills above the station level and all valleys below it (Bible, 1962). A method of doing this is described by Hammer (1939), which makes use of a specially designed transparent graticule divided into zones by circles, these being subdivided into compartments by radial lines. The total effect is the sum of each compartment and each compartment is computed by its average height. The gravity effect of a single sector is calculated from the following formula:

$$\delta g_{\text{sec}}(r, \theta) = G\rho\theta \{ (r_0 - r_i) - (r_i^2 + h^2)^{1/2} - (r_0^2 + h^2)^{1/2} \} \dots (29)$$

where G is the universal gravitational constant, q is sector angle in radians, $h = e_s - e_a$, e_s is the station elevation, e_a is the average elevation in the sector, and r_0 and r_i are the outer and inner sector radii. The terrain correction δg_T is the sum of the contribution of all the sectors:

$$\delta g_T(r, \theta) = \sum_r \sum_\theta \delta g_{\text{sec}}(r, \theta) \dots (30)$$

The method employed recently for calculating terrain corrections using digital computers have been devised by (Kane, 1962; Bott, 1959). These generally use a scheme of digitizing the topography by writing the elevations at points on a uniform grid so that the same grid points can be used for all stations.

When both the latitude and elevation corrections (including terrain correction, when warranted) have been applied, the residual is called the Bouguer anomaly. This is the small component of g which is supposed to have its origins at a shallow depth, and

the primary data from which a geological interpretation of their source is attempted. In the geodetic sense the anomaly is a single numerical value for any individual observation and is the difference of the observed value from a theoretical or calculated value based on certain assumptions about the form of the gravity field over the earth as a whole.

4.4.1 THE FREE - AIR ANOMALY

The free - air anomaly is the simple effect of a station at a higher elevation being farther away from the Geoid. It is called free air because the theoretical anomaly is calculated as if the gravity measurement were at the elevation of the station but without taking into account the attraction of material between that elevation and the sea level. Thus,

$$\Delta g = g_{obs} + 0.3086h - \gamma\phi \dots\dots\dots (31)$$

4.4.2 THE BOUGUER ANOMALY

The Bouguer anomaly is the difference between the measured value at the point of observation and the theoretical value calculated for that elevation or water depth and the appropriate density of the earth's materials. The Bouguer anomaly calculated ignoring topographic effects is known as simple Bouguer anomaly (S.B.A).

That is,

$$S.B.A = g_{obs} + 0.3086h - 0.1119h - \gamma\phi \dots\dots\dots (32)$$

When all corrections, the free air, Bouguer, together with terrain corrections are applied to the observed gravity, the resulting anomaly obtained by subtracting the standard

theoretical gravity value at the given latitude is called Complete Bouguer anomaly (C.B.A) and is given by

$$C.B.A = g_{obs} + 0.3086h - 0.1119h - \delta g_T - \gamma_\phi \dots \dots \dots (33)$$

where δg_T is the terrain correction

4.4.3 ISOSTATIC ANOMALY AND ISOSTATIC CORRECTION

According to the theory of isostasy introduced by C.E.Dutton (1889) anomalous loads such as mountain ranges are supported by the extra buoyant forces supplied by an excess of material of lighter density at the base of the crust that displaces the denser substratum, where unusual deficiencies in crustal matter such as the great ocean trenches are prevented from buckling upward by reduction in the normal buoyant forces caused by a deficiency of lighter material at the base. These give rise to variations of crustal thickness (roots and antiroots) which is the cause of the broad anomalies in the Bouguer gravity at the surface.

Calculations of isostatic effects can be made on either principle (Airy's or Pratt's) to determine the magnitude of the required deficiency in density under areas of high topography. The method of isostatic correction is similar to that of terrain correction. When a correction of isostasy is applied to the Bouguer gravity, the residue is usually called the ISOSTATIC ANOMALY.

