AN ASSESSMENT OF GRAVITY CORRECTIONS

SOR YANGO
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BY

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A PROJECT SUBMITTED IN PARTIAL FULLFILLMENT OF THE REQUIREMENT FOR THE AWARD OF THE DEGREE OF BACHALOR OF TECHNOLOGY (B.TECH) IN APPLIED GEOPHYSICS, TO THE DEPARTMENT OF PHYSICS, FACULTY OF SCIENCE, ABUBAKAR TAFAWA BALEWA UNIVERSITY, BAUCHI.

2015

DECLARATION

I hereby declare that this project was written by me and it was a record of my own research work carried out. It has not been presented before in any higher institution for the award of a degree. All references made to related literatures have been duly acknowledged.

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Sor yango Date

Student

The above declaration is confirmed by

________________________ 14-01-2015

Prof. O.K. Likkason Date.

Project supervisor

Certification

This is to certify that this research work titled "an assessment of gravity corrections" carried out by Sor yango, 07/20183u/1 meets the requirements and regulations governing the award of a degree of Bachelor of technology in Applied Geophysics of Abubakar Tafawa Balewa university and is approved for its contribution to knowledge and literacy presentation.

14-01-2015

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Date

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To the lover of my soul through Christ Jesus, I will forever remain thankful for how far you've brought me.

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To Dennis Offodum; just because.

DEDICATION

To Othniel M. Sor

TABLE OF CONTENTS

LIST OF FIGURES

LIST OF TABLES

ABSTRACT

Gravity data bases are being revised to improve accuracy, versatility and coverage. The currently adopted approach to correct observed gravity data for geophysical purposes includes several approximations. These were originally used to reduce computational effort, but have remained standard practice, even though the required computing power is now readily available. The measured gravity data are highly affected by topography and elevation of the station, instrumental drift, tidal effects and latitude variation of the earth. The interpretation of data proceeds from simple meter corrections to corrections that rely on increasingly sophisticated earth models. Important details of the reduction equations continue to be refined and debated (e.g. LaFehr, 1991a, 1991b, 1998.) The most striking revision is the use of the internationally accepted terrestrial ellipsoid for the height datum of gravity stations rather than the conventionally used geoid or sea level. The use of the revised procedures is encouraged for gravity data reduction because of the widespread use of the global positioning system in gravity fieldwork and the need for increased accuracy and precision of anomalies and consistency with national databases. Anomalies based on the revised standards should be preceded by the adjective "ellipsoidal" to differentiate anomalies calculated using heights with respect to the ellipsoid from those based on conventional elevations referenced to the geoid. An example is given to illustrate this point.

(LaFehr, T. R., 1991a, standardization in gravity reduction: Geophysics, 56, 1170-1178.

LaFehr, T. R., 1991b, An exact solution for the gravity curvature (Bullard B) correction: Geophysics, 56, 1179-1184

LaFehr T. R., 1998, On Talwani's "Errors in the total Bouguer reduction": Geophysics, 63, 1131-1136)

CHAPTER ONE: Introduction

1.1 Background of the Study

Current processing procedures as described in geophysical textbooks for reducing observed gravity data to anomaly form generally were formalized in the 1920s and 1930s. These procedures were dictated by the accuracy requirements of the gravity surveys, survey objectives and limitations in computational power, terrain databases and absolute gravity accuracies. Surveys were of a local nature permitting numerous simplifying assumptions in the procedures that minimized the computational requirements.

Despite the simplifying assumptions, these procedures, with minimal modification continue to be used in local surveys for a variety of explorations that require high accuracy. As more accurate gravity anomalies have become of interest, modifications to reduction procedures have been investigated (e.g., Oliver and Hinze,1986;LaFehr,1991a,b,1998; Chapin,1996a; Talwani,1998; Li and Gotze, 2001) and implemented on a limited basis but they have not been used generally for national gravity databases. These investigations were basically triggered due to the availability of improved terrain and geoid databases, enhanced computational power and increased use of global position system (GPS) technology to establish locations and heights of gravity stations.

The revised methodologies used in preparing the principal facts of the gravity observations and corrections of the data to gravity anomalies are based on internationally accepted procedures, protocols, equations, and parameters but in several respects differ significantly from the standard procedures described in most textbooks and used in many databases.

The gravitational field of the earth varies from one part of the earth to another. The field varies in a regular way with higher values at the poles and decreases in value towards the equatorial belt. This variation is regarded as the normal component of the gravity field. If this normal component is removed from the gravitational field, an anomalous field remains that depend on the morphology of the surface, structural features within the earth and variations in densities. Earlier reviews (LaFehr, 1980; Paterson and Reeves, 1985; Hansen, 2001) document the continuous evolution of instruments, field operations, data processing techniques, and methods of interpretation and referred to unpublished works to help provide an accurate understanding of the usefulness of gravity methods.

A steady progression in instrumentation (torsion balance, a very large number of land gravity meters, underwater gravity meters, shipborne and airborne gravity meters, borehole gravity meters, modern versions of absolute gravity meters, and gravity gradiometers) has enabled the acquisition of gravity data in nearly all environments from inside boreholes and mine shafts in the shallow earth's crust, to the undulating land surface, the sea bottom and surface, in the air, and even to the moon and other planets in our solar system. This has required a similar progression in improved methods for correcting for unwanted effects (terrain, tidal, drift, elevation, and motion-induced) and the parallel introduction of increased precision in positioning data acquisition.

The fundamentals of interpretation are the same today as they were 25 years ago, but the advent of GPS and the era of small, powerful computers have revolutionized the speed and utility of the gravity method. Over the past decades, software has evolved from running on mainframes to Unix-based workstations and has now migrated to personal computers. With the availability of software running on laptop computers, data are acquired automatically and even processed and interpreted routinely in the field during data acquisition. There is therefore a great need for a deeper understanding of gravity data corrections so as to ensure an increase in accuracy as well as speed.

1.2 Statement of the Problem

The bulk of the corrections carried out are not carefully understood nor applied hence, affecting greatly the anomaly field database. This research work seeks to examine these corrections in detail to minimize possible errors or oversight as the case may be.

1.3 Aim of the Study

The sole aim of this work is to explore/revise in great detail the various processes involved in effectively correcting gravity data.

1.4 Objectives

The objectives of this project work are as follows;

a) To explore TEMPORAL based gravity corrections

b) To explore SPATIAL based gravity corrections.

c) To apply (a) and (b) to a test example.

1.5 Scope and Limitation of the Study

This research work shall cover a review of the corrections applied to gravity data and the procedures leading up to acquiring a total gravity anomaly field without going deep into interpretation.

1.6 Definition of Terms

Base Station: An observation point used in geophysical surveys as a reference to which measurements at additional points can be compared.

Gravimetry: the measurement of gravity or gravitational acceleration especially as used in geophysics.

Altimeter: an instrument usually an aneroid barometer that measures atmospheric pressure and changes determining height above ground or any other reference datum based on the fall of atmospheric pressure accompanying an increase in altitude.

Gals: the Gal (for Galileo) is the cgs unit for acceleration .one Gal is equal to one centimeter per second squared. ($1 cm/s^2$)

Milligals: one milligal is one thousandth of 1 Gal because variations in gravity are very small, units for gravity surveys are generally in Milligals (mgal).

Anomaly: a variation from what would normally be expected at that point.

Loop: a pattern of field observation that begins and ends at the same point with a number of intervening measurements. Such a pattern is useful in correcting for drift in gravimeter observations altimeter.

