

UNIVERSITY OF GHANA

APPLICATION OF SEISMIC REFRACTION SURVEY METHODS IN FOUNDATION EVALUATION IN PARTS OF LEGON-ADENTA AREAS, GREATER ACCRA REGION



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DECLARATION

This dissertation is the result of the research work undertaken by Kyeremeh Ampaabeng Kwame in the Department of Earth Science, University of Ghana, under the supervision of Dr. Thomas Armah.

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DEDICATION

This dissertation is dedicated to the Almighty God, my supervisor Dr. Thomas Armah, my parents Mr. Samuel Dwaah and Dinah Kyeraa and to the needy in society.



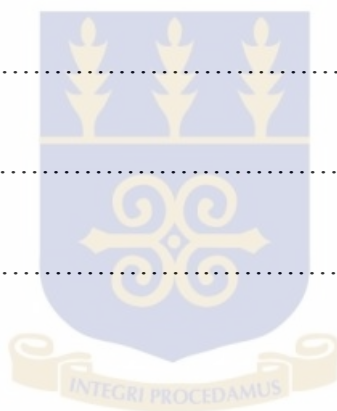
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ABSTRACT

This dissertation sought to find out the use of seismic refraction to determine the depth of the bedrock beneath unconsolidated sediments at Legon in the Accra Metropolitan Area of the Greater Accra Region of Ghana. Seismic refraction which is a traditional geophysical method commonly used to determine the depth to bedrock, competence of bedrock, depth to water table or depth to other seismic velocity boundaries was used to carry out the survey. Since the early 1980s, seismic refraction method has been increasingly used in shallow environmental and engineering site characterization works. This work is based on generation of direct compressive wave using a near surface impulsive energy which propagates through the soil media and it is refracted along stratigraphic boundaries. The signal generated undergoes refraction based on the interface boundary encountered media through which it is propagated. The refracted signal return to the ground surface and time of arrival is recorded by geophones. Four major tasks were carried out in determining the depth of the bedrock beneath unconsolidated sediments. The first part was using arrival time at geophones and the distance from shot points to geophones to plot time – distance graph to obtain intercept on time axis. The second was inferring two different layers due to the two different velocities obtained as the waves propagate through the soil. Also models were generated to observe the two layers as well as the depth of the second layer and thickness of the first layer. Finally the velocities in the layers were used to infer the nature of the layers. These major tasks are carried out and interpreted to produce results that will determine the depth and ability of the subsurface to support foundations of structures.

CHAPTER ONE

1.0 INTRODUCTION

Seismic refraction method is the most commonly conducted geophysical survey for engineering investigation. Seismic refraction provides engineers and geologist with the most basic of geological data through simple procedures with common equipment (Milsom, 2003). It is most commonly used where bedrock is less than 500 ft below the ground surface. Seismic refraction is defined as the travel path of a sound wave through an upper medium and along an interface and then back to the surface (Dobrin, 1976). Refraction surveys are widely used to study water tables and for engineering purposes, the consolidated layers of the ground surface and also in determining near- subsurface for deep reflection traces. The travel time of seismic wave is usually only a few tens of milliseconds and there is a little separation between arrivals of different types of waves that have travelled by different paths. In seismic refraction surveys, only the first arrivals which are always of a P-wave can be picked with any confidence. Ideally the interface studied in small refraction survey should be shallow, roughly planar and dip less than 15° and wave velocities must increase with depth at each interface. The first arrival at the surface will come from successively deeper interface as distance from shot point increases. In shallow refraction works, it is often sufficient to consider the ground in terms of dry overburden, wet overburden and weathered and fresh bedrock. It is very difficult to deal with more than three interfaces (Milsom, 2003).

The P-wave velocity of a dry overburden is sometimes as low as 350 m/s, the velocity of sound in air, and is seldom more than 800 m/s. There is usually a slow increase with depth which is

almost impossible to measure followed by an abrupt increase to 1500 m/s to 1800 m/s at water table (Milsom, 2003).

Fresh bedrock generally has a p-wave velocity of more than 2500 m/s but it is likely to be overlain by a transitional weathering where the velocity, which may initially be less than 200 m/s usually increase steadily with depth and the accompanying reduction in weathering.

In seismic refraction technique, an impulsive seismic source creates a sound wave which travels through the earth. Several options are available for the impulsive source depending on the depth of travel and sufficiently consolidated of the medium of travel. The most common type of source in seismic investigation for site study is sledge hammer or strike plate. Other sources include explosives, shot gun shells detonated in shallow auger holes and various mechanical devices that shake the ground or drop large weights. The types of sources used could also be dependent on the signal versus noise ratio in the survey area. Noise can come from vehicular traffic, people and animals walking near by geophones, electrical current in the ground (electromagnetic interference in the ground which affects the geophone cables), low-flying air craft or any sound source. Generally the noise could be overcome by using large source, which effectively increase the signal. Filtering on the seismograph could also reduce noise (Musgrave, 1967).

The impulsive source generates seismic wave that travel through the subsurface. When wave front reaches a layer of higher velocity (eg bedrock) a portion of the energy is refracted and travel along the interface between the lower layer and high velocity layer (refractor) at a velocity determined by the composition of the refractor. The energy of the propagating wave leaves the refractor and returns to the surface. The angle of refraction of the propagating wave depends on the composition of the refractor and the material of the layer the refractor is in contact with (Hawkins, 1961).

When seismic wave returns to the surface, its arrival is detected by series of geophones and recorded on a seismograph which are interpreted and analyzed for further works. The geological information gained from the seismic investigation can be used in hydro geologic assessment of ground water pollution sites and the surrounding area. The interpretation of seismic data can indicate changes in lithology or stratigraphy, geologic structure, or water saturation (water table) (Dobrin, 1976). Seismic methods are commonly used in determining depth and structure of geologic and hydro geologic units (depth of bedrock or water table), estimating hydraulic conductivity, detecting cavity or voids, determining structure stability, detecting fractures and faults zones and estimating rippability (Coffeen, 1978). The choice of method depends upon the information needed and the nature of study area. A seismograph records the arrivals of the refracted seismic wave with respect to time. These waves are detected at the surface by small receivers (geophones), which transform the mechanical energy of the wave to electric voltages. These voltages are relayed along cable to seismograph which records the voltage output verse time. Each seismograph has different capabilities to handle data that is dependent on a number of channels in the seismograph. Each channel records responds of geophones. The series of geophones are placed along the line at a set interval known as spread. This interval depends on the desired resolution and depth of exploration in order to detect the primary refractor (bedrock). It is necessary to spread geophones three to five times the depth of the overburden. For a normal seismic survey a series of four to six shots are enough for each spread, one at each end or beyond the ends and changes in layers of the overburden material are determined by intermediate shot points which increase the accuracy of the depth to bedrock interpretation. Seismic refraction data obtained is then interpreted graphically or with the aid of a computer. There are multiple

interpretation schemes for seismic refraction data depending upon the method and desired results (Redpath, 1973).

1.1 BACKGROUND OF STUDY

Many structures have demolished while others have sunk into the earth due to the poor consolidation of the sub layer (Sharma, 1997). The combined effect of prolong weathering under warm, humid conditions followed by differential erosion, physical and chemical modification particularly under arid or semi-arid conditions have led to different consolidation of layers near the ground sub-surface, (Godman, 1989). When rocks are exposed to atmospheric conditions they slowly break down. This process is known as weathering. The principal types are mechanical weathering and chemical weathering.

Mechanical weathering causes disintegration of rocks into smaller pieces by exfoliation or decrepitation (slaking). The chemical composition of the parent rock is not or slightly altered. Mechanical weathering can result from the action of agents such as frost action, salt crystallization, biogenic process (plant and animals), unloading of rock masses (sheet jointing), wind glaciers, temperature changes (freezing and thawing), streams and moisture changes (cycle of wetting and drying). Mechanical weathering is very active on high mountains with cold climate. As unloading takes place discontinuities may develop parallel to the surface of the rock outcrops. The rock outcrops appear to be spalling off like layers of giant onion. The rock mass is divided into block or sheets, a few centimeters thick near the ground surface and becoming thicker with depth until it fades out completely at depth of about 50 m (Kehew, 1995).

Chemical weathering creates new minerals in place of the ones it destroys in the parent rock. As rocks are exposed to the atmospheric condition near the ground surface, they react with components of the atmosphere to form new minerals. The most important atmospheric reactants are oxygen, carbon dioxide and water. Other reactants are available in polluted air. Examples include acid rain problems associated with the release of sulfuric acid from coal fire power plant, sulfur dioxide and smoke emission and nitrogen oxides from vehicle exhaust. Minerals in rocks sometimes dissolve completely through certain chemical reactions which bring about weathering. For instance limestone dissolve by meteoric water which contains dissolved carbon dioxide. This result in the formation cavities called dolines or karsts and geologic hazard called sink holes. Because of impurities in limestone a red residual soil remain at the limestone surface called terra rossa (Sowers, 1975). Underground cavities can also be formed in gypsum because of its large solubility (Bruune, 1965). Hydrolysis of acid in rock mineral results in acidic weathering. Feldspars are transformed by hydrolysis as they react with hydrogen ions to form various products including clay minerals. This phenomenon is responsible for the degradation of granite and other plutonic rocks to materials that resemble more dense soil than rocks. After weathering has taken place, the weathered material are transported by running water to low land areas. As these transported materials are deposited, they form layers over existing layer bring about poor consolidation of these particles and increasing the number of horizons above the bedrock. One needs to know the depth and much about these consolidated particles within the layers before putting up any structure such as roads and buildings to avoid sinking and collapsing of the structure

The knowledge about the depth and consolidation of soil particles call for investigating the sub-surface using seismic refraction.

1.2 LOCATION AND CLIMATE

The study area is Legon in the Accra Metropolitan Area of the Greater Accra Region of Ghana which forms part of the areas along the coastal belt with Eastern Region in the north, Central Region in the west, Volta Region in the east and the Atlantic Ocean in the south. This is shown in figure 1.1. The Accra Metropolitan Area has a total land size of two hundred square kilometers and is made up of six sub metros namely Okaikoi, Ashiedu Keteke, Ayawaso, Kpeshie, Osu Klotey and Ablekuma. The southern boundary of the metropolis of Accra Metropolitan Assambly shares boundaries with the Ga District Authority (GDA), Tema Municipal Authority (TMA) and the Gulf of Guinea (Accra Metropolitan Assembly, 2006).

The Accra Metropolitan Assembly lies in the savanna zone. There are two rainy seasons. The first begin in May and ends in mid-July. The second season begins in mid-August and ends in October. There is very little variation in temperature throughout the year. The mean monthly temperature ranges from 24.7⁰C in August to 28⁰C in March with annual average of 26.8⁰C. As the area is close to the Equator, the daylight hours are particular uniform throughout the year. Relative humidity is generally high varying from 65% in the midafternoon to 95% at night. The predominant wind direction in Accra is from the WSW to NNE sectors. Wind speed normally range between 8 km/h to 16 km/h. High wind gusts out with thunder storm activity which passes in squall along the coast (Accra Metropolitan Assembly, 2006).

There is evidence to suggest that the vegetation of the metropolitan areas has been altered in the more recent past century by climatic and other factors. Much of the metropolitan area was believed to have been covered by dense forest of which only a few remnant trees survived. A climatic change combined with the gradient of the plains and cultivation has imposed vegetation

structures similar to those on the southern shale, Sudan and Guinea Savannah all of which lie north of the Accra plains. There are three broad vegetation zones in the metropolitan area which comprise shrub lands, grass land and coastal land. Only the shrub land occurs more commonly in the western outskirts and in the north towards the Aburi hills. It consists of dense cluster of small trees and shrubs which grow to an average height of five meters. The grasses are a mixture of species found in the under growth of forests. They are short and rarely grow beyond one meter. Ground herbs are found on the edge of the shrub. They include species which normally flourish after fire. The coastal zone comprises two vegetation types, wetlands and dunes. The coastal wetlands zone is highly productive and important habitat for marine and terrestrial mainly bird life. Mangroves, comprising two dominant species, are found in the tidal zones of all estuaries and lagoons. The grasslands have important primary protection role in providing nutrients for prawns and juvenile fish in the lagoon systems (Accra Metropolitan Assembly, 2006).

The dune lands have been formed by a combination of wave action and wind. They are mostly unstable but stretch back several hundreds of meters in places. There are several shrubs and grassland species which grow and play important role in stabilizing dunes. Coconut and palms grow well in this zone providing protection and an economic crop. Most of these coconuts were planted since 1920s but it is established that over 80% of those plantations have disappeared as a result of felling disease and coastal erosion. The loss of these trees is one of the principal reasons for the severity of erosion in some areas. In addition to the natural vegetation zones, a number of introduced trees and shrubs thrive in the metropolitan area. Neems, mangoes, cassias, avocados and palms are the prominent trees in the area landscape.

Apart from mangroves and salt marsh grasses, which grow in the intertidal zone, sea grasses or attached algae also occur mainly in rocky areas and wave cut platforms.

These areas have increased as a result of erosion exposing the underlying bedrock - especially to the east of Tema.

Ocean floor sea grasses are confined to a few sheltered areas of the coastline and the lagoons. The ocean floor regime is too unstable to support large areas of sea grass.



Figure 1.1 The map of Greater Accra Region of Ghana.

1.3 GEOLOGY OF STUDY AREA

Ghana falls mostly within the West Africa Craton which stabilized in the early Proterozoic (2000 Ma) during the Eburnean Orogeny. The orogeny also stabilized the Zaire Craton and affected vast western Africa and neighbouring regions in South America that were conterminous with the Eburnean tectonothermal province. Outside South America, the West Africa Craton is the second largest region in Africa where lower Proterozoic rocks are extensively preserved. These early Proterozoic rocks comprise extensive belts of metamorphosed volcanic and sedimentary rocks exposed in Ghana, Burkina Faso, Niger and Cote d'Ivoire. Recent review of the geology of Ghana by Kesse (1985), present a good treatise on specific minerals and rocks resource available in Ghana and Leub et al (1990), present a significant different stratigraphic interpretation for the Birimian system in Ghana, stressing lateral lithologic continuity and facies changes within the group. He believed that some of the granitoids possess significant potential for gold mineralization.

Ghana can be divided geologically into several distinct terranes. Among these are;

An early Proterozoic terrane (Birimian system) which hosts most of the country's mineral deposits and occupies the western and northern parts of the country.

The Tarkwaian System, a distinctive sequence of clastic sediments within the Ashanti, Bui and Bole-Navrongo Belts.

The Voltaian Basin, in which are preserved the late Precambrian to Paleozoic sediments that mantle the craton.

The Dahomeyan System, occupying the easternmost part of Ghana

A Pan-African mobile belt, the Togo and Buem Formations, separated from the Birimian terrane by a prominent topographic feature known as the Akwapim-Togo range.

Phanerozoic Sedimentary rocks

Intrusive rocks

The area under study falls within the Togo unit which represents the cover rock of the basement Dahomeyan gneisses. The Dahomeyan system is a part of the second major tectono-stratigraphic terrane in Ghana; it underlies eastern and southeastern Ghana. The Dahomeyan is the easternmost rock group in Ghana and differs significantly from other rocks in Ghana in that it is composed of high grade metamorphic rock. The system consists of four lithologic belts of granitic and mafic gneiss. The mafic gneisses are relatively uniform oligoclase, andesine, hornblende, salite and garnet gneisses of igneous parentage and generally tholeiitic composition (Holm, 1974). The granitic gneisses interlayer with the mafic gneiss and are believed to be metamorphosed volcanoclastic and sedimentary rocks. Persistent bands of nepheline gneiss in the system appear to be metamorphosed calc-alkaline igneous rocks (Holm, 1974).

Recent works by Attoh (1998); Attoh et al. (1997) and then Agbossoumonde et al. (2004) consider the Dahomeyan as consisting of various structural units based on age and tectonics. The principal litho-tectonic units identified are: (i) quartzo-feldspathic and augen-gneisses referred to locally as the Ho gneisses; (ii) a suture zone of distinct mafic and ultramafic rocks, and (iii) a granitoid gneiss-migmatite assemblages east of the suture zone.

The Ho gneisses represent the deformed edge of the West African Craton and includes augen gneisses and mylonitic gneisses most of which are hornblende-rich and often sheared and at some localities, for example Anyirawase-Tsibu pass and Klave, overthrust the Togo unit. Doleritic and dioritic intrusives are common along contact zones with the Togo cover rocks and may represent late stage magmatism associated with the Pan- African orogen. The Ho gneiss recorded ages of 2176 ± 44 Ma on whole rock Rb-Sr isochron (Agyei et al., 1987).

The present acceptable nomenclature therefore presents the Dahomeyan broadly as consisting of different structural units. The gross structure of the Dahomeyan is that of alternating northeast trending lithologic belts with moderate dip to the southeast. Along the western boundary of the belt of the gneisses are in fault contact and over thrust onto rocks of the Buem-Togo Belt.

The rocks of the Togo unit outcrop on the Akwapim ranges in Ghana, from the mouth of the Densu River near Accra on the Atlantic coast to the boundary with the republic of Togo. The hills continue eastwards as the Togo Mountains in the republic of Togo and as the Atacora Mountains in the republic of Benin. However, the term 'Togo', 'Akwapimiam' and 'Atacora' all refer to the same unit (Aicard, 1957). The Ho-Abutia hills and a few other hills to the east of the Akwapim range are considered as outliers of the Togo.

The Togo unit represents cover rocks of the basement Dahomeyide gneisses. The contact between the Togo and the Dahomeyan is to the east. The unit grades from the east to west from phyllite and chlorite schist upwards into quartzite, micaceous quartzite and sandstone. The geology of the Togo unit has been described by Junner (1940). The main lithologies within the Togo unit are quartzites, schists, phyllites and phyllonites. Minor amounts of shale and siliceous limestone occur within the Togo unit. The quartzites range from being thickly to thinly foliated and flaggy; with the foliation planes often marked by flattened quartz grains and muscovite. Recrystallized quartz is often strained especially in areas where they are in contact with the basement Dahomeyan rocks. Tourmalinization of the quartzites is often observed in these contact zones. Phyllites and phyllonites occur at the areas of thrust contact between the Togo and the Buem unit and then Dahomeyan unit or as thin intercalations in the quartzites.

Generally the quartzites and the phyllites within the Togo unit are intensely deformed, and mostly occur as craton verging recumbent folds with regionally pervasive sub-horizontal

foliation. Regarding the age of the Togo unit, Atttoh et al. (1997) reported $40\text{Ar}/39\text{Ar}$ dates from muscovite in the quartzites in Ghana to be $579.4 \pm 0.8\text{Ma}$, and 600 km north in the Republic of Togo, muscovite in the rocks give a spectrum with minimum ages being $608.1 \pm 1.2\text{ Ma}$.

1.4 STATEMENT OF THE PROBLEM

Foundation of structures transmit load from the structures to the earth. The foundation design is based on the load characteristics of the structure and the properties of the soil or the bedrock at the site. The primary considerations for the foundation supports are the bearing capacity, settlement and ground movement beneath the foundations. Bearing capacity is the ability of the soil to support the loads imposed by the building or structures. The load transmitted from structural foundations to the soil compresses the soil so much one needs to determine the soil parameters through field and lab testing eg consolidation test, triaxial shear test, vane shear test, stiffness test, standard penetration test etc before erecting any structure (Bowles, 1988).

Compressive strength is the capacity of the soil material to withstand axially directed compressive forces. The most common measure of compressive strength is uniaxial compressive strength or unconfined compressive strength. Usually the compressive strength of soil material defines the ultimate stress. It is one of the most important mechanical properties of the soil used in the design and analysis of foundation of structures (Budhu, 2007).

In addition, site investigation will always include subsurface sampling and laboratory testing of the soil samples retrieved. The digging of test pits and trenching (particularly for locating faults and slide plane), cone pick may also be used to learn about the soil condition at depth.

However these testing are time consuming and are limited to particular points in the site.

Despite all these several methods of studying the subsurface, several structures still sink into the earth due to the loose nature of the subsurface. This sinking of structures has led to the tilting of several tall structures, such as the Leaning Tower of Pisa. Several of these sinking and tilting of structures has prompted scientist to begin taking a more scientific-based approach to examining the subsurface (Bowles, 1988). The use of geophysical exploration technique (measurement of seismic wave (pressure, shear and Raleigh waves)) will help to study and estimate the properties of the soil in a larger area and the depth of the bedrock in order to determine the ability of the subsurface to support the load from structures is the main aim of this dissertation.