In general the isostatic anomaly maps produced show very much smaller anomalies than the Bouguer maps. Except in reconnaissance surveys over broad and anomalous regions, however, the isostatic

the regional properly from the observed data. Where the contours at a distance from a local anomaly are quite regular, it is possible to take out the regional trend by drawing lines which connect the undisturbed contours outside the area within which the anomaly is confined. Where the smoothed contours cross contours of observed gravity, the differences between the two, which had discrete values at each intersection, are marked and themselves contoured. the resulting map gives residual gravity. The residual profile plotted below the observed cross section was obtained simply by subtracting the estimated regional value from the observed gravity at all points along the profile. The graphical methods have the advantage that all the available geological information from an area can be put to use in drawing the regional. This is particularly important where the survey covers an area that is smaller than that of the major structural feature governing the regional trends.

On the other hand, with the analytical methods of determining residual gravity, numerical operations on the observed data make it possible to isolate anomalies without such a great reliance upon the exercise of judgment in carrying out the separations. Such techniques generally require that gravity values be spaced in a regular array, and templates are designed so that values can be interpolated from maps on a uniform grid. Four analytical methods in common use:

1. The direct calculation of residuals
2. The determination of second derivative
3. Polynomial fitting
4. Downward continuation

A good compromise between the two approaches is computer modeling. This involves setting up various models of the geology, all compatible with established information, residual maps are rapidly generated for each model by the computer.

The removal of regional effects is one of the two important problems in gravity interpretation. The other is obtaining information from the regional anomaly on the structural configuration and density distribution of its source.

4.5.2 INTERPRETATION

The interpretation problem usually is finding the mass distribution for the residual anomaly. Isogal maps look very like topographic contour maps. They show circular, elongated and irregular area of high and low gravity. They may also show linear belts of steep gradients which are not necessarily associated with any of the features just mentioned. It is possible merely from inspection of the map to make a tentative qualitative interpretation if something is known about the geology. Depending on their relief and the distance between axes, this may be interpreted as being indicative of structural features or trends which may be attributed to deformation in sediments or to density contrasts in the basement or both. Gravity highs are in many areas associated with anticlines or with horst blocks, both being structures which bring older, denser rocks nearer the surface. In other regions gravity highs may be due to the presence of heavy basic intrusions. Conversely sedimentary basins and relatively light acid intrusions usually produce gravity lows. The belts of steep gradients are produced by vertical contacts between rocks of different density such as may occur across fault planes.

If an interpretation is to be quantitative to any degree at all, there must be density information from measurements or, more commonly, from inferences based on the general nature of the rocks, which can be learned from existing geological studies or maps of the area, if any. Quantitative interpretation means finding out the position, size, and shape of the gravitating mass through analysis of its potential field. Almost all interpretation of gravity data is by indirect methods. From inspection of the isogal maps, taking into consideration all other information about the region, a possible model of the structure is devised and its gravity effect calculated. The observed and calculated anomalies are compared, the model then progressively modified and its anomaly recomputed until a reasonable fit between the two is obtained. If the geological data are scanty it may not be possible to do more than calculate a range of approximate solutions, but even to be able to set limits to the possibilities can be most useful.

When choosing a model it is usual to make the simplest of geometrical approximations as the gravitational attraction of a number of simple forms can be easily calculated from graphs and standard formulae. Since the postulated geometrical shape can at best be only a crude approximation to the real structure, the fit between the calculated and observed anomalies, though often surprisingly good, is unlikely to be 'perfect', i.e. within the limits of observational error, but the procedure does enable an estimate of dimensions and depth to be made quickly. In many instances where there is little geological control there may well be little point in pursuing the interpretation further, since it is highly unlikely that a more complicated model will be a better approximation to the true structure, however exact the fit between the observed and calculated anomalies.

Spheres, cylinders and slabs are the usual geometrical forms chosen when making preliminary interpretation. Since gravimeters respond only to changes in the vertical component of the gravitational anomaly, formulae are required giving the change in the vertical component of the attraction at points on the surface. Normally the chosen formulae express the variation in the attraction along a traverse at the surface across the body. If the body is elongated in a horizontal direction (i.e. is two dimensional) the traverse direction is taken normal to the direction of elongation.