Drift: a gradual change in the gravimeter reading caused by the gravitational attraction of the moon and the earth.

Profile: a graph or drawing that shows the variations of one property usually as ordinate with respect to another property.

Observed Gravity: gravity readings observed at each gravity station after corrections have been applied for instrument drift and earth tides.

Bouguer Correction: gravity effect of the mass of the earth between the vertical datum and the observation site.

CHAPTER TWO: Literature Review

2.1 Principles of Gravity Survey

Gravity survey is the measurement of the gravitational field at a series of different locations over an area of interest. The objective in exploration work is to associate variations with differences in the distribution of densities and hence, rock types. Gravity survey also measures the small variations in the pull of gravity over the earth's surface and makes a map of profiles of these changes depending on the volume and layout of data acquired. This helps geologists/geophysicists understand where the dense and light rocks are beneath the surface. The goal of gravity survey is to locate and describe subsurface structures from the gravity effect caused by their anomalous densities. Most commonly, gravimeter measurements are made at a network of stations spaced according to the purpose of the survey.

In environmental studies, a detailed high-resolution investigation of the gravity expression of a small area requires small distances of a few meters between measurement stations. In regional gravity surveys as used for the definition of hidden structures of prospective commercial interest, the distance between stations maybe several kilometers .if an area surveyed is not so large, a suitable site is selected as base station (or reference station) and the gravity differences between the surveyed sites and this site are measured . In a gravity survey on a national scale, the gravity differences maybe determine relative to a site where the absolute value of gravity is known.

2.2 Gravity Field of the Earth

Gravity method is governed or built on a well-known law of elementary physics-the Newton law of universal gravitation which states that the mutual force (F) of attraction between two particles or bodies is directly proportional to the product of their masses (M) and inversely proportional to the square of their distance apart.

Thus,

$$F = \frac{GM^1M^2}{r^2} \quad \text{...} (2.1)$$

Where G is a constant of proportionality (the gravitational constant)

M^1 and M^2 are particle masses and "r" is the center-center distance between the particles.

Newton's second law of motion is also used to establish other expressions such as the relation between force, F, mass, m, and acceleration, a. [F=ma]

Because the earth is approximately spherical and because the mass of the sphere can be treated as though all of it were concentrated at a point to the center, any object with mass M_0, resting on the surface of the earth will be attracted to the earth. If the object is lifted a short distance above the earth and allowed to fall, it will do so with a gravitational acceleration "g" given by

$$g = \frac{F}{M_0} \approx GM_e/R^2 \quad (2.2)$$

Where M_e is the mass of the earth and "g" is the acceleration which is a function of both the mass of the earth M_e and the distance R to its center.

Most gravity variations associated with geologic bodies in the outer several kilometers of the earth's crust are measured in mGals. The maximum gravity difference between the earth's normal field and that actually observed on the surface and corrected for latitude and altitude is of the order of several hundred mGals. This difference known as the gravity anomaly reflects lateral density variations in rocks extending to a depth of several tens of kilometers. In the real sense, acceleration due to gravity described in equation (2.2) is not the total magnitude of gravity. The other component is the normalized centrifugal force due to rotation of the earth about its polar axis. The total gravity field of the earth is therefore the resultant of gravitational force and centrifugal force per unit mass.

2.3 Measurement of Gravity

There are basically two types of gravity measurements:

Absolute Gravity: this corresponds to the determination of the absolute magnitude of gravity at any place. In this case, the value of gravity if found is the value at that point only without referring to the gravity value at any other point. The absolute measurement of gravity is usually carried out at a fixed installation by the accurate timing of a swinging pendulum or of a falling weight. The measurement of the absolute value of gravity is difficult and requires complex apparatus and a lengthy period of observation in the laboratory. The absolute gravity value can be determined using two (2) methods:

i. **Swinging pendulum method:** a simple pendulum consisting of a heavy weight is suspended at the end of a thin wire. A stiff wire metal or quartz rod about 50cm long is attached to a movable mass. The period of the pendulum is measured for oscillations about one of the pivot, the distance (l) between the pivots is then measured accurately. The absolute gravity is then calculated from the relation $g = \frac{4\pi^2 l}{T^2}$ where g is the acceleration due to gravity, l is the length of the wire and T is the period of oscillation

ii. **Falling body method:** the principle of this method involves timing the fall of a drop of body fluid of known size through a definite distance in a mixture of non-miscible fluids. This mixture of low viscosity and specific gravity is then gotten by adjusting the proportions of the fluids.

This methods are used are used to determine the absolute values of gravity at a network of worldwide sites such as the national geophysical laboratory in UK or National Bureau of Standards in USA. [International Gravity Standardization Network, 1971, IGSN71].

Relative Gravity: the relative measure of gravity is the difference of gravity between locations, it's simpler and is the standard procedure in gravity surveying. Basically, all regional and small scale gravity surveys adapts this method ,measuring differences in the gravitational acceleration with reference from a Base station of already existing absolute gravity values.

2.4 Data Acquisition

In order to detect a target (a dense body or fault), gravity readings must be taken along traverses that cross the location of the targets. Its expected size will determine the distance between readings or station spacing with larger station separations for large targets and small separations for smaller targets. [It is advisable to model the expected anomaly mathematically before conducting field work. from this, along with the expected instrument accuracy, an estimate can be made of the anomaly size and the required station spacing].

Surveys are conducted by taking gravity readings at regular intervals along a traverse that crosses the expected location of the target. However, in order to take into account the expected drift of the instrument, one station (a local base station) must be located and has to be reoccupied every half to 1 hour or so (depending on the instrument drift characteristics) to obtain the natural drift of the instrument. These repeated readings are performed because even the most stable gravity meter will have their readings drift with time due to elastic creep within the meter's springs and also to help remove the gravitational effects of the earth tides readings have to be taken at a base station. The instrument drift is usually linear and less than 0.01 mgal/hour under normal operating conditions.

Since gravity decreases as elevation increases, the elevation of each station has to be measured with an error of no more than about 3 cm. Readings are taken by placing the instrument on the ground and levelling it. This may be automatic with some instruments. In addition to obtaining a gravity reading, a horizontal position and the elevation of the gravity station must be obtained. The horizontal position could be either latitude and longitude or the x and y distances (meters or feet) from a predetermined origin. The required elevation accuracy for detailed surveys is between 0.004 and 0.2 m and to obtain such accuracy requires performing either an electronic distance meter (theodolite) survey or a total-field differentially corrected global positioning survey (GPS).

Gravity data acquisition can be done over different platforms…

- *i.* ***Land Operations***
- *ii.* ***Underwater Operations***
- *iii.* ***Sea-Surface Operations***
- *iv.* ***Airborne Operations***
- *v.* ***Satellite-derived Gravity***

The basic concept is to get the variation of gravity of the earth surface. The differences in preference of an acquisition platform varies largely from speed to geologic environment to convenience. Each platform has its distinct advantage over others.

2.5 Gravity Instrumentation

A gravity meter or gravimeter measures the variations in the earth's gravitational field. The variations in gravity are due to lateral changes in the density of the subsurface rocks in the vicinity of the measuring point. Because the density variations are very small and uniform, the gravimeters have to be very sensitive so as to measure one part in 100 million of the earth's gravity field (980 gals or 980,000 mGals) in units of mGals or microgals. Gravity sensors fall into one of two categories; absolute or relative gravity sensors.