1.5 OBJECTIVE

- Determining the depth of the bedrock beneath unconsolidated sediments within a large area and identifying the properties of the unconsolidated sediments and the competence of the bedrock to support structures using the Intercept-time Method (I.T.M) and General Reciprocal Method to development 2D models for the study.

CHAPTER TWO

LITERATURE STUDY

There are several geophysical methods used in studying the subsurface. These methods measure the physical properties of the subsurface to infer the underneath. Measurements in geophysical surveys are made in the field. Field theories of gravity, magnetic and electromagnetic work, and even particle fluxes and seismic wave fronts could be described in terms of radiation fields. The fields used in geophysical surveys may be natural ones (eg the earth's magnetic or gravity fields) but may be created artificially as when alternate currents are used to generate electromagnetic fields. This leads to a broad classification of geophysical methods into passive and active types respectively.

In geophysical survey, signals are always generated from various sources depending on the method used. Usually many readings are recorded from the regional background before interpretation can begin. Interpretation tends to concentrate on anomalies thus on the differences from a constant or smoothly varying background. Geophysical anomalies take many forms, for example a massive sulphide deposit containing pyrrhotite would be dense magnetically and electrically conductive. The result of a survey along traverse lines are often presented in the profile form and fed into the computers to be plotted. This type of presentation is particularly helpful in identifying anomalies due to manmade or natural features. There are various geophysical methods of studying the subsurface and these methods are discussed below (Milsom, 2003).

Differences in rock density produce small changes in the earth gravity field than can be measured using portable instruments known as gravity meter or gravimeter. The gravitational constant G has a value of $6.67 \times 10^{-11} \text{Nm}^2\text{kg}^{-2}$. Gravity fields are equivalent to acceleration for which the S.I unit is the m/s^2 . This is conveniently large for geophysical work and gravity unit. The earth's gravity field is almost the same as that of a sphere having the same average radius and total mass but increase slightly towards the poles. The relationship between normal and sea level gravity and the latitude is described by the international gravity formula adopted in 1967.

$$g_{\text{norm}} = 9780318.5 + 51629.27 \sin^2 \lambda + 229.5 \sin \mu \lambda \text{ (Nettleton, 1976).}$$

In gravity work, more than any other branch of geophysics, large and calculable effects are produce by source which is not of geological interest. These effects are removed by reductions involving sequential calculations of a number of recognized quantities. In each case the sign of reduction is opposite to that of the effect it is designed to remove. A positive effect is one that increases the magnitude of the measured field. The major reductions carried out are Latitude, Free-air and Bouguer corrections (Blaricom, 1992).

Although magnetic method is governed by the same fundamental equation, magnetic and gravity survey are very different. The magnetic properties of adjacent rock masses may differ by several orders of magnitude rather than a few percent. A body placed in a magnetic field acquires a magnetization which if small is proportional to the field. The susceptibility of a rock usually depends on its magnetite content. Sediments and acid igneous rocks have small susceptibilities whereas basalts, dolerites, gabbros and serpentinites are usually strongly magnetic. Weathering generally reduces susceptibility because magnetite is oxidized to magnemite and remanently magnetized hematite. In geophysics the terms north and south used to describe polarity are

replaced by positive and negative. The direction of magnetic field is conventionally defined as the direction in which a unit positive pole would move but, since all things are relative, geophysicist give little thought to whether it is north or south magnetic pole (Hammer, 1939).

In a magnetic survey absolute numerical readings are obtained at a touch of a button with proton and caesium magnetometer. Error is minimize when sensors are on 3m poles than when, as with fluxgates are held close to the body. At the start of each survey day, the diurnal magnetometer is set up. The first reading of the field magnetometer is taken at the base or sub-base and it is made at the same time reading is being taken manually or automatically. The field readings are taken twice and small differences are ignored but large differences are investigated. The readings are then recorded into note books.

The field interpretation allows areas needing infill to be identified and then revisited immediately and at little cost. Good interpretation requires profiles which preserve all the detail of the original readings, and contour maps, which allows trends and patterns to be identified. Fortunately, the now almost ubiquitous laptop PC has reduced the work involved in contouring (Milsom, 2003).

In the radiometric method the radioactivity of rocks is monitored using gamma-ray scintillometers and spectrometers. Although most radiometric instruments are developed with uranium search in mind, other uses among regional geological mapping and correlation are available.

Spontaneous radioactive decay produces alpha, beta and gamma radiation. Alpha and beta rays are actually particles; gamma rays are high-energy electromagnetic waves which can be treated as if composed of particles (Milsom, 2003).

An alpha particle consists of two protons held together by two neutrons to form a stable helium nucleus. The emission of alpha particle is the main process in radioactive decay resulting in a decrease of four in atomic mass and two atomic numbers. The particles have large kinetic energy but rapidly slow down due to collision with other atomic nuclei. The average distance travelled in solid rock before this occurs is measured in fractions of millimeter.

Beta particles are electrons ejected from atomic nuclei they differ from other electrons only in having higher kinetic energy and so cease to be identifiable after being slow down by multiple collisions. Energy is lost most rapidly in collision with other electrons. In solids or liquids the average range of a beta particle is measured in centimeters (Turner et al., 1993).

Gamma rays are electro-magnetic rays with frequencies so high that they are best regarded as consisting of photons with energies proportional to the frequencies. The average range of gamma rays is generally considered to start about 0.1 MeV. Gamma rays are electrically neutral and therefore photons penetrate much greater thickness of rocks than do either alpha or beta particles and are consequently the most geophysical useful form of radiation. Gamma rays provide radiation on the presence of unstable atomic nuclei. Ground radiometric surveys turned to be rather frustrating operations because of the shielding effect of even thin layers of rock or soil. It is very easy to overlook radioactive minerals in rocks that are only patchily exposed at the surface. Reliance on stations placed at uniform distance along the Travers maybe unwise and so the observer needs to be aware of the environment. Accurate radiometric data can be obtained only by occupying stations long enough for statistical variation to average out. Because gamma rays are strongly absorbed by both rocks and soil comprehensive notes are taken during radiometric surveys. Departures from 2π geometry are always noted together with details of soil cover areas where rocks cannot be seen. Attempts are made to decide whether overburden

was transported into the place or developed in situ to estimate its thickness. Readings are taken with time constant or count period and recorded. Field techniques are varied in the course of the survey and the location of the sensor is specified (Clark, 1996).

Many geophysical surveys rely on measurement of voltages and magnetic field associated with electric current flowing in the ground. The current could be sustained by natural oxidation reduction reaction or variations in ionospheric magnetic field or can be generated artificially. The current is being made to flow by direct injection, capacitive coupling or by electro-magnetic induction. Surveys involving direct injection are through electrodes at the ground surface and are generally referred to as direct current survey. Current that is driven by electric fields acting either through electrodes or capacitatively is sometimes called galvanic. Surveys in which the current are made to flow inductively are referred to as electro magnetics or EM surveys. Natural current methods include the use of natural potential (self-potential and induced polarization) methods, EM surveys using local source and VLF and CSAMT surveys which use plane waves generated by distance transmitters (Milsom, 2003).

The resistivity method measures the ground resistivity by simultaneously passing current and measuring voltage between a single pair of grounded electrodes. The contact resistance depends on things such as ground moisture and contact area which may amount to thousands of ohms. The problem can be avoided if the voltage measurements are made between a second pair of electrode using high impedance voltage. Such a voltmeter draws virtually no current and the voltage drops through the electrodes is therefore negligible. A geometric factor is needed to convert the readings obtained with these four electrode arrays to resistivity. The result of any single measurement with any array could be interpreted as due to homogeneous ground with a

constant resistivity. The geometric factor used to calculate this apparent resistivity can be derived from the formula; $V = \rho I / 2\pi a$ for electric potential (V) at distance (a) from a point electrode at the surface of the uniform half-space (homogeneous ground) of resistivity ρ . The current (I) may be positive (if into the ground) or negative (Barker, 1981).

Arrays are usually chosen at least partly for their depth penetration which is almost impossible to define because the depth to which a given fraction of current penetrates depends on the layering as well as the separation between the current electrodes. Voltage electrode positions determine which part of the current is sampled and the penetration of the Wenner and Schlumberger arrays are thus likely to be very similar for total array lengths. The quantitative determination of the resistivity change would require much greater expansion for any array there is also an expansion at which the effect of a thin horizontal layer of different resistivity in otherwise homogeneous ground is the maximum. It is expected that much greater expansion is needed in this case than is needed simply to detect an interface. There is usually time while distant electrodes are being moved to calculate and plot apparent resistivity. Minor delays are in any case better than retaining with an interruptible result and field plotting. Simple interpretation can be carried out using two layer type curves on transparent materials. Ideally, this process is controlled using auxiliary curves to define the allowable positions of the origin of the two layer curves being fitted to the later segment of the field curve. Step-by-step marching was the main interpretation method until about 1980's. Computer based interactive modeling is now possible even in field camps and gives more reliable results (Millsom, 2003).

SP survey method is used for some near surface ore bodies that are readily detected by other electrical methods produce no SP. SP survey can be carried out by using two electrodes separated by a small constant distance normally 5 or 10m to measure average field gradient. IP surveys method are perhaps the most useful of all geophysical methods in sub surface survey being the only one responsive to low grade disseminated mineralization. There are two main mechanisms of rock polarizations (time domain and frequency domain) and there are three main ways in which polarization effects can be measured (time domain transmitters, time domain receivers and decay curve analysis) or (frequency domain transmitter, frequency receiver and face measurement (Corwin et al, 1979). The method used to display IP data vary with the array. Profiles or contour maps are used for gradient arrays, while dipole-dipole is always presented as pseudo-sections.

Electromagnetic (EM) which is a source of noise in resistivity and I P surveys is the basis of the number of geophysical methods. These were originally mainly use in search of conductive sulphide mineralization but are now being increasingly used for area mapping and depth sounding. Because a small conductive mass within a poorly conductive environment has greater effect on induction than on DC resistivity discussions of EM methods tend to focus on conductivity, the reciprocal of resistivity rather than on resistivity itself. Conductivity is measured in mhos per meter or in siemens per meter. There are two limiting situations. In the one eddy currents are induced in a small conductor embedded in an insulator producing a discrete anomaly that can be used to obtain information's on conductor locations on conductivity (Milson, 2003). In other, horizontal currents are induced in a horizontally layered medium and their effect at the surface can be interpreted in terms of apparent conductivity. Most real situations involve combinations of layered and discrete conductors, making greater demands on

interpreters and sometimes on field personnel's. Wave effects are important only on frequencies above 10kHz, and the methods can otherwise be most easily understood in terms of varying current flow in conductors and varying magnetic fields in space where the change in the inducing magnetic field is produced by the flow of sinusoidal alternating current in a wire of coil, the method is describe as continuous wave electromagnetic (CWEM).Alternatively transient electromagnetic (TEM) methods may be used, in which the changes are produced by abrupt termination of current flow(Milson, 2003).

CWEM and TEM methods are theoretically equivalent but have different advantages and disadvantages because the principal sources of noises are quite different.

VLF and CSAMT/MT make use of high power military communication transmissions in the 15 kHz - 25 kHz band termed very low frequency (VLF) by radio engineers. These waves have frequencies higher than those used in conversional geophysical works, but allow electromagnetic surveys to be carried out without local transmitters. Natural electromagnetic radiation covers a much broader range of frequencies. Longer wave links (lower frequencies) are generally due to ionospheric micropulsation, while much of the radiation in audible range id generated by distant thunderstorm activity. These later signals form the basis of audiomagnetotellurics (ATM) methods and mineral exploration and resistivity depth sounding. Because seferic signal strengths vary considerably with time, methods have been developed of producing signal similar to natural ones using control source (CSAMT) (Fraser, 1969).

Dip-angle data from VLF can be awkward to contour and dip angle maps, on which conductors are indicated by steep gradients, may be difficult to access visually. VLF results tend to 8be rather noisy being distorted by minor anomalies due to small local conductors and electrical

interference. This noise could be reduced by filtering. The parameters commonly measured in CSAMT surveys are the ratios of E_x to H_y and the phase differences between them. The amplitude ratio is used to calculate a quantity known as the cagniard resistivity (Fraser, 1969).

Programs can be run on PCs to carry out one dimensional and two dimensional inversions of cagniard resistivity to actual resistivities. To estimate the resistivity at a given depth, the cagniard resistivity must be obtained down to at least three times that depth. Phase differences are used mainly for investigating small sources, rather than layering. For example it may be possible to see both the top and the base of a buried source using phase measurement.

Radar methods use the reflections of short bursts of electromagnetic energy spanning a range of frequencies from about 50% to 50% above some specified central frequency. A typical 100MHz signal has a significant content of frequencies as low as 50MHz and as high as 150MHz. The development of Ground Penetrating Radar derives from the use of radio echo-sounding to determine ice thickness from which it was only a short step to studies of permafrost. It is now widely used to study the very shallow subsurface at landfill, constructions and archaeological sites (Milson, 2003).

A GPR system consists of a control and recorder unit (CRU) linked to receiver and transmitter units, each of which is in turn linked to an antenna. Metal wires are an efficient conductor of alternating current at radar frequencies, and signals to and from the CRU are normally transmitted by optical fibres. Receivers and transmitter antennae may be separated or integrated into a single module. Separability is desirable even if the spacing is to be kept constant because

of its optimum value depends on environmental and target depth as well as on frequency and transmitter size.

GPR data are recorded digitally and need extensive processing. The reduction in size and increase in capacity of small computers has made it possible to process in the field, sometimes using the field instruments themselves. The high repetition rates possible with radar systems allow large number of signals to be recorded at each transmitter or receiver set-up and to be stacked to reduce the effect of random noise. After stacking, the data are passed through a low-cut filter to remove noise due to inductive effects and limitations in instrument frequency response, and a high-cut filter to eliminate noise spikes. The processed GPR is recorded as a series of digital values equally spaced in time. It can be displayed either as a simple curve or by the variable area method in which excursions on one side of the zero line are shaded. The diffraction patterns provide velocities which vary only with depth and also sought to reveal the presence of limited, usually highly conductive targets (Milsom, 2003).

Seismic methods are the most effective and most expensive of all the geophysical techniques used to investigate layered media (Musset et al, 2000). A seismic wave is acoustic energy transmitted by vibration of rock particles. Low energy waves are approximately elastic; leaving the rock mass unchanged by their passage, but close to a seismic source the rock may be shattered and permanently distorted. Estimating the thickness of geological layers is a common requirement for many engineering, environmental and geological tasks. Common invasive techniques such as drilling, cone push, etc provide information at points, but geophysical methods must be used to build a more complete understanding of how layer thickness varies under a line. Seismic refraction is commonly used to obtain such information. The seismic

refraction is a quantitative method as it produces depths of various geological layers as well as the seismic velocities of the various layers. The seismic velocities help in the interpretation of geological layers. The geophysical property that is being measured here is the seismic velocity. In the seismic refraction survey, the kinds of wave that are of most importance are the P-wave and the S-wave (Milsom, 2003). The P-waves propagate at a higher velocity and are therefore recorded as first picks or breaks of the seismic refraction waves that propagate through the earth material.

The S-waves refractions evaluate the shear wave generated by the seismic source located at a known distance from the array. The waves are generated by horizontally striking an object on the ground surface to induce the shear waves (Milsom, 2003).

A Seismic refraction wave is properly described in terms of wave fronts, which define the points that the wave has reached at a given instant. However, only a small part of a wave front is of interest in any geophysical survey, since only a small part of the energy returns to the surface at points where detectors have been placed. It is convenient to identify the important travel path by drawing seismic rays to which the law of geometrical optics can be applied, at right angle to the corresponding wave front (Hawkins, 1961).

When a Seismic wave encounters an interface between two different rock types of different densities, some of the energy is reflected and the remainder continues on its way at a different angle i.e. is refracted. Refraction is governed by Snell's law which relates the angle of incidence and

refraction to the seismic velocities in the two media: $\frac{\sin i}{\sin r} = \frac{V_1}{V_2}$.

That is the ratio of wave velocity in one to the wave velocity in another medium is constant for the two different media.

This law describes the behavior of waves as they travel through media of different densities. It helps to know the velocity as well as the direction of the wave as it encounters an interface (Milsom 2003).

If V_2 was greater than V_1 and if $\sin I_c = V_1/V_2$, then the refracted ray will travel parallel to the interface at velocity V_2 . This is critical refraction. After the waves have undergone critical refraction, some of the waves return to the ground surface as a head wave and is represented by rays which leave the interface at a critical angle. This is illustrated in figure 2.1. The head wave travels through the upper layer at velocity V_1 but because of its inclination, appeared to move across the ground at the V_2 velocity with which the wave front expands below the interface. The head wave therefore eventually overtakes the direct wave despite the longer travel path. The cross-over distance for which the travel time of the direct and refracted waves is expressed by the equation

$$X_c = 2d \sqrt{(V_2 + V_1) / (V_2 - V_1)}$$

Where

X_c is the cross-over distance

d is the perpendicular depth of the subsurface

V_1 is velocity in the first layer

V_2 is velocity in the second layer

This equation forms the bases of a simple method of refraction interpretation. X_c is always more than double the interface depth and larger if the depth is large or the difference in velocities is small. The critical time obtained by dividing the critical distance by the direct wave velocity is

also sometimes used. If more than one interface is involved, the ray that is critically refracted at the lower most interface leave the ground surface at an angle is given by $\sin i_n = V_1/V_n$

Thus the angle with which the energy leaves the ground surface for ultimate critical refraction at a deep interface depends only on the velocities in the upper most and lower layers involved and not on the velocities in between the layers (Milsom, 2003).

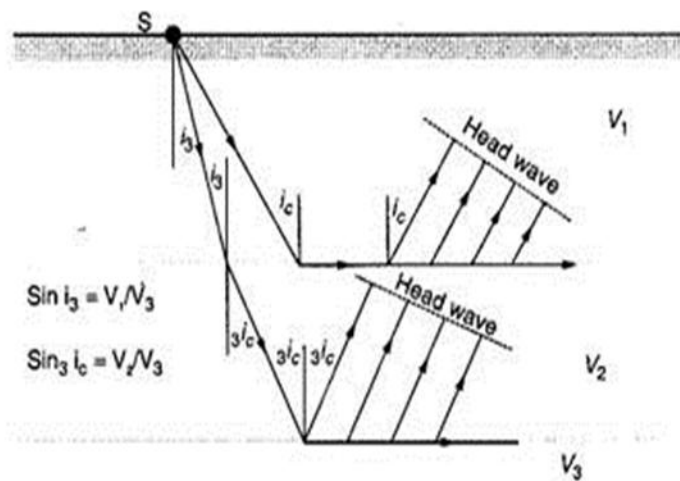


Figure 2.1 A critical refraction at two interfaces: $\sin i_c = V_1/V_2$

Traditional interpretation of seismic refraction data has used a concept of layered horizons or zones where each horizon had a discrete seismic velocity. Subsurface geology is traditionally visualized as a layered media when results from seismic refraction survey are applied to geotechnical engineering application. Interpretation methods based on the refraction of the first portion of seismic wave have been known for many years (Jakosky, 1940). The advent of hand-held calculators in the 1970,s and the personal computers by 1980,s as well as the development of seismograph for civil engineering use has made seismic refraction a practical geotechnical

exploration tool for more than two decades. Recently, vast increasing computing power has now made computation-intensive software tools practical that could provide very different kinds of subsurface interpretation. These software tools could help answer two basic questions of seismic refraction investigations. The first is the depth of investigation of the seismic survey and secondly alternative interpretation available when subsurface geology is not modeled well as a traditional layered media (Redpath, 1973).

Several methods are available for determining the depth of the bedrock or thickness of the first layer. The cross-over distance along the refractor equation as stated before could be express as d in terms of the other parameters. Where

$$d = X_c/2 \left\{ \sqrt{\frac{V_2 - V_1}{V_2 + V_1}} \right\}$$

All the letters have their usual meaning.