CHAPTER 5

5. DESCRIPTION OF ANOMALIES

5.1 INTRODUCTION

The previous geophysical works discussed in chapter 3 were aimed at studying specific geological problems related to crustal and upper mantle structures of the Ethiopian rift system. The major objectives of the present research work is also reduction and preparation of all available gravity data and interpret the produced gravity maps in terms of the geologic and tectonic features of the study region, and compare the results obtained with the previous works aforementioned. Though no gravity observations were made, the task were accomplished with data, compiled from different sources and reprocessed following the procedures in gravimetry as applied to geophysics.

5.2 GRAVITY DATA PREPARATION AND PROCESSING

Gravity data were compiled from previous surveys conducted in the study region.

The sources include:

- the Ethiopian institute of geological surveys (EIGS) and
- the Geophysical Observatory of A.A.U (Dr. Abera's Ph.D work).

Theoretical gravity values were computed for all stations of both data sets (already being reduced), using the formula of Geodetic reference system 1967 (GRS67). In order to keep the consistency between the two data sets, reprocessing and re-evaluation was inevitable with the same theoretical formula, and were checked if they have

homogeneously been connected to the IGSN71 datum. The observed gravity values for some stations in the data set obtained from EIGS were not available. These stations were ignored, since it was necessary to start with the observed gravity value for the recalculation of the anomalies. A compilation of the two data sets between the defined latitudes and longitudes resulted in a total of 3000 stations. All stations occupied have been referred to the IGSN71 (Morelli et al., 1971) by subtracting 14.91 mGals. The primary base station used was of the Geophysical Observatory (977452.16) located at Addis Ababa University.

The standard corrections discussed in chapter 4 are made, to determine a reliable Free - air and Bouguer anomalies. These data was reduced to sea level with a uniform density of 2.67gm.cm^3 . The reduction process was performed using a FORTRAN program (Alemu, 1989) based on the above density.

As one of the objectives of this paper is to establish a standard gravity network, the compilation and preparation of these data in a standard format was completed and is ready to be used it for detailed survey. Terrain corrections have not been taken into account and only Simple Bouguer anomalies were calculated. The estimated mean terrain effect is about 2 mGals. This estimate was based on values calculated along two traverses that cross the Aluto mountain (2335m), rising about 700 m above the rift floor by Dr. Abera Alemu (1992). This value is used to apply for all stations and treated as a systematic error in computing the overall mean square error of the gravity anomalies.

where $m_{\Delta g}$ is the standard error (due to random errors) and $s_{\Delta g}$ is the systematic error (due to neglect of terrain effect) (Sjöberg, 1990) of the gravity anomalies.

Assuming the mean crustal density 2.67 g/cm^3 used here is correct, the standard error $m_{\Delta g}$ can be computed from the law of propagation of errors as:

$$m_{\Delta g}^2 = m_g^2 + (1/R(\partial g/\partial \varphi)m_\varphi)^2 + (\partial \Delta g/\partial h)^2 m_h^2 \dots \dots \dots (36)$$

where $m_g = \pm 0.85 \text{ mgal}$, $m_\varphi = \pm 500 \text{ m}$ ($\gg 0.27$ minutes of arc) and $m_h = \pm 10 \text{ m}$ are the maximum standard error estimates reported for both data sets (Abiy, 1989; Alemu, 1992) in determining the observed gravity (observation error), the geographic latitude and the elevation of the gravity point respectively.

$$1/R(\partial g/\partial \varphi) = 1/R(\partial g/\partial \varphi) = 0.812 \sin 2\varphi \text{ mGal/km} (37)$$

where $R = 6371229 \text{ m}$ is the mean radius of the earth.