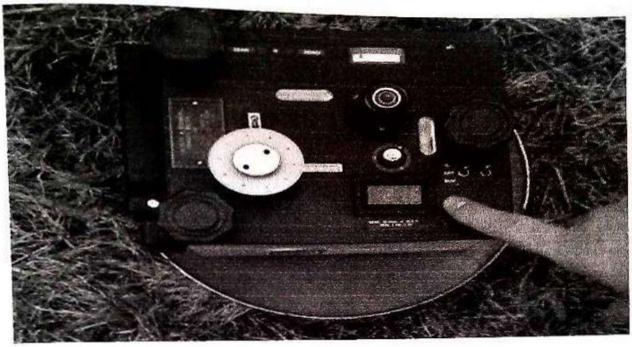

Fig 2.1: Top view of a relative gravimeter

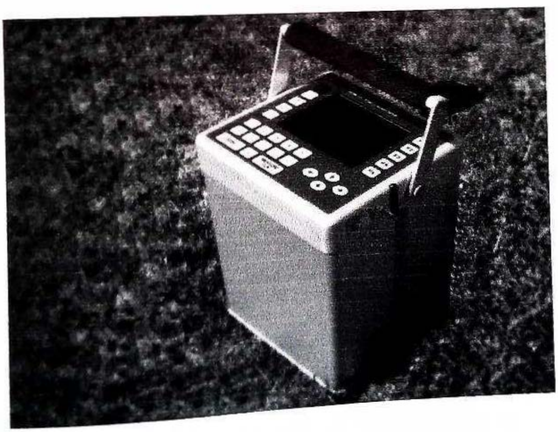

Fig 2.2: side view of a modern gravimeter

An absolute gravity instrument measures the true value of gravity each time it makes a measurement. A relative gravity instrument measures the difference in gravity between successive measurements. A relative instrument is all that is usually required for most exploration measurements. In general, absolute gravity instruments are typically far more expensive, much bigger, take much longer to make a high precision measurement, and usually require more knowledge and skill to use than do relative gravity instruments.

The historical advancement of gravity instrumentation has been driven by a combination of increased precision, reduced time for each measurement, increased portability, and by a desire for automation and ease of use. There have been hundreds of different designs of gravity sensors proposed or built since the first gravity measurements were made. Given the relative size and importance of gravity exploration compared to seismic exploration, it is impressive to realize that there are about 40 different commercially available gravity sensors (Chapin, 1998) and about 30 different gravity sensor designs that have either been proposed or are currently under development.

Even so, there are only four general types of gravity sensors that have been widely used for ground based exploration at different times. They are the pendulum, the free-fall gravimeter, the torsion balance, and the spring gravimeter.

2.5.1 Pendulums

For more than 2 millennia, the widely accepted theory of gravity was described by Aristotle (384 – 322 B.C.) such that the velocity of a freely falling body is proportional to its weight. Then in 1604 Galileo Galilei, using inclined planes and pendulums, discovered that free fall is a uniformly increasing acceleration independent of mass. In 1656 Christian Huygens developed the first pendulum clock and showed that a simple pendulum can be used to measure absolute gravity. In order to make an absolute measure of g to a specified precision, the moment of inertia (I), the mass, the length (h), and the period of the pendulum must be known to the same degree of precision, where the desired precision is better than one part in a million. Even well into the 20th century it was virtually impossible to measure h or I with any great precision. Consequently, using a pendulum to make an absolute measurement gave precisions on the order of about 1 Gal. But it is easier to use a pendulum as a much higher precision relative instrument by measuring the difference in gravity between two locations using the same pendulum. This method was used for the next 160 years as scientists began mapping the position dependency of gravity around the earth.

2.5.2 Free-fall gravimeter

Free-fall gravimeters have advanced rapidly since they were first developed in 1952. The method involves measuring the time of flight of a falling body over a measured distance and both the measurements of time and distance are tied directly to internationally accepted standards. The method requires a very precise measure of a short time period, which only became possible with the introduction of the quartz clock in 1950's. The first free-fall instruments used a white light Michelson interferometer, a photographic recording system, a quartz clock, and the falling body was typically a 1 m-long rod made of quartz, steel, or invar, which was dropped over several meters. The final value of gravity was obtained by averaging over many drops, typically 10 to 100.

2.5.3 Torsion balance

Starting in 1918 and up to about 1940, the torsion balance gravity gradiometer, developed by Baron Roland van Eõtvös in 1896, saw extensive use in oil exploration. It was first used for oil prospecting by Schweydar (1918) over a salt dome in northern Germany and then in 1922 over the Spindle-Top salt dome in east Texas. By 1930 about 125 of these instruments were in use in oil exploration world-wide. The torsion balance is used to measure the gradients of gravity and differential curvature. It was capable of detecting gravity differences on the order of 1 mGal and took about one hour to make a measurement. Two equal masses, about 30 grams each are located at different heights with a vertical separation of about 50 cm. One of the masses is attached to one end of a rigid bar, about 40 cm-long, and the other mass is suspended from the other end of the rigid bar by a fiber. The rigid bar is suspended at its center of mass by a torsion fiber with a small mirror attached to the fiber in order to measure the rotation of a light beam. The balance bar rotates when a differential horizontal force acts on the two masses, which happens when the earth's gravitational field in the neighborhood of the balance is distorted by mass differences at depth such that the horizontal component of gravity at one mass is different from that at the other mass. The torsion balance was very sensitive to terrain and could only be used in relatively flat areas. These instruments became obsolete with the development of spring gravity meters.

2.5.4 Spring gravimeters

Spring gravimeters measure the change in the equilibrium position of a proof mass due to the change in the gravity field between different gravity stations. The vast majority of these instruments use an elastic spring for the restoring force, but sometimes a torsion wire is used.

John Hershel first proposed using a spring balance to measure gravity in 1833. But it was not until the 1930s when demands of oil exploration, which required that large areas be surveyed quickly, and advances in material science permitted the development of a practical spring gravimeter. The simplest design is the straight-line gravimeter, which consists of a proof mass hung on the end of a vertical spring and is a relatively insensitive devise. Straight-line gravimeters are used primarily as marine meters. The first successful straight-line marine gravimeter was developed by A. Graf in 1938 (Graf, 1958) and was manufactured by Askania. LaCoste and Romberg (L&R) also manufactured a few straight-line marine gravimeters.

In order to obtain higher resolution required for land gravimetry, a more sophisticated spring balance system involving a mass on the end of a lever arm with an inclined spring was developed. The added mechanical advantage of the lever arm can increase sensitivity by a factor of up to 2000. The key to the L&R sensor was the zero length spring, invented by LaCoste (LaCoste, 1934). The zero length spring made relative gravity meters much easier to make, to calibrate and to use (LaCoste, 1988). The L&R gravity sensor can make routine relative gravity measurements to an accuracy of about 20 μGals without corrections for instrumental errors (Valliant, 1991) and, when great care is taken in correcting for both instrumental and external errors, down to 1-5 μGal in the field and 0.2 μGal in a laboratory (Ander et al., 1999).

2.6 Application of Gravity Survey

Gravity measurements are used at a wide range of scales and for a wide range of purposes. On an interstellar scale, understanding the shape of the gravity field is critical to understanding the nature of the space-time fabric of the universe. On an exploration scale, gravity has been widely used for both mining and oil exploration, and even at the reservoir scale for hydrocarbon development.

The gravity method is one of the most effective means for mapping the subsurface geology. It is particularly useful in differentiating rock types which are not distinguishable by virtue of their magnetic and electrical properties. In petroleum exploration, gravimetric surveys are second in importance only to seismic surveys. They are commonly employed on a regional basis in advance for seismic programs. Gravity method can often map the distribution of massive sulphide base metal deposits in a reasonably direct fashion (seigal, 1995) because of the relatively high density metallic sulphides. The mapping of subsurface cavities of either natural or manmade origins is important from a geotechnical standpoint because the gravity method can determine their location and approximate dimensions.