Another method that could be used to determine the depth of the bedrock is reciprocal time interpretation. This is defined as the time taken for a seismic energy to travel between the two long-shot positions. The difference between the reciprocal time (t_R) and the sum of the travel time t_A and t_B from the two long –shot to any given geophone G is given by the expression,

$$t_A + t_B - t_R = 2D/F$$

Where D is the depth of the refractor beneath G and F is the depth conversion factor. If there is only a single interface D which is equal to the thickness d of the upper layer and F is equal to $V_{1,2}$ From the figure 2.2, the sum of the travel times from S_1 and S_2 to G differs from the reciprocal time, t_R taken to travel from S_1 to S_2 by the difference between the time taken to travel QR at velocity V_2 and QGR at velocity V_1 (Redpath, 1973).

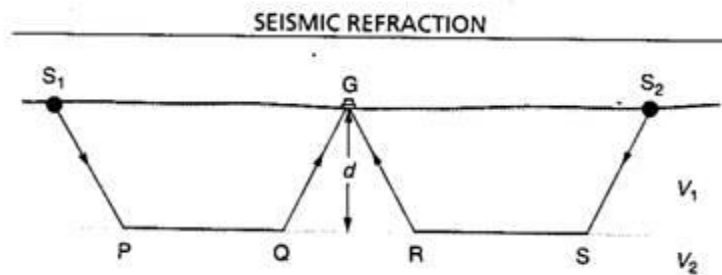


Figure 2.2 Reciprocal time interpretation

One other method for determining the depth of the bedrock is to use the time – distance plot intercept method. The data extracted consisting of time arrival and distance from shot point at each geophone were plotted as time-distance plot with time on the vertical axis and distance on the horizontal axis. Best fit lines are drawn through the points. The number of straight lines obtained show the number of layers identified in the survey. The reciprocal of the gradient from the line of best fit through the plotted data represents the wave velocity through a given medium. This means that the reciprocal of the slope on the direct wave arrival line represent the velocity of the wave in the upper layer and that in the refracted arrival wave line represent the velocity of the wave in the lower layer.

These are done for all the data on a spread and are plotted on a single sheet that has the working area. The extrapolation of the upper line cuts the time axis at a point termed as intercept time.

The time intercept equation is expressed as

$$t_i = 2d\sqrt{(V_2^2 - V_1^2)}/V_1V_2$$

Where the letters

t_i is the time intercept

d is the perpendicular depth

V_1 is velocity of wave in the first medium

V_2 is velocity of wave in the second medium

The depth could be express in terms of the other parameters as

$$d = t_i/2 \{ \sqrt{V_1 V_2 / \{V_2^2 - V_1^2\}} \}$$

Depths estimated from the refraction survey were related to geophones and the shot point elevations which were therefore measured to obtain a true picture of the subsurface refractor. Furthermore, the depth determined is the perpendicular distances to interfaces from shot point or geophones. The intercept-time equations require the true value of V_2 to be used. However a wave that travels down-dip does not only has to travel further at velocity V_2 to reach more distance geophones but also further at a small velocity V_1 in the upper layer (Hawkins, 1961).

In figure 2.3, the refracted energy from S_1 arrives later at B than at A not only because of the greater distance travelled along the refractor but also because of the extra distance d_1 travelled in the low velocity layer. Energy from S_2 arrives earlier at P than would be predicted from the time of arrival at Q by the time taken to travel d_2 at velocity V_1 .

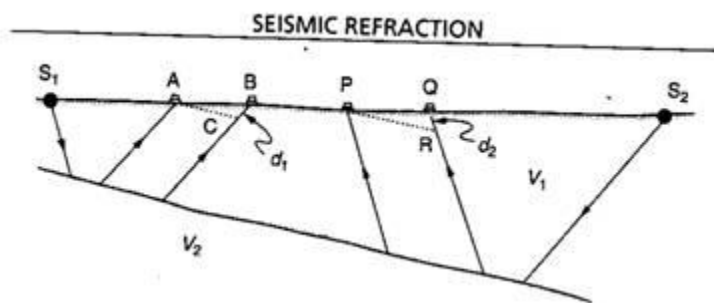


Figure 2.3 Refraction at a dipping interface.

It therefore arrives later with a low apparent velocity. The reverse is a true shooting up-dip when arrivals at further geophones may actually precede those at the nearer ones. The slope of the line through the refracted arrival on a time – distance plot depends on the dip angle α , according to

$$V_{app} = V_2 / (1 + \sin \alpha)$$

The thickness of soil layers of proposed foundations is used as the main reference in the design of the structure and is needed as appropriate information (method and technology) during the foundation construction. These parameters can be obtained from the common methods such as the drilling and the standard penetrating testing (SPT). However, these tests are time consuming and have to be done at a widely spaced interval. As a result, the most critical point in terms of thickness and strength are sometimes not tested. Thus the reliable and fast nondestructive testing (NDT) capable of measuring the thickness, stress and stiffness (bearing capacity) is desirable. One such NDT, which is based on waves propagate is the seismic refraction method. The method uses several approaches such as time- distance plot, reciprocal time interpretation, 2D models etc to find the P-wave velocity and the thickness of the layered profile (Terzaghi et al, 1996). The P-wave velocity is an effective parameter for determining the stiffness or bearing capacity of materials.

Seismic refraction being a very good method for studying depth and horizons of layer has some limitations.

First-arrival refraction work uses only a small proportion of the information contained in the seismic traces and it is not surprising that interpretation is subject to severe limitations. These are especially important in the engineering work in low- velocity-layer studies only a time delay estimate is sought and short shots alone are often sufficient (Milsom, 2003).

A refractor that does not give rise to any first arrivals is said to be hidden. A layer is likely to be hidden if it is much thinner than the layer above and has a much lower seismic velocity than the layer below. Weathered layers immediately above basement are often hidden. The presence of a

hidden layer can sometimes be recognized from second arrivals but this is occasionally possible in parts because refracted waves are strongly attenuated in thin layers.

A layer may also be hidden if the head wave that it produces does arrive first over some part of the ground surface if there are no appropriate located geophones. Concentrating geophones in critical regions can sometimes be useful but the need to do so will only be recognized if preliminary interpretations are being made on daily basis (Hawkins, 1961).

If the velocity decreases at an interface, critical refraction cannot occur and no refracted energy returns to the surface. Little can be done about these blind interfaces unless vertical velocities can be measured directly. Thin high-velocity layers such as perched water table and buried terraces often create blind zones. The refracted waves within them lose energy rapidly with increasing distance from the source and ultimately become undetectable. Much later event may then be picked as first arrival producing discontinuities in the time distance plot (Milsom, 2003). Seismic refraction method is a commonly used traditional geophysical technique to determine depth-to-bedrock, competence of bedrock to support structures (Redpath, 1973).

CHAPTER THREE

3.0 METHODOLOGY

The research covers an area in Legon in the Accra Metropolitan Area of the Greater Accra Region of Ghana. Seismic refraction work was done on this area to bring out primary data. A geophysical seismic refraction method was used along four transverses. The survey designed was dependent on the site and was planned so that the geometry of geophone spread would allow the target to be resolved. A primary limitation of refraction method on many sites is that long refraction traverses are sometimes required. In such situations the subsequent spread are always established such that it overlaps with the previous spread for clear resolution. The spacing of the geophone stations within a spread even though could be varied from several meters to tens of meters depending on the depth of the geological layer and the resolution required, the spacing for the geophone stations within a spread for the survey was pegged at five meters (5 m). Shot points were extended along the entire traverse length. The arrival times of the waves from shot points were picked by geophones and sent along cables to seismograph. The information was then fed to a computer to plot time-distance graph and also develop 2D models for interpretation. From the time-distance graph, slopes were found from the best fit lines to determine the velocities of the media involved. Again the intercept on the time axis was used to evaluate the thickness of the first layer as well as the depth of the second layer which is the bedrock. The generated models also give a two dimensional pictorial view of the major horizons or layers with the depth probed.

3.1 FIELD DATA COLLECTION

In the study area four traverses were established. These traverses range in length between one hundred and twenty five meters (125 m) and two hundred and forty meters (240 m). The

traverses were labeled L1, L2, L3 and L4. Along each traverse, twenty four (24) geophones were planted at sampling interval five meters (5m) leading to a shot spread length of one hundred and twenty meters (120m).

For such spread, seismic signals from hammer impact on a metal plate were generated at five different points noted as stations. These arrays of shots for each spread were conducted to enhance the data interpretation.

The shots were made at stations of 0 m, 32.5 m, 62.5 m, 92.5 m and 125 m with an impulsive hammer. The impulsive hammer generated mechanical waves which travelled through the earth. As the wave travelled through the first layer of earth and encountered another layer of different density, the wave underwent a critical refraction. The refracted wave travelled along the refractor and returns to the surface of the earth as head wave with an apparent velocity of the first medium on the surface. The arrival of this mechanical wave on the ground surface was received by receivers (geophones) which converted the mechanical energy into voltages. These voltages were transported along cables to seismograph and these were recorded as arrival time and distance from shot points.

For a traverse that was longer than one hundred and twenty meters (120m), another spread was established to survey the extra length. In such situations the subsequent spread were always established such that it will overlap with the previous spread.

The first traverse was an example of such situation; the traverse length was two hundred and forty meter (240m) which was more than one hundred and twenty meters. The subsequent spread was established to overlap the first spread in order to obtain a good resolution for interpretation. This is illustrated in figure 3.1

In these spreads layout style sufficient information on the direct wave and reasonable coverage on the refractor is obtained due to the fact that the length of the spread was about three times the cross over distance. The spread length was eight times the expected refracted depth.

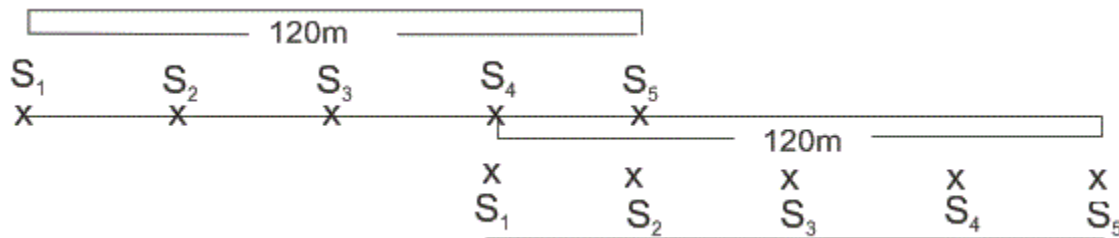


Figure 3.1 Overlapping of a longer traverse spread.

The second, third and fourth traverses had a length of one hundred and twenty meters and so no subsequent spreads were established. The twenty four geophones were planted on them at five meter interval with the five shot points on the single spread

The center shots were useful if there were considerable difference in the interpretation at the opposite ends of the spread and if these seem to imply different numbers of refractors. They made it possible to obtain a more reliable estimate of the velocity along an intermediate refractor or to monitor the thinning of intermediates that was hidden at end of the spread by refraction from a greater depth (Clark, 1996).

3.2 DATA ANALYSIS

Picking of time arrival on refraction records relied subjectively on estimates of first break positions and could be difficult at remote geophones where the signal to noise ratio is poor. Some of the late peak and troughs in the same wave train were likely to be stronger and it was

sometimes possible to work back from these to estimate the position of the first break. The trace beyond the first break was affected by many other arrivals as well as by late parts of the primary wave train, and these modified peak and trough locations. The short points and their arrival picks for each of the traverse are illustrated below.

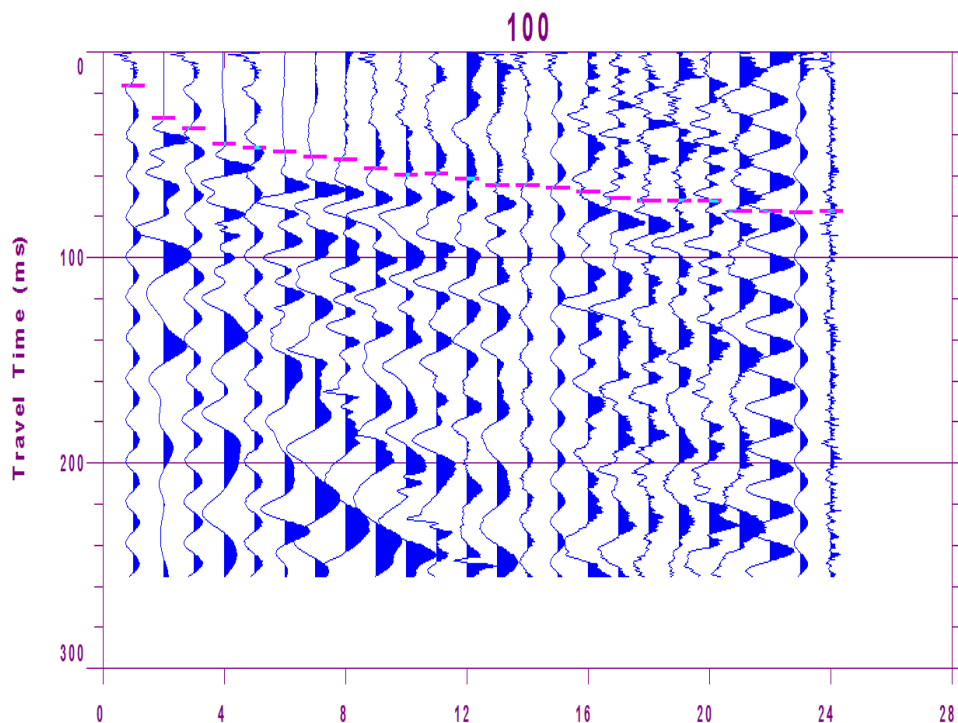


Figure3.2 First shot point with arrivals picked by geophones on the first travers L1

The first shot point was at 0 m along the length of the first traverse L1 and the closer geophones receive the arrival signals first than the farther ones as indicated by the red marks as shown in figure 3.2.

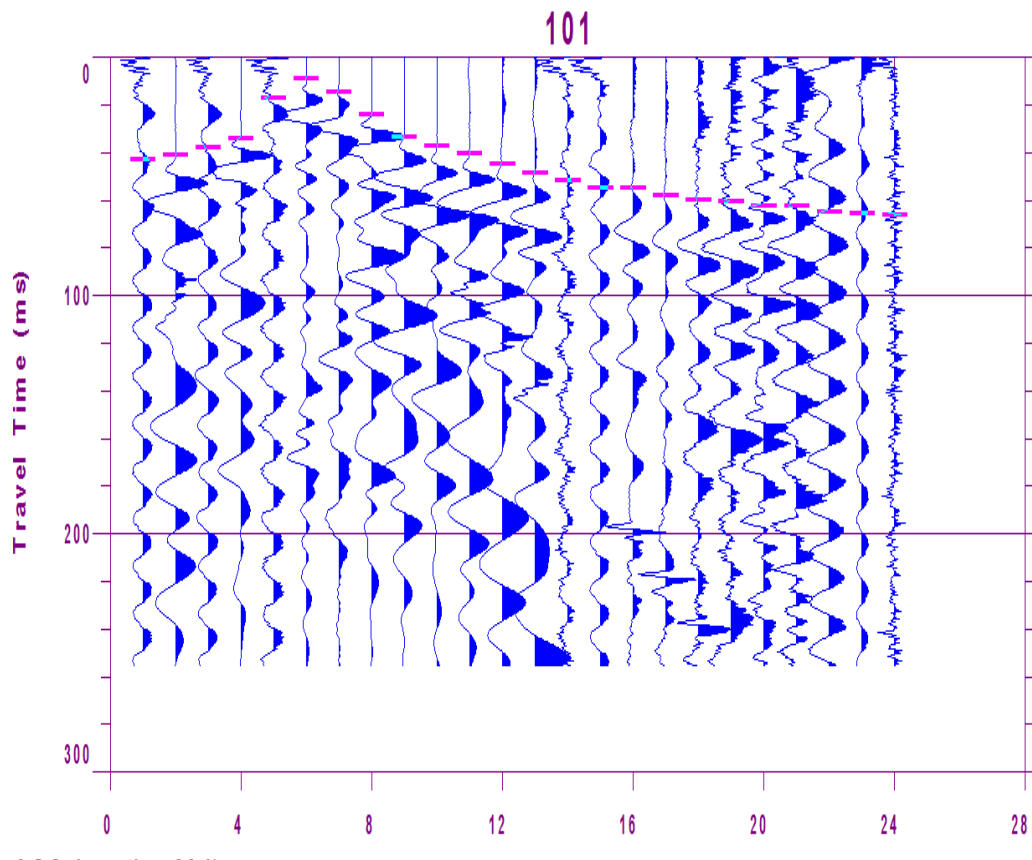


Figure3.3 Second shot point with arrivals picked by geophones on the first traverse L1

From figure 3.3 shows the recorded trace due to the second shot point was at 32.5 m along the length of the first traverse L1. This was shot between the fifth and sixth geophones.

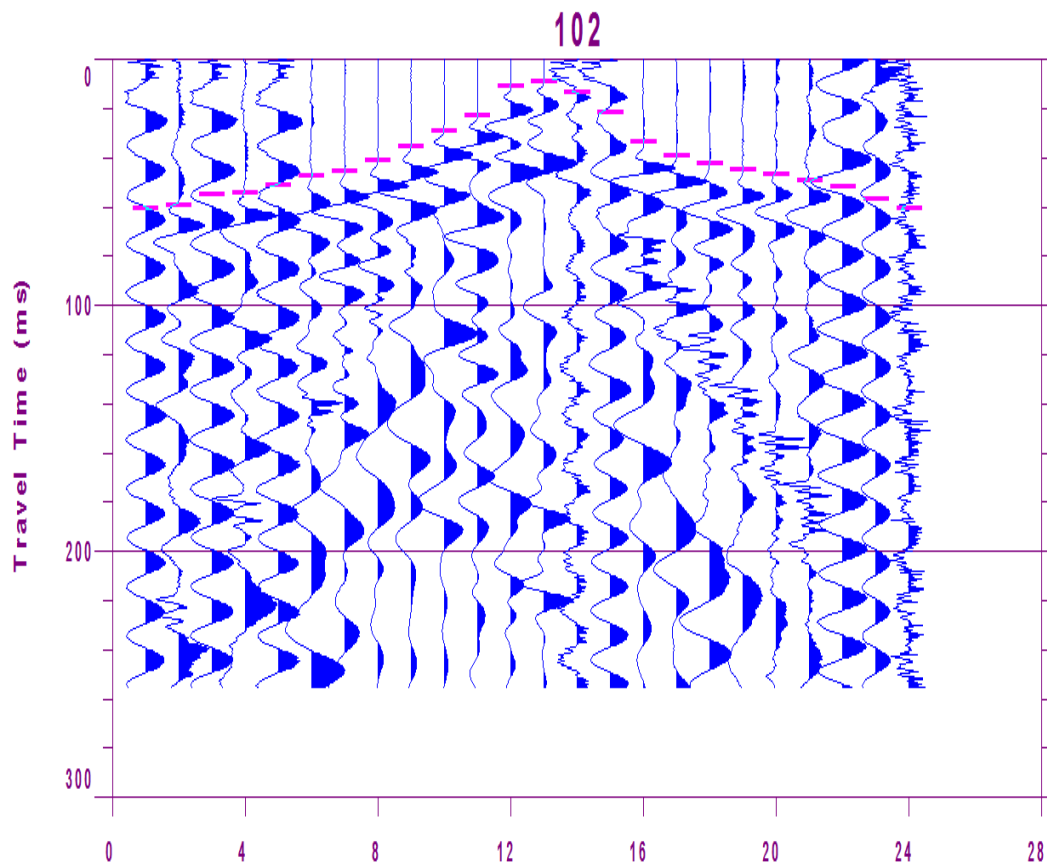


Figure3.4 Third shot point with arrivals picked by geophones on the first travers L1

The third shot was impacted between the twelve and thirteenth geophones. This shot point was at 62.5 m along the length of the first traverse L1. The closer geophones picked the arrival signals first than the farther ones at just at where the crest and trough were clear as indicated by the red marks. The crest and trough above the red marked lines indicate noise which could be problems from the gadgets. The noise interference was ignored following the extrapolation as seen in figure 3.4.