For the mean latitude, $\varphi \approx 10^\circ$ of the study area, the gravity variation is about 0.03 mGal for each 120 meters traveled in a N-S direction. This means that in order to obtain an anomaly accuracy of 1 mGal , the position must be known to at least 4000 meters . For the positional error $m_\varphi = \pm 500 \text{ m}$ ($\gg 0.27'$ of arc) we adopted in fixing the latitude φ from the available topomap of the region, the error estimate in the normal gravity amounts to $1/R(\partial g/\partial \varphi)m_\varphi = 0.14 \text{ mGal}$ for $\varphi = 10^\circ$

The maximum error estimate in the Bouguer anomaly corresponding to our adopted maximum elevation error $m_h = \pm 10 \text{ m}$ of a gravity station therefore amounts to $(\partial \Delta g/\partial h)m_h = \pm 1.12 \text{ mGal}$.

Using the formula $\sigma_{\Delta g}^2 = m_g^2 + (1/R(\partial g/\partial \phi)m_\phi)^2 + (\partial \Delta g/\partial h)^2 m_h^2 + s_{\Delta g}^2$, the total standard error $\sigma_{\Delta g}$ of the gravity anomalies Δg is:

$$\sigma_{\Delta g} = \pm ((0.85)^2 + (0.14)^2 + (1.12)^2 + (2.0)^2)^{1/2} = \pm 2.5 \text{ mGal.}$$

The overall accuracy of the Bouguer anomaly values (assuming the correct density has been chosen) is therefore expected to be around ± 2.5 mgal.

5.4 THE GRAVITY MAPS

5.4.1 FREE - AIR ANOMALY

There is a close correlation between the pattern of the free air anomaly (Fig. 8) and the topography of the study region. However, an overall correlation can not be obtained between the anomalies and the subsurface geological and structural features. Thus, it is limited to support in the interpretation of the Bouguer anomalies.

The free air anomaly values vary from -20 to 0 mgals in the main Ethiopian rift, from -20 to 30 mgals in the Afar depression, and from 0 to 110 mgals in the Plateaus. The highest value of the anomaly occurred at the north western part of the survey area with an amplitude of 110mgal, at the elevation of Mt. Guna (4231m). The other maximum value of 90 mgal occurred on the western plateau is located north of Addis Ababa around Debreberhan.

On the eastern plateau, on the Arsi-Bale massif a maximum of 70 mgal occur, and on southeast of the Afar depression, on the Harar plateau, Mt. Gara muleta is marked with high positive free air anomaly. The minimum value of the free air anomaly occurred on the western margin of the Afar, in the Borkena graben. A positive value reaching 30 mgal in the Afar depression may be related to the Gabilema range located west of lake Abbe.

The curvilinear nature of the main Ethiopian rift at about 8°N latitude is clearly evident. The free air anomaly map also reveals the E-W oriented latitudinal Yerer fault where the rift escarpment is displaced and runs northeast.

5.4.2 BOUGUER/RESIDUAL ANOMALIES

The Bouguer anomaly map produced (Fig.9) reveals an overview of the topographic features and underlying crustal structures. The distinction between the rift and the adjacent plateaus and the regional structural features were depicted from the map on the basis of the general shape and wavelengths of the anomalies. However, anomalies on the Bouguer map have no appreciable correlation with the structural and geological features of shallow origin. Separation of the regional and residual anomalies from the Bouguer gravity map and their qualitative interpretation led to defining the anomalies related to deep structures, and local anomalies of shallow origin respectively.

As shown in the previous works of Mohr and Gouin (1967/68) the rift valley is in gravity 'high' as compared to the plateaus. The relatively positive anomaly of the rift shown on the Bouguer map strikes approximately NNE-SSE between latitudes 7° and 9°N

and curves to NE-SW orientation towards the Afar depression. A series of local positive and negative anomalies are superimposed on the relatively positive anomaly which are related to local variations in geological structure.

In general, there seems a close correlation between the Bouguer anomaly and topography of the study region, where minimum gravity values are located in areas of maximum elevation and vice versa. The minimum gravity value with amplitude of -270 mgals occur around Debreberhan. Other local minima with amplitudes between -250 and -210 mgals are located on the western plateau. The eastern plateau is also marked by a gravity 'low' with a minimum value of -250 mgals.