On the reservoir scale, borehole gravity meters (BHGM) have been used extensively to detect porosity behind pipes or to measure more accurate bulk density for petrophysical uses.

The use of gravity rapidly expanded in both mining and hydrocarbon exploration for any targets for which there was a density contrast at depth, such as salt domes, ore bodies, structure, and regional geology. The gravity method is sometimes applied to specialized shallow applications, including archeology and detecting shallow mine adits and faults in advance of housing developments.

CHAPTER THREE: Materials and Methods

3.1 Materials

3.1.1 Treatment and analysis of gravity data

Gravity data reduction is a process that begins with a gravity meter reading at a known location (the gravity station) and ends with one or more gravity anomaly values at the same location. The gravity anomaly values are arrived at through corrections to remove various effects of a defined earth model. The basic reduction of gravity data has not changed substantially during the past 75 years. What has changed is the speed with which the computations can be done. In the late 1950's, Heiskanen and Meinesz (1958) maintained that barely more than one rough-mountain station a day can be reduced by one computer. In 1958 a "computer" was an individual having this job title. Today, with digital terrain data and electronic computers, full terrain and isostatic corrections can be done in seconds.

The reduction of gravity data proceeds from simple meter corrections to corrections that rely on increasingly sophisticated earth models. Important details of the reduction equations continue to be refined and debated (LaFehr, 1991a, 1991b, 1998; Chapin, 1996; Talwani, 1998).

3.1.2 Data planes and the mean sea level

The geographic coordinates of gravity observation sites are given in units of degrees of longitude and latitude except for stations of local surveys in which the earth is assumed to be flat and horizontal distances between observations are measured in Cartesian coordinates. Traditionally, the vertical datum for gravity stations is the geoid or (mean) sea level because surface elevations are given with respect to sea level. However, globally there is a difference of ±100 m in the height between the geoid and the ellipsoid, which is the basis of the theoretical gravity. To avoid errors arising from varying regional or national horizontal datums, the International Terrestrial Reference Frame (ITRF), in conjunction with the 1980 Geodetic Reference System (GRS80) has proposed the use of the ELLPSOID as a horizontal datum. On a global basis, use of local datums may lead to errors in position of up to 1 km. In the context of locating gravity stations, the differences between the recent realizations of the 1984 World Geodetic System (WGS84) and the ITRF are negligible.

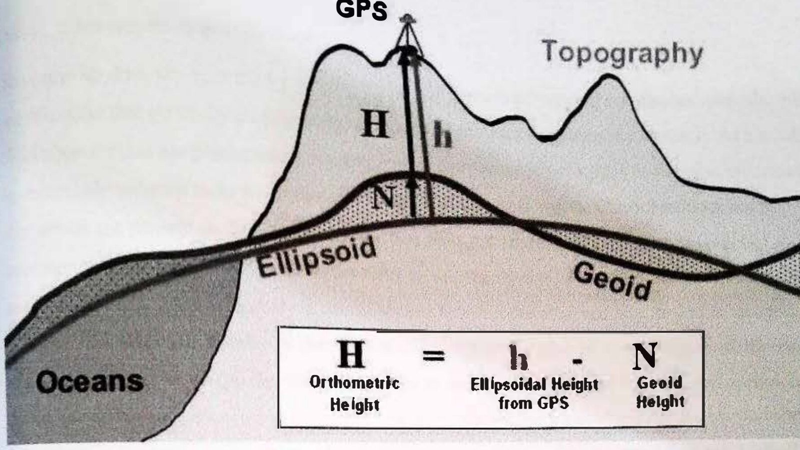

Fig 3.1: Geophysical reference datum.

TheWGS84 datum is used in specifying horizontal location in the GPS, and the ITRF is the coordinate system for satellite altimetry-derived gravity data sets over the oceans. The precise WGS84 coordinates agree with the internationally accepted ITRF coordinates within 10 cm. there is a long wavelength error in the current procedures that is largely eliminated by using the height relative to the GRS80 ellipsoid rather than to (mean) sea level. In gravity corrections and gravity anomalies, the elevation has been used routinely. The main reason is that, before the emergence and widespread use of the Global Positioning System (GPS), height above the geoid was the only height measurement we could make accurately (i.e., by leveling).The GPS delivers a measurement of height above the ellipsoid. In principle, the ellipsoid height rather than the mean sea level height should be used because a combination of the latitude correction estimated by the International Gravity Formula and the height correction is designed to remove the gravity effects due to an ellipsoid of revolution. In practice, for minerals and petroleum exploration, use of the elevation rather than the ellipsoid height hardly introduces significant errors across the region of investigation because the geoid is very smooth. The ellipsoid height is the sum of the elevation relative to the geoid (msl) and the geoid height relative to the ellipsoid.

3.2 Methods

3.2.1 Corrections to gravity data

Raw gravity data are affected by a wide variety of sources of varying amplitudes, periods, and wavelengths that generally mask gravity variations of geologic or geophysical interest. As a result, field observations are processed to minimize these extraneous effects. This conversion procedure is commonly referred to as correction or reduction of the gravity data. Correction does not imply that errors are present in the data, and reduction does not suggest that the data are reduced to a common vertical datum; but both terms refer to the conversion of raw gravity observations to anomaly form. A gravity anomaly is the difference between the observed gravity and the modeled or predicted value of gravity at the station. The observed value is a conversion of the raw gravimeter measurement to the absolute gravity at the station, corrected for temporal variations using ties to stations of known gravity; the modeled or theoretical value of gravity at a station takes into account planetary and topographical gravitational effects. These corrections are basically divided into TEMPORAL and SPATIAL based corrections.

Temporal based variations are changes in the observed variation that are time dependent. These factors cause variations in acceleration that would be observed even if the measuring instrument is not moved. Spatial or space based variations are changes in acceleration that are space dependent. The factors responsible for spatial variations range from elevations, topography, etc. Basic data reduction requires knowledge of the station location, measurement time, the meter constant and theoretical gravity.

3.2.2 Instrumental drift

These are changes in the observed acceleration due to changes in the response of the gravimeter overtime. This is mainly caused by the slow creep of the spring, though temperature and pressure changes have been shown to contribute to it. The effect is known as drift and has to be corrected for. The correction is often made by starting and finishing a set of observations at the same point having noted the time at which all measurements were taken. The drift which is usually positive is the difference between the last and first observations. It is proportional to time. Basically, the drift rate varies a little from day to day and during the course of a single day. The first step in drift correction is a return to the base station or base station reoccupation every few hours is necessary during data acquisition. Figure 3.2 is a graphical representation of a typical base station reoccupation.

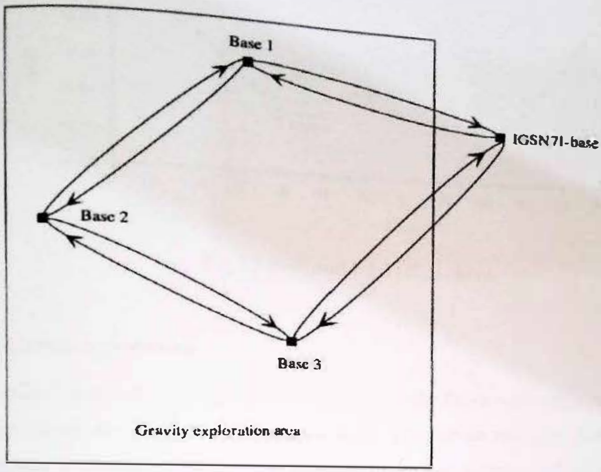

Fig 3.2: base station reoccupation during data acquisition.