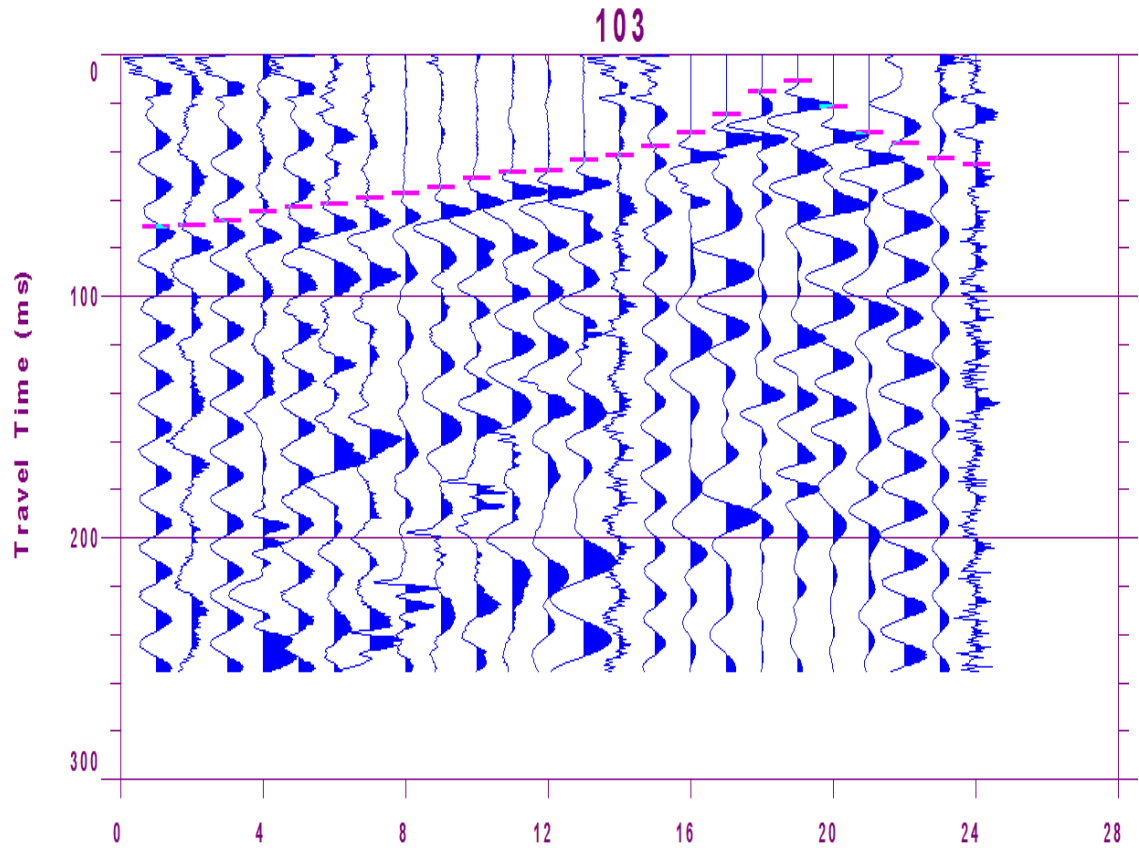


Figure 3.5 Fourth shot point with arrivals picked by geophones on the first traverse L1

With the geophones at their fixed positions along the spread as shown in figure 3.5, the fourth shot was impacted between the eighteenth and nineteenth geophones. This shot point was at 92.5 m from the first shot position along the length of the first traverse L1.

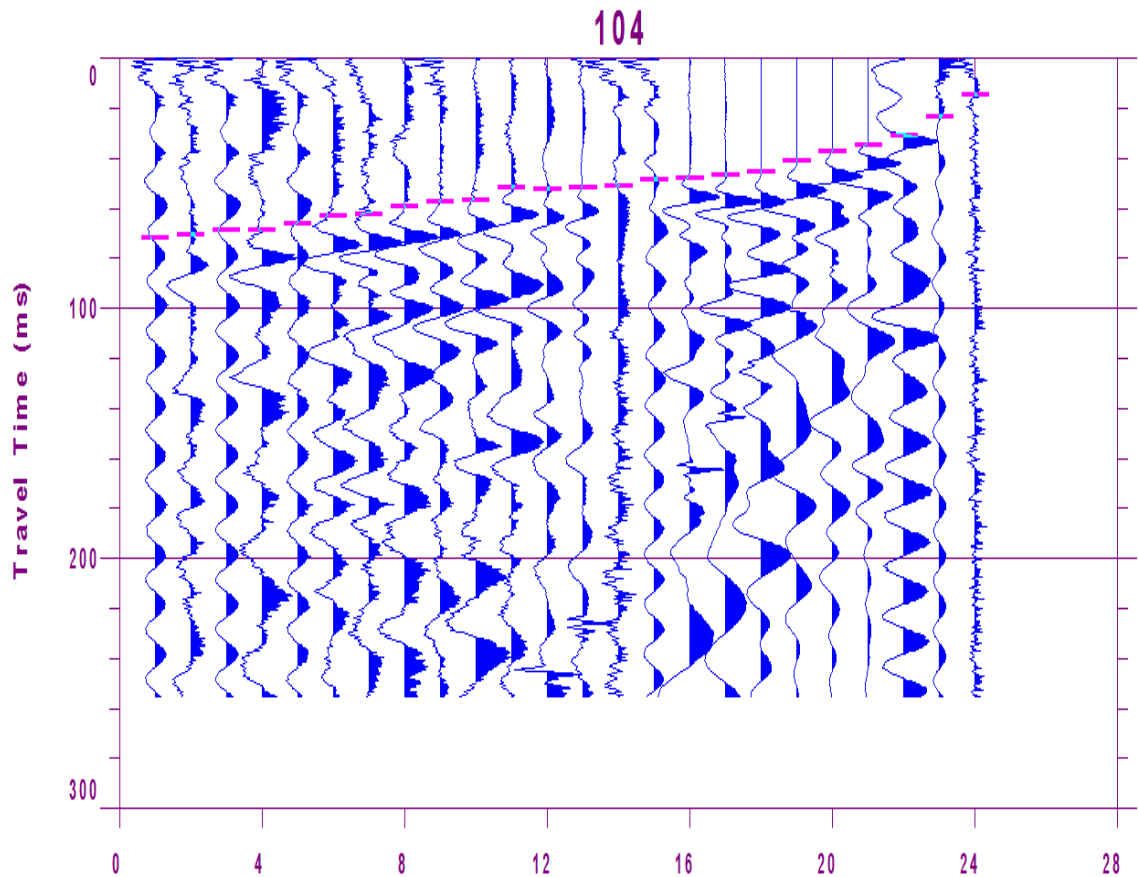


Figure 3.6 Fifth shot point with arrivals picked by geophones on the first traverse L1

The fifth shot point was at 125 m along the length of the first traverse L1 from the first shot point as shown in figure 3.6. This shot point was just beyond the twenty fourth geophone. This meant that the spread had ended. Since the traverse was more than 125 m, there was the need to create an overlap spread to continue the survey for other shot positions. The first geophone picked the longest time of arrival since it was the farthest from the shot point position.

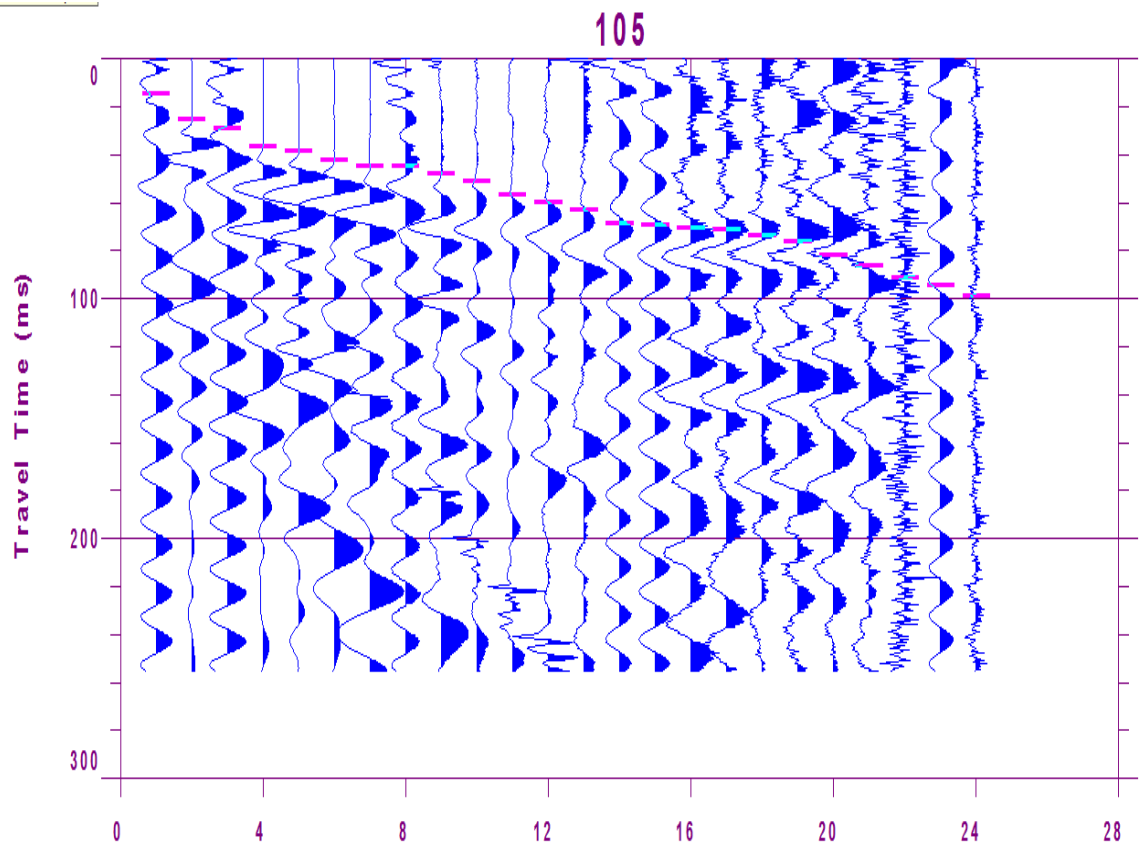


Figure 3.7 Sixth shot point with arrivals picked by geophones on the first traverse L1

From figure 3.7, the sixth shot point was at 110 m along the length of the first traverse L1 but was the first shot along the overlap spread. The shot point was at 110 m behind the fifth shot point 125 m as a result of the overlap. The shot point was 0 m which was just five meters in front of the first geophone on the overlap spread.

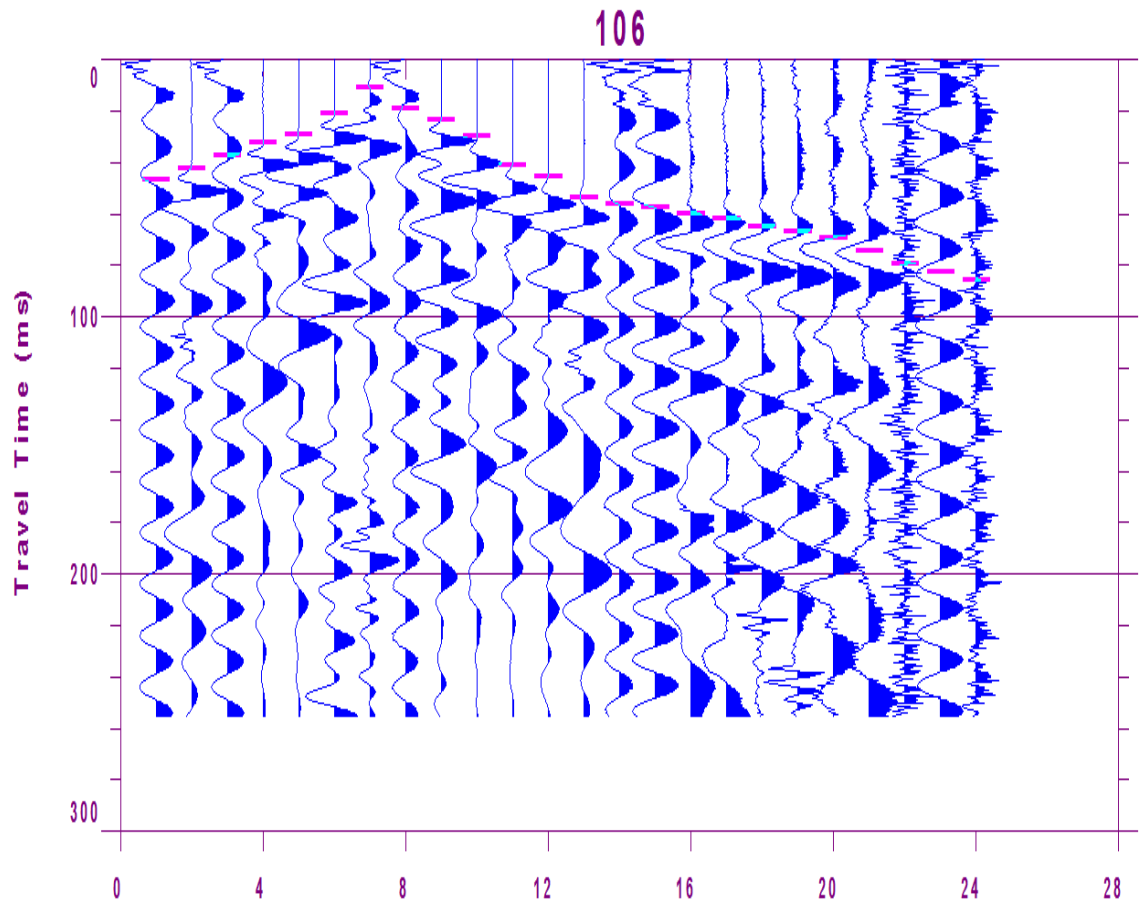


Figure 3.8 Seventh shot point with arrivals picked by geophones on the first traverse L1

The seventh shot point was at 142.5 m along the length of the first traverse L1 on the overlap spread. The shot point was between sixth and seventh geophones on the overlap spread. The closer geophones receive the arrival signals with less interference than the farther ones as indicated by the breaks above the red marks in figure 3.8. The interference is as a result of internal problem within the system.

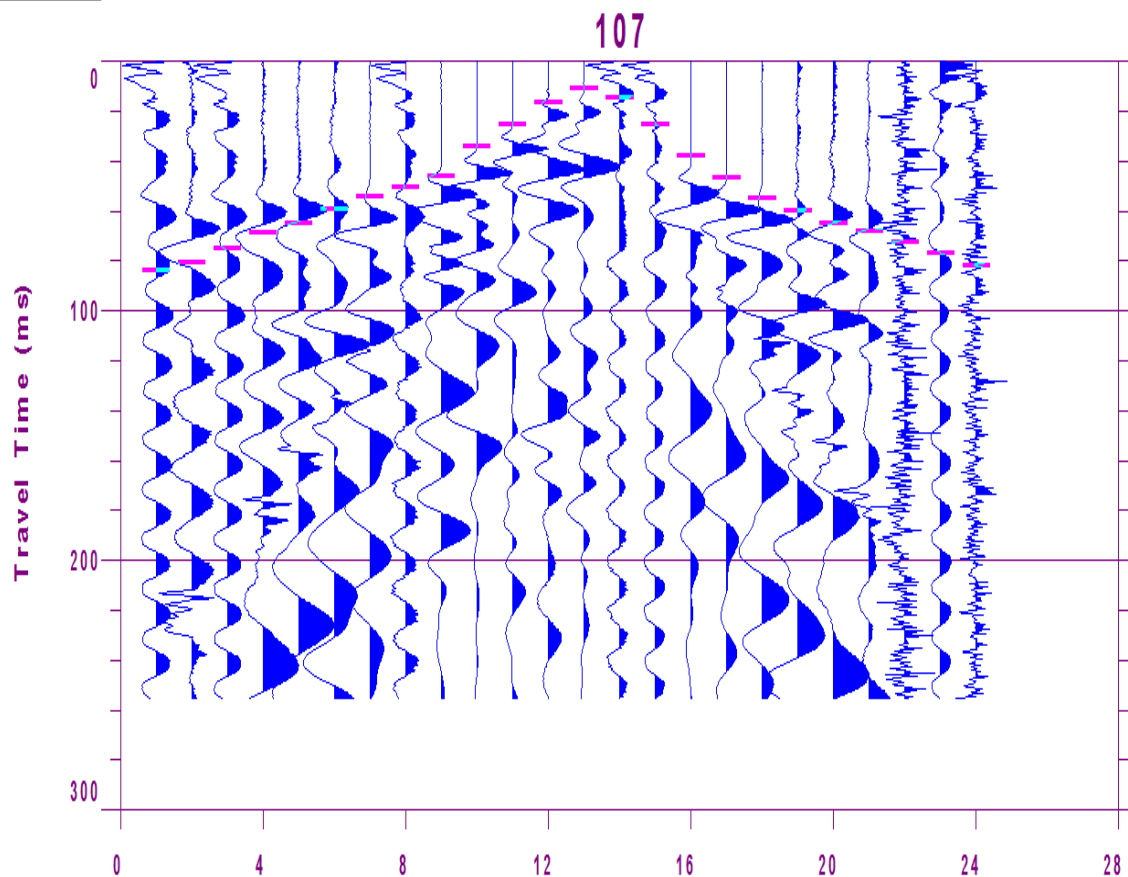


Figure 3.9 Eighth shot point with arrivals picked by geophones on the first traverse L1

The eighth shot point was between the forty second and forty third geophones at 172.5 m along the length of the first traverse L1 from the first shot point 0 m. The shot point was between the twelfth and thirteenth geophones along the second spread as shown in figure 3.9. One disadvantage of the longer spread length is wave attenuation resulting in very late arrivals at distant geophone positions.

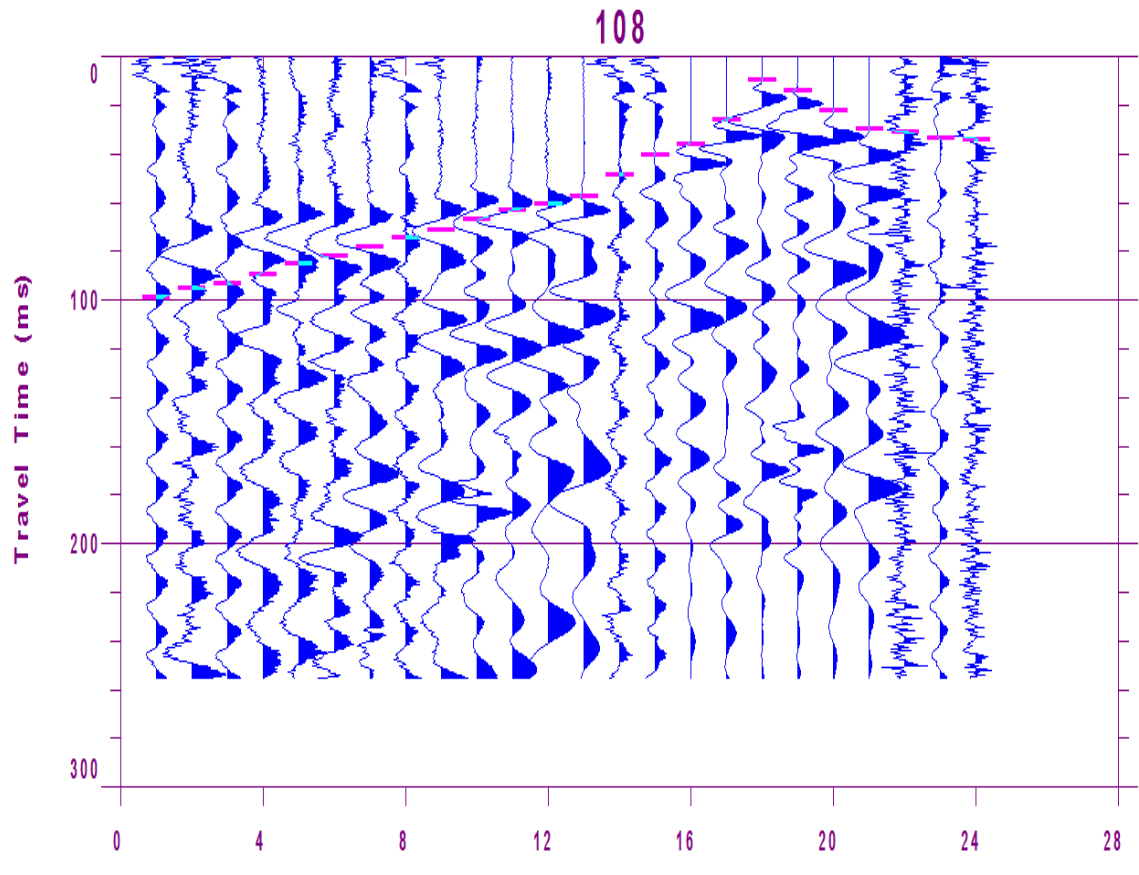


Figure 3.10 Ninth shot point with arrivals picked by geophones on the first traverse L1

The ninth shot point is at 202.5 m along the length of the first traverse L1. The shot point was between the eighteenth and nineteenth geophones on the overlap spread. From figure 3.10 it is observed that a lot of noises were dealt with as the extrapolation was followed.