On the study region considered here, the intensity of gravity increases from -190mgal in the main Ethiopian rift to -50mgal in the Afar depression. The pattern of a series of anomaly belts separated by small gravity highs show the general trend of the rift axis, following the axis of the Wonji fault belt.

Gravity gradient increases eastward from the western border and reaches maximum at the boundary of the western Ethiopian plateau and the Afar where dipping of the margin towards the depression starts. The Borkena graben is marked with a relatively 'high' gravity extending SSE from 11° to 10.5° N, abutting the western plateau escarpment with a high gravity gradient running N-S at about 39° E longitude. Southward, the Robi graben extends to latitude 10° N with approximately the same gravity 'high'.

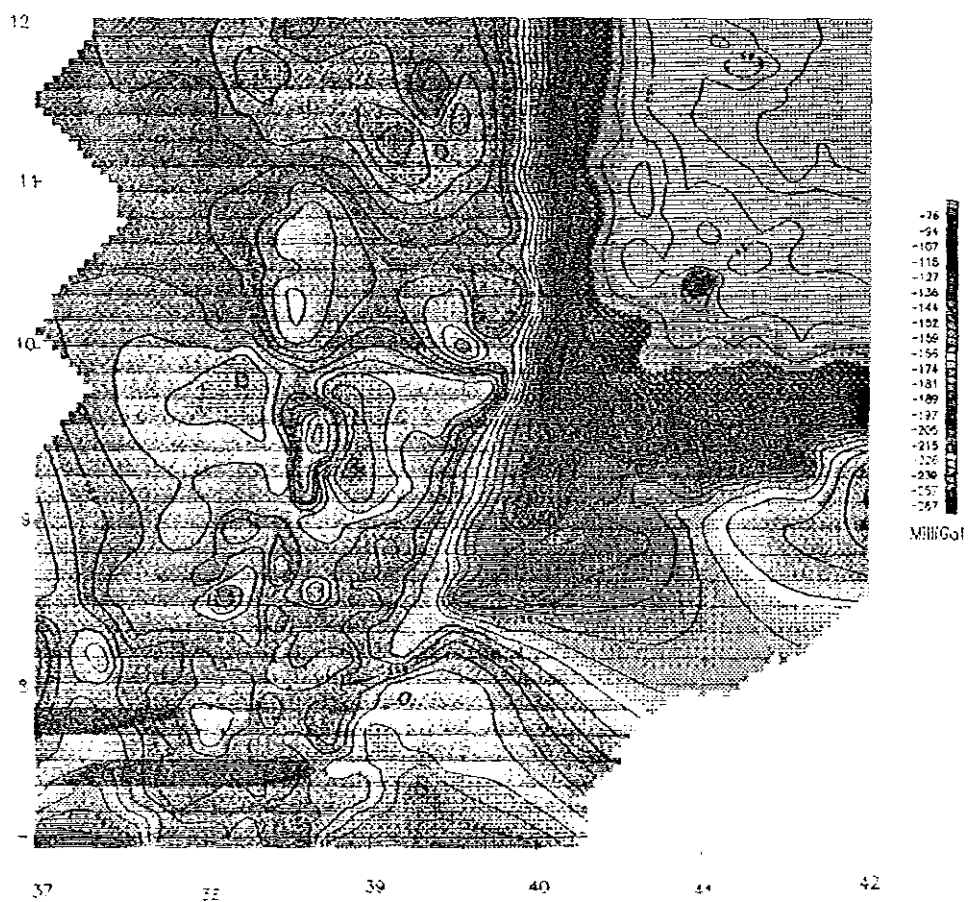


Fig. 9. Bouguer Anomaly Map

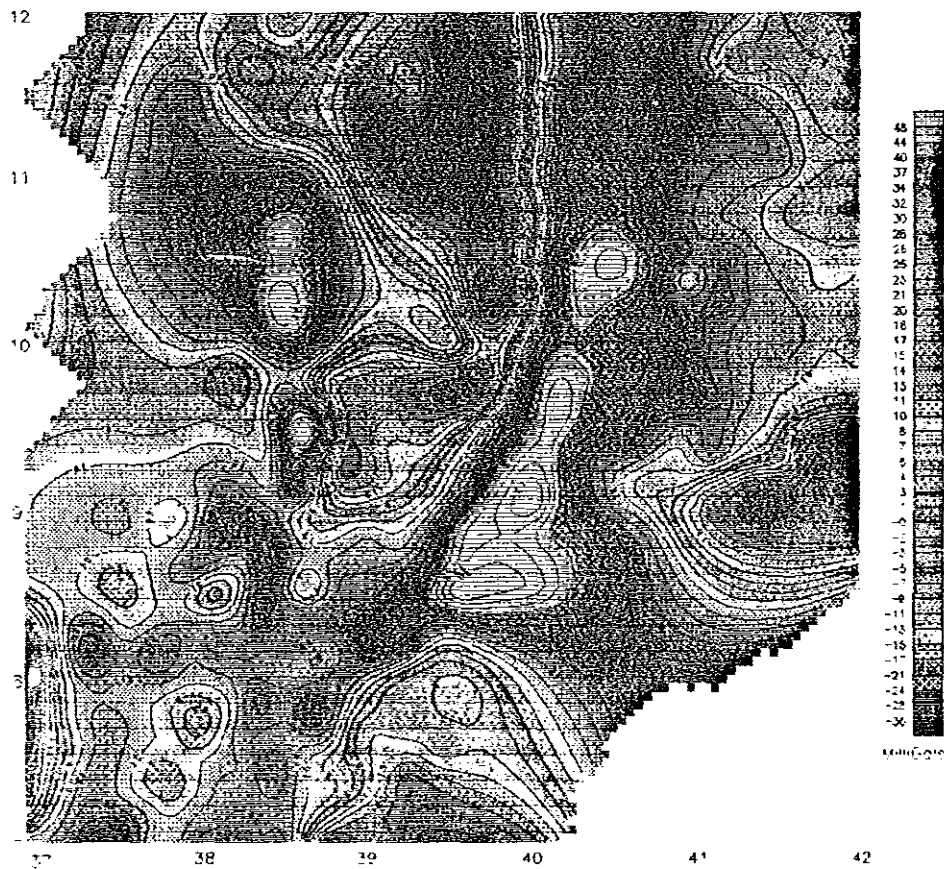


Fig.10. Residual Anomaly Map

The two gravity gradients, one trending N-S on the western boundary of Afar and the other on the southeastern boundary with NE-SW orientation, converge at the opening of the Afar depression. The rift floor of Afar is marked by gravity 'high', beginning from the Mt. Guraghe-Chilalo and Yerer - Gugu cross rift lineament where the junction between Afar and the main Ethiopian rift seems to lie.