The mathematical model often used is derived from simple linear drift assumptions. For a linear drift condition, the drift rate, Þ is given by

Þ= g_2-g_1/ t_2-t_1

Where g_2 and g_1 are the observed gravity readings for the last and first stations and t_1, t_2 the corresponding times of the stations respectively.

3.2.3 Tidal effect

In additional to instrument drift, gravity measurements made at the same location will vary with time due to tidal effects. This is dominantly due to the gravitational pull of the sun and moon. The magnitude can generally be predicted accurately and removed from the measured gravity data. Drift and tidal corrections are usually combined and taken together. Fig 3.3 is a typical drift curve combined with the tidal effect.

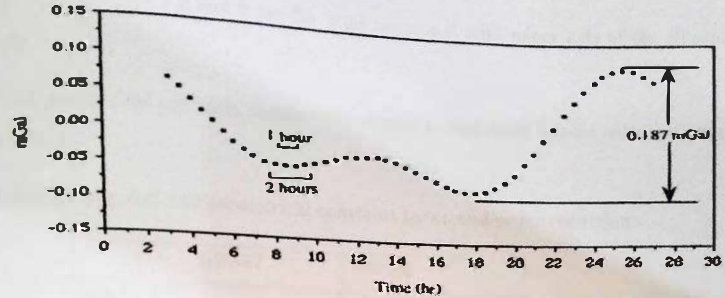

Fig 3.3 drift and tidal effect curve.

3.2.4 Elevation correction

Historically, the elevation or height correction is called the free-air correction and is based on the elevation above the geoid or the mean sea level. The free-air reduction accounts for gravity observations not made on the vertical datum surface. This is accounted for using the vertical gravity gradient as if the observation were made in free-air a distance H above or below the vertical datum surface. It is essentially a correction to the observed gravity for the inverse-distance-squared decay of gravity on moving away from the Earth. The linear approximation, based upon a spherical Earth model is often expressed as $FAC = -\frac{2g}{R}H$ where g is the mean gravity on an assumed spherical Earth of radius R. The frequently adopted numerical value of the free-air gradient is 3.086μm/sec²/m. (Telford, etc. 1990).The free-air reduction is added to observed gravity for observations above the vertical datum surface and subtracted for those below.

The spherical approximation is however considered inadequate because the figure of the earth is more accurately represented by an oblate ellipsoid. Therefore, a second order correction based upon a Taylor expansion of normal gravity above the earth, is used. This second-order free-air reduction is derived in Featherstone (1995) as: $\Delta gF = \frac{2g}{a}(1 + f + m - 2f\sin^2)H - \frac{3g}{a^2}H^2$,

Where f is the geometrical flattening of the mean Earth ellipsoid, and m is the geodetic parameter, which is the ratio of gravitational and centrifugal forces at the equator. g is normal gravity or latitude correction as given by the IGF, IAG OR WGS and

$f = a - \frac{b}{a}$, $m = \frac{\omega^2 a^2 b}{Gm}$ a and b are the semi major and semi minor axis of the ellipsoid respectively.

Physical and geometrical constants required to compute second-order free-air reduction are as shown in table 3.1

Table 3.1 physical and geometrical constants for second-order reduction

	GRS67	GRS80
a(m)	6378160	6378137
F	0.00335292371299	0.00335281068118
M	0.00344980143430	0.00344978600308

The difference between the linear and second order free-air reductions reaches a maximum of -4.986μm/sec² at H≈ 8848m and a latitude of θ=27°58'.

3.2.5 Bouguer correction

The simple bouguer correction accounts for the gravitational attraction of the layer of the earth between the vertical datum (mean sea level) and the station. This correction in milligals traditionally is calculated assuming the earth between the vertical datum and the station can be represented by an infinite horizontal slab given by:

$$Bc = 2\pi G\rho h$$

Where G is the gravitational constant given as $6.673 \pm 0.001 \times 10 \wedge -11$ m³/kg/s², h, is height to a reference datum, ρ is the density and is dependent on the material making up the spherical cap. In local surveys, this value is determined by the density of the geological materials between the survey and the local vertical datum. In regional and continental databases, a mean density is used for the spherical cap, typically 2670kg/m³ for the solid earth, 1027kg/m³ for sea water, 1000kg/m³ for fresh water and 917 kg/m³ for ice. The simple Bouguer slab is only a first approximation to a spherical cap having a thickness equal to the station elevation and the standard density. The curvature correction, adds the remaining terms for the gravity effect of the spherical cap to the theoretical gravity. Recent publication of an exact solution for the curvature correction (LaFehr, 1991b) has brought new attention to this second step in Bouguer reduction. In this revised procedure, to account for the effect of the curvature of the earth, the horizontal slab equation is replaced by the closed-form formula for a spherical cap of radius 166.7km. That is, $\Delta\, gbc = 2\pi G\rho(\mu h - \lambda R)$ where μ and λ are dimensionless coefficients defined by LaFehr (1991), R is the radius of the earth $(Ro + h)$ at the station where Ro is the mean radius of the earth.

A terrain-corrected Bouguer anomaly is called a complete Bouguer anomaly, where the terrain represents the deviations from the uniform slab of the simple Bouguer correction and the spherical cap of the curvature correction. An excess of mass due to terrain above the station reduces the observed gravity as does a deficiency of mass due to terrain below the station. An exception occurs when airborne gravity is being reduced to the level of the ground surface (as opposed to the flight surface). In this case terrain corrections can have either sign in rough topography.

3.2.6 Terrain correction

The terrain correction accounts for variation in the observed gravitational acceleration caused by variations in topography near each observation point. An excess of mass resulting from terrain above the station reduces the observed gravity as does a deficiency of mass resulting from terrain below the station. The application of terrain corrections to gravity data minimizes rugged topographic effects that may cause errors in the gravity anomalies of tens of milligals. Efficient and effective use of these corrections requires comprehensive digital elevation models and computational power that is only now becoming generally available. In rough topography, the magnitude of terrain corrections can exceed 10 mGal, and their accuracy is limited by the ability to estimate inner zone terrain corrections precisely in the field and the quality of the digital elevation model. Generally, a single density is used for terrain corrections. Methods using variable surface density models have been proposed by Vajk (1956) and Grant and Elsaharty (1962).

3.2.7 Eõtvös correction

The attraction of the earth at a point fixed with respect to the earth is reduced by the centrifugal force related to the earth's rotation. This implies that the angular velocity of an observer moving east is always greater than for an observer remaining stationary with respect to the earth's surface and consequently , gravitational attraction will be slightly increased for an observer moving in a westerly direction. This motion related effect called the Eötvös effect most be accounted for in gravity measurements made on moving platforms such as ships or aircraft. The Eötvös correction is given by

$$g_e = 7.503 v \cos\lambda \sin\alpha + 0.004154 v^2$$

Where v is in knots, α Heading with respect to true north, λ is the latitude, g_e is in mgal.

The Eötvös correction can reach significant magnitudes in applications involving moving platforms. Errors in heading or velocity can produce errors in the reduced gravity measurements that are similar in magnitude to anomalies caused by typical crustal sources. The often neglected Eötvös correction is the limiting factor in the precision of shipborne and airborne surveys.