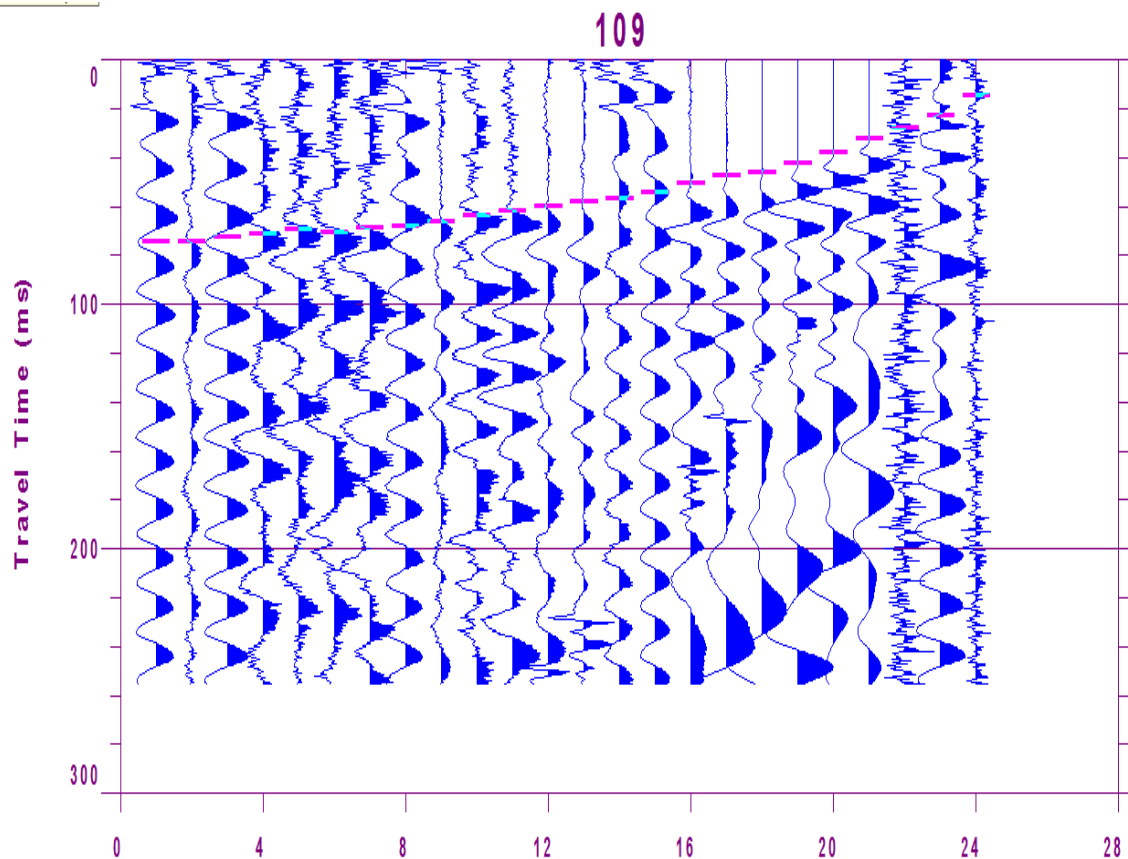


Figure 3.11 Tenth shot point with arrivals picked by geophones on the first traverse L1

From figure 3.11 the last and tenth shot point along the first traverse L1. The shot point was just beyond the twenty fourth geophone along the overlap spread. The shot point was at 235 m along the length of the traverse.

The process of picking arrival times were repeated along the second, third and fourth traverses.

3.3 Time – Distance Plot

The record data consisting of the arrival times at the various geophones and the geophones distances from the shot points were plotted as x-y scatter of distance against time as shown in figure 3.12. This was done for both forward and reverse shots. Lines of best fit were then drawn through the plots to determine the average slopes relating to the reciprocal of the speed of travel of the waves. The plots displayed a unique characteristic of a single change in direction or slope of the plotted points. These indicate that the shallow subsurface investigated in all the areas are characterized by a two layer-unit with contrasting wave velocity.

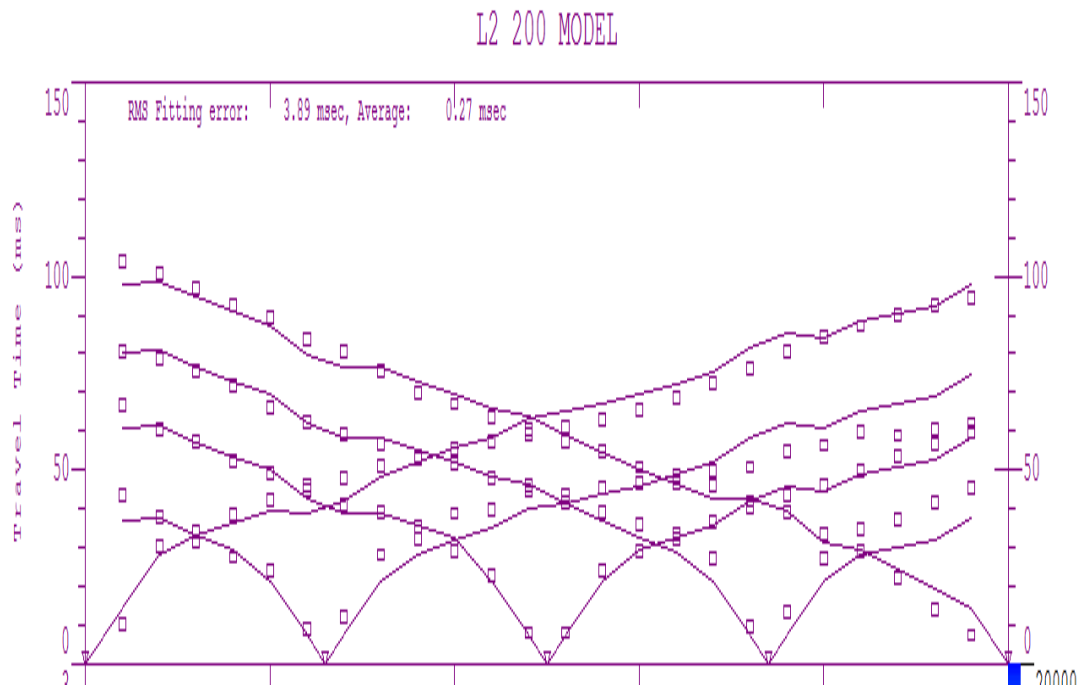


Figure 3.12 Time-distance plot for forward and reverse shots

Considering the figure 3.13 below if a wave is generated from a point A and travels through the upper medium with velocity V_1 and undergoes a critical refraction at B, such that it travels along the interface in the second layers with velocity V_2 . The refracted wave covers a distance BC along the interface and return to the earth surface at D. If the perpendicular depth from A to B is z , then z could be calculated from the deduced equation below

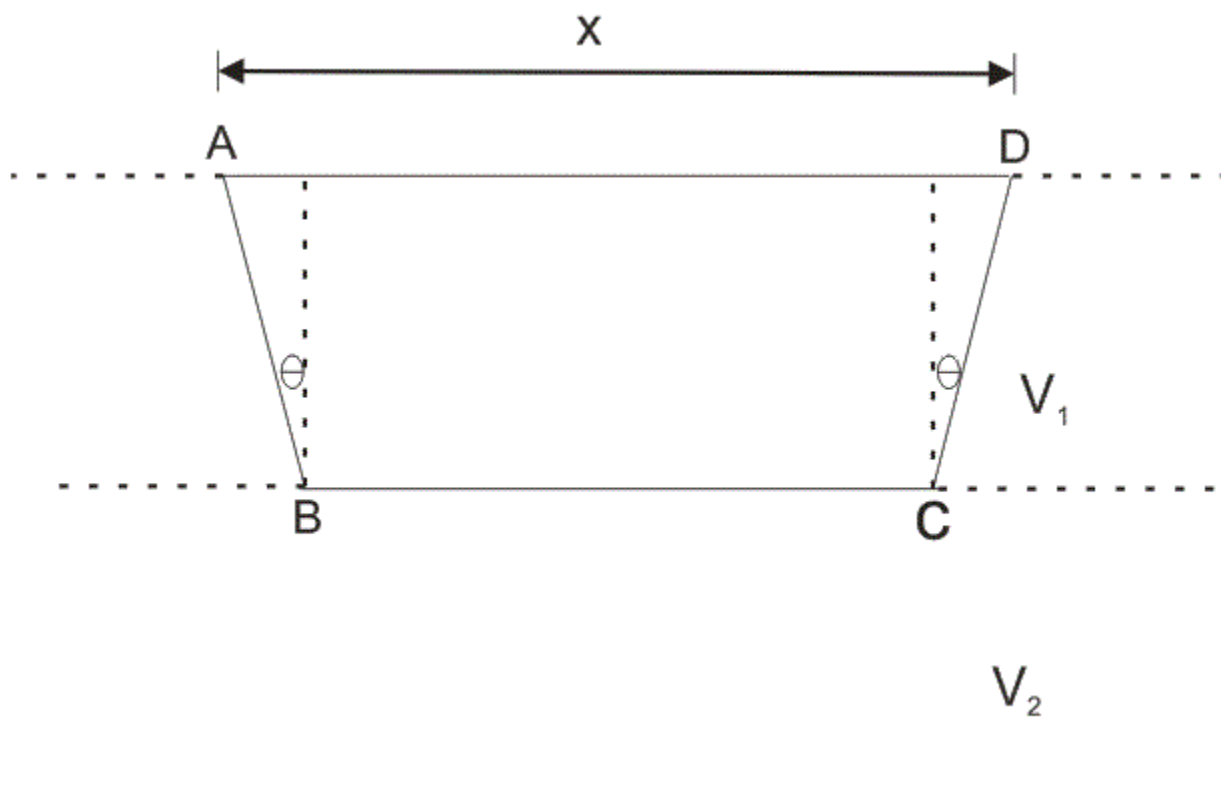


Figure 3.13 A ray undergoing critical refraction between two different media

If t is the time of travel from A to D then the time t ;

$$t = AB/V_1 + BC/V_2 + CD/V_1$$

$$t = AB + CD/V_1 + BC/V_2$$

Since $AB = CD$

$$t = 2AB/V_1 + BC/V_2$$

$$t = 2z/V_1 \cos \theta + \{x - 2z \tan \theta\}/V_2$$

$$t = x/V_2 + 2z/V_1 \cos \theta - 2z \tan \theta/V_2$$

$$t = x/V_2 + 2z \{1/V_1 \cos \theta - \tan \theta/V_2\}$$

At the intercept time $t = t_i$

$$x/V_2 = 0$$

$$t_i = 2z \{1/V_1 \cos \theta - \tan \theta/V_2\}$$

$$z = V_1 V_2 t_i \cos \theta / 2 \{V_2 - V_1 \cos \theta \tan \theta\}$$

This is used to evaluate the depth to the top of the second layer

From the four traverses, the time-distance plots for both forward and reverse shots were plotted to find the velocity of the wave in the upper and lower media and also depth to particular points along the second layer from the first layer as shown in figures 3.14, 3.15, 3.16 and 3.17.