The residual gravity map (Fig.10) characterizes the general pattern of the Bouguer map, but with small local anomalies appearing after the removal of the regional. The two main gradients separating the Afar depression from the plateaus can be readily seen on both maps. Particularly, in the Afar depression the Bouguer map shows a trend of elongated maxima running northwards parallel to the main escarpment, with increasing values, which according to Makris and Ginzburg (1985) was associated with the crustal thinning towards the coast of the Red sea, towards a zone of high crustal attenuation with the Moho revealed to lie at the depth of 5Km from seismic refraction data (Berckhemer et al., 1975).

On the western Ethiopian plateau, a succession of local positive and negative anomalies occur west of the Borkana-Robi marginal graben, parallel to the gradient trending N-S. A wide positive anomaly belt occur on the western plateau adjacent to the main Ethiopian rift about 37°E longitude, and gravity values decrease slowly towards the top of the western margin and then rise up slowly towards the rift floor where local anomalies are mainly related to rift volcanic centers and associated rocks.

The rift lakes basin is limited north by the uplifted Meki - Awash watershed and south by the upwarped margin of lake Shalla. The gravity gradient shown on the residual

map, northeast of lake Zway, confirms the important Meki - Awash watershed 'transverse arch' probably accompanied by shallow depth as interpreted by Gouin and Mohr (1964).