3.2.8 Isostatic correction

The isostatic compensation correction is the gravity effect derived from a geologic model based on the theory of isostasy that regional topographic variations are compensated by density changes of the lithosphere such that at some depth, the earth is in hydrostatic equilibrium. As a result of isostasy, a strong inverse correlation exists between regional terrain and Bouguer gravity anomalies. To minimize the effect of these regional subsurface variations on anomalies, their gravitational response is modeled with the simplifying assumption that crustal thickness varies directly with local topography. This isostatic compensation correction is calculated in a manner similar to the terrain effect using the Airy-Heiskanen model (Heiskanen and Vening Meinesz, 1958)

3.2.9 Latitude correction

The latitude correction is intended to eliminate the centrifugal acceleration that affects observed gravity, and which is a function of latitude. It also accounts for the oblate elliptical shape of the Earth (fig.3.4). The latitude correction is usually calculated from an international gravity formula (IGF), whose constants are based upon the mean Earth ellipsoid adopted by the International Association of Geodesy (IAG). This mean Earth ellipsoid is chosen such that its defining physical and geometrical parameters closely model those of the Earth (Chovitz, 1981). The most recent mean Earth ellipsoid adopted by the IAG is the Geodetic Reference System 1980 or GRS80 (Moritz, 1980), and which supersedes the Geodetic Reference System 1967 (GRS67).

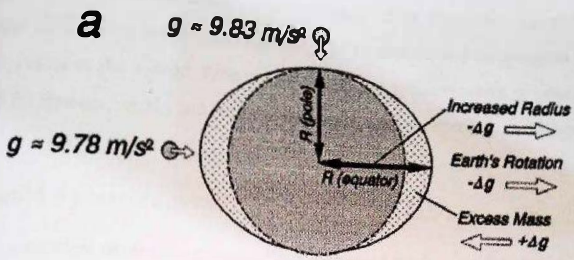

Fig. 3.4 shape of the earth.

Geodetic reference system formulae refer to theoretical estimates of the earth's shape. From these estimate, the international gravity formulae is obtained amongst several others over the years. The first internationally accepted IGF was that of 1930 given as $g\theta = 9.78049(1 + 0.0053024\sin^2\lambda - 0.0000059\sin^2 2\lambda)$. This was found to be in error by about 13mgals. With the advent of satellite technology, much improved values were obtained. This was followed by the 1967 reference formula as

$$g\theta = 978031846(1 + 0.0053024 sin^2\lambda - 0.0000058 sin^2 2\lambda) \ldots\ldots\ldots\ldots\ldots (3.1)$$

Where λ is the latitude.

The international adapted reference system of 1980 leading to the world geodetic system 1984 (WGS84) is given as

$$g\theta = 9.7803267714(1 + \frac{00193185138639 sin^2\lambda}{\sqrt{1 - 0.00669437999013 sin^2\lambda}}) \ldots\ldots\ldots\ldots\ldots (3.2)$$

This reference formula is subtracted from the observed gravity data and thus correcting for variation of gravity with latitude.

3.2.10 Gravity anomalies

Anomalies are the difference between the observed gravity, typically its absolute value corrected for temporal and spatial variations, and the modeled or theoretical gravity at the site of the observation. Different types of anomalies reflect variations in the components used in defining the modeled gravity at the station. Free-air anomalies are measurements or observed gravity values corrected for free-air, terrain and latitude variations. Free-air anomaly is mathematically expressed as

$$\Delta gFA = observed\ gravity - lattitude + free\ air\ correction \ldots\ldots.. (3.3)$$

Bouguer anomalies are simply measurements corrected for free-air, Bouguer slab, terrain, isostatic and latitude. It is expressed as

$$\Delta gBA = lattitude\ correction + free\ air\ correction \pm bouguer\ correction + terrain\ correction + isostatic\ correction \ldots\ldots\ldots\ldots (3.4)$$

3.2.11 Regional –residual gravity separation

This is a common first step in data processing /interpretation. Anomalies of interest are commonly superposed on a regional field caused by sources larger than the scale of study or too deep to be of interest. Historically this problem was approached along two lines by either using a simple graphical approach (selecting manually data points to represent a smooth regional field) or by using various mathematical tools to obtain the regional field. Many of the historical methods are still in common use today. The graphical approach initially was limited to analyzing profile data and to a lesser extent, gridded data. The earliest nongraphic approach used a regional field defined as the average of field values over a circle of a given radius with the residual being the difference between the observed value at the center of the circle and this average. (Griffin 1949). Henderson and Zietz (1949) and Roy (1958) showed that such averaging was equivalent to calculating the second vertical derivative except for a constant factor but had the advantage of being able to identify smaller anomalies. Several methods have since been proposed with acclaimed criticism by various earth scientists. It seems that significantly better results can be obtained by using appropriate statistical geological models rather than by attempting to adjust band-pass filters parameters manually as proposed by many. There is no single right answer for how to highlight one's target of interest via Regional-Residual separation methods. Its major advantage however is that it gives a regional component of the gravity field similar to the one obtained from a graphical separation.

3.3 Test Example

Some data collected along a north-south gravity profile with distances measured from the south end of the profile whose latitude is 51°12'24"N was as shown in Table 3.2. The lacoste-Ramberg gravimeter was used for the survey. Before, during and after the survey, readings were taken at a base station, BS where the value of gravity is 981144.22 mGal. This was done in order to monitor instrumental drift and to allow the absolute value of gravity to be determined at each observation point.

Table 3.2: Test Example: Gravity Data Obtained

STATION	TIME	DISTNCE	ELVTN	READNG
BS	805			2934.2
1	835	0	84.26	2946.32
2	844	200	86.85	2941.08
3	855	400	89.43	2935.76
4	903	600	93.08	2930.45
1	918			2946.58
BS	940			2934.95
1	1009			2946.32
5	1024	800	100.37	2926.6
6	1033	1000	100.91	2927.91
7	1044	1200	103.22	2920
8	1053	1400	107.35	2915.16
1	1111			2946.58
BS	1145			2935.28
1	1214			2946.26
9	1232	1600	110.1	2911.52
10	1242	1800	114.89	2907.29
11	1300	2000	118.96	2904
12	1315	2200	125.18	290.56
1	1330			2946.32
BS	1400			2935.56

CHAPTER FOUR: RESULTS AND DISCUSSION

4.1 Results

The gravity correction to the test example is shown in table 4.1 below. The density of 2.68 g/cm^3 was adopted for the bouguer correction. A series of sections illustrating the variation in topography, observed gravity, free air anomaly and bouguer anomaly along the profile are drawn. The normal or latitude dependent gravity was gotten using various latitudes and tabulated. Bouguer and free air gravities were gotten and hence, their anomalies.