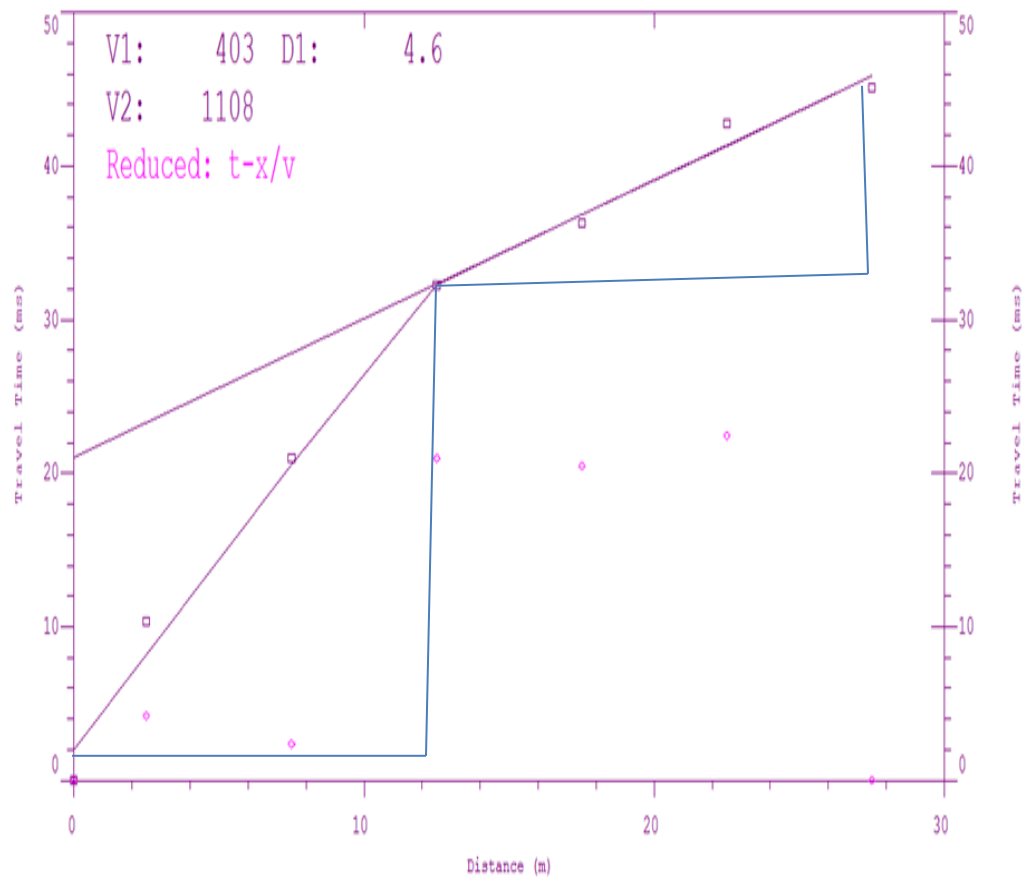


Figure 3.14 Time - distance plot from the first traverse L1

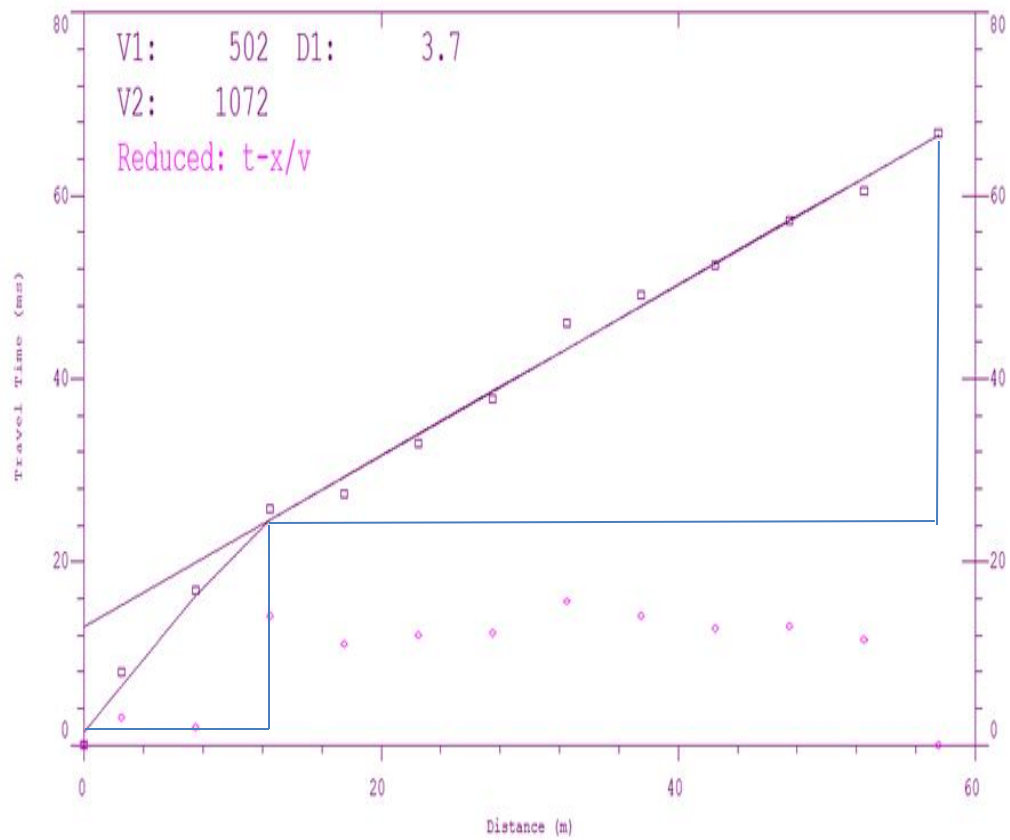


Figure 3.15 Time - distance plot of the second traverse L2

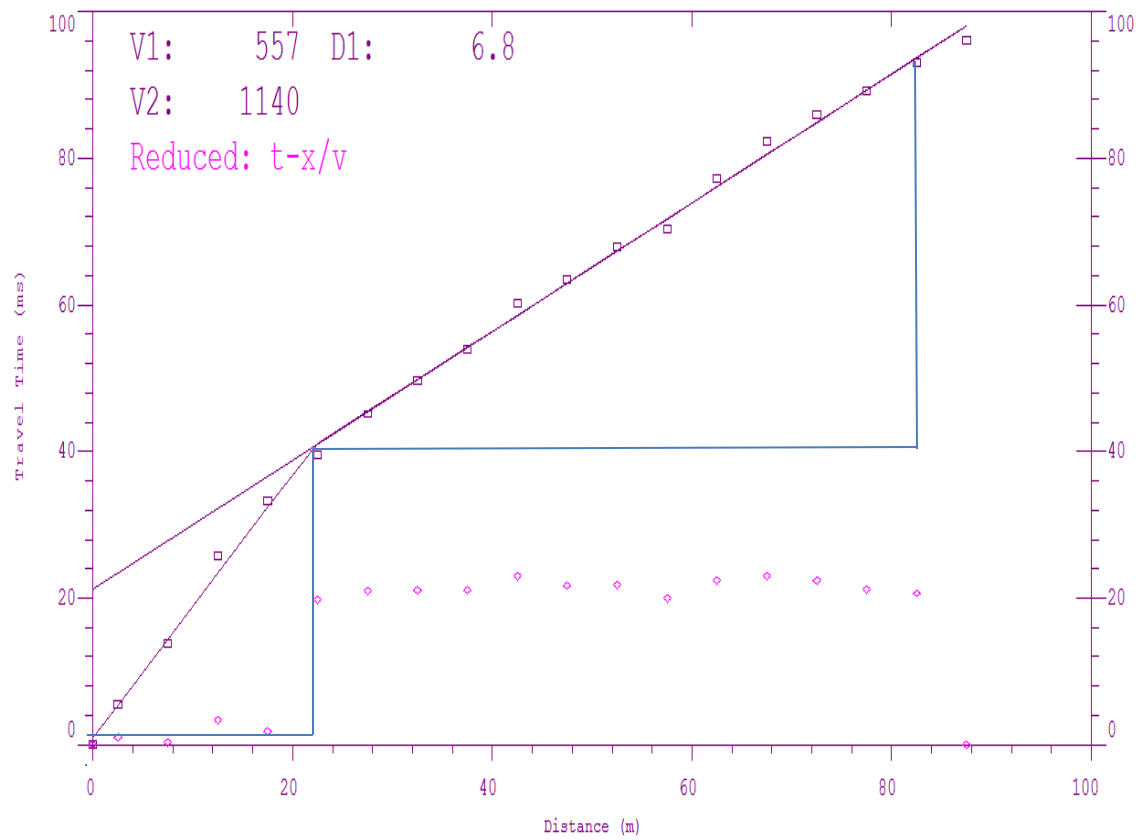


Figure 3.16 Time - distance plot of the third traverse L3

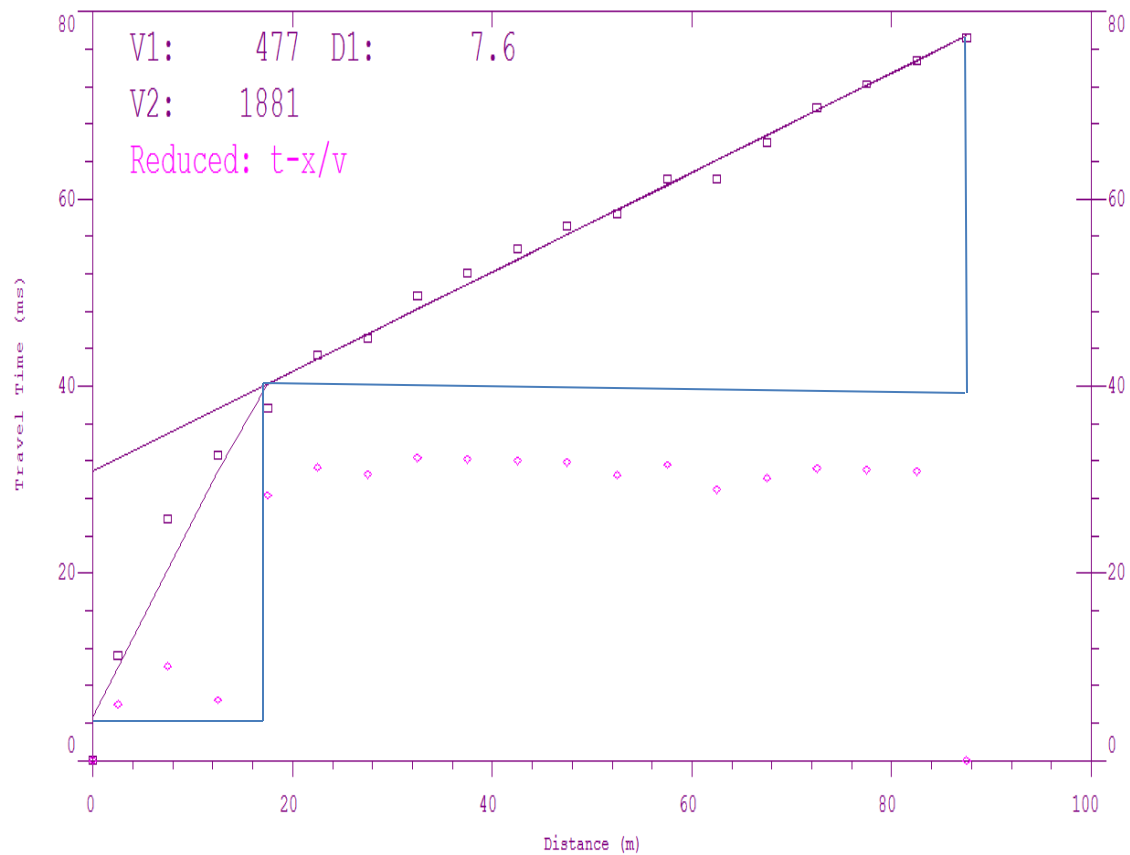


Figure 3.17 Time – distance plot of the fourth traverse L4

Considering the plot from the first traverse L1 as shown in figure 3.14,

Taking the inverse of the slope of the lower line which represents the velocity of the wave in first layer;

$$1/V_1 = \Delta T / \Delta D$$

$$1/V_1 = \{30.2 - 0.2 / 12 - 0\}$$

$$1/V_1 = \{30 / 12\}$$

$$V_1 = 0.4$$

The velocity V_1 of the wave in the first layer was 400 m/s

Taking the inverse of the slope of the upper line which represents the velocity of the wave in second layer;

$$1/V_2 = \Delta T / \Delta D$$

$$1/V_2 = \{44.5-32 / 27.5-13.65\}$$

$$1/V_2 = \{12.5 / 13.85\}$$

$$V_2 = 1.108$$

The velocity V_2 of the wave in the second layer was 1108 m/s

The intercept time is 21 ms.

The intercept time is the time at which the back extrapolated refracted arrival line cuts the time axis. Intercept times were conventionally obtained by drawing best-fit lines through the refracted arrival time (Milsom, 2003). For a single refractor, it is related to the velocity and the refractor depth by the equation

$$t_i = 2d\{\sqrt{(V_2^2 - V_1^2)} / V_1 V_2\}$$

If V_2 is much larger than V_1 , the critical angle is then almost 90° and the delay suffered by the refracted ray in travelling between the surface and the refractor is close to double the vertical travel time.

Inferring the depth from the time intercept equation;

$$d = t_i\{V_1 V_2\} / 2\{\sqrt{V_2^2 - V_1^2}\}$$

$$d = 0.021\{400*1108\} / 2\sqrt{\{1108^2 - 400^2\}}$$

$$d = 0.021\{443200\} / 2\sqrt{1067664}$$

$$d = 9307.2 / 2066.56$$

$$d = 4.50\text{m}$$

The depth at 120 m from shot point 0 m to a point on the surface of the second layer was 4.50 m.

These methods used the reciprocal of the slopes of the lines of best fits to find the velocities of the waves through the two layers and intercept time equation to find the depth to various points along the surface of the second layer from the top of the first layer and also from the original shot position. These were recorded in the table 3.1, Where

V_1 is the velocity of the wave in the first layer.

V_2 is the velocity of the wave in the second layer.

t_i is the intercept time on the time-distance plot.

d is the depth to the surface of the second.

z is the distance from the origin 0 m along the surface of the upper layer.

This was repeated along the other traverses in the table 3.2, 3.3 and 3.4 where the velocities of the waves in the first and second layers together with the intercept time from the time-distance plot were used to infer depth at various points along the interface. The distances from the original shot position along the surface of the upper layer (z) and the corresponding depth (d) to the surface along the second layer recorded in the tables below were plotted on $z - d$ plot for each traverse as shown in figures 3.18, 3.19, 3.20 and 3.21 below.

Table 3.1 Calculated depths to the surface of the second layer from the top of first layer along the traverse L1

z	13	20	42	50	72	84	100	108	112	120	128	132	156	168	188	192	210	225
V_1	454	389	756	582	590	498	403	380	394	586	395	323	655	421	513	510	458	823
V_2	2890	1681	3297	1811	1706	2092	1108	1611	1619	2874	2910	1995	1688	1017	1152	1413	3306	2971
t_i	39	27	41	30	25	31	22	21	22	30	15	29	32	28	32	37	35	47
d	9.0	5.4	15.9	9.2	7.9	7.9	4.9	4.1	4.5	9.0	3.0	47.5	11.4	6.5	9.2	10.1	8.1	20.1

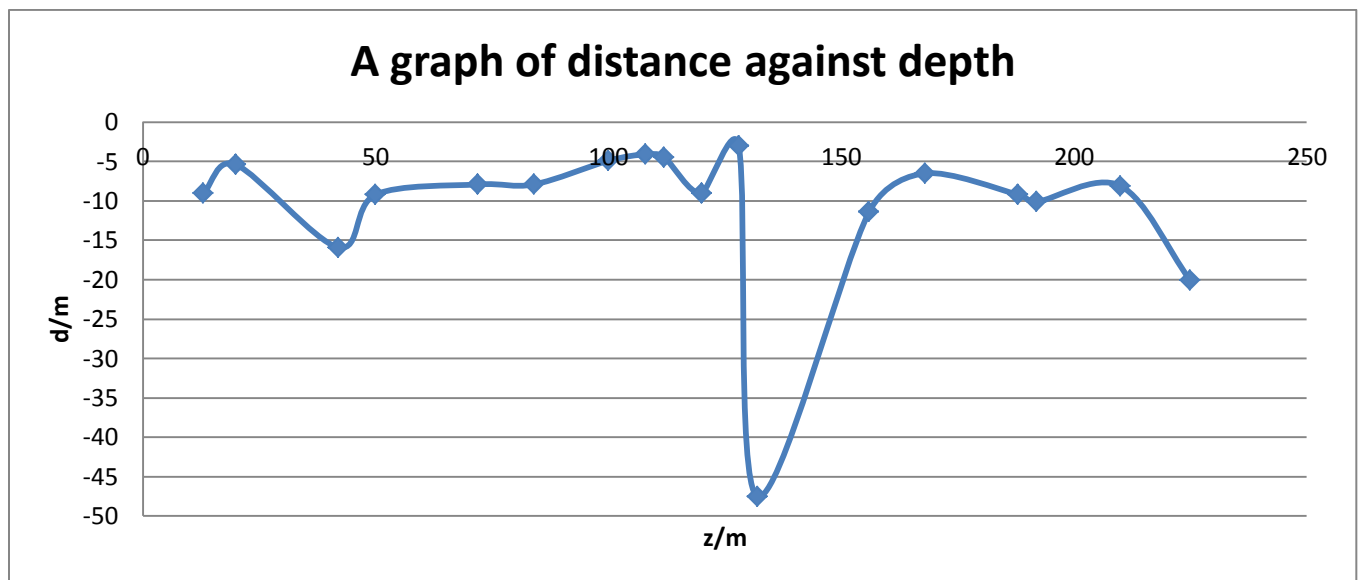


Figure 3.18 A z – d plots for depth and distance from ITM of traverse L1

Table 3.2 Calculated depths to the surface of the second layer from the top of the first layer along the traverse L2

z	10	26	48	55	70	86	100	106
V_1	332	313	355	335	316	279	287	691
V_2	1770	1015	2870	1197	1424	1545	1159	1361
t_i	26	15	32	18	20	24	22	15
d	4.4	2.5	5.7	3.1	3.2	3.4	3.3	6.0

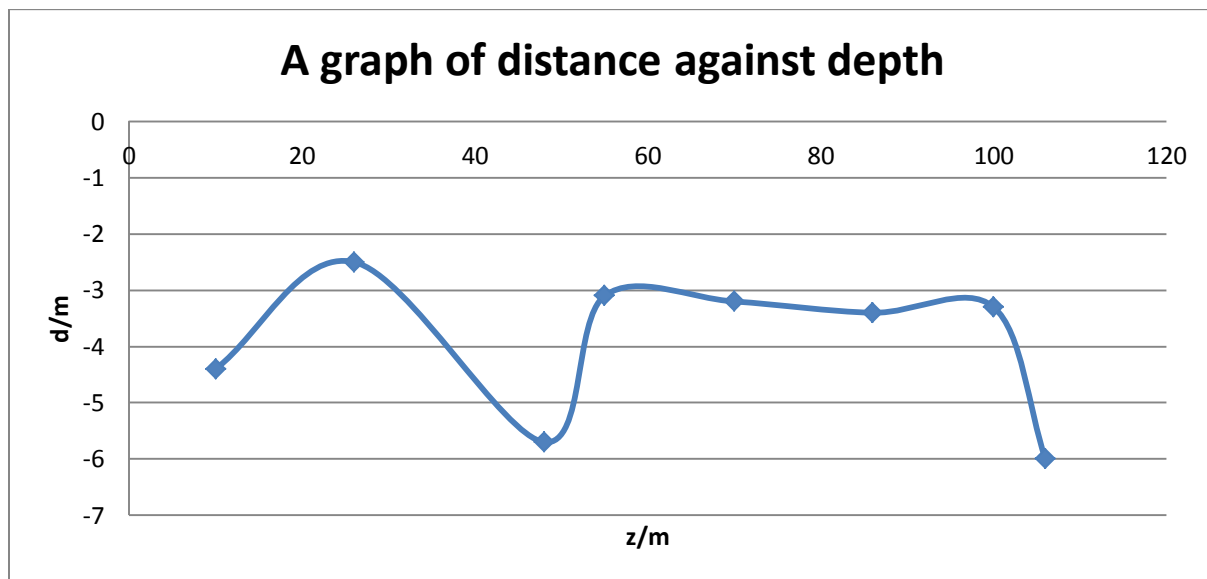


Figure 3.19 A z - d plots for depth and distance from ITM of traverse L2

Table 3.3 Calculated depths to the surface of the second layer from the top of the first layer along the traverse L3

z	5	28	43	59	75	81	95	113
V_1	185	182	557	129	718	875	97	657
V_2	999	814	1140	1007	802	1133	896	1308
t_i	23	13	22	18	14	14	23	25
d	2.2	1.2	7.0	1.2	11.3	9.6	1.1	9.5

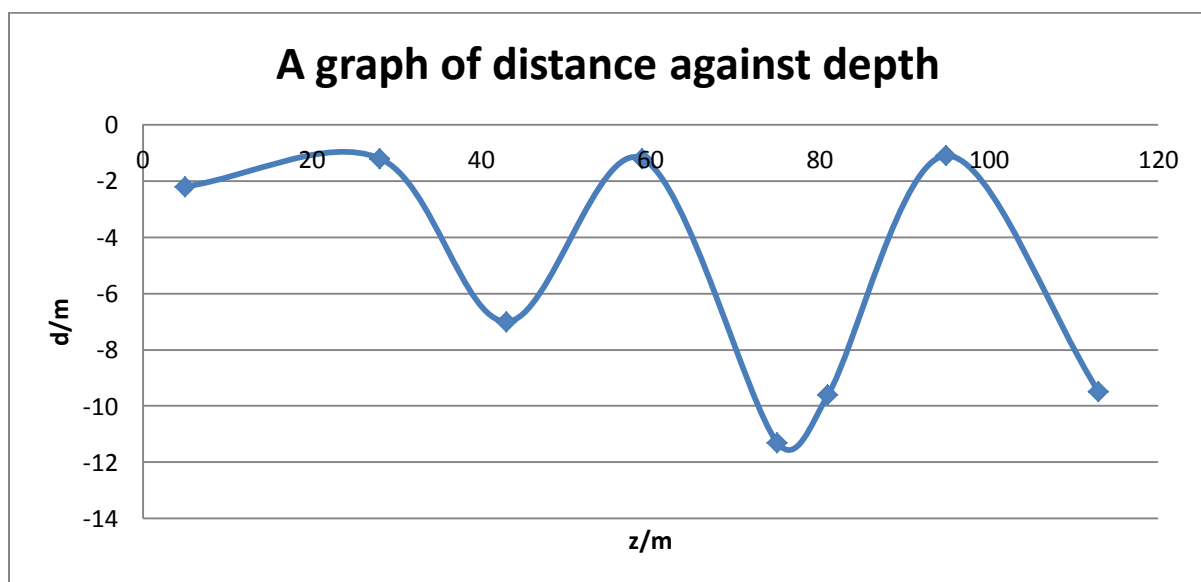


Figure 3.20 A $z - d$ plots for depth and distance from ITM of traverse L3

Table 3.4 Calculated depths to the surface of the second layer from the top of the first layer along the traverse L4

z	13	23	45	52	67	81	98	105
V_1	696	293	477	304	352	389	279	686
V_2	1686	658	1881	1337	1232	1298	1031	1542
t_i	47	15	13	21	18	25	20	30
d	18.0	2.5	3.2	3.3	3.3	5.7	2.9	11.5

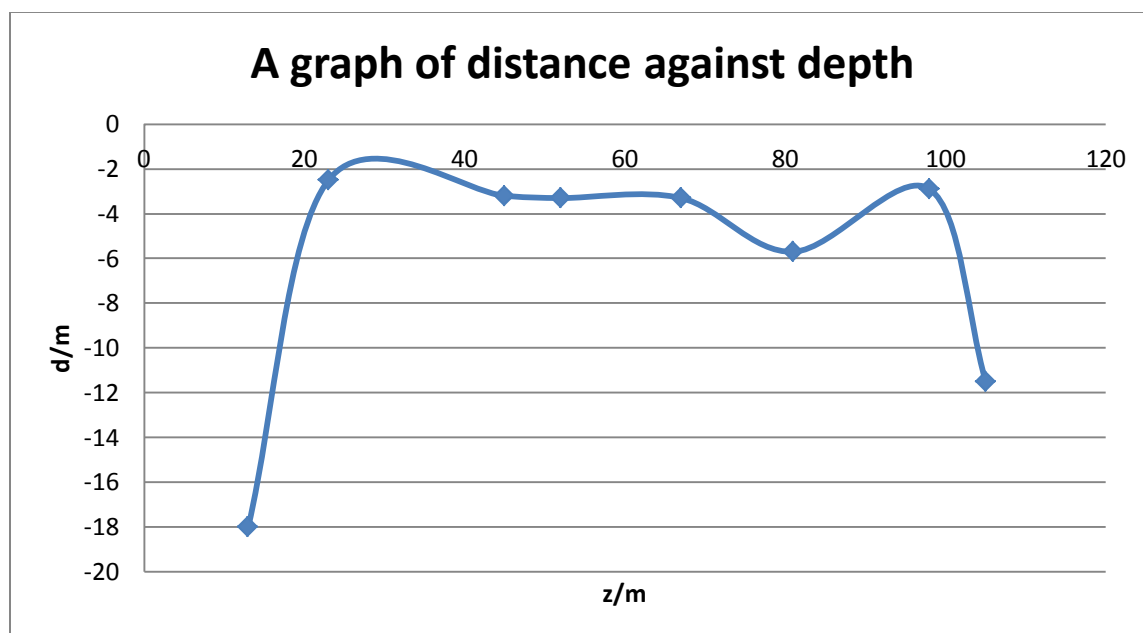


Figure 3.21 A z - d plots for depth and distance from ITM of traverse L4

3.4 MODELS

The data obtained from the surveys which were time of arrival from shot points to geophones and distances from shot points to geophones were used to generate 2D models using the Generalized Reciprocal Method (G.R.M). These models showed two layers. The velocity the upper layer was always lower compared to that in the lower layer. The differences in velocities of the waves in the upper and lower layers show that the layers are of two different media. The possibilities of these media are either dry or saturated with water among other possibilities. The differences in velocity help to know the material properties of the subsurface in terms of density, lithification, porosity etc.

The GRM calculated the refractor depth for each geophone location using overlapping refraction arrival times from forward and reverse shots along each traverse. The five multiple shot points along the survey traverses permitted the interpretation the changing interface depth along the second layers and layered velocities. The GRM generated the 2D or layered model with continuity the refractor surface across the traverses.

Figure 3.22 shows a compiled travel time curves, velocity model and a depth section for the traverse L1 generated using the G.R.M. Two velocity layers were modeled along the traverse L1. The velocity model L100 indicated a velocity increase in the velocity of the second layer. The velocities with which the seismic wave travels through the first and second layers are 420 m/s and 1839 m/s respectively. The depth of the second layer which is the bedrock varies along its length. This ranges between five meters to twelve meters deep along its length. This gives it an undulating surface. The G.R.M was used to generate the models for the other traverses L2, L3 and L4.