5.4.3 REGIONAL ANOMALY

The regional anomaly map (Fig.11) which was extracted from the Bouguer map shows smoothed large scale regional features. The regional gravity field indicates the transition from the plateaus to the rift floor. There is an inverse correlation between the topographic relief and the gravity field in the study region. This is associated with isostatic compensation of the topographic relief by low density material at depth (Makris et al., 1975). The minimum regional anomaly of magnitude -230 mgals occur on the western plateau with its center located west of Addis Ababa. This anomaly is thought to be caused by density contrasts at the crust-mantle boundary. Although Makris has indicated the location of this minimum anomaly north of Addis Ababa, the present work shows its occurrence is west of Addis Ababa. This difference in location might probably be due to high distribution of data used in the work as compared to his data.

Regional values increase from the center of the minimum anomaly, both towards the northeast and southwest direction. The maximum values occurring in the Afar depression could be associated with the thinning of crust and further towards the Red Sea coast. The minimum anomaly striking NNW-SSE extends for about 180Km and has a width of 83Km. This may be associated with the uplifted region that forms part of the Ethiopian Dome where the effect of the anomalous mantle could be maximum. It is now generally accepted that the long wavelength gravity anomaly in east Africa is due to the contribution of the anomalous mantle.

6. DISCUSSION, CONCLUSION AND RECOMMENDATION

The gravity anomaly map shows maxima and minima that are separated by contour lines of equal gravity. The maxima correspond to the central parts of rock masses that have a higher density in contrast with that of the surroundings. The qualitative interpretation of the gravimetric maps is to identify and locate geological units their boundaries and structure.

On the western Ethiopian plateau, the 'low' about Debretabor coincides with the topographic effect of Mt. Guna. Another 'low' which seems to be a continuation of the Guna 'low' and separated by small gravity 'high' could be associated with the elevation at a volcanic center south of Mt. Guna. The area around these volcanic mountains is overlain by extensive alkaline to transitional basalts often forming shield volcanoes, trachytes and phonolites of Oligocene - Miocene. The high between the two lows occurring about 11.5° latitude could not be evidenced by the surface geology. This may probably be attributed to E-W faulting extended from the southeastern margin of the lake Tana basin.

Immediately southwards, south of Dejen a small gravity 'high' is marked, where the Free air values reach about zero. This high is not evident from the surface geology, but it could be associated with the topographic effect on both the northern and southern ends which are marked by gravity lows related to high elevations. The low on the northern side occur at a shield volcano while on the southern side the low about Gebregurcha is associated with the quaternary volcano, mt. Selale.

North of 11°N latitude, the marginal graben forming the outer tectonic boundary of Afar against the Ethiopian plateau is not marked by a strong gravity 'low'. Mohr and Rogers (1967) associated the absence of this low around Dessie with a very thick succession of the Trap series basalts. These basalts are extensively exposed between Kombolcha and Woldia.

The Residual map shows a wide positive anomaly striking N-S on the western plateau between latitudes 7° and 9°N on the Gibe basin. The surface geology of the area is marked by the late Eocene - late Oligocene trachytes, ignimbrites, rhyolite and tuff with minor basalt. The free air anomaly show negative values over this region which correspond to the low elevation of the area. The local maxima that occur north west of Welkite could possibly be attributed to the flood basalts of middle Miocene covering the area.

On the western margin of the main Ethiopian rift, the gravity low at about latitude 8.5°N, coincides with the elevation at Mt. Gurage. On the opposite side of the rift, the eastern plateau exhibits a low at Mt. Chilalo where the line across the rift, joining the two mountains form the known Chilalo - Gurage cross rift lineament. About 38°E longitude three small lows occur, one at MT Zquala and the other two being related to fields of small basaltic vents.