Table 4.1: Gravity Readings Corrected For Temporal and Spatial Based Variations of GRS67

station	TIME	DISTANCE	ELVTN	READNG	INTER.RD	DRIFT	OBSERVD	ADJUSTED	FAC	BC	FAG	BG	NORMALG.	FA	BA
BS	805			2934.2	2934.2	0	981144.2	981144.2					981144.22		
1	83	0	84.26	2946.32	2934.43	11.89	981156.1	981156.2	26	-9.46	981182.1	981172.65	981176.55	5.56	-3.9
2	844	200	86.85	2941.08	2934.5	6.58	981150.8	981150.9	26.8	-9.75	981177.6	981167.85	981182.97	-5.37	-15.12
3	855	400	89.43	2935.76	2934.59	1.17	981145.4	981145.4	27.59	.10.04	981173	981162.94	981195.11	-22.13	-32.17
4	903	600	93.08	2930.45	2934.65	-4.2	981140	981140	28.72	-10.45	981168.7	981158.22	981204.83	-36.16	-46.61
1	918			2946.58	2934.72	11.86	981156.1	981156.2					981176.55		
BS	940			2934.95	2934.95	0	981144.2	981144.2					981144.22		
1	1009			2946.32	2934.27	12.05	981156.3	981156.2					981176.55		
5	1024	800	100.37	2926.6	2934.32	-7.72	981136.5	981136.6	30.97	-11 27	981167.5	981156.2	981214.53	-47.06	-58.33
6	1033	1000	100.91	2927.91	2934.34	-6.41	981137.8	981137.9	31.14	-11.33	981169	981157.62	981224.22	-55.27	-66.6
7	1044	1200	103.22	2920	2934.37	-14.37	981129.9	981129.9	31.85	-11.59	981161.7	981150.11	981233.9	-72.2	-83.79
8	1053	1400	107.3	2915.16	2934.39	-19.23	981125	981125	33.12	-12.06	981158.1	981146.05	981240.89	-82.78	-94.84
1	1111			2946.58	2934.44	12.14	981156.4	981156.2					981176.55		
BS	1145			2935.28	2935.28	0	981144.2	981144.2					981144.22		
1	1214			2946.26	2934.26	12	981156.2	981156.2					981176.55		
9	1232	1600	110.1	2911.52	2934.29	-22.77	981122	981122	33.97	-12.37	981155.9	981143.55	981253.25	-97.33	-109.7
10	1242	1800	114.89	2907.29	2934.32	-27.03	981117.2	981117.2	35.45	-12.91	981152.6	981139.73	981262.91	-110.27	-123.18
11	1300	2000	118.96	2904	2934.35	-30.35	981113.9	981113.9	36.71	-13.36	981150.6	981137.22	981272.56	-121.98	-135.34
12	1315	2200	125.18	2900.56	2934.38	-33.82	981110.4	981110.5	38.63	-14.06	981149	981134.97	981290.08	-141.05	-155.11
1	1330			2946.32	2934.41	11.91	981156.1	981156.2					981176.55		
BS	1400			2935.56	2935.56	0	981144.2	981144.2					981144.22		

4.2 Discussion

4.2.1 Gravity data

Fig 4.1 shows a gravity profile gotten by plotting the values of the bouguer anomaly against the station distances as obtained from the test example. It is one of the many results expected in gravity surveys particularly when plotted this way. Such plots reflect the effects of two major factors. The regional field is represented in this plot and the residual (or local) field is the field outcome of the process. The residual field is usually the field of interest.

The separation of the two fields (regional and residual) from the observed field (bouguer gravity anomaly plotted here) is a process that is necessarily interpretational. The rather simple way of doing this is to visually smooth the curve in a picture of the field that is not complex. For example, the field of Figure 4.1 is the one where the regional field is dominant and a simple surface gives this field while the deviations from this surface reflects the content of the residual field. The residual field is sloping at the rate of -0.05mGal/m.

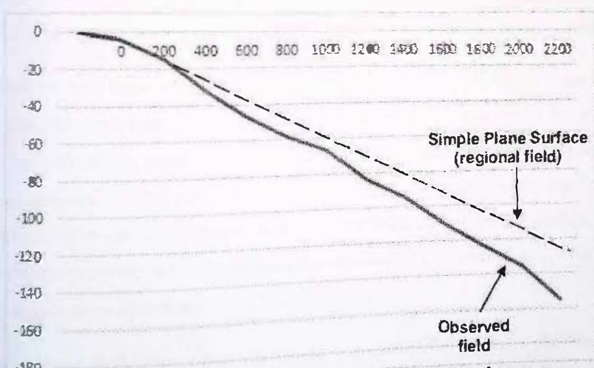

Fig 4.1 Bouguer gravity anomaly plotted against profile distance.

4.2.2 Interpretation procedures of gravity data

Anomalies in gravity method are widely used to provide information on structures located beneath the earth's surface. Different rock types which occur in, and beneath the earth's crust have different densities, so that mass is by no means uniformly distributed in the outer part of the earth. Over areas of mass excess observed gravity is greater than normal and there is a tendency towards positive anomalies. The aim of geophysicist is to deduce from the pattern of anomalies the location and form of the structures which produces the disturbance in gravity. Normally if the mass distribution is of primary interest, it will be the bouguer anomalies which are considered and these will be plotted in the form of profiles or contour maps. Modelling gravity data in profile form is useful in the calculation of the depth of various features and can be done by either forward or inverse algorithms. The attraction of relatively local features is often seen only as a minor distortion of the pattern due to some major structures (local and regional) ,classified generally as local and regional bodies. The large scale geologic structures are not usually of interest in routine exploration. The gravitational acceleration produced by these large scale features is referred to as the regional gravity anomaly. The second contribution is caused by small scale structures and the gravitation due to it is referred to as the local or residual gravity anomaly.

There are many methods of regional-residual bouguer gravity anomaly separation. The methods range from graphical estimates (visual smoothening) to mathematical estimates (polynomial fitting, analytical continuation, wave number filtering, second vertical derivatives etc.).

There are two basic characteristics of the gravitational field which make a unique interpretation impossible. The first is that the measured value of "g" and hence the reduced anomaly at any station reflects the superimposed influence of many mass distribution. The attraction of relatively local features is often seen only as a minor distortion of the pattern due to some major structure. Interpretation can only proceed after the contributions of different bodies are isolated. This effect is always present, but it becomes most serious in the case of geophysical prospecting, where extremely local structures are of interest.

The second difficulty is that gravity as a potential field, shares the fundamental ambiguity of inverse boundary value problems common to all potential fields. For a given distribution of anomalies on or above the earth's surface, an infinite number of mass distribution can be found which would produce them. At first, the interpretation problem appears hopeless. However, geological reasonableness will often rule out whole classes of solutions and other information such as the probable density or depth of the source of the field may lead one to the most likely mass distribution.

Ideally, the final outcome of data interpretation is a physical property map known as the isogal map. Isogal maps look a lot like topographic maps. They show circular elongated and irregular areas 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 and qualitative interpretation if something is known about the geology. Gravity highs are in many areas associated with anticlines and indicates the presence of heavy basic intrusions.

For an indirect approach to gravity interpretation, the interpreter must have available a selection of forms whose attraction can be computed. Indirect interpretation involves four steps:

i. construction of a reasonable model
ii. Computation of its gravity anomaly.
iii. Comparison of computed with observed anomaly
iv. Alteration of model to improve correspondence of observed and calculated anomalies.

The process is thus iterative and the goodness of fit between observed and calculated anomalies is gradually improved. Bodies of complex geometry in two or three dimensions are not so simply dealt with and in such cases it is advantageous to employ techniques which perform the iteration automatically. The most flexible of such methods is nonlinear optimization (Al-chabi, 1972). All variables (body points, density contrasts, and regional field) may be allowed to vary within defined limits.

For many years, the accepted method of interpretation was to assume various simple shapes for the source of an anomaly, compute their effects at the surface and modify them until a fit with the observed field is obtained. The achieving of a fit indicated only that the selected model was a possible solution. This cut and try process of interpretation is often called the indirect method of interpretation. In spite of its lack of elegance, shapes are readily available and a quick comparison with the observed anomaly is possible. In the case of preliminary surveys, with observations of limited number or uncertain accuracy, this procedure may be all that is justified. On the other hand, if in a certain area there is complete coverage, with stations of high accuracy, it may be desirable to employ a more direct approach. Methods have been developed in recent years to extract information on the mass distribution by mathematical operations on the observed field. These methods which generally require the use of high speed computers cannot reduce the fundamental ambiguity mentioned ab ve. Some parameters of the unknown structure must be assumed at the start. Under the assumptions, they attempt to extract the maximum information from the field.