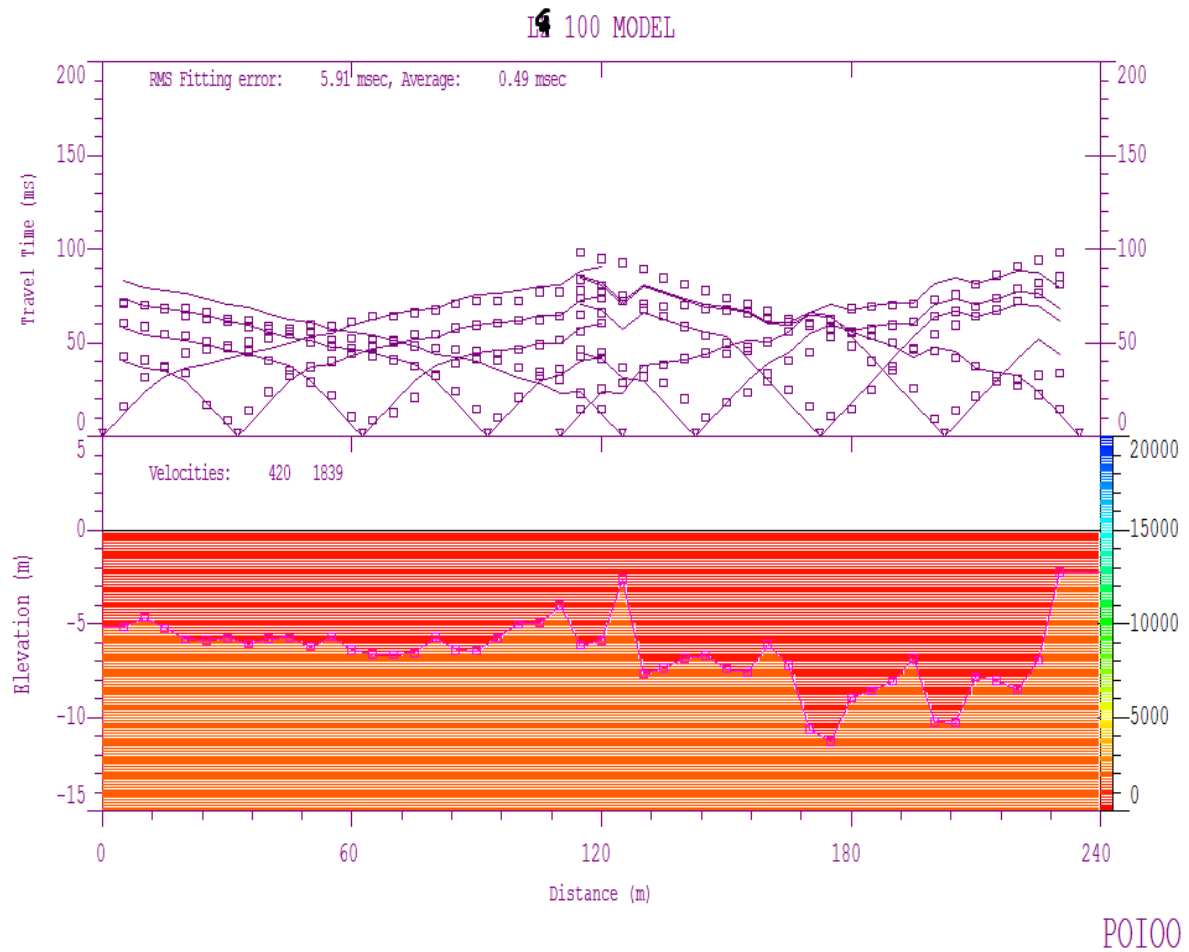


Figure 3.22 2D model generated from the traverse L1

The L200 model which was the model generated by the G.R.M from the second traverse L2 has the second layer at a depth varying between three meters (3 m) and five meters (5 m) along its length as shown in figure 3.23. The velocity with which the seismic wave propagates through the upper layer is about 356 m/s and that of the second layer is about 1490 m/s.

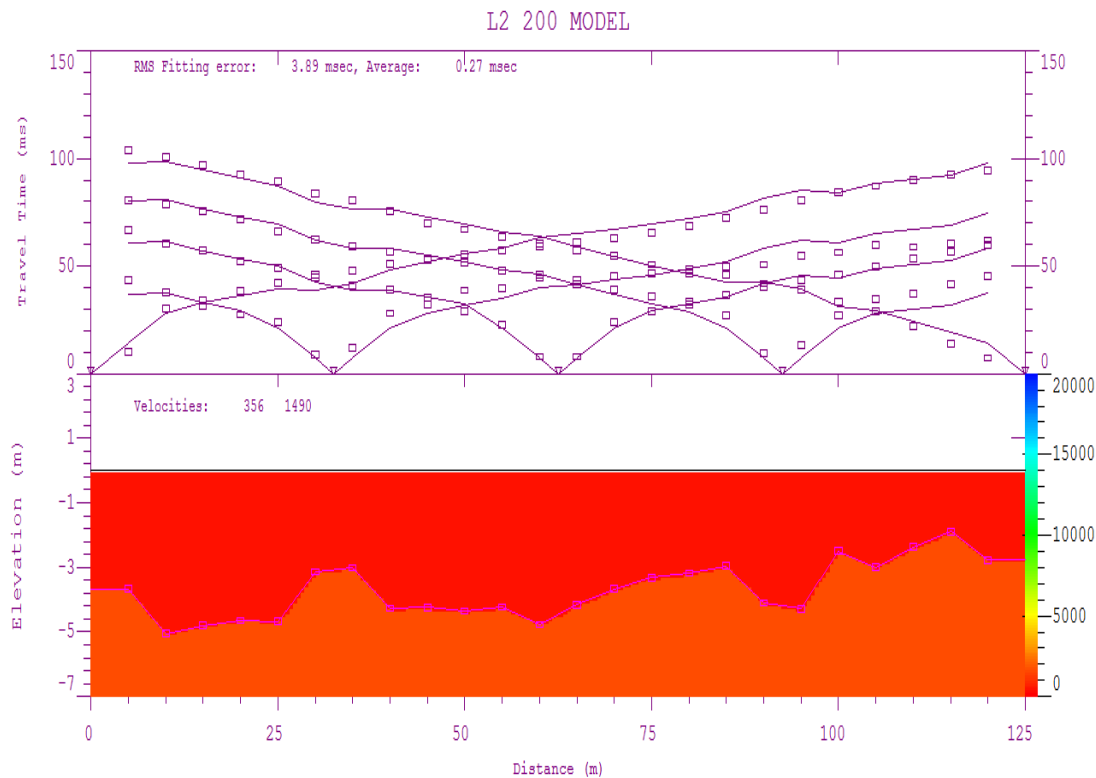


Figure 3.23 2D model generated from the traverse L2

The third model M 300; also produced from the processing of the profile L3 data by the GRM had its second layer having an undulating surface due to the different depths along its length. The second layer had a depth ranging from about three meters (3 m) to twelve meters (12 m) deep as in figure 3.24. The velocity with which the seismic wave propagates through the upper layer is about 558 m/s and that of the second layer is about 1344 m/s.

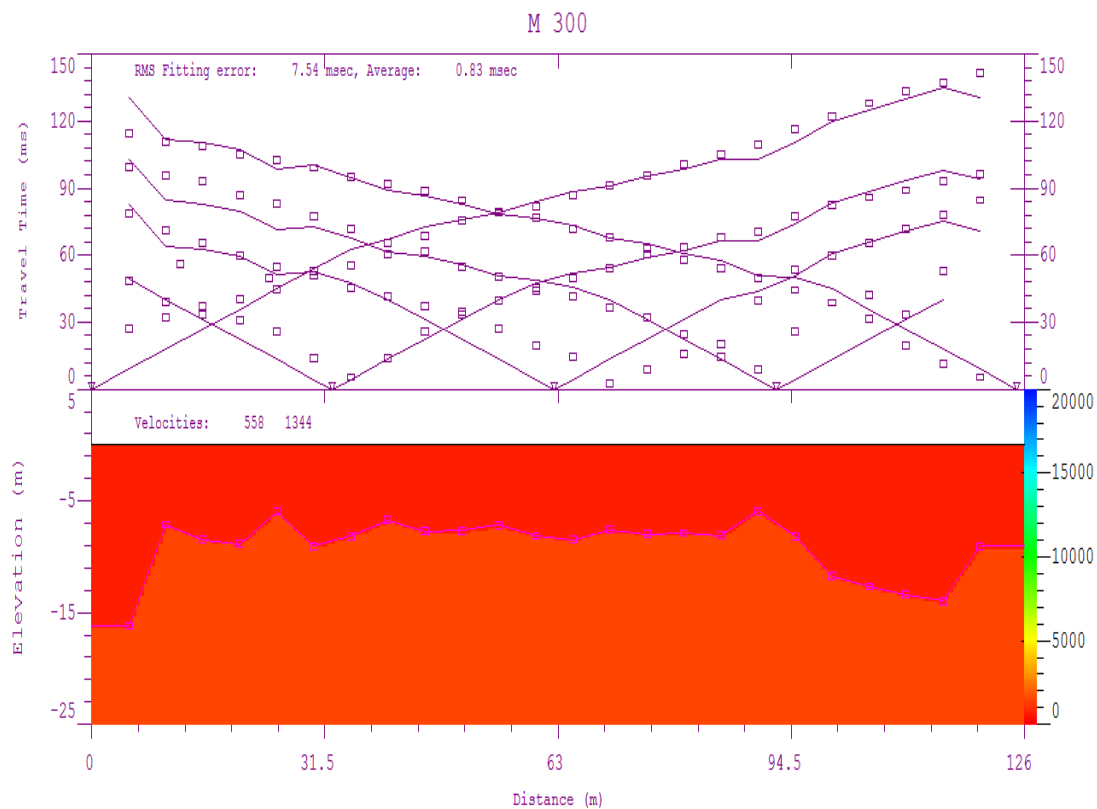


Figure 3.24 2D model generated from the traverse L3

The G.R.M was also used to generate the forth model L400 which was obtained from the profile L4 as shown in figure 3.25. The model had its second layer having an undulating surface. The velocities with which the seismic wave travels through the first and second layers are 401 m/s and 1292 m/s respectively. The second layer had a varying depth from about 3.5 m to 16.5 m.

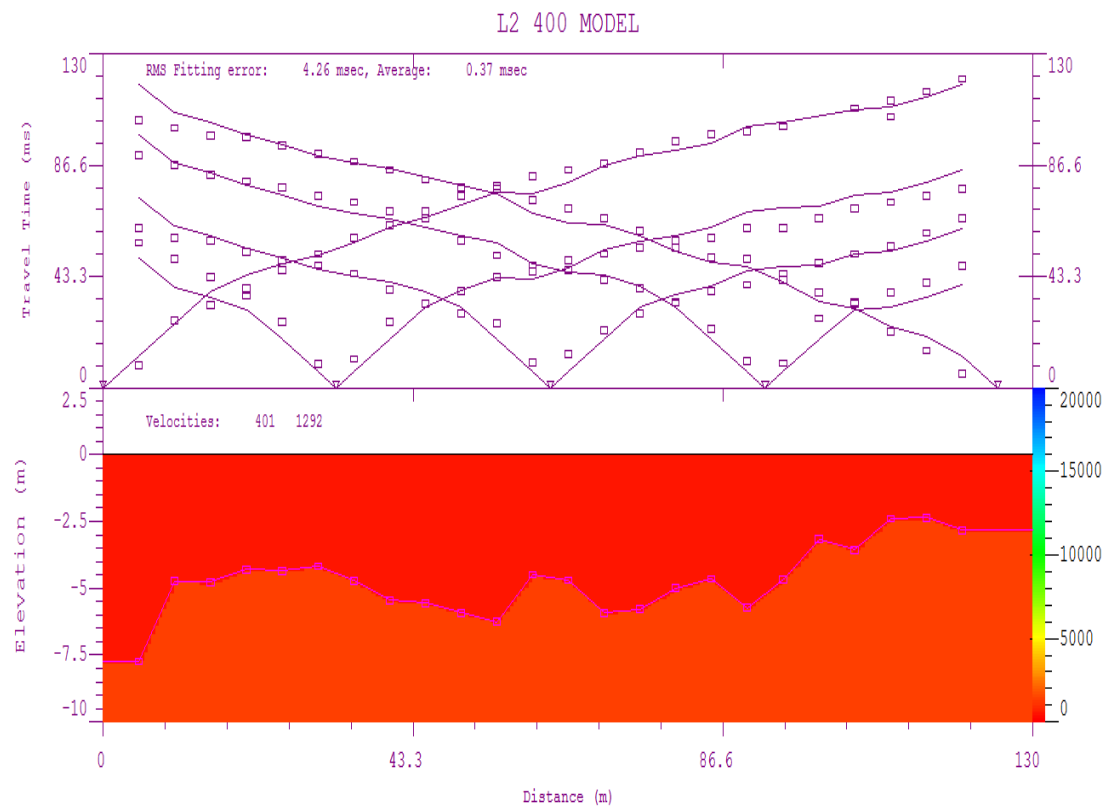


Figure 3.25 2D model generated from the traverse L4

CHAPTER FOUR

4.0 RESULTS AND DISCUSSION

The work done was to use seismic refraction survey in the foundation evaluation. Reliable estimate of the strength and deformation characteristics of rock masses are required for almost any form of analysis used for the design of slopes, foundation and underground excavation. The primary considerations for any foundation support are the bearing capacity, settlement and ground movement beneath the foundation. The bearing capacity of any site depends on the physical properties of the soil and the bedrock at the site

In areas where the soil provides sufficient strength for the support of structures, it is used as the bed for foundation. In such areas, compatibility of the soil is very strong and so stiffness of the soil increases with decrease in permeability. Field placement work often has a specification requiring a specific degree of compaction or alternatively specific property of the compacted soil.

Traditional interpretation of seismic refraction data has used a concept of layered horizons or zones where each horizon has a discrete seismic velocity. For the purpose of this survey, two primary interpretation methods were considered; traditional layered interpretation which is represented by the intercept-time method (ITM) and the use of General Reciprocal Method (G.R.M) to generate 2D models. Both methods provided seismic velocity typically determined from the first arrivals of compression wave (p-wave), as part of the results of investigation. The use of compression wave in seismic refraction method prevented straight forward detection and interpretation of lower velocity layers or zones (hidden layers) underling higher velocity layers.

The time-distance plot assumed that the subsurface materials were of layers or zones and that each layer had a different uniform velocity. Lateral changes in velocity, especially in shallow portions of the subsurface profile interpretation were detected by interpreting data between adjacent or center shot along the profile. The two uniform straight best fit lines on the time-distance plot showed that the subsurface had two layers. Velocity is calculated as distance traveled divided by time elapsed but since the plot was from a time – distance plot, the inverse of the slope of the lines of best fits gave the velocities of the two media or layers identified.

Depth investigation is a basic parameter that is frequently misunderstood or not adequately quantified. This is especially true for engineers and managers who do not understand geophysics but depend upon seismic refraction work. From the time-distance plot, the intercept time for both forward and reverse shot as produced from the back-extrapolated refracted arrival intercept on the time axis were substituted into the intercept-time equation to calculate the depth of the interface at different points along its length. This method limited the idea of the depths at points along the surface between the depths at the points where the intercept time equation revealed. The time-distance plot helped to achieve the target of identifying two layers of the subsurface and the thickness of the first layer or depth of the bedrock at particular points. The 2D models that were developed from the G.R.M also gave a pictorial view of the subsurface along its whole length. From the 2D models, the thickness of the first layer or depth of the principal refractor at every point along its length was revealed. The survey showed that the thickness of the first layer range between five meters to sixteen meters.

Comparing the depth to the refractor at various points along its length from the original shot position 0 m, produced by the two methods thus the I.T.M and the G.R.M, it was realized that the depth produced by the I.T.M at certain points were not accurate.

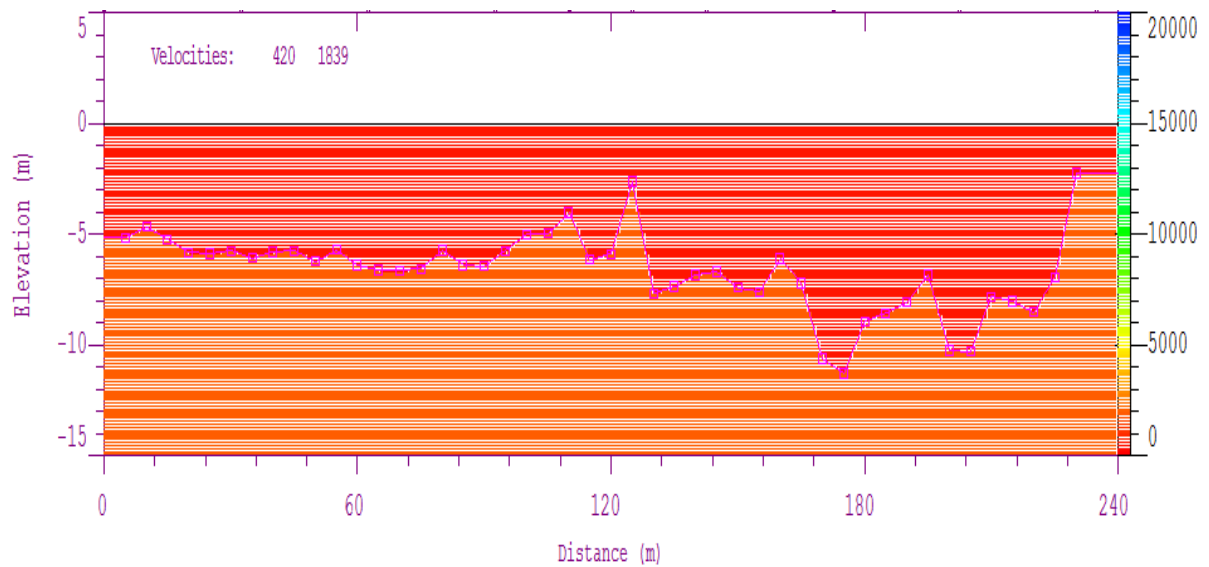
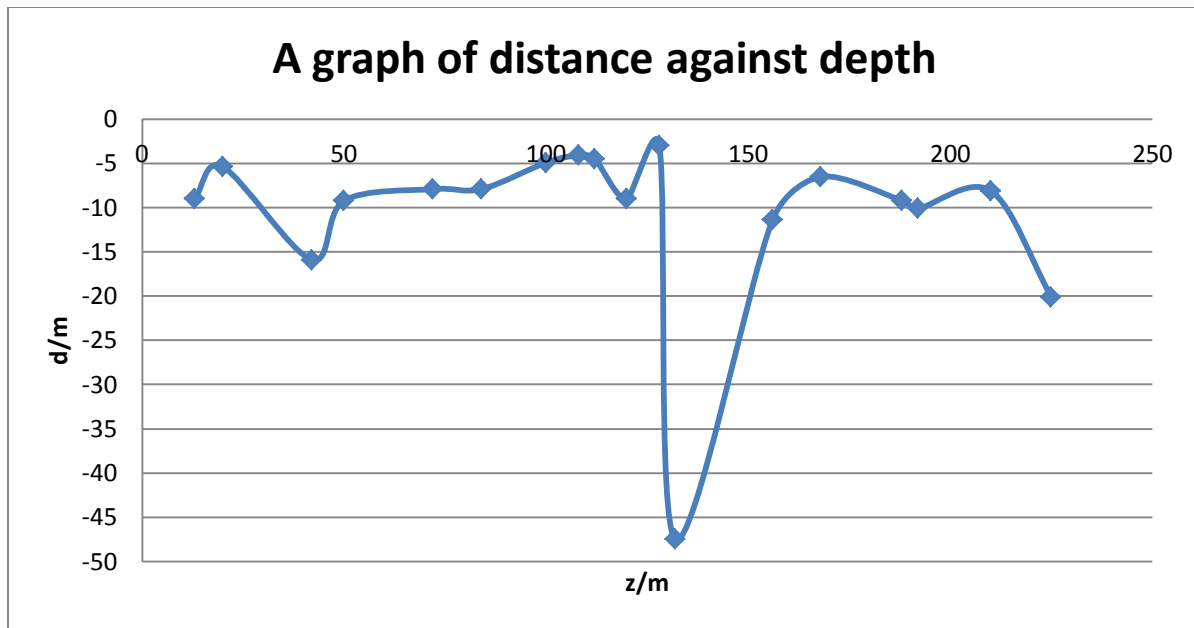


Figure 4.1 Depths from the first traverse produced from I.T.M and G.R.M.