Local positive and negative anomalies are shown on the residual map with amplitudes between -60 and 50 mgals, where amplitudes in the rift floor is generally low. The anomalies in the rift floor are associated with the axial zone of recent faulting the Wonji fault belt, trending NE-SW in the main Ethiopian rift and NNE-SSW in the Afar

depression. The relative positive anomaly of the MER is associated by the mass deficiency providing isostatic balance of the regional uplift to compensate for the relatively lower elevation of the rift floor as it was concluded by Gouin (1972).

Gravity 'highs' occur immediately north of the major cross rift transcurrent faults. The gravity highs over the wonji fault belt is further confirmed having a sudden gradient over the western more active side of the belt. This high is attributed to basaltic injection of a thin crust affected by tension as it was suggested by Gouin and Mohr (1964). The highs over the Wonji fault belt in the MER form a pattern in such a way to reflect the structure of the rift. This belt of intense, fresh faulting running along the center of the floor of the MER (see Mohr, 1960) passes along the northern end of lake Langano, along the eastern shores of lake Shalla, at the Corbetti volcano and along the northwest shores of lake Awasa.

Afar is generally characterized by a small negative Bouguer value, with some local maxima and minima superimposed in association with local disturbances. A local gravity minima located about Dubti could probably be associated with the surface geology of that area marked by Pliocene - Pleistocene alkaline basalts and trachytes. The high Bouguer gradient on the eastern side of the anomaly assumes the existence of a fault or fissure striking N-S. The decrease of gravity values from west to east across the gradient indicates an increase in elevation which probably suggests the presence of a difference in elevation, possibly related to the E-W oriented Gamari range.

The maximum gravity value occurs east of sardo where refraction seismic results (Berckhemer et al., 1975) show that the strongest crustal attenuation occurs in this area

having a minimum thickness of only 14 Km. South of lake Abbe the intense E-W fault belt is characterized by a gravity 'high'.

CONCLUSIONS

The gravity data, reduced to the IGSN71 datum, is presented in a standard format as one of the objectives of this thesis work is so. The availability of this data will provide researchers with a considerable information of the gravity field of the study region and could be applied for further research programs intended to be conducted in the region.

A qualitative interpretation of the compiled Bouguer map, using the existing geological information and a limited seismic refraction results as constraint, revealed the inverse correlation between the topographic relief and the Bouguer anomalies. local minima and maxima superimposed on the broad negative regional anomaly show the existence of structural units having density contrasts with the surrounding. The existing geological features and some isolated structural units identified in the previous works are further confirmed.

The main conclusions are:

- The long wavelength negative Bouguer anomalies that characterize the regional structure is disturbed by the relatively high anomaly belt which clearly identifies the structural feature of the rift distinguished from the plateaus. This anomaly belt is thought to be due to the existence of an anomalous mantle at shallow depth.
- Gravity values over the main Ethiopian rift are generally 'low' as compared to that of the Afar depression by a magnitude of about

- 140 mGals. The gravity high over Afar which indicates the strong attenuation of the crust flooring the depression. This is confirmed by seismic refraction results taken along profiles in Afar (Berckhemer et al., 1975), that crustal thickness reaches 14 Km at the triple junction, around Sardo.
- The regional gravity values are related with the deep structural features. The pattern of the isomGal anomalies on the regional anomaly map indicate the transition from the plateaus with higher Bouguer masses above sea level, towards the Red sea via the Afar depression. The Afar depression is a continental crust, forming a platform between continental and oceanic rifts as was suggested by previous researchers.
- Part of the Ethiopian dome is located on the western plateau at about the latitude of Addis Ababa. It is marked by minimum regional anomaly indicating the existence of a low density material beneath the crust.
- The existence of axial high anomaly belt along the Wonji fault belt, its extension towards central Afar following the fault belt and its occurrence coinciding with recent silicic volcanoes of the rift is further confirmed.

In this paper, the general structural and geological features of the study region are described by the qualitative interpretation of the gravity maps. But it is obvious that this kind of interpretation fails to give some critical information regarding the dimensions of deep structures underlying the crust. Therefore it is herein suggested that more refraction seismic surveys to be carried out all over the region to provide researchers with the necessary information and generate crustal models that produce the observed gravity anomalies.

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