Several techniques, very popular in the interpretation of magnetic anomalies, can easily be adapted to gravity data.

A. Forward Modeling

The publication by Talwani et al. (1959) of equations for computing gravity anomalies produced by two-dimensional bodies of polygonal cross section provided the impetus for the first use of computers for gravity modeling. The two-dimensional sources were later modified to have a finite strike length (Rasmussen and Pederson, 1979; Cady, 1980), and this led to publicly available computer programs for 2½-D gravity modeling (Wehring, 1985; Saltus and Blakely, 1983, 1993). Three dimensional density distributions were initially modeled by Talwani and Ewing (1960) using thin horizontal polygonal plates. Plouff (1975, 1976) showed that the use of finite thickness horizontal plates was a practical and preferable alternative. Right rectangular prisms (Nagy, 1966) and dipping prisms (Hjelt, 1974) remain popular for building complex density models, especially as inexpensive computers become faster. Barnett (1976) used triangular facets to construct three-dimensional bodies of arbitrary shape and compute their gravity anomalies, whereas Okabe (1979) used polygonal facets.

Parker (1972) was the first to use Fourier transforms for the calculation of 2-D and 3-D gravity anomalies from complexly layered two-dimensional models. Because the gravity anomaly is calculated on a flat observation surface above all the sources, this approach is particularly well suited to modeling marine gravity data. Fourier methods can provide an alternative to spatial domain approaches for modeling simple sources such as a point mass or uniform sphere, a vertical line mass, a horizontal line mass, or a vertical ribbon mass (Blakely, 1995). Blakely (1995) also presented theory and computer subroutines for computing gravity fields of simple bodies in the spatial domain, including a sphere, a horizontal cylinder, a right rectangular prism, a 2-D body of polygonal cross section, and a horizontal layer. Today forward gravity modeling is often done using commercial software programs based on the theoretical papers and early software efforts mentioned above, but incorporating inversion algorithms and sophisticated computer graphics. A relatively recent development in forward modeling is the concept of "structural geophysics" (Jessell et al., 1993; Jessel and Valenta, 1996; Jessell, 2001; Jessell and Fractal Geophysics, 2002), in which a layered earth model having specified physical properties is subjected to a deformation history involving tilting, faulting, folding, intrusion, and erosion. The resulting gravity field is computed using deformed prisms based on the model of Hjelt (1974).

B. Gravity Inversion

For the purpose of this review, inversion is defined as an automated numerical procedure that constructs a model of subsurface geology from measured data and any prior information independent of data. Quantitative interpretation is then carried out by drawing geologic conclusions from the inverted models. A model is either parameterized to describe so rce geometry or the model is described by a distribution of a physical property, such as density contrast or magnetic susceptibility. The development of inversion algorithms naturally followed these two directions. Bott (1960) first attempted to invert for basin depth from gravity data by adjusting the depth of vertical prisms through trial and error. Danes (1960) uses a similar approach to determine the top of salt. Oldenburg (1974) adopted Parker's (1972) forward procedure in Fourier domain to formulate an inversion algorithm for basin depth by applying formal inverse theory. A number of papers followed on the same theme by extending the approach to different density-depth functions or imposing various constraints on the basement relief (e.g., Pedersen, 1977; Barbosa et al., 1997; Chai and Hinze, 1988; Reamer and Ferguson, 1989; Guspi, 1992).

Recently, this general methodology has also been used extensively in inversion for base of salt in oil and gas exploration (e.g., Jorgensen and Kisabeth, 2000; Nagihara and Hall, 2001; Cheng et al., 2003). A similar approach has also been used to invert for the geometry of isolated causative bodies by representing them as polygonal bodies in 2D or polyhydronal bodies in 3D (Pedersen, 1979; Moraes and Hansen, 2001), in which the vertices of the objects are recovered as the unknowns. Alternatively, one may invert for the density contrast as a function of position in subsurface. Green (1975) applies the Backus-Gilbert approach to invert 2D gravity data and guides the inversion by using reference models and associated weights constructed from prior information. In a similar direction, Last and Kubik (1983) guides the inversion by mimimizing the total volume of the causative body, and Guillen and Menichetti (1984) choose to mininuze the inertia of the body with respect to the center of the body or an axis passing through it. While these approaches are effective, they are limited to recovering only single bodies. Li and Oldenburg (1998) formulate a generalized 3D inversion of gravity data by using the Tikhonov regularization and a model objective function that measures the structural complexity of the model. A lower and upper bound are also imposed on the recovered density contrast to further stabilize the solution. A similar approach has been extended to the inversion of gravity gradient data (Li, 2001; Zhdanov et al., 2004). More recently, there have been efforts to combine the strengths of these two approaches. Krahenbuhl and Li (2002, 2004) formulate the base of salt inversion as a binary problem and Zhang et al. (2004) take a similar approach for crustal studies.

Interestingly, in the last approaches the genetic algorithm (GA) has been used as the basic solver. This is an area of growing interest, especially when refinement of inversion is sought with constraints by increased prior information.

C. Geologic Interpretation

Observed gravity anomalies are direct indications of lateral contrasts in density between adjacent 'vertical columns' of rock. These columns extend from the earth's terrestrial or sea surface to depths ranging from say 10 meters to more than 100 km. The gravity surveying process, in fact, measures any and all lateral density contrasts at all depths within this column of rock. Data filtering, enables us to isolate portions of the gravity anomaly signal that are of exploration interest. These target signatures can then be interpreted together with ancillary geologic information to construct a constrained shallow earth model.

CHAPTER FIVE: SUMMARY AND CONCLUSIONS

5.1 Summary

Modified reduction procedures are incorporated into gravity databases, thereby improving the accuracy and geophysical utility of gravity anomalies. These revisions employed slightly the use of a single horizontal datum for locating gravity stations and the internationally accepted terrestrial ellipsoid as the vertical datum for stations to avoid problems originating in local or regional datums. Changes in the reduction procedures minimize errors attributable to terrain, earth curvature, second-order vertical gradients in gravity, atmospheric mass effects, and differences in the normal gravity and station height datums. Their consistent use improves the accuracy and precision of gravity anomalies, especially in their long-wavelength components.

5.2 Conclusion

The most significant difference in these assessment is the use of the ellipsoid as the vertical datum rather than conventionally used sea level. This leads to minor, but measurable, differences in the absolute values of anomalies, which may cause confusion with previous gravity data. Thus, it is advisable to informally recognize that an anomaly calculated using the revised procedures is termed an ellipsoidal gravity anomaly. The revised procedure is recommended for future gravity reductions and should be used to recalculate anomalies in existing data sets.

5.3 Recommendation

Since gravity method is the most effective means for mapping subsurface geology, the following recommendations are made:

I. Global Position System (GPS) should be used for height measurements rather than the altimeters whose readings are not accurate.

II. Since modified correction procedures are incorporated into the standards of national and regional data bases to improve the accuracy, coverage and geophysical utility, it is important to involve undergraduate students in establishing Base stations throughout the federation.

III. Students should maximize the enhanced computational power and increase interest in gravity methods to upgrade gravity instrumentation for the future of gravitational exploration method.

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