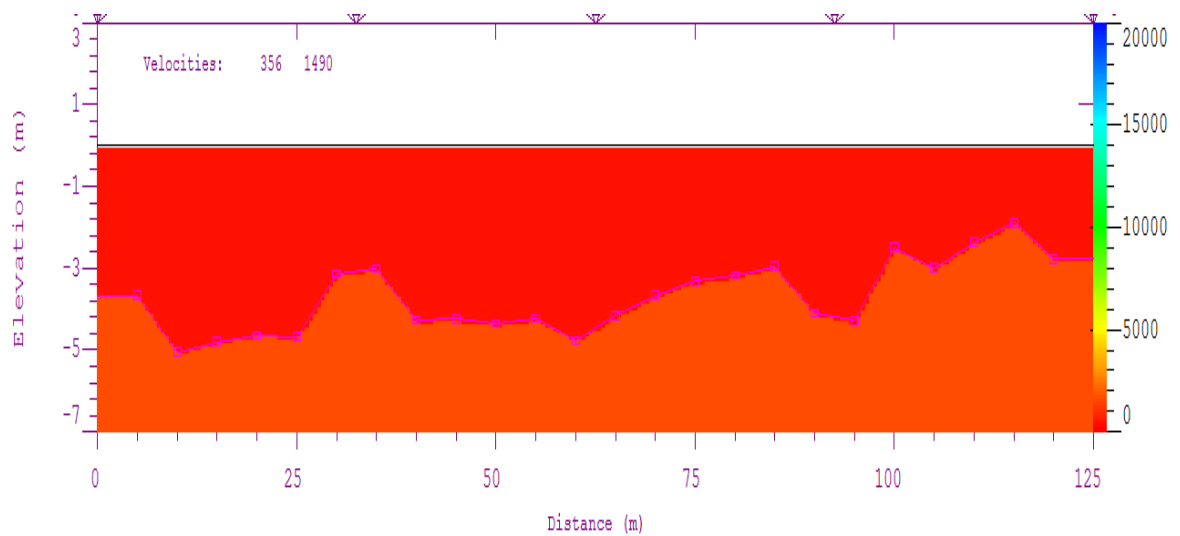
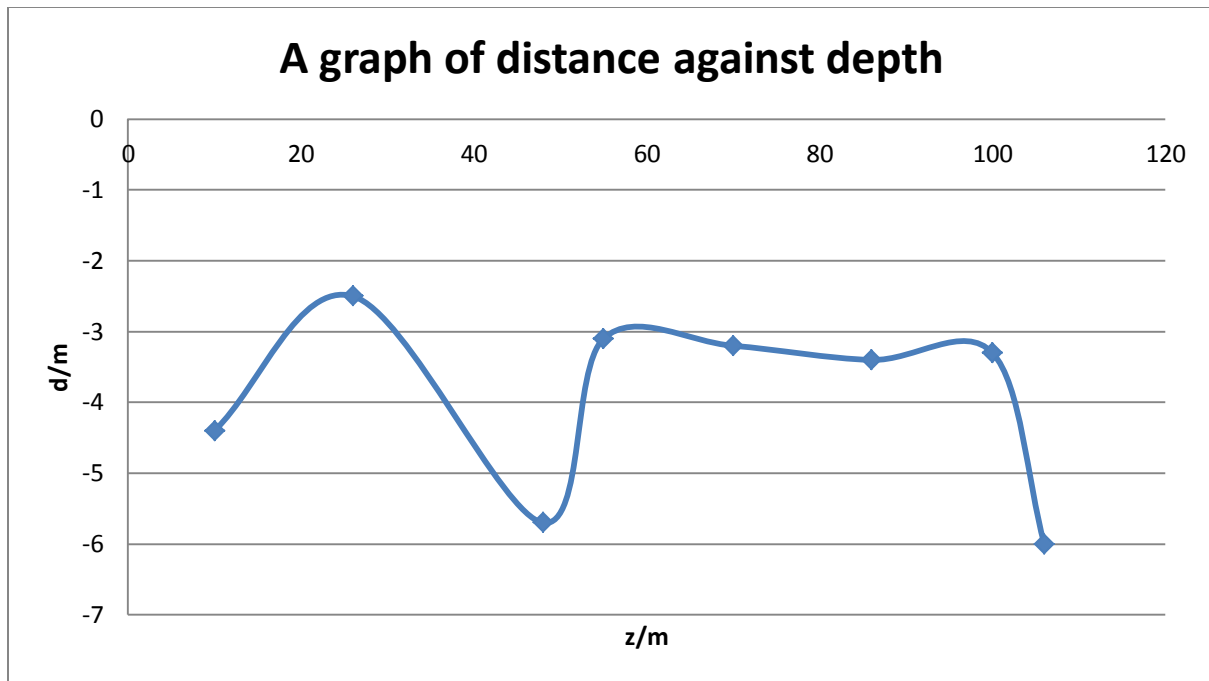


Figure 4.2 Depths from the second traverse produced from I.T.M and G.R.M.

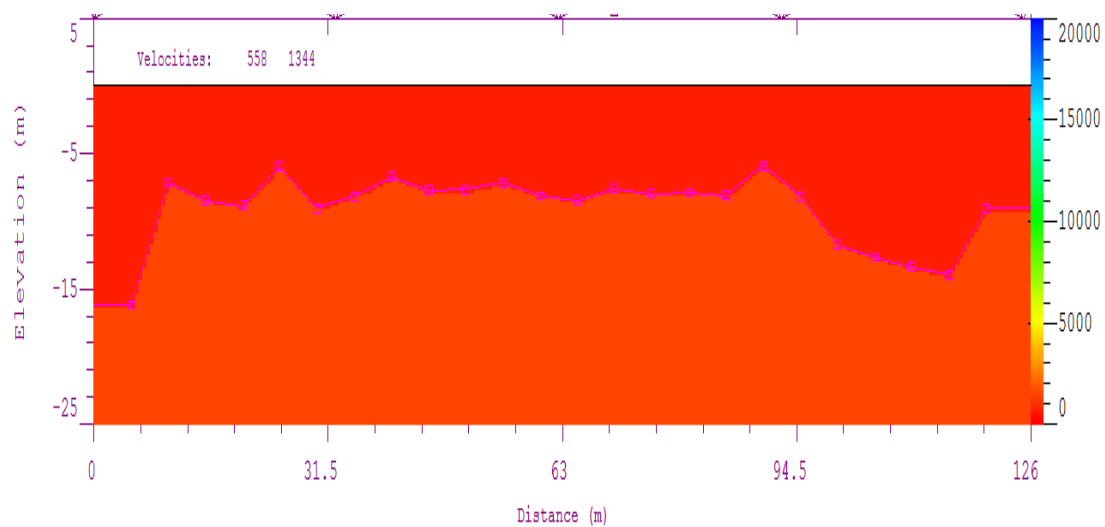
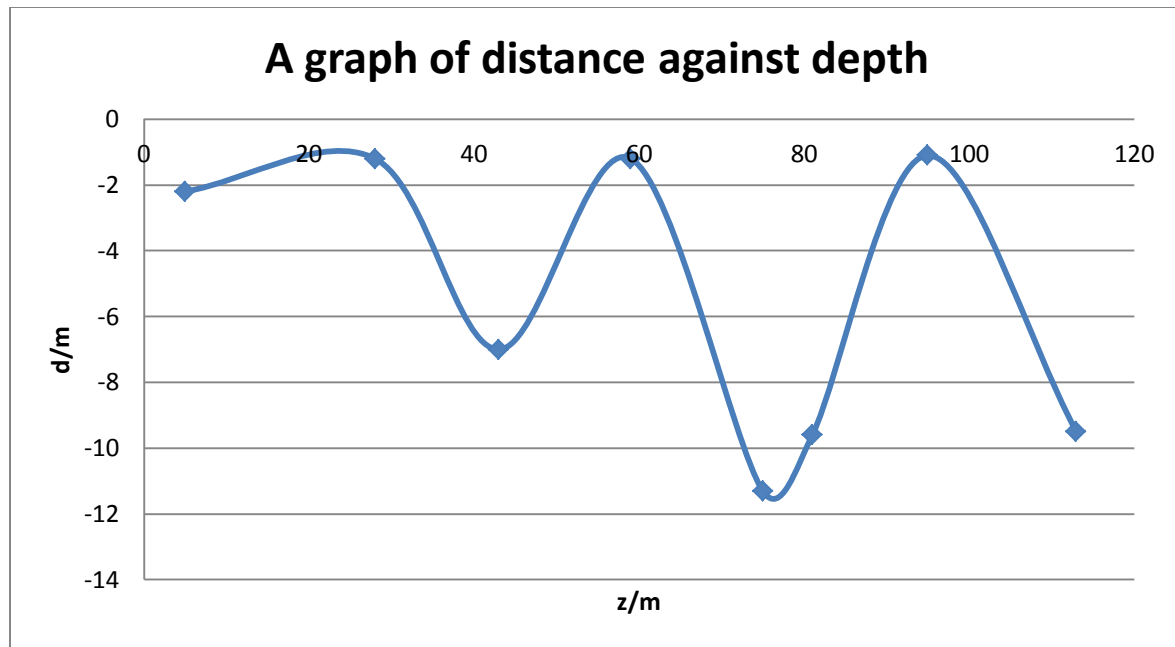


Figure 4.3 Depths from the third traverse produced from I.T.M and G.R.M.

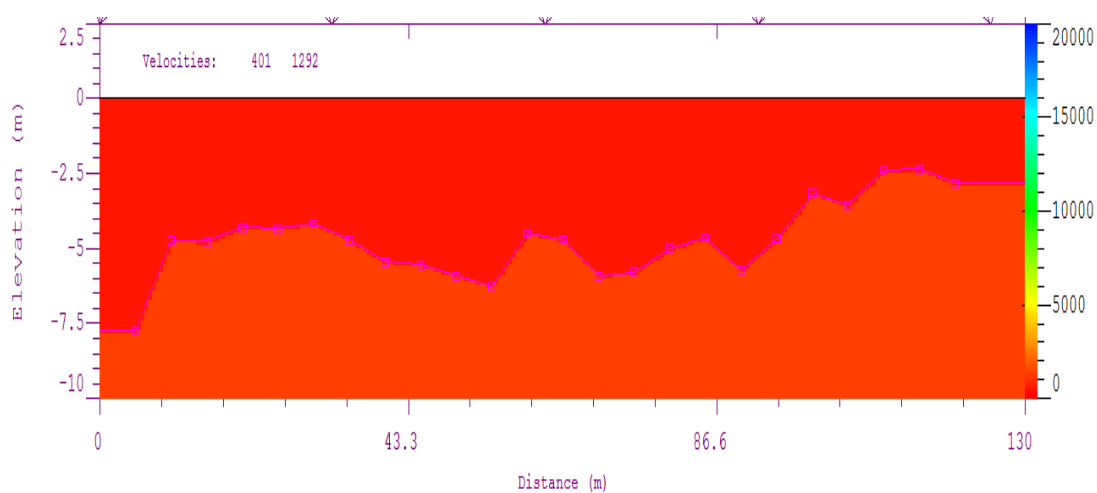
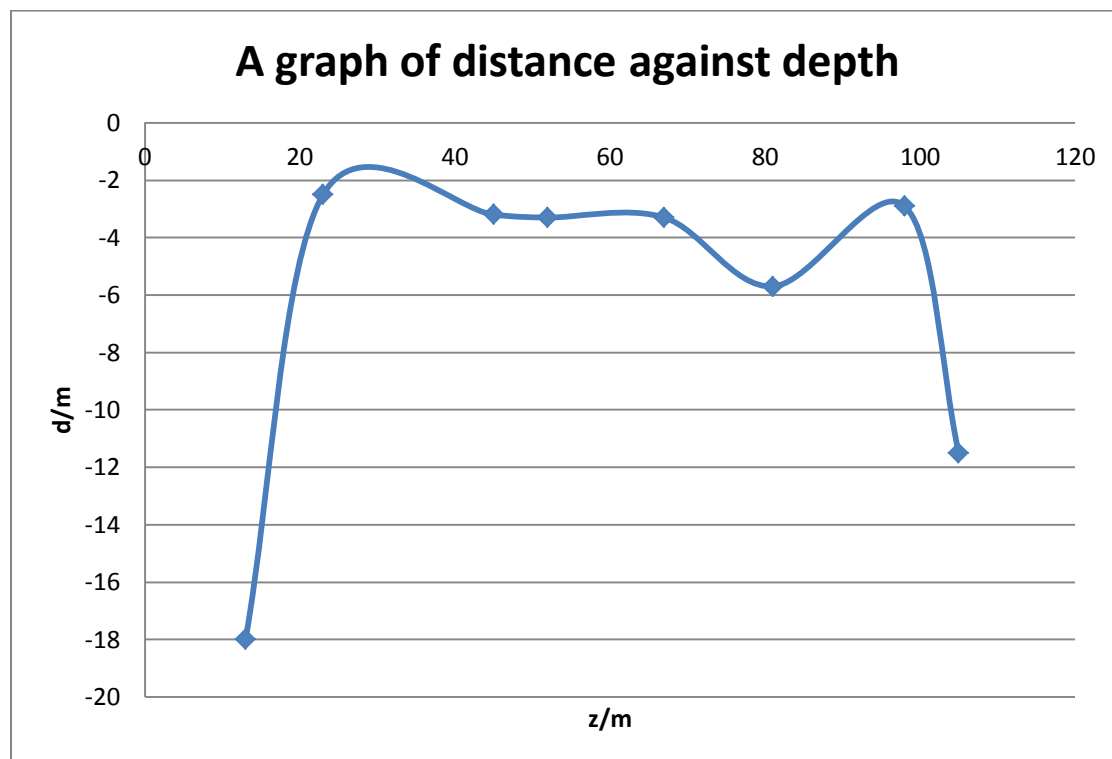


Figure 4.4 Depths from the fourth traverse produced from I.T.M and G.R.M.

This was observed when the graph produced from the z-d plot of the distance from the original shot point and depth at various points along the refractor obtained from the I.T.M were compared. For instance when depths from the first traverse were inferred at positions of 20 m, 50 m, 150 m and 200 m from the origin along the traverse, the I.T.M produced respective depths of 5 m, 8 m, 16 m, 8 m, whilst the G.R.M 2D model of the second traverse at the same distances also showed respective depths of 6.2 m, 6.5 m, 5.0 m, 7.5 m and 12 m as compared in figure 4.1. The two methods produced different depths but the difference was not that much. When the depths produced by the I.T.M and G.R.M were compared on the second transverse in figure 4.2 at positions of 20 m, 40 m, 60 m, 80 m and 100 m, the I.T.M produced depths of 2.5 m, 4.5 m, 2.8 m, 3.4 m and 3.2 m respectively whilst the G.R.M produced respective depths of 4.8 m, 4.6 m, 4.6 m, 3 m and 3.4 m. The compared depths of the refractor from the two methods was done along the third traverse at distances of 20 m, 40 m, 60m, 80 m and 100 m as in figure 4.3 and the I.T.M produced depths of 1 m, 6.2 m, 1 m, 9.5 m, and 1m respectively whilst the G.R.M showed respective depths of 9 m, 7.5 m, 7 m, 9 m and 13 m. The two methods produced completely different depths but since the G.R.M gave a 2D model of the layers, one could really see the depth of the refractor. This difference in depth produced by the two methods showed that the I.T.M could not be fully relied on. The depth comparison of the two methods was repeated along the fourth traverses at distances of 20 m, 40 m, 60 m, 80 m and 100 m from the original shot point as in figures 4.4 and the two methods, I.T.M showed respective depths of 4 m, 3 m, 3 m, 7.5 m and 3 m whilst the G.R.M produced respective depths of 4.5 m, 5.5m, 5 m, 5 m and 3.5 m. This also showed differences in depth at various positions along the traverse.

The inverse of the slopes calculated from the best fit lines on the time –distance plot showed that the two layers had different velocities and that the lower layer had a higher velocity than the upper layer. The 2D model also showed that the two layers had different velocities and confirmed that the upper layer velocity was lower than that of the second layer. The velocities ranged between 400 m/s to 880 m/s.

The difference in velocities in the layers of the subsurface was indicates that the layers were different in terms of certain properties. Since compressional or p-waves are mechanical waves they always require a material medium to pass through. P-wave velocity measures the travel speed of longitudinal wave in materials. The velocity measurement provided a correlation to the physical properties of the medium through which it travels.

The high velocity in the second layer meant that the grains of the second layer (bedrock) had probably a high degree of compaction. A well lithificated rock generally has high velocity as the grains are in good contact. Also the high compaction among the grains of the second layer made it have a high density that result in an increase in velocity of the wave in the bedrock. Again compressional wave velocity is strongly dependent on the effective stress in the medium. For a rock that is buried in the earth, the confining pressure is the pressure of the overlying rock or layer. In general, velocity rises with increasing confining pressure and then levels off to the terminal velocity when the effective pressure (pressure difference between the confining pressure and pore pressure) is high.

The high velocity in the lower layer may also be associated with high water saturation of the layer. When these velocities in the two layers are compared, one could say that the upper layers could be loose, unsaturated with water but saturated with air and more permeable. This showed that they had little ability to resist pressure and stress.

These soil types could not serve as the base to foundations of structure.

In areas of shallow bedrocks, most foundations may bear directly on the bedrock. The areas of deeper bedrock with soft overlying soil, deep foundations are used to support structures directly on the bedrock. From the models it was observed that the velocities of the waves through the lower layers thus bedrock were generally higher. The higher velocities in the lower layers meant that there was probably low porosity with more water saturation in the bedrock. The low porosity with more water concentration contributes to the fact that the grains of the bedrock are cemented together by post depositional chemical process given the bedrock a higher density.

The combined effect of decrease porosity and high lithification cause the observed increase of velocity with depth of burial of the bedrock. The increase in lithification which result into reduction in porosity and made the bedrock massive and thereby having a high density

These properties of low porosity, high lithification and density in the second layer (the bedrock) make it able to resist load from structures and thereby are able to support foundations of these structures.

CHAPTER FIVE

5.0 CONCLUSION AND RECOMMENDATION

5.1 CONCLUSION

The work done in the defined region showed the use of two methods to infer the subsurface in terms layered surfaces, nature of horizon and depth. These methods were the I.T.M and the G.R.M. From the survey I wish to conclude that two layers were unveiled. These were the upper layer and the lower layer.

The velocities of the waves in the upper layers were as low as seen on the time – distance plot and this meant that they were not good beds to support the foundation of structure. These low velocities were due to the fact that the upper layers were probably too porous and that they were saturated with so much air. Again it was clear from the lower velocities through the upper layer that the degree of lithification was probably very low. This also meant that compatibility was also very poor. Also these lower velocities in the upper layer could result from the fact that the upper layer was saturated with air meant that the internal pressure within the upper layer was low and could not resist much stress. These properties which are believed to be possessed by the upper showed that the soil of the upper layer is not a good bed to support foundations. Then again the models showed that the upper layers were thick with reference to the depth of the bedrock.

The second layer on the other hand which form the bedrock had a higher wave velocity as compared to the first layers as seen from the time – distance plot. This could be due to the fact that the bedrock was less porous and well saturated with water. This could also mean that the

bedrock's internal pressure was higher making it able to resist much stress from structures. The degree of lithification was probably higher when it was compared to the upper layer leading to higher compatibility. The depth of the bedrock showed that it was deeply buried. Again the second layer had a higher density which makes it massive and a very good bed for foundation support.

With respect to the depth of layers, the I.T.M could not always give a true reflection of depth along a traverse when it was compared to depth along a traverse from the 2D models of the G.R.M.

5.2 RECOMMENDATION

From the models it was realized that the upper layer could be too porous and so much saturated with air. This meant that any small pressure exerted on it will cause it to collapse downwards. This showed that when the foundation of any structure is mounted on the top layer, there is the likelihood that the structure may sink like the Tower of Pisa or even collapse

On the other hand, the second layer showed low porosity with more water saturation giving it much internal pressure. The layer showed a high level of lithification with compatibility when compared with the upper layer. The second layer had a higher density and was massive. These properties which are presumed to be exhibited by the second layer make it able to resist much pressure. Based on this it could be realized that the second layer forms a good bed for foundation support. However I wish to recommend that other geotechnical methods should always be applied to support this investigation. Again from the survey I wish to recommend that the I.T.M must not only be used to investigate depth as it does not always give a true reflection of depth along a traverse as was seen from the depth calculated from the time intercept equation.

From all the survey and analysis carried out, I which to recommend that whenever any structure is to be put at any place within the survey domain, the upper layer has to be graded off before the foundation could be erected. This must be done so that it could have a solid base on the bedrock to prevent it from future sink or collapse.